76th ANNUAL FIELD CONFERENCE
OF PENNSYLVANIA GEOLOGISTS

Geology of the Pennsylvanian-Permian in the Dunkard Basin

September 29 - October 1, 2011
Washington, PA

Hosts: Pennsylvania Geological Survey
Pittsburgh Geological Society

Guidebook for the
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GEOLOGY OF THE PENNSYLVANIAN-PERMIAN IN THE DUNKARD BASIN

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Cover: Vik Skema, in search for the ever elusive *Callipteris conferta*.

Cartoons: John A. Harper

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IN MEMORIAM
VERONICA A. REYNOLDS
(1969-2008)

My best friend, Veronica Reynolds, passed away in December 2008 due to pancreatic cancer. She was 39 years old. We met at Allegheny College (Meadville, PA) in August 1987, and we lived next door to one another in the same dormitory during freshman year. Veronica and I became fast friends, and were roommates for the remaining three years of our time at Allegheny, she a chemistry major and me a geology/environmental science double major. In fact, I credit much of my current appreciation of, and interest in, chemistry to Veronica.

Although we graduated in 1991 and went our separate ways, we kept in touch over the years. Veronica earned an M.S. in organic chemistry from the University of Minnesota, then moved back to her home state of New York to work as an earth science teacher. After the 9/11 attacks on our country, Veronica was hired as a criminalist with the New York Police Department, testing anthrax samples in the NYPD laboratories in Jamaica, NY.

Looking for a change, Veronica relocated to Reno, Nevada, a few years later, where she accepted a position as a criminalist, conducting laboratory work on blood alcohol samples and providing expert witness testimony in DUI cases.

I visited Veronica in May 2008 to celebrate her birthday, and we not only saw the local sites in Reno but also took a day trip to Lake Tahoe (see photo). Little did either of us know that this would be the last time we saw each other; Veronica passed away seven months later.

In loving memory of my best friend, I have established an annual FCOPG student scholarship. What better way to honor Veronica's memory than to sponsor a deserving college student's attendance at this annual event. The recipient of the first annual Veronica Reynolds Memorial Scholarship is Michele Cooney of Pittsburgh, PA.

Kristin M. Carter, P.G.
Chief, Petroleum and Subsurface Geology Section
Pennsylvania Geological Survey

STRATIGRAPHY OF THE DUNKARD GROUP IN WEST VIRGINIA AND PENNSYLVANIA

Nick Fedorko and Viktoras Skema

INTRODUCTION

The Permo-Pennsylvanian Dunkard Group strata are the youngest in the Appalachian Basin, preserved in the deepest part of the Pittsburgh-Huntington Synclinorium in western West Virginia, eastern Ohio, and southwestern Pennsylvania. A small area is also mapped in a syncline in western Maryland, although the authors speculate that strata found here may be older, belonging to the underlying Monongahela Group (Figure 1). This paper focuses on the stratigraphy of the Dunkard Group in West Virginia and Pennsylvania. Martin (1998) provides a comprehensive discussion of the Dunkard Group in Ohio as well as in West Virginia and Pennsylvania.

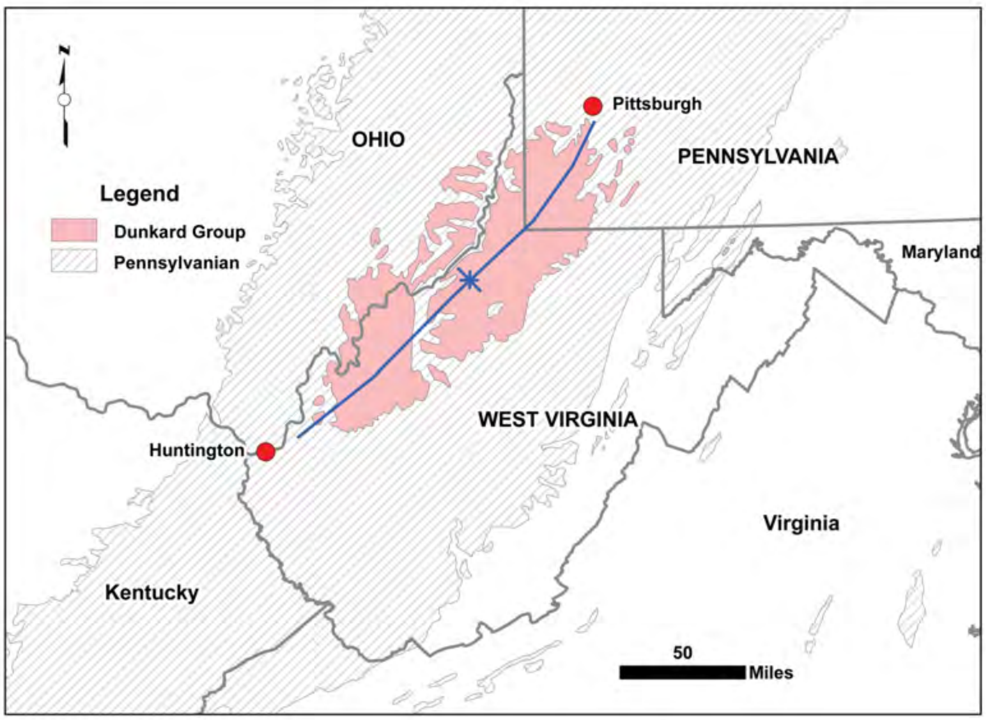
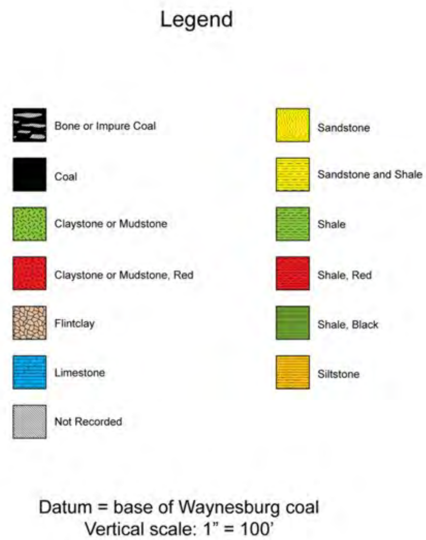


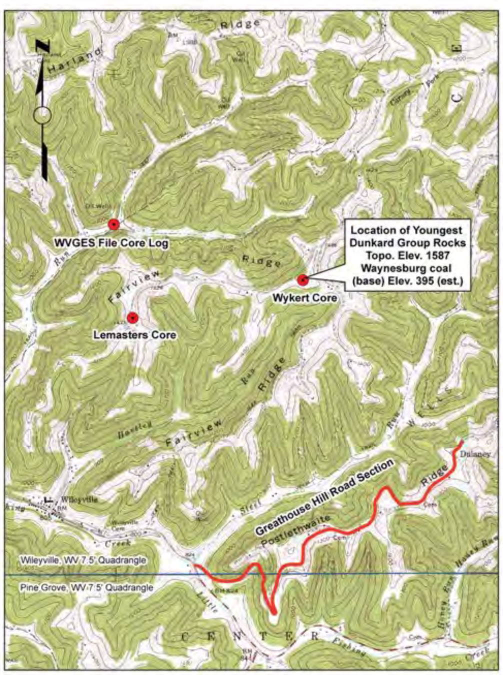
Figure 1. Area underlain by the Dunkard Group rocks in the northern Appalachian basisn (from King and Beikman, 1974).

Maximum thicknesses of Dunkard Group strata are found beneath the highest ridges nearest the synclinorium axis in Wetzel County, West Virginia and Greene County, Pennsylvania (Figure 2). In West Virginia approximately 1,190 ft (363 m) of Dunkard Group

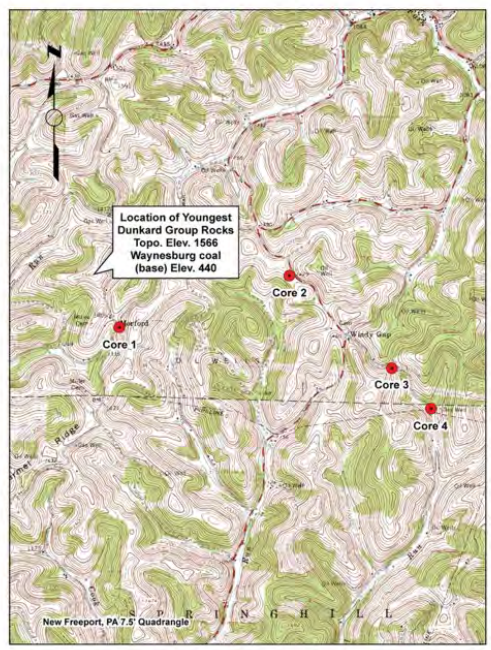
Fedorko, Nick and Skema, Viktoras, 2011, Stratigraphy of the Dunkard Group in West Virginia and Pennsylvania, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 1-25.

Figure 2A. Composite sections of the maximum Dunkard Group rocks in WV and PA. **See p. 3 for the locations and legend.**





Location of data used to construct the composite of the thickest Dunkard Group section in West Virginia near Wileyville, Wetzel County. Lemasters and Wykert core logs from Clendening (1974). Driller's log of sub-drainage core from West Virginia Geological and Economic Survey (WVGES) files. Greathouse Hill Road section (modified) from Cross and Schemel (1956).



Location of data used to construct the composite of the thickest Dunkard Group section in Pennsylvania near Windy Gap and Morford, Greene County. Driller's logs of cores 1, 2, and 3 courtesy of Consol Energy. Geologist's log of core 4 from the Pennsylvania Geological Survey. Uppermost part from I.C.White (1891, p. 24), measured section in Aleppo Township of exposures along the Dunkard Fork of Wheeling Creek from the source (assumed to be near Morford) northwestward to Crab Apple Creek near the West Virginia line.

Figure 2B. Legend for Figure 2, p. 2, and maps showing the locations of data used to construct the composite sections.

strata are preserved beneath Fairview Ridge near Wileyville, Wetzel County. The maximum section in Pennsylvania, beneath a ridgetop northwest of Morford and near Windy Gap, is estimated to be approximately 1,125 ft (343 m) thick. The Pennsylvania section may be thinner, but given southward section thickening and facies changes, this section may contain younger rocks in the summit (Stevenson, 1876). However, correlation of the two sections in Figure 2 suggests the top of the West Virginia section may be slightly younger. These are not insignificant stratigraphic sections given that the underlying Pennsylvanian strata--including the Pottsville through Monongahela Group in this region total an estimated 1,350 ft (412 m).

STRATIGRAPHY

The Dunkard Group includes all the strata from the base of the Waynesburg coal bed or its stratigraphic equivalent to the top of the highest exposures (Berryhill and Swanson, 1962). The Dunkard Group is comprised of interbedded black, gray, green, and red shale, gray, green and red claystone and mudstone, gray and green siltstone, gray and green lithic and micaceous sandstone, nonmarine limestone, and generally thin coal beds. Calcareous material is common throughout the Dunkard Group either disseminated throughout beds or as nodules. The Dunkard Group is divided into the Waynesburg, Washington, and Greene Formations (Berryhill and Swanson, 1962) (Figure 3).

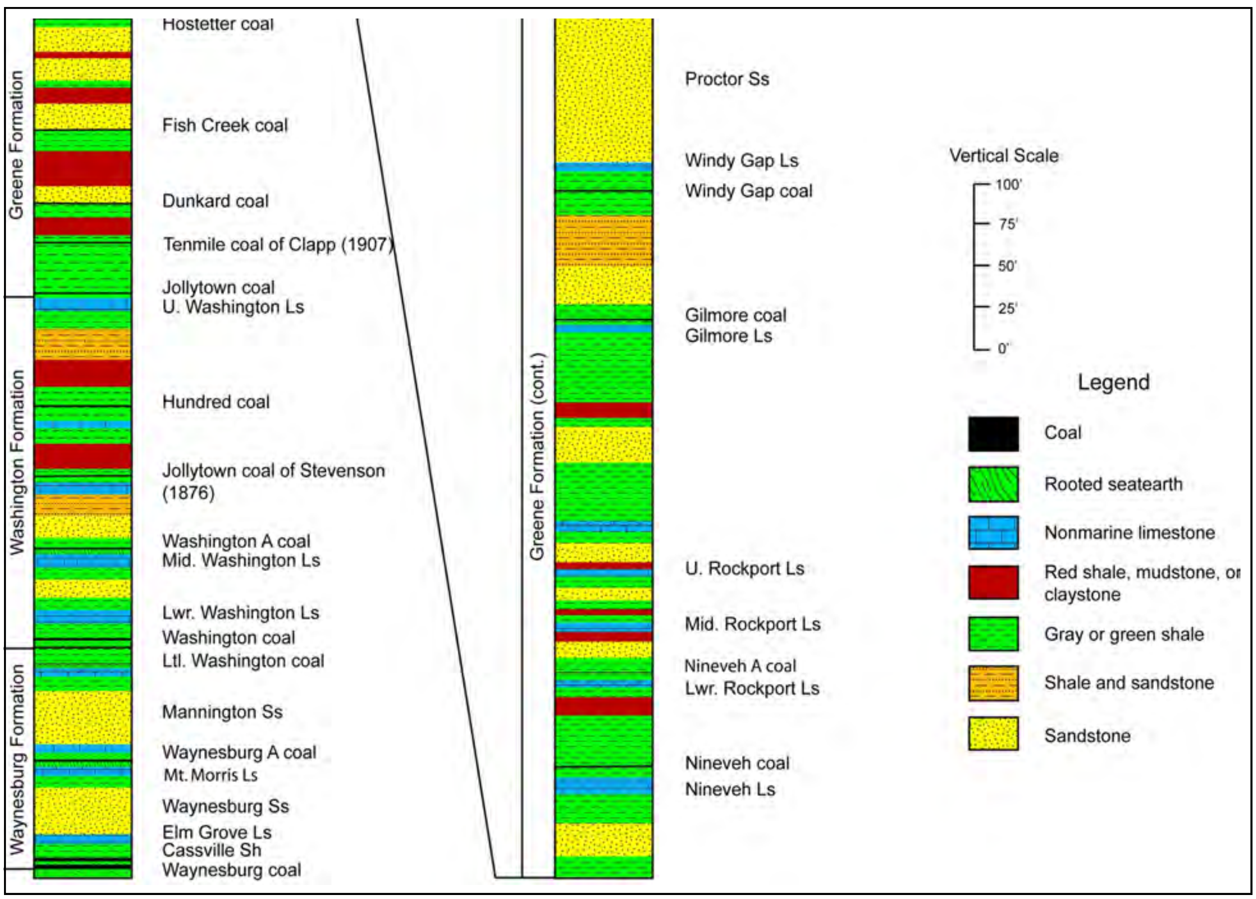


Figure 3. General section of the Dunkard Group in northern WV and southwestern PA constructed from core logs from the West Virginia Geological and Economic Survey (WVGES) and Cross and Schemel (1956, modified).

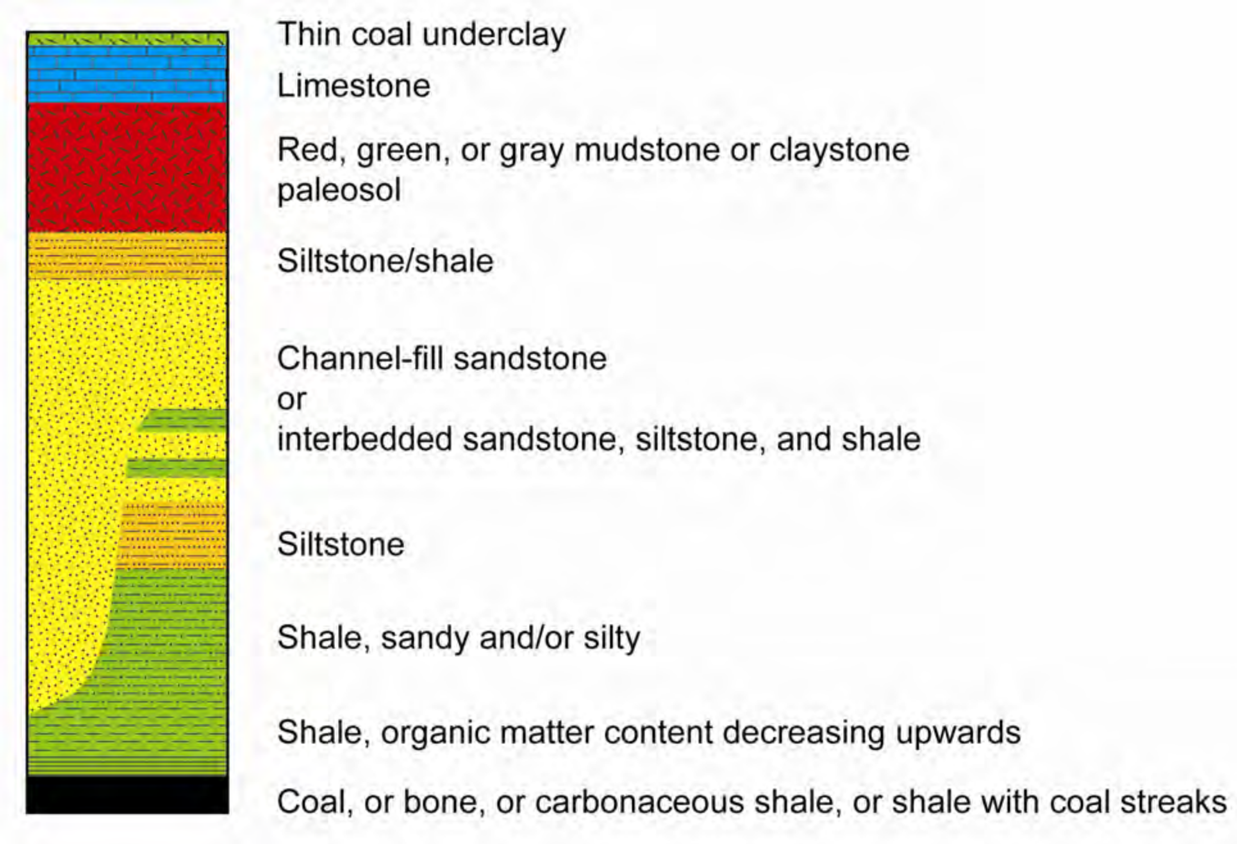


Figure 4. Idealized depositional cycle of the Dunkard Group (modified from Beerbower, 1961).

As throughout the underlying Pennsylvanian, these rocks were deposited cyclically. Beerbower (1961) described the rock types commonly found in the Dunkard Group arranged in a typical cycle or cyclothem (Figure 4). A coal bed marks the base of a cycle. The best development of coal is found in the lower part of the Dunkard Group with the Washington coal bed generally considered to be the youngest minable bed in the Appalachians. Nevertheless, resources in the Washington coal bed are economically marginal as historical mining has mainly taken place during periods of high demand and artificially high prices. The bed is high in ash content as a result of shale laminae within the coal, multiple and variable numbers of noncoal partings, and layers of impure coal. The arrangement of interbedded coal and shale is highly variable laterally.

Coals above the Washington are generally less than 3 ft (0.9 m) thick and are usually thinner, especially in the Greene Formation (Figure 5). The coals are usually high in ash yield and sulfur content (Eble et al, this volume). They have little or no economic value. These coal beds are frequently represented laterally by thin accumulations of bone coal, carbonaceous shale, shale with coal streaks, or dark gray to black shale (Figure 6). However, these coal beds and their horizons can be traced regionally. Dunkard Group coal beds are frequently overlain by shale, usually darkened with organic matter, which decreases with distance above the coal bed. Coal bed roof shale beds frequently contain fossil plant compressions and impressions, although plant fossils are also found in shale not associated with coal and more rarely in limestone beds (see Blake and Gillespie, 2011 for further information concerning the plant fossils and their significance.) Cross section A-A’, included on the CDROM of this guidebook, illustrates the lateral variability and continuity of the coal beds as well as the nature of the other lithologies.

Figure 5. The Upper Washington Limestone and Jollytown coal at Hundred, WV. The resistant shale bed overlying the Jollytown coal contains abundant fossil material including fish parts. Note the thin underclay beneath the coal and the shale overlying the "fish bed". Photo by Vik Skema.

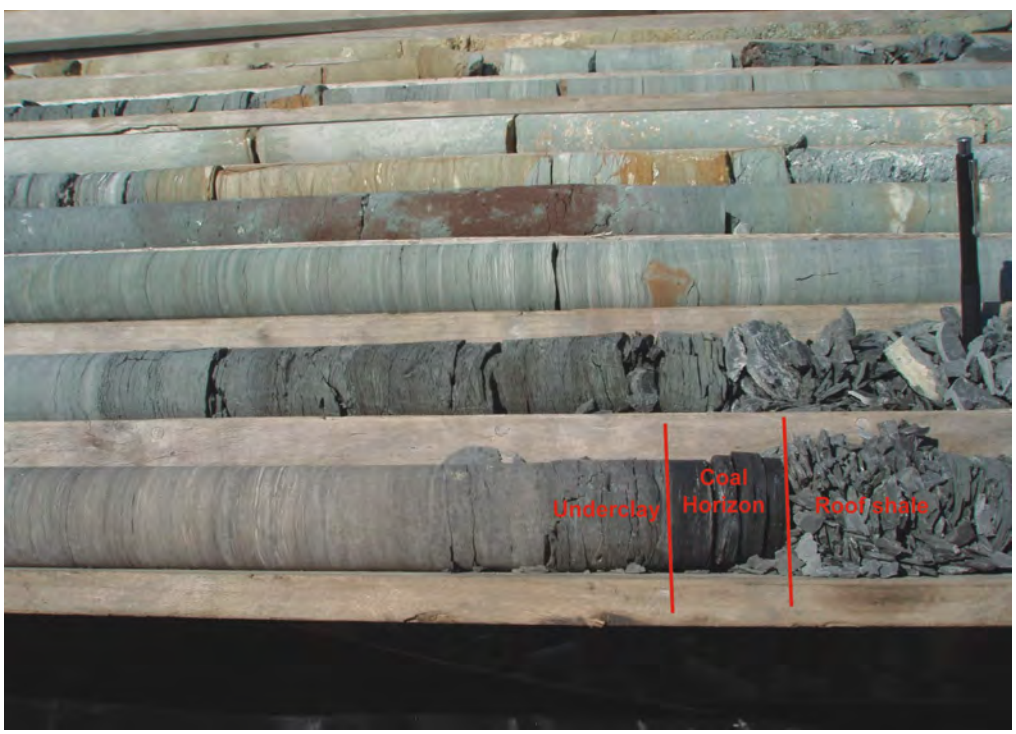


Figure 6. Organically enriched zone representing a Greene Formation coal bed about 400 ft (122 m) above the Washington coal bed. Photo taken in Greene County, PA, by William DiMichele.

Figure 7. Mitch Blake of the WVGES at Hundred, WV. The cut illustrates a cycle from the limestone up into the sandstone. Photo by Bill DiMichele or Vik Skema.

Figure 8. An example of a thick, multi-story channel sandstone in the Dunkard Group. This is the Hundred sandstone at the type locality about 1.5 mi (2.4 km) west of Hundred, WV on US Rt. 250. The people are (left to right) William Grady, Nick Fedorko, and Vik Skema. Photo by William DiMichele.

As the shale grades upward, it may become silty and sandy, may be lighter gray or green in color and grade into a siltstone or sandstone. Overlying units may be sequences of interbedded shale, siltstone and sandstone, representing landscape positions proximal to active streams (Figures 4 and 7). Multi-story, thick channel-form sandstone bodies with undulating, erosive bases may also directly overlie coal beds and roof shale (Figures 4 and 8). In some instances, streams cut through the coal and shale sequence, with the resulting sand body replacing those beds. Martin (1998) presented a comprehensive discussion of sandstone deposition in the Dunkard Group.

Sandstones may grade upward back into siltstone and gray, green, or red shale in response to waning energy gradients associated with meandering channel systems (Figures 4 and 9). In many instances, red, green or gray mudstone or claystone paleosols develop in this parent material. Calcareous nodules and disseminated calcareous material are common in these beds. Slickensides (vertic features) and redoximorphic mottling are also common, indicating periods of wetting and drying (Figures 4, 9, and 10).

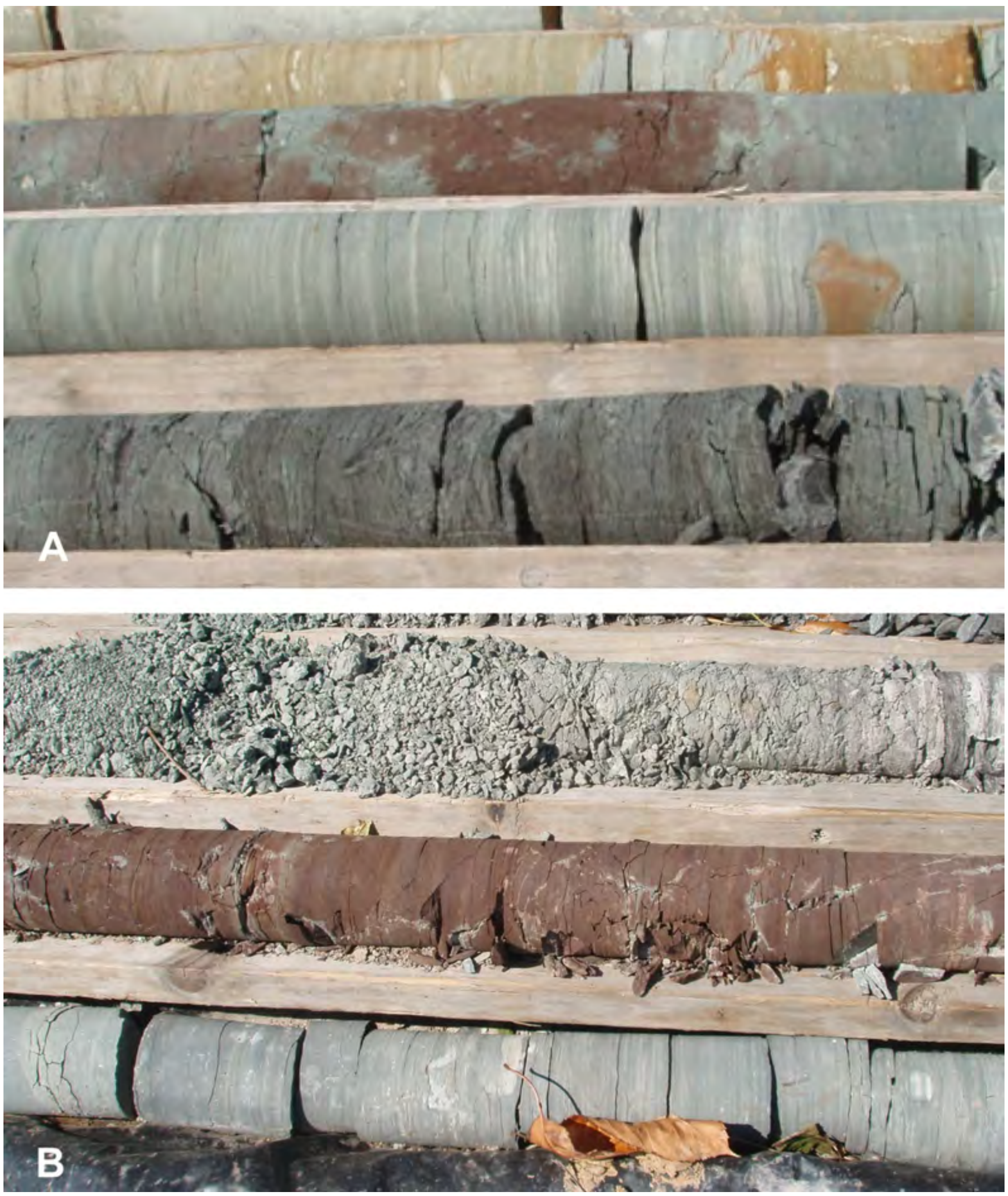


Figure 9. A – Core showing a gray-green shale with thin sandstone laminae (middle row). Bottom row shows a dark green-gray subfissile shale with faint reddish mottling. The third row up shows a red and gray[green mottled mudstone paleosol. B – Core of red (middle) and gray-green shale (bottom), common in the Dunkard Group. Note also the gray-green mudstone paleosol in the top row exhibiting typical angular blocky structure. Photos taken in Greene County, PA, by William DiMichele.

Figure 10. Coarse structured, gray-green mudstone paleosol beneath the Middle Rockport limestone. Note the slickensides, calcareous nodules, and coarse red mottles. Photo taken on Greathouse Hill Road near Wileyville, Wetzel County, WV by

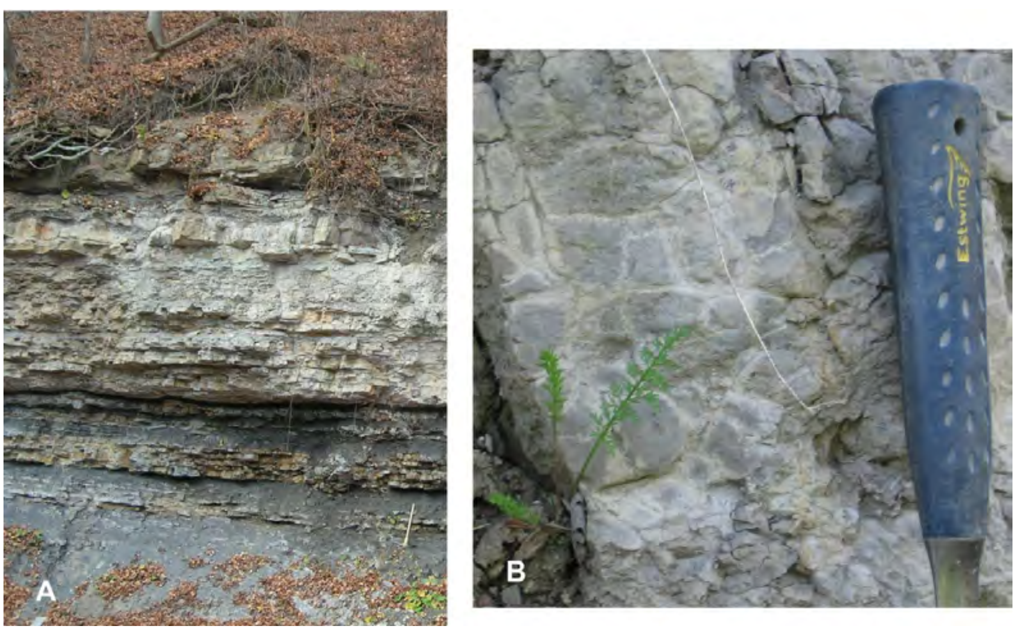


Figure 11. A – Typical sequence of limestone interbedded with shale and mudstone exhibited in the Lower Washington limestone on Roberts Ridge Road near Moundsville, Marshall County, WV. Note the 3-ft (0.9-m) long mattock for scale. B – Closeup of one of the beds in A showing brecciated/nodular fabric. Photos by Nick Fedorko.

Overlying these paleosols and underlying the start of the next cycle are nonmarine lacustrine limestone beds (Figure 4). Individual limestone beds rarely exceed 5 ft (1.5 m) in thickness and are typically micritic. They often exhibit features of subaerial exposure such as desiccation cracks and fractures and nodular or brecciated fabrics. Limestone beds are frequently interbedded with argillaceous limestone, calcareous mudstone, and calcareous shale, comprising limestone sequences (Figure 11) exceeding 5 ft (1.5 m) in thickness. Separating the limestone from the coal of the overlying sequence are thin mudstone or claystone seat earths, usually thin and poorly developed (Figures 4, 5, and 6).

Though cyclicity of coals and other lithologies is obvious throughout the Pennsylvanian and Permian strata of the Dunkard Basin, the arrangement of repeating lithologies is seldom as described in the typical cycle such as set forth by Beerbower (1961) (Figure 4). Transition between the various lithologic units of Dunkard Group cyclothems is often gradational, melding lithologies of adjacent units. As an example, the siliciclastic partings in coals coarsen and thicken with the introduction of fluvial silt and sand. The total thickness of the Washington coal complex increases from a minimum of 5 ft (1.5 m) in the northern end of the basin to a maximum of 36 ft (11 m) thick in north central Greene County, where sandstone interbeds are best developed (see descriptions for Stops 2 and 4, for more detailed information concerning the Washington coal complex).

The Waynesburg A coal sometimes has limestone above and below it, and in places even contains an ostracod-bearing limestone parting (see description for Stop 2), indicating that smaller cycles can occur within the larger cycle.

Variance from the typical cyclothem is evident from difference in position within the basin -scale depositional system and to differences induced locally by depositional environments. This is very evident in the case of the fluvial component of the cycle. Thick channel-form sandstone bodies are confined to linear belts. The areas of the basin farthest removed from these fluvial channels received only fine clay material from flooding events. For example, the Washington Formation in the extreme northern end of the basin clearly had a dominant lacustrine depositional environment for an extended time, barely showing any effects of fluvial processes (Figure 12).

Paleosols, so apparent where developed in the fine material at the top of waning fluvial deposits, are barely discernable in the lake deposits of the north (Figure 12). The paleosol beneath the Washington coal complex is one of the better developed and thickest soil profiles of the Dunkard Group, but in places in the lake environment of the north, it is very thin and weakly developed. In addition, in the northern, limestone-dominated section an ostracod-bearing limestone is frequently below the paleosol of the typical cycle (Figure 4). Cecil et al. (2011) discuss further the autocyclic and allocyclic controls on the deposition of cycles in the Dunkard Group.

Invertebrate fauna in the Dunkard Group is comprised mainly of ostracodes, *Estheria* (conchostracans), gastropods, pelecypods, and spirorbid worm tubes found in the shale and limestone. Fish remains, such as teeth, scales and bones, as well as remains of amphibians and tracks, and rare reptilian remains are also found (Cross and Schemel, 1956). All of these remains are considered nonmarine. Shale within the Washington coal bed has yielded fossil *Lingula* sp. in areas of eastern Ohio and in the bordering northern panhandle of West Virginia, the only hint of distant marine influence to be documented.

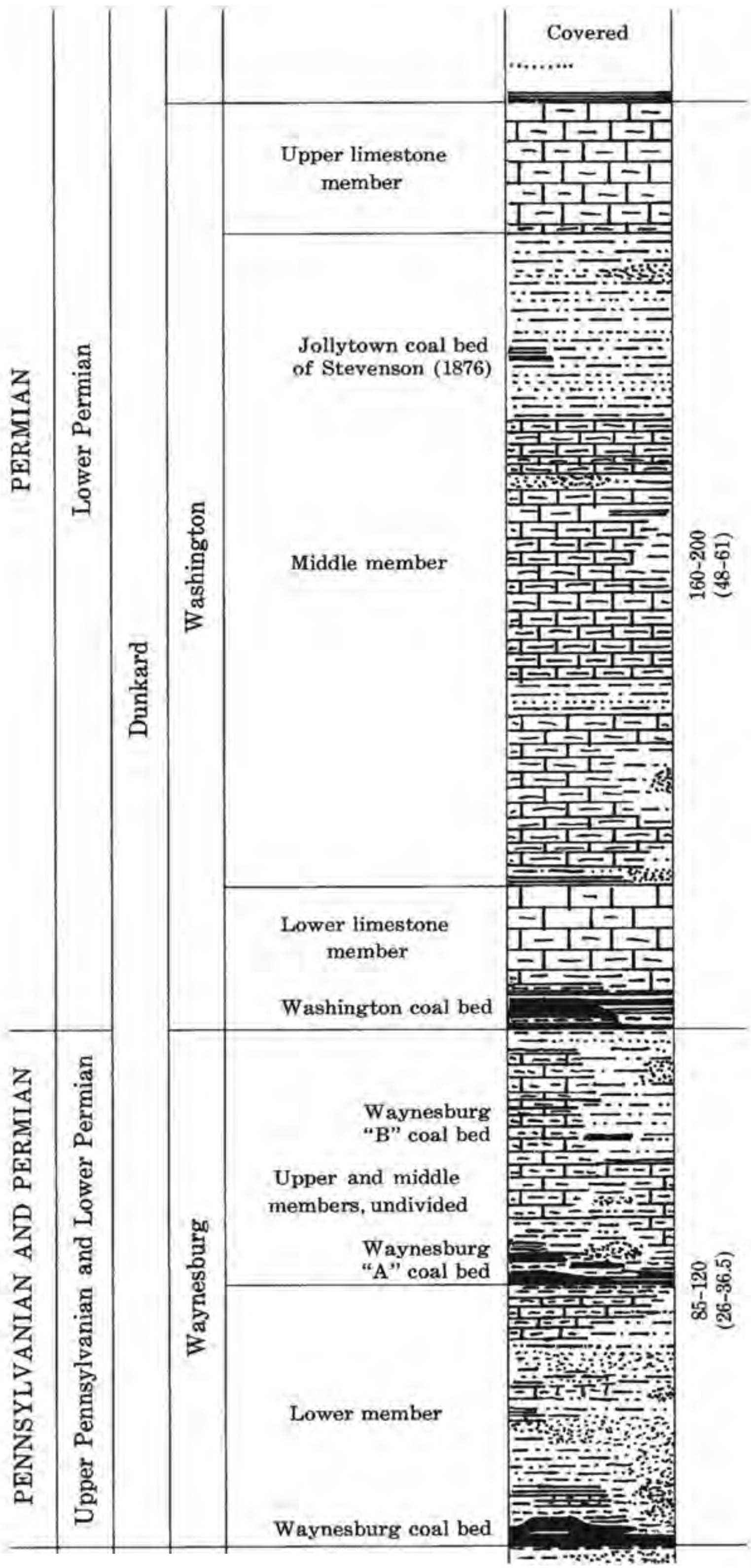


Figure 12. General section of the Dunkard Group in Washington County, PA (Schweinfurth, 1976) illustrating the predominance of limestone facies and the paucity of siliciclastic facies as compared to the same stratigraphic interval to the south.

Well documented features of the regional stratigraphy of the Dunkard Group and underlying Monongahela group are the north-to-south facies changes and section thickening (Stevenson, 1876; White, 1891 and 1903; Arkle, 1959, 1969, and 1974). Both sections of rock thicken southward and in the west-central area in West Virginia (Figure 1) and are dominated by sequences of interbedded red shale, green shale, red mudstone paleosols, siltstone, and sandstone (Figures 13 and 14). Limestone and coal beds are absent or rare, although the Rockports limestones were named from occurrences in the red facies area in west-central West Virginia. Calcareous material is present as nodules, streaks, and as fine material disseminated throughout units. Thicker red paleosols may be coeval with down-dip limestone sequences (Figure 14). Moving northward, coal beds and limestone appear in the sections, redbeds become fewer. At the northern extreme of the Dunkard Group in Washington County, Pennsylvania, redbeds are reportedly absent and limestone and coal beds are most abundant and well developed (Figure 12) (Stevenson, 1876). Figure 15, taken from cross section A-A' illustrates these south-to-north facies changes in the Dunkard and Monongahela groups.

Arkle (1959) described these facies as the gray (northern) and red (southern) and transitional (intertonguing area between the

Figure 13. Core of rocks of the Monongahela and Dunkard groups in the area of the "red facies". This is one of the core holes shown at the southern end of cross section A-A'. Photo taken in Jackson County, WV, by Nick Fedorko.

Figure 14. Roadcut illustrating "red facies" cycles in the Washington Formation comprised of sandstone, siltstone, red and green shale, and red-dominated mudstone paleosols. Coal and limestone are absent, except for the Washington(?) coal, which is present lower down south of this cut. Photo taken on new Rt. 50 bypass around Parkersburg, Wood County, WV at the exit ramp before crossing the Ohio River. Photo by Nick Fedorko.

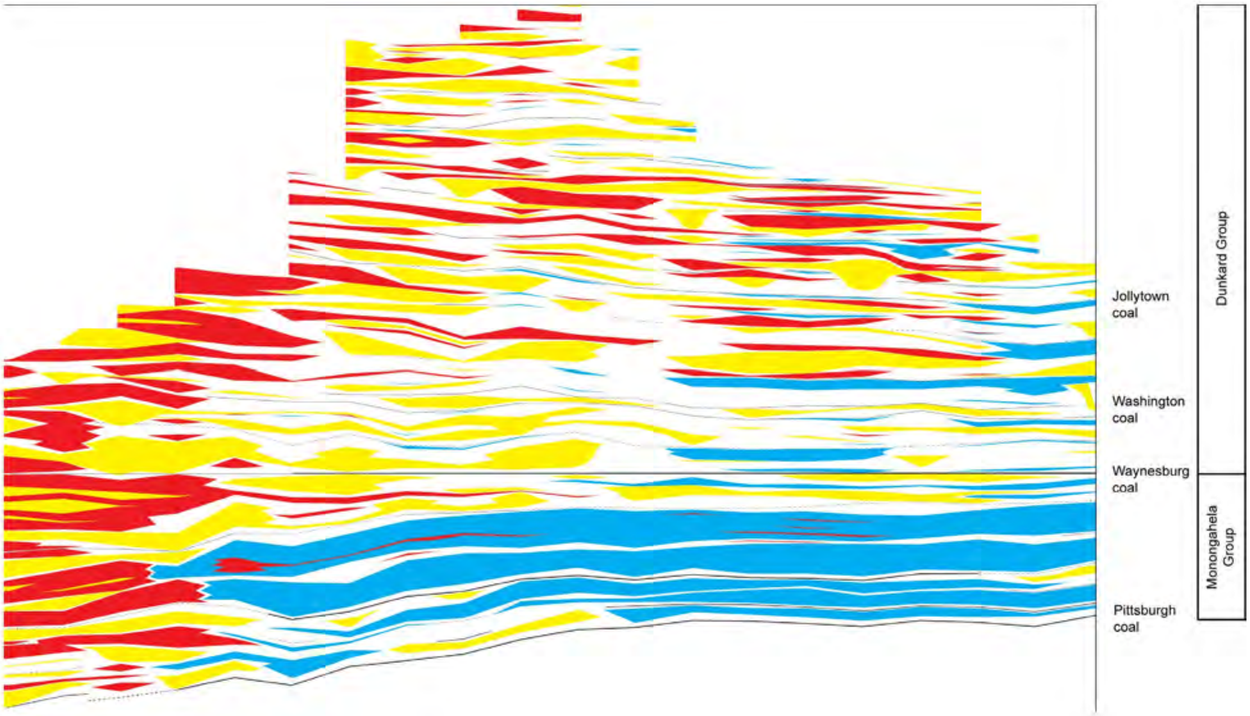


Figure 15. Distribution of facies in the Monongahela and Dunkard groups taken from cross section A-A'. Coal beds are black lines, limestone is blue, sandstone is yellow, redbeds are red, and white is shale.

gray and red facies). Later, Arkle (1969 and 1974) interpreted the gray facies as the lacustrine-swamp facies and the red facies as the alluvial facies. Martin (1998) identified these facies provinces (south to north) as upper fluvial plain, lower fluvial plain, and fluvial-lacustrine deltaic plain. Current indicators in sandstones and the arrangement of facies of the Dunkard Group indicate a north to northwest paleoslope with sediment source in the southeast with some evidence on the northeastern margin for subordinate sediment input from a northern, cratonic source (Martin, 1998). See Cecil et al. (2011) for discussion of modern analogues and the role of climate in the deposition of the Dunkard Group.

STRATIGRAPHIC NOMENCLATURE

The first investigation of these rocks was undertaken by Henry D. Rogers (1839) (Figure 16A). His geologic investigation to that time only reached as high as a coal bed and prominent sandstone exposed in the area of Waynesburg, Pennsylvania he identified as units XXVIII and XXX respectively, now known as the Waynesburg coal bed and Waynesburg Sandstone. He acknowledged that younger strata to the west had not yet been investigated. Rogers (1839) placed all of the strata from the level of the Ohio River at Pittsburgh (middle Conemaugh Group) to the highest known at that time in what he called the Pittsburg Series (Figure 17).

H.D. Rogers continued to investigate the geology of Pennsylvania and in 1858 created a 5-fold subdivision of the bituminous coal-measures of the state. The classification was based on shared lithological characteristics, particularly the presence or absence of merchantable coal beds. The strata from the *top* of the Waynesburg coal bed to the top of the highest exposures he called the Upper or Newer Coal-Shales or Upper Barren Group, reflecting the paucity of coal and the lack of merchantable coal beds. He estimated that 900 to 1000 ft (274 to 305 m) of

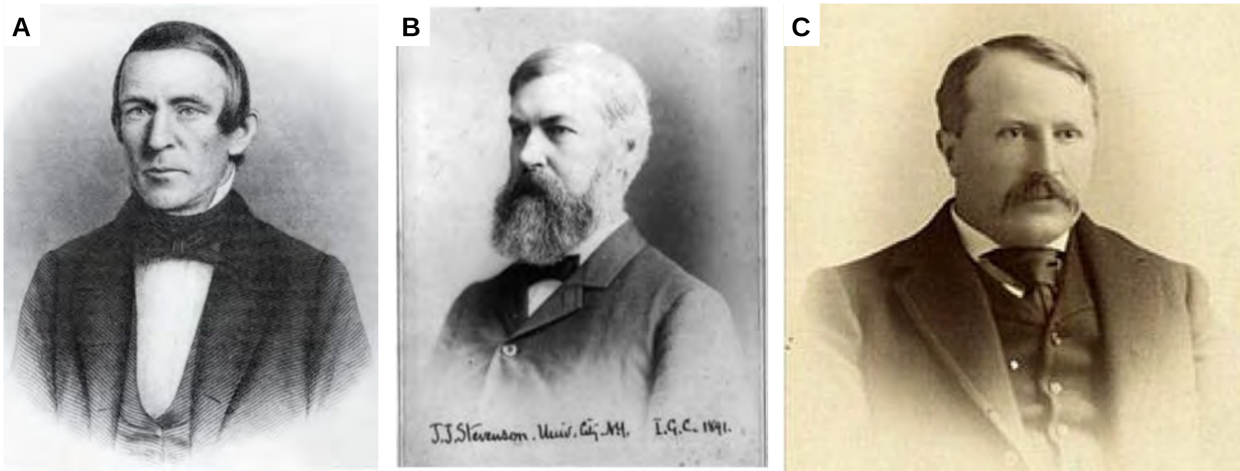


Figure 16. Photographs of the pioneers who first mapped and described the geology of southwestern Pennsylvania and adjacent areas. A—Henry Darwin Rogers, first State Geologist of Pennsylvania and director of the First Pennsylvania Geological Survey. B—John James Stevenson, Professor of Geology and Natural History at New York University and principal investigator of the bituminous coal fields of western Pennsylvania for the Second Geological Survey of Pennsylvania. C—Israel Charles White, assistant to Professor Stevenson for the Second Geological Survey of Pennsylvania and first State Geologist of West Virginia.

these strata occurred in Greene County, Pennsylvania, including 4 or 5 thin coal beds seldom exceeding 2 ft (0.6 m) in thickness, a number of thin limestone beds, 2 to 5 ft (0.6 to 1.5 m) thick, with the bulk of the rocks being sandy shales and flaggy, micaceous sandstones. He did not name any of the units. More detailed sections eluded Rogers due to meager exposures. Rogers (1858) was the first to speculate that these rocks might be Permian-age:

> The last and highest rock observed is a mass of grey and buff micaceous sandstone, 110 feet thick, occupying the highest ground between the head-waters of Ten-mile and Wheeling creeks. Here, if anywhere in the Great Appalachian Coal-Field, the geologist should institute a search for Permean fossils identical with the species lately found in Illinois and Kansas, above the productive Coal-measures.

Geologic investigations in western Pennsylvania were continued by John J. Stevenson (Figure 16B) under what is known as the Second Geological Survey of Pennsylvania. J.J. Stevenson authored Report K, entitled *Report of Progress in the Greene and Washington Districts of the Bituminous Coal-Fields of Western Pennsylvania*, in 1876. In the intervening years since H.D. Rogers' work, Stevenson, aided in the field by a young geologist named Israel Charles White, scoured the district for insight into the youngest rocks now known as the Dunkard Group. Stevenson (1876) used the 5-fold classification system introduced by Rogers (1858) with modification, assigning the name *Upper Barren Series* to these rocks and moving the lower boundary upward from the top of the Waynesburg coal bed to the top of the Waynesburg Sandstone. He further divided the series into the *Washington County Group* and *Greene County Group.* The Washington County Group extended from the top of the Waynesburg Sandstone to the top of the Upper Washington Limestone while the Greene County Group included all the rocks above this limestone (Figure 17).

Stevenson (1876) and his field aide I.C. White gained considerable knowledge of the

Upper Barren Series stratigraphy. Stevenson (1876) named numerous coal beds, a few sandstone beds, a few limestone beds, and assigned roman numerals to 14 limestone beds. In the Washington County Group he named (ascending) the Waynesburg A coal bed, Limestone Ia, the Waynesburg B coal bed, Limestone Ib, the Little Washington coal bed, the Washington Sandstone, the Washington coal bed, the Lower Washington Limestone (Limestone II), Limestone III, the Washington A coal bed, Middle Washington Limestone (Limestone IV), the Jolleytown (sic) coal bed, Limestone V, and the Upper Washington Limestone (Limestone VI). In the Greene County Group he named Limestones VII, VIII, IXa, IXb, the Dunkard coal bed, the Fish Creek Sandstone, Limestone X, the Nineveh coal bed, Limestones XI, XII, and XIII, the Gilmore Sandstone, and Limestone XIV. In addition, he recognized an unnamed coal bed below Limestone VIII and documented coal locally in the place of bituminous shales at several positions in the Greene County Group.

Stevenson (1876) also noted the occurrence of "fish beds", i.e. thins beds of carbonaceous shale containing abundant fossil material comprised of fish parts, small crustaceans, and frequently plant fossils on top of two limestones, an observation of stratigraphic significance. He reported one such bed as occurring above the Upper Washington Limestone and, where found about 1 mi (1.6 km) above the village of Jollytown as being so carbonaceous as to form a thin coal bed. He likewise noted such a fish bed on top of the Lower Washington Limestone that also sometimes formed a thin coal bed. Lack of exposures, especially those of significant stratigraphic extent frustrated and hampered the investigation.

I.C. White (Figure 16C) left the Pennsylvania Geological Survey and returned to his native West Virginia. In 1891, White authored *Stratigraphy of the Bituminous Coal Field in Pennsylvania, Ohio, and West Virginia* published by the U.S. Geological Survey as Bulletin 65. White (1891) renamed the Upper Barren Series (Stevenson, 1876) the *Dunkard Creek Series* for exposures along that stream flowing along the Pennsylvania-West Virginia border eastward into the Monongahela River a few miles north of the state line. He moved the boundary back down to the top of the Waynesburg coal bed where Rogers had placed it in 1858. White also placed the Pennsylvanian-Permian boundary at the same position based on the Permian affinities of plant fossils found in the Waynesburg coal roof shale as reported on by Fontaine and I.C. White in 1880. White (1891) did not use or discuss the division of these rocks into the Washington County and Greene County groups as set forth by Stevenson (1876) (Figure 17).

White (1891) provided additional detail on the stratigraphy of these beds, gave names to additional units, and gave geographic names to many of the limestone beds assigned roman numerals by Stevenson (1876). The new units named by White (ascending) were as follows with the equivalent limestone of Stevenson (1876) in parentheses: Cassville Plant Shale (Waynesburg coal roof shale), Mount Morris Limestone (no equivalent), Colvin's Run Limestone (Ia), Blacksville Limestone (III), Marietta Sandstones, Jollytown Limestone (VII?), Hostetter coal, Nineveh Limestone (X), Nineveh Sandstone, Windy Gap coal bed, and Windy Gap Limestone (XIV). White's stratigraphy contradicted Stevenson's in places, for example, White placed the Jollytown coal bed on top of the Upper Washington Limestone rather than between it and the Middle Washington Limestone. This issue and others will be discussed in a later section.

White (1903) in a report on coal in West Virginia again reviewed the general stratigraphy of the youngest rocks of the Appalachian basin and now referred to them as the *Dunkard Series*, dropping the word "creek" from the name (Figure 17). No new named units or insight into the

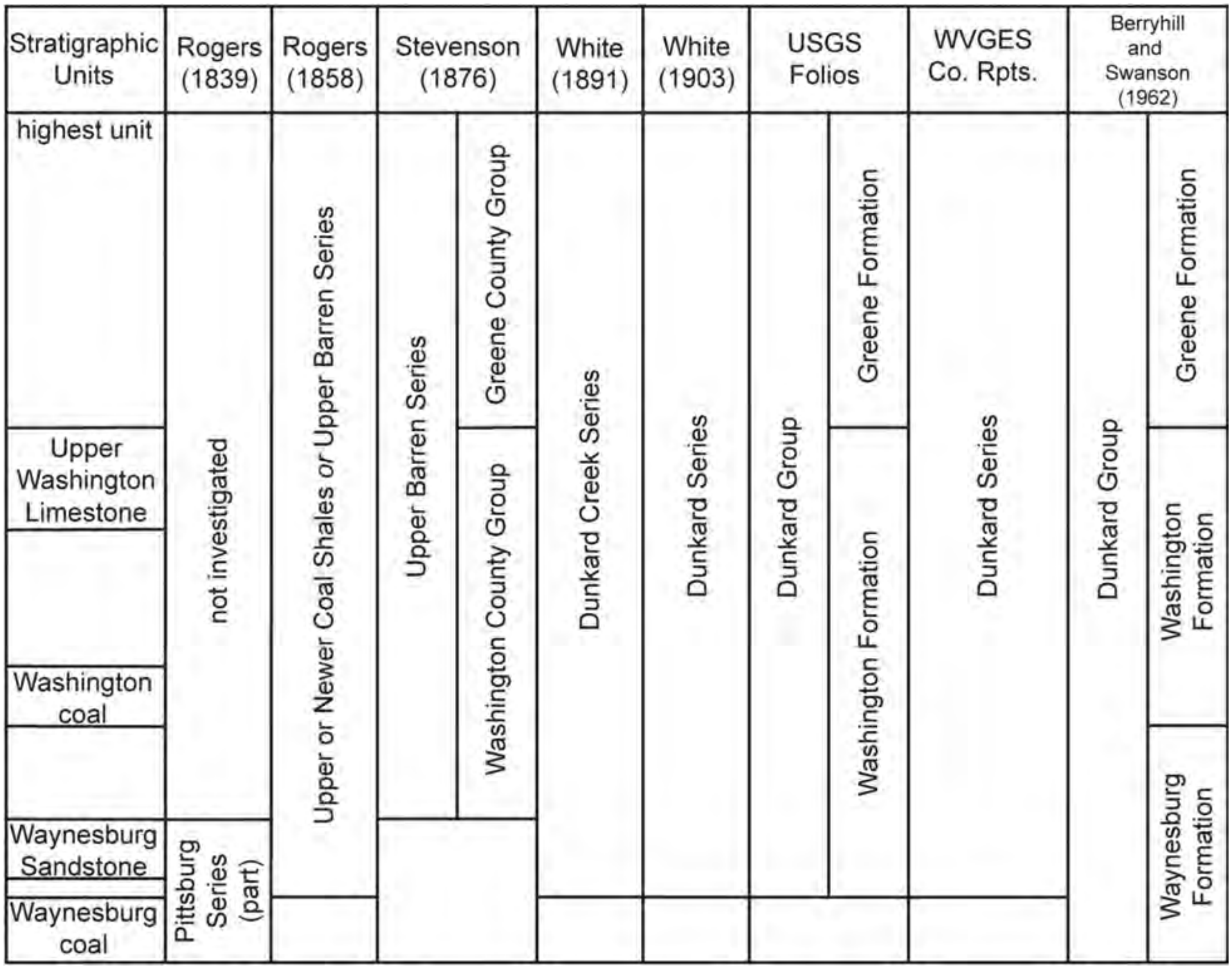


Figure 17. Summary of the nomenclature historically applied to the rocks above the Waynesburg coal bed, now the Dunkard Group.

stratigraphy of the beds is offered in this report. Inexplicably, White (1908) reverted to the name Dunkard Creek Series when discussing the coal resources of these rocks in a supplemental coal report.

In the early twentieth century, geologists of the United States Geological Survey began producing geologic folios of 15-minute topographic quadrangle areas in Pennsylvania. The first of these relevant to the Dunkard Group, was the geologic folio of the Waynesburg Quadrangle, authored by Ralph W. Stone and published in 1905. Stone (1905) followed the lead of White (1891) in using the name *Dunkard*, but changed it to the *Dunkard Group*. He also adopted Stevenson's 1876 divisions of the Dunkard, but dropped the word 'county' and elevated them to formation status so the Dunkard was now divided into the Washington and Greene formations. He also adopted the lower boundary of Rogers (1858) and White (1891) placing the base of the Dunkard Group and the Washington Formation at the top of the Waynesburg coal bed. The boundary between the Washington and Greene formations remained at the top of the Upper Washington Limestone, as Stevenson (1876) had defined (Figure 17). This same nomenclature was followed by the authors of geologic folios published after 1905.

Stone (1905) did not introduce any new names for stratigraphic units, although he did recognize the discrepancy in the position of the Jollytown coal bed as used by Stevenson (1876) and White (1891), using the name as defined by Stevenson (1876).

The Waynesburg geologic folio was followed by publication in 1907 of the Amity Quadrangle geologic folio, authored by Frederick G. Clapp. Like Stone (1905), Clapp used the name Jollytown for a coal bed below the Upper Washington Limestone, sensu Stevenson (1876). Importantly, he observed a fish bed above the Upper Washington Limestone on Cemetery Hill near Washington, PA accompanied by an 18-in thick coal bed 5 ft (1.5 m) above the Upper Washington Limestone. He did not name this coal bed. Clapp (1907a) did name a coal bed found about 30 ft (9 m) above the Upper Washington Limestone the *Tenmile* for exposures in the vicinity of Tenmile Creek. He stated that this is the bed referred to as Jollytown by White (1891). In 1907 Clapp also authored the Rogersville Quadrangle geologic folio, covering much of the southwest corner of Greene County, PA. In this folio, Clapp (1907b) states that the Tenmile coal bed is 50 to 100 ft (15 to 31 m) above the Upper Washington Limestone.

The Claysville Quadrangle geologic folio, north of the Rogersville Quadrangle and largely in Washington County, PA was authored by M.J. Munn in 1912. Like Stone (1905) and Clapp (1907a and b), Munn used the name Jollytown for the coal as defined by Stevenson (1876). Munn (1912) also observed bituminous shale with thin layers of coal 3 to 15 ft (0.9 to 4.6 m) above the Upper Washington Limestone, which he referred to as the *Upper Washington* coal bed, originally named by Stevenson (1907), and indicated it was correlative with the Jollytown coal of White (1891). Munn placed the Tenmile coal bed of Clapp (1907a) 35 to 70 ft (11 to 21 m) above the Upper Washington Limestone and 15 to 30 ft (4.6 to 9 m) above what he named the Donley Limestone from the village of Donley, PA.

I.C. White was instrumental in establishing the West Virginia Geological and Economic Survey in 1897 and was named the first State Geologist. Publication of various geologic reports, some already cited, followed soon after. Important to this discussion is the publication of reports on the geology of a county or group of counties, commonly referred to in shorthand as the *County Reports*. The reports relevant to the Dunkard Group are, in chronological order, *Ohio, Brooke, and Hancock Counties*, Grimsley (1907), *Marshall, Wetzel, and Tyler*, Hennen (1909), and *Marion, Monongalia, and Taylor*, Hennen and Reger (1913). All of these reports use *Dunkard Series* for these rocks. These reports and others published later introduced many additional stratigraphic names in the Dunkard Group and are the source for many of those presented in Figure 3 that have not already attributed.

Lastly, Berryhill and Swanson (1962) authored a paper entitled *Revised Stratigraphic Nomenclature for Upper Pennsylvanian and Lower Permian Rocks, Washington County, Pennsylvania*. In this report, Berryhill and Swanson retained the name Dunkard Group but moved the lower boundary to the *base* of the Waynesburg coal bed. They did this to conform to the thinking about cyclothems at that time. They argued that coals in this part of the section in southwestern Pennsylvania were for the most part the only key mapping beds and that “the basic mapping unit for field classification, therefore, includes the several rock types between coal beds, and the basic unit represents a sedimentary cycle”. They gave these units the status of members, and since …..”the basal unit of a member as here defined is a coal bed, it follows that the basal unit of a formation must also be a coal bed” (Berryhill and Swanson, 1962). Formation boundaries were placed at the base of the major coal beds, with the coal name being assigned to the formation. Following this principle, the rocks from the base of the Waynesburg coal bed to the base of the Washington coal bed were named the *Waynesburg Formation*. They then redefined the Washington Formation as extending from the base of the Washington coal bed to the top of the Upper limestone member (Upper Washington Limestone)--not a coal, the one exception to their rule, and retained the Greene Formation as previously defined (Figure 17).

Berryhill and Swanson (1962) considered the use of geographic place names for numerous units of different lithology confusing and in violation of the code of stratigraphic nomenclature, (the name Waynesburg has been used for six separate stratigraphic units, three of which are beds having different lithologies), so for the Dunkard Group they only retained coal bed names and then further divided the formations into informal lower, middle and upper members (e.g. Figure 12). The group and formation stratigraphic nomenclature of Berryhill and Swanson (1962) has been adopted in Pennsylvania but not in West Virginia. Among geologists working in this section in both states, the names of many individual stratigraphic units are still commonly used.

STRATIGRAPHIC NAME CONFLICTS

In the course of working with the Dunkard Group rocks and reviewing the literature, some differences in how names have been applied to some units in West Virginia and Pennsylvania have been recognized. The first of these to be discussed is the Jollytown coal bed.

The Jollytown Coal Bed

As has been noted in the previous discussions, the application of the name Jollytown to a coal bed in the Dunkard Group has been problematic. The name came from exposures of a coal in the vicinity of the village of Jollytown, in southeastern Gilmore Township, Greene County, Pennsylvania. Stevenson (1876) first used the name, although he spelled it *Jolleytown.* He stated (Stevenson, 1876, p. 48) that the coal occurs 20 to 75 ft (6 to 23 m) *below* (author emphasis) the Upper Washington Limestone "which Mr. White [I.C. White], in his notes on Gilmore Township of Greene County, has named the *Jolleytown coal*."

In the section on the geology of Gilmore Township, Stevenson (1876, p 111) wrote, "At Jolleytown, just below the forks of the stream, the *Jolleytown coal* is seen crossing the road at the upper end of the village near the hotel." In previous paragraphs, Stevenson (1876, pp. 110-11) described an exposure at the mouth of Negro Run, an estimated 0.9 mi (1.5 km) northwest of the previous exposure, that showed 4 ft (1.2 m) of Upper Washington Limestone overlain by 6 in (15 cm) of fossiliferous black shale (fish bed) which in turn is overlain by 13 in (33 cm) of coal, but he did not apply a name to this bed. However, in 1891, White (p. 35) referred to the same Negro Run section and to the 13-in (33-cm) thick coal as the Jollytown coal bed. Hennen (1909, p. 197) showed a section of 11 in (28 cm) of Jollytown coal underlain by 6 in (15 cm) of shale and then 3 ft (0.9 m) of Upper Washington Limestone exposed in the bed of Town Run in the village of Jollytown east of both of the above sections (Figure 18).

Clendening (1974) visited Jollytown, Pennsylvania for the purpose of sampling the Jollytown coal bed as part of a palynological study of the Dunkard Group. He reported that the Jollytown coal bed in Town Run was 24 ft (7 m) above the culvert under the main road, which the Wadestown 7.5-minute topographic map shows to be at estimated elevation 1005 ft (306

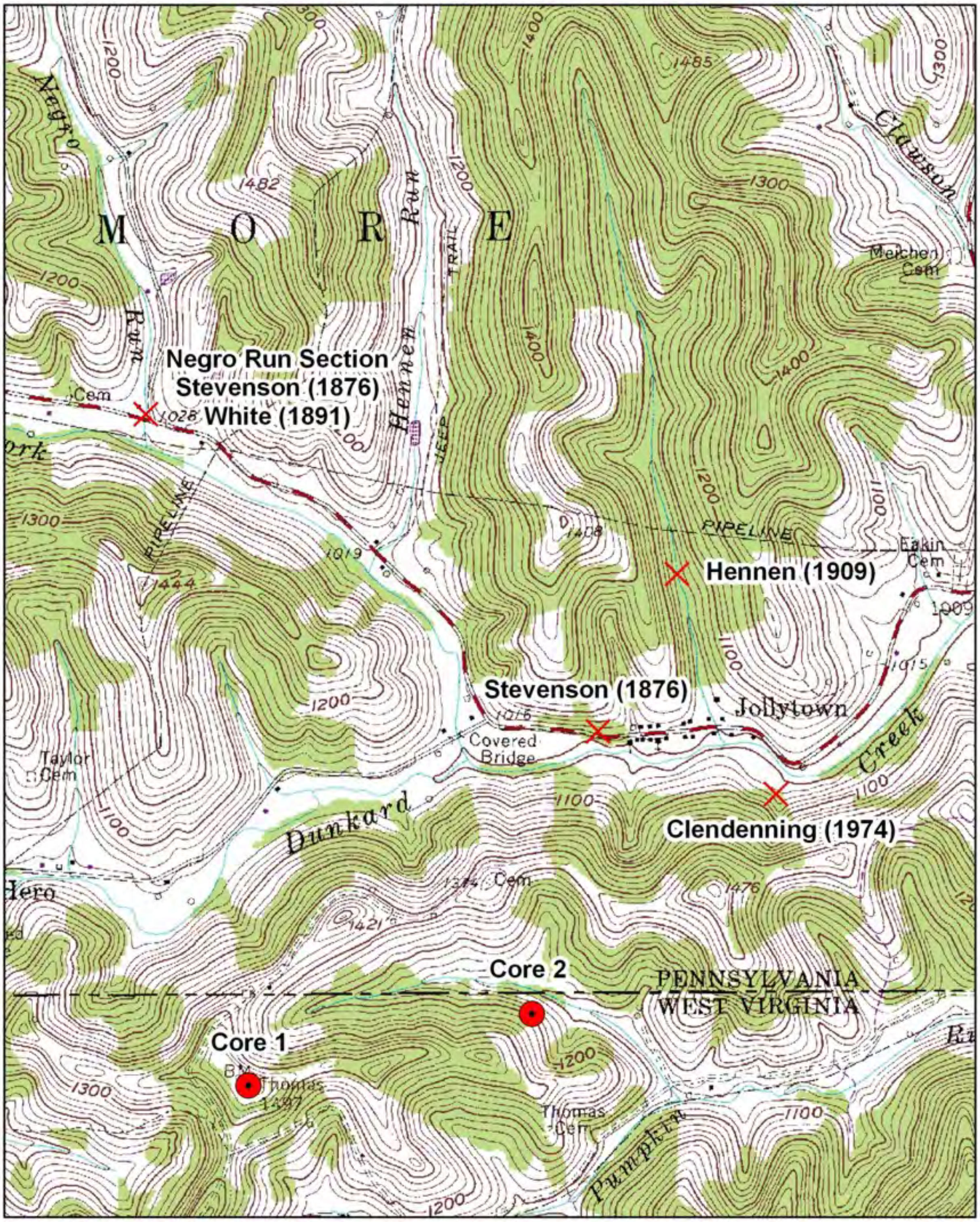


Figure 18. Jollytown, PA area showing the locations of Jollytown coal sections and core holes penetrating the Jollytown coal. Wadestown 7.5-minute topographic quadrangle. Scale 1:18,000.

m), yielding an estimated elevation of the Jollytown coal of 1029 ft (314 m). The road at the upper end of the village as described by Stevenson (1876) runs just above the 1020-ft (311-m) topographic contour. The road junction at Negro Run is at elevation 1028, but the relationship of the coal to this elevation is unknown In addition, Clendening (1974) examined the Jollytown coal bed 900 ft (274 m) southeast of the junction of Town Run and the road on the opposite side of the Pennsylvania Fork of Dunkard Creek and reported that it occurs 30 ft (9 m) above the floodplain or an estimated elevation of 1025 to 1030 ft (312 to 314 m) (Figure 18). The evidence is convincing that the coals noted in all of the above exposures are one-and-the-same and are by definition, the Jollytown coal bed.

Core records support the conclusion that the coal cropping out at Jollytown Pennsylvania and referred to as the Jollytown coal bed by both Stevenson (1876) and White (1891) *is* above the Upper Washington Limestone, not below it. Core holes 1 and 2, drilled south of Jollytown in West Virginia (Figure 18) indicate the presence of a coal on top of a limestone at elevations of 1010 and 1012 ft (308 to 309 m) respectively (Figure 19) which fit the descriptions of the Jollytown coal and Upper Washington Limestone as reported by both White (1891) and Hennen (1909). The difference in elevations reported above is readily attributable to dip. In addition, the cores clearly show that these are the only coal and limestone present within the interval that would be exposed in Jollytown and the thickness of the coal in core 1 correlates well with the thicknesses reported by White (1891) and Hennen (1909). The cores also establish the relationship of the Jollytown coal bed and underlying Upper Washington Limestone to under- and overlying key beds.

The Jollytown Coal of Stevenson (1876)

A coal bed or equivalent carbonaceous facies does occur in the stratigraphic position below the Upper Washington Limestone and above the Middle Washington Limestone (see cross section A-A') described by Stevenson (1876) as the Jollytown coal at Jollytown, Pennsylvania. For that reason, we refer to this coal as the *Jollytown coal of Stevenson (1876)*. Core log 1 (Figure 19) shows the horizon of the Jollytown coal of Stevenson (1876) represented by a thin shale with coal layers about 80 ft (24 m) below the Upper Washington Limestone. This horizon is absent in core 2. Stevenson (1907) continued to refer to this coal as Jollytown and described it as being 25 to 40 ft (7.6 to 12 m) below the *Franklin Limestone* (Limestone V of Stevenson, 1876). His mistake is understandable, since the coal below the Upper Washington Limestone is usually present throughout much of Greene and Washington Counties, whereas the coal directly above the limestone is spotty and often no more than black shale except in extreme southern Greene County and the adjacent part of West Virginia.

In West Virginia, the name *Jollytown A* has been used for this bed coal, possibly stemming from use by Cross and Arkle (1951) and Cross and Schemel (1956). The name Jollytown A was first used by Stauffer and Schroyer (1920) in Ohio for the coal above the Upper Washington Limestone and stated by the authors to be the Jollytown coal of White (1891). They used Jollytown A for this coal specifically to distinguish it from the underlying Jollytown coal of Stevenson (1876) and eliminate confusion. By using the name Jollytown A for the Jollytown coal of Stevenson (1876) in West Virginia, Cross and Arkle were perhaps implying that Stauffer and Schroyer (1920) were incorrect in correlating this coal in Ohio with the Jollytown coal of White (1891), although this is not implicitly stated.

Figure 19. Core 1 and core 2 showing the Jollytown coal and the Upper Washington limestone and the stratigraphic relationships to over- and underlying key beds just south of Jollytown, PA (see Figure 18). Geologist's core logs courtesy of Consol Energy.

The Hundred Coal

Another thin, discontinuous coal bed occurs in the interval below the Upper Washington Limestone and above the Jollytown coal of Stevenson (Figure 3). The coal was named the *Hundred* coal bed by Hennen (1909) from exposures about 1.5 mi (2.4 km) west of Hundred, West Virginia. Beneath the Upper Washington Limestone a thick, channel-fill sandstone found in this area was named the Hundred Sandstone by Hennen (1909). A short distance below this bed is the Hundred coal bed. Interestingly, Hennen (1909) noted that this was the only place this coal bed was observed. Cross section A-A' bears this out, as the horizon of this coal bed represented by carbonaceous material is present only in holes 15 and 16.

The Washington A Coal and the Middle Washington Limestone

Stevenson (1876) recognized three limestone beds between the Washington coal bed and the coal bed he called the Jollytown. These are (ascending) the Lower Washington Limestone (Limestone II), Limestone III, and the Middle Washington Limestone (Limestone IV). Stevenson (1876) named a coal above Limestone III and separated by shale the *Washington "a"* coal bed, now referred to as the Washington A coal bed. Sections on pp. 104, 105, and 107 (Stevenson 1876) show the Washington A coal to be 55 to 75 ft (17 to 23 m) above the Washington coal bed.

White (1891) named Limestone III of Stevenson (1876) the Blacksville Limestone from exposures in Blacksville, Monongalia County, West Virginia. He indicated that the Washington A coal is 70 to 80 ft (21 to 24 m) above the Washington coal bed, presumably above the Blacksville Limestone but his section on p. 102 does not show the Blacksville Limestone , but rather 60 ft (18 m) of shale and sandstone between the Lower Washington Limestone and the coal bed. Stevenson (1907) did place the Washington A coal bed in the interval between the Blacksville and Middle Washington limestones. Grimsley (1907) cited White's section of 1891 for the stratigraphy of the Dunkard Group. Hennen (1909) also cited White's section of 1891, but inexplicably in subsequent measured sections for the report area and in the descending list of stratigraphic units, he showed the Washington A coal bed above the Middle Washington Limestone, although at an interval consistent with White (1891). Hennen and Reger (1913) in a general section for the Dunkard Group (p.p. 165-166) showed the Washington A coal bed to be above the Middle Washington Limestone and 107 ft (33 m) above the Washington coal bed, suggesting that these authors were possibly using the name for the Jollytown coal of Stevenson (1876). Stauffer and Shroyer (1920) speculated this as well. The name Blacksville Limestone does not appear in Grimsley (1907), Hennen (1909), Hennen and Reger (1913), or in any other subsequent work in West Virginia.

Presently in West Virginia, the Washington A coal bed is considered to be above the Middle Washington Limestone following Hennen (1909) and Hennen and Reger (1913) as indicated on Figure 3 and on cross section A-A'. However, this may be in error, and as a limestone bed correlative with Limestone III of Stevenson (1876) or the Blacksville Limestone of White (1891) does occur below this coal and a limestone bed correlative with the Middle Washington Limestone of Stevenson (1876) and White (1891) does occur below the Jollytown coal of Stevenson (1876).

The Tenmile Coal

As related previously, Clapp (1907a) named a coal bed found about 30 ft (9 m) above the Upper Washington Limestone the *Tenmile* for exposures in the vicinity of Tenmile Creek. However he was mistaken in asserting that this is correlative with the bed referred to as Jollytown by White (1891). Cross section A-A' illustrates that the Tenmile coal bed is a separate and distinct bed found a short distance above the Jollytown coal bed.

The Dunkard Coal

Stevenson (1876) first used the name *Dunkard* for a coal bed in the Dunkard Group, however he reported rather divergent stratigraphic positions for the bed. In his general section for the Greene County Group (Greene Formation) taken from Centre Township (Stevenson, 1876, p. 35) he identified a coal bed about 180 ft (55 m) above the Upper Washington Limestone as the Dunkard coal. In this same section, he showed an unnamed, "local" coal about 65 ft (20 m) above the Upper Washington Limestone. In a section in southwestern Greene County (Stevenson, 1876, p. 36) he identified a coal about 110 ft (24 m) above the Upper Washington Limestone as the Dunkard coal, while leaving coal beds at about 45 and 187 ft (14 and 57 m) above the Upper Washington Limestone unnamed. Finally, in a section west of Jollytown Stevenson (1876, p. 36) identified a coal 65 ft (20 m) above the Upper Washington Limestone as the Dunkard coal bed. In current useage, this latter coal bed is identified as the Dunkard as illustrated in cross section A-A'. The coal bed Clapp (1907b) identified as the Tenmile in the Rogersville 15-minute quadrangle is probably the Dunkard coal of current useage. Likewise, the coal bed 35 to 75 ft (11 to 23 m) above the Upper Washington Limestone that Munn (1912) identified as the Tenmile may also be the Dunkard coal bed of current useage.

CONCLUSIONS

The stratigraphic names applied to beds above the Dunkard coal bed as shown on Figure 3 and cross section A-A' have not been reconciled with the definitions and descriptions of the original workers but have been applied as they have been used by the investigators in the intervening years. There is no doubt that additional conflicts in applications of names to beds will occur in the future considering the lack of surface exposure of the Dunkard Group rocks, especially of long, continuous sections. This is especially true of the Greene Formation rocks. This condition continues to frustrate geologists of today as much as it did those of the past. Closely spaced core records are the best hope of better understanding the stratigraphy of the Dunkard. Access to those cores and core logs depends on the good will and cooperation of those in the energy industry, like Consol Energy who supplied many of the records shown in cross section A-A', some used in Figure 2, and both those used in Figure 19. For that, we are very appreciative.

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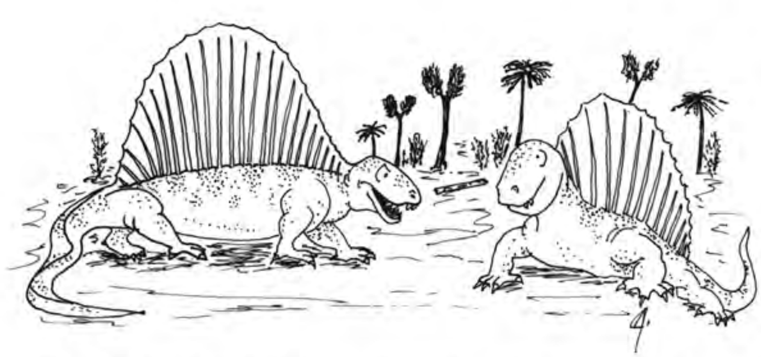
GREAT MOMENTS IN GEOLOGIC HISTORY
Part 12: The Permian



I'm tellin' ya, Sid, the way things have been goin' in the Paleozoic, I'm glad I won't be around for the Mesozoic!

AUTOCYCLIC AND ALLOCYCLIC CONTROLS ON THE ORIGIN OF THE DUNKARD GROUP

Cecil, C. Blaine, William DiMichele, Nick Fedorko, and Vik Skema

INTRODUCTION

Beerbower (1964) developed the concepts of autocyclic and allocyclic controls on sedimentation and stratigraphy on the basis of his work on the origin of the Dunkard Group and the voluminous previous work on the origin of Pennsylvanian cyclothems (Beerbower, 1961). According to Beerbower (1964) autocyclic processes include the redistribution of energy and materials within a sedimentary system such as stream meandering, channel avulsion, delta switching, etc. In contrast, allocyclic processes include changes in energy and materials within a sedimentary system induced by processes external to the sedimentary system. Allocyclic processes include eustatic, tectonic, and climatic change. Beerbower (1969, p. 1843) summarized his concepts as follows:

> In a general sedimentary model, cyclic deposits may be regarded as autocyclic or allocyclic (Beerbower, 1964, p. 32). The former are generated changes in sedimentary environment inherent in the sedimentation process, for example, delta switching. The latter are independent of particular depositional events, are generated outside the depositional unit and include tectonic, eustatic, and climatic cycles. Therefore, full interpretation of an alluvial cyclic deposit requires separation of autocyclic and allocyclic phenomena and isolation of the several particular causes of cyclicity.

The concepts of autocyclic and allocyclic processes are perhaps the most powerful diagnostic methods available to stratigraphers and sedimentologists for the analysis of the origin of sedimentary rocks because these concepts provide a comprehensive, integrated, diagnostic, analytical framework (e.g., Busch and Rollins, 1984). Sequence stratigraphy, which has come into vogue amongst stratigraphers during the past two decades, evaluates the eustatic variable of allocyclic analysis while overlooking climatic and tectonic change as well as autocyclic processes. An example of the short comings of a single variable sequence stratigraphic model can be illustrated by the sequence stratigraphy of coal-bearing sequences that were deposited under a humid climate while contemporaneous eustatic changes in an arid climate result in eolianites and carbonates, a totally different lithostratigraphic response to the same eustatic event (Cecil et al., 2003a).

Although autocyclic processes may result in cyclic deposition in that they repeat, they are without a predetermined frequency. Consequently, they are aperiodic because they do not represent a determinate period of time. The same can be said for allocyclic tectonic change. The sedimentary response to autocyclic processes tend to be rather local and they generally do not produce regionally mappable units. In contrast, allocyclic changes in precipitation and sea level may be the result of deterministic processes that are periodic, such as orbital forcing. Thus,

Cecil, C. B., DiMichele, William, Fedorko, Nick, and Skema, Vik, 2011, Autocyclic and allocyclic controls on the origin of the Dunkard Group, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 26-45.

periodic changes in sea level and (or) precipitation led to predictable sedimentary cycles such as the well known Pennsylvanian cyclothems (e.g., Busch and Rollins, 1984).

Deposition of the Dunkard Group has been attributed to fluvial-deltaic processes including aggradation on an alluvial plain and lacustrine deposition (e.g., Beerbower, 1961, 1964, 1969; Berryhill and Swanson, 1962; Berryhill, 1963; Martin, 1998). Autocyclic processes that controlled the origin of the Dunkard Group have been explicitly evaluated and summarized by Beerbower (1969) and implicitly by Martin (1998). Beerbower (1969) attributed alluvial plain aggradation in the southern region of the Dunkard Group to autocyclic processes. Because autocyclic process controlled aggradation in the alluvial plain, marker beds are rare and geologic mapping of alluvial plain time-stratigraphic units is exceedingly difficult. However, allocyclic processes were the predominant control on the stratigraphy of the Dunkard Group in the lacustrine basin where marker beds such as coal bed horizons, limestone horizons, and regional-scale paleosols are rather widespread and mappable. Allocyclic processes are further evaluated herein.

Relative Importance of Autocyclic Processes

As pointed out by Beerbower (1969), autocyclicity was the predominant control on alluvial plain deposition. Alluvial plain deposition is most extensively developed in the southern and southeastern part of the basin in central West Virginia. The alluvial plain environment is interpreted herein as a low gradient alluvial fan where the gradient is estimated to have been approximately 5.28 m per km (1 ft per mi). Anastomosing streams likely crisscrossed the aggrading fan. Autocyclic channel avulsion was common, resulting in channel incision as new streams flowed down the regional gradient of the fan. Root traces commonly occur in channel fills, levees, and over-bank deposits in the basin center indicating low-flow conditions and prograding alluvial plain aggradation when lacustrine conditions were absent (Figure 1).

Relative Importance of Allocyclic Processes

Changes in precipitation were the predominant direct allocyclic control on Dunkard Group lithostratigraphy. Variations in precipitation controlled variations in terrestrial organic productivity, lacustrine base levels, basin-scale weathering, water table and pedogenesis, sediment supply, soil moisture, and sedimentary geochemistry. Tectonics controlled accommodation space and basin configuration. Tectonic controls on basin configuration may have included occasional tectonic development of a silled basin, which contributed to lacustrine accommodation space. Although the presence of sills has not been documented, they can be inferred. As will be discussed subsequently, tectonically induced rain shadow effects may have contributed to the period-scale climatic drying trend that began during deposition of the Late Pennsylvanian Monongahela Group (Cecil et al.,1985; Cecil. 1990). In contrast to climate and tectonics, eustasy had little or no recognizable effect on the stratigraphy of the Dunkard Group as fully marine conditions have never been documented within the ~366 m (~1,200 ft) of Dunkard strata. One incursion of brackish water may be indicated by an occurrence of *Lingua* in a dark shale parting within the Washington coal complex (see Stop 2 description). The known spatial distribution of the *Lingula*-bearing bed is confined to a relatively small geographic area in the vicinity of the Ohio River and its tributaries in Marshall County, West

Figure 1. Root penetrations nearly destroyed the original bedding in channel fill sandstone, levee, and over-bank deposits near the basin center. Important sedimentological features indicated here include the replacement of lacustrine conditions by alluvial plain progradation and aggradation, and rizomorphs indicative of aquatic vegetation demonstrating low-flow conditions in all depositional environments. Photo is approximately 52 m (170 ft) above the Waynesburg coal bed in the Brick Hill section near Moundsville, WV.

Virginia and Belmont County, Ohio (Cross et al., 1950). Of the three-allocyclic processes, climate was the predominant control on lithostratigraphy and will be discussed in more detail.

Climate and Climate States

Climate refers to average long-term weather (Bates and Jackson, 1987, p. 125). A climate state, as used herein, refers to a climate that prevailed in a region for an extended period. Thus, major shifts in climate are referred to as a change in state. Use of climate state terminology is preferable to such terms as greenhouse or ice house because climate state terminology does not implicitly or explicitly imply cause. The duration of paleoclimate states are defined in Table 1.

Carboniferous and Permian paleoclimate states in the Appalachian basin appear to have persisted at multiple time scales (Bush and Rollins, 1984; Cecil et. al., 1985, Cecil, 1990), ranging from tens of millions of years to a few thousand years (Table 1). Climate variations shorter than one thousand years are assigned herein to variations in weather. The hierarchy of time scales in Table 1 provides a convenient method of expressing the duration of prevailing climate states without implying causes or controls.

The three predominant components of climate are temperature, wind, and precipitation. Most workers use mean-annual precipitation to estimate and describe paleo-precipitation (e.g., Retallack, 1990). However, seasonality of rainfall (Table 2) is a more

Table 1. Duration of paleoclimate states using terminology modified from glacial and interglacial time scales. (Adapted from the AGI Glossary of Geology).

Time scale	Duration
Period	10 My
Epoch	1 My
Stage	10^2 Ky
Age	10 Ky
Stage	1 Ky
Weather	< 1 Ky

quantitative measure of climate as a control on pedogenesis, sediment supply, sedimentation, and stratigraphy (Cecil and Dulong, 2003). The climate classification in Table 2 is based on the number of months in a year that rainfall exceeds evapotranspiration, a classification that implicitly indicates a degree of monsoonal atmospheric circulation. However, it is possible, and even probable, that an ambient climate may have a relatively weak annual rainy season when rainfall does not exceed evapotranspiration. In the latter case, the climate would be designated as arid using the classification in Table 2. Although an arid climate designation is technically correct, given the physical parameters specified in Table 2, plants, soils, and sediment yield would respond and record weak seasonality in precipitation. The predominant paleoclimate states that controlled deposition of Dunkard strata were determined by seasonality of precipitation, even if there was a weak annual rainy season.

Climate and Soil Moisture

Climate seasonality is one of the principal factors that controls pedogenesis and soil properties, including soil moisture (Table 2) (Cecil and Dulong, 2003). Since the appearance of land plants, soil moisture must have been a first-order control on plant ecology and the production of terrestrial organic matter. Terrestrial organic matter is important in soil genesis because the amount of organic matter is a predominant control on soil pH and Eh which, in turn, are key controls on pedogenesis. Thus, paleosol properties can be used to estimate soil moisture, even though it is not possible to measure soil moisture directly, as is possible in

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Table 2. Annual precipitation regimes, the degree of seasonality, and probable soil moisture based on the number of consecutive wet months per year in regions that were warm, had a low elevation, and a low surface gradient. By definition, a month is wet when rainfall exceeds evapotranspiration (modified from Cecil, 2003, p. 3-4, and Cecil and Dulong, 2003).

Number of wet months	Precipitation regime	Degree of seasonality	Probable soil moisture
0	Arid	Aseasonal	Aridic
1-2	Semiarid	Minimal	Aridic to Ustic
3-5	Dry subhumid	Maximum	Ustic
6-8	Moist subhumid	Medial	Udic
9-11	Humid	Minimal	Aquic
12	Perhumid	Aseasonal	Peraquic

Table 3. Soil orders and their probable soil moisture regimes (see Cecil and Dulong, 2003 for more details).

Soil order	Soil moisture	Comment
Aridisols	Aridic	Mostly dry throughout the year
Molisols, Alfisols, Vertisols	Ustic	Dry throughout most of the year
Vertisols	Udic	Moist for a significant part of the year
Oxisols, Ultisols, and Spodosols	Aquic	Moist throughout most of the year
Histosols	Peraquic	Water logged most or all of the year

modern soils. A qualitative estimate of soil moisture regimes for various soil orders is presented in Table 3.

AUTOCYCLIC AND ALLOCYCLIC CONTROLS ON DEPOSITIONAL SYSTEMS

Late Paleozoic Paleoclimate

On the basis of changes in paleoclimate indicators in Paleozoic strata, examples of both abrupt and gradational shifts in period-scale paleoclimate states can be found in the Central Appalachian basin (Cecil et al., 2004, p. 85, fig. 10). For the Permo-Carboniferous, a 10 My humid period that began in the late Devonian continued through the early Mississippian but ended abruptly at approximately the end of the Tournaisian. A 15 My early Mississippian humid period was punctuated by stage-scale climate and sea-level cycles. Subsequently, period -scale sea-level rise was accompanied by a 19 My middle Mississippian arid period (Visean) across North America. The Visean aridity began to gradually end at approximately the beginning of the Serpukhovian when eustatic sea level fall exposed the top of the Greenbrier Limestone (Alderson Limestone Member, southeastern West Virginia), coeval with the onset of a dry-subhumid climate. The karsted Greenbrier surface is overlain by siliciclastics that were delivered to the basin by an increase in fluvial sediment supply in response to the shift from Visean aridity to the increase in seasonal rainfall in the early Serpukhovian.

In the central Appalachian basin, the climate of the subsequent 8 My late Mississippian (Serpukhovian) became increasingly moist, progressing from Visean aridity to dry-subhumid in the earliest Serpukhovian to moist-subhumid by the end of the Serpukhovian (Cecil, 1990; Cecil et al., 2004) as global sea level continued to fall. Increasingly humid conditions are indicated by the late Serpukhovian by the development of humid climate paleosols including thin but mappable coal beds in late Serpukhovian strata. The moist-subhumid climate ended abruptly at approximately the Mississippian-Pennsylvanian boundary when the dry-subhumid climate of the latest Mississippian (latest Serpukhovian) changed to an Early Pennsylvanian (Bashkirian) humid to perhumid climate (Cecil et al., 1985; Cecil, 1990; Cecil et al., 2004). The Early Pennsylvanian humid climate persisted for approximately 6 My. The humid climate was coeval with maximum high-latitude continental ice volume and a maximum low stand in sea level that produced the mid-Carboniferous eustatic event (Saunders and Ramsbottom, 1986), and an unconformity of global proportions. The 6 My humid period led to chemical weathering and erosion that produced erosional surfaces, residual deposits, and karsting not only in much of the Appalachian basin but across the North American craton. Mid-Carboniferous residual weathering deposits include, but are not limited to, the following: a) Mercer clay in Pennsylvania, b) Olive Hill clay in Kentucky, c) Cheltenham clay on the Ozark dome and adjacent plareau in Missouri, d) residual chert ("chat" term used by the Kansas Geological Survey) in southeastern Kansas, and e) the Molas Formation in Colorado. Paleo-karsting is well developed on Mississippian limestones in Kentucky, the Madison Limestone in Colorado, and the top of the Guernsey Limestone (Madison Limestone equivalent) in Wyoming and Montana. Erosional surfaces at the Mid-Carboniferous boundary are common in numerous basins across the craton. Although there may have been a modest east-to-west drying gradient, the humid climate of the Early and early Middle Pennsylvanian persisted across the craton until near the end of the Bashkirian (late Atokan) when continental ice began to melt, global sea level slowly rose, accompanied by gradual temporal drying as documented in the Appalachian basin (Cecil et al., 1985; Cecil, 1990). Spatial and temporal Atokan sea level rise and gradual continental flooding is indicated by the onset of widespread sedimentation across the cratonic unconformity well beyond the extent of Morrowan strata (McKee and Crosby, 1975). Morrowan strata appear to be confined to cratonic depressions of various origins.

Period-scale gradual drying continued, interrupted by stage-scale climate and sea level cycles, as sea level gradually rose and continental ice sheets melted until the Kasimovian (Missourian). Maximum drying is indicated in the Appalachian basin at about the Kasimovian-Gzhelian (Missourian-Virgilian) boundary. Maximum drying is also indicated at the same time in Arrow Canyon, Nevada, by evaporative gypsum rosettes in Missourian age limestones. This was also a time of maximum highstand in sea level and minimal ice volume (Rygel et al., 2008). The Ames marine zone, the last known fully marine incursion into the Appalachian basin (early to middle Kasimovian; Edmunds, 1996, Lebold and Kammer, 2006), may represent the maximum high stand during the Late Pennsylvanian. The absence of subsequent sea level transgressions make it nearly impossible to identify linkages among sea level, climate, and high -latitude ice volume for either period- or stage-scale climate systems for the remainder of the Pennsylvanian and Permian. However, available paleoclimate proxies indicate that precipitation gradually decreased throughout deposition of the Dunkard Group, possibly for more than 10 My. Unlike the Carboniferous, where abrupt lithostratigraphic changes document changes in period-scale climate states, there is no clear stratigraphic evidence indicative of a distinct climate period during deposition of the Dunkard Group. Nor is there obvious evidence in the Appalachian basin to suggest eustatic changes in sea level or ice volume, even though stage-scale sea-level changes occurred in the Early Permian of Kansas and elsewhere across the craton. Rather, it is clear that the climate-drying trend that began in the later Middle Pennsylvanian into the late Pennsylvanian (Cecil, 1990; Roscher and Schneider, 2006) continued, interrupted intermittently by stage-scale shifts in climate state, throughout deposition of the Dunkard Group. There is no discernable evidence indicating that there were significant climatically induced changes in temperature during Dunkard deposition. If so, the equitable temperatures in the Appalachian basin can be attributed to an equatorial paleogeography (see paleogeographic reconstructions by Scotese, 2001).

During the Pennsylvanian, maximum precipitation in the tropics was associated with stationary low pressure systems that developed during maximum ice volume and maximum lowstands, regardless of the duration of climate states (Cecil, 1990). Peat precursors to coal formed during stage-scale, low stand, humid conditions. As Pennsylvanian ice sheets melted, rising sea level flooded cratonic basins while the low-stand humid climate and atmospheric low pressure doldrums gave way to high pressure and cross-equatorial surface winds. The high pressure systems induced wind-driven circulation in the shallow epeiric seas, which shut down dysoxic conditions and the deposition of organic-rich mud (black shale) while turning on the carbonate factory (Cecil et al., 2003a). Climate drying and monsoonal circulation were maximized during interglacials and maximum high stand conditions, regardless of paleolatitude across the craton (Cecil, et al., 2003a ; Cecil, et al., 2004). During high stands, open marine limestones were deposited off shore while nonmarine limestones were deposited contemporaneously in continental lacustrine systems. However, linkages among ice volume, sea level, and paleoclimate during deposition of the Dunkard remain equivocal because of the absence of marine incursions into the Appalachian basin. Regardless, the importance of climate as a control on Permo-Carboniferous stratigraphy cannot be overstated.

Terrestrial Organic Productivity

Papers by DiMichele et al., Blake and Gillespie, and Eble et al. (2011) present a comprehensive overview of the state of knowledge of the paleobotany of the Dunkard Group. Thus, the observations present herein are limited to field evidence for apparent climate controls on stratigraphic variations in terrestrial organic productivity. Temporal variations in terrestrial organic productivity were probably controlled by both period- and stage-scale climatic variations in rainfall. In general, it appears that terrestrial organic productivity was diminished during deposition of Dunkard strata relative to the preceding Pennsylvanian because of period-scale climate drying. This is especially true for the Greene Formation where plant fossils are relatively uncommon and coal beds are relatively thin, discontinuous, and impure. Coal bed horizons above the Washington coal complex at the base of the Washington Formation are little more than spatially recognizable centimeter-scale coal bed horizons or coaly streaks generally overlying poorly developed underclay paleosols (see Fedorko and Skema, 2011, for the stratigraphy of coal beds in the Dunkard Group). Root traces occur commonly throughout aggraded alluvial plain deposits, including some channel sandstones, suggesting the following: a) aquatic plants were common, b) streams were shallow, and c) stream flow was low.

Sedimentary Geochemistry

The significance of the sedimentary geochemistry of the Dunkard Group is perhaps best illustrated by a comparison with the Morrowan Pocahontas and New River formations, and the Atokan Kanawha Formation (Bashkirian into the Moscovian). These formations contain the low-sulfur bituminous coal beds of southern West Virginia (for additional lithostratigraphic details see Cecil et. al., 1985, p. 17, tab. 4). Calcareous strata are extremely rare in Early and early Middle Pennsylvanian strata, being found only in the occasional concretion that formed in rocks deposited under marine influence, most often in the early Middle Pennsylvanian Kanawha Formation. In contrast, nonmarine limestones are common in the lower part of the Dunkard, and Dunkard strata become increasingly calcareous up-section, particularly in the Greene Formation. Nonmarine limestones grade into coeval petrocalcic paleo-Vertisols in the alluvial plain reflecting both a paleosol catena (paleotopography) and dry climatic conditions of calcium carbonate deposition, be it lacustrine or pedogenic carbonate.

Much of the iron in Early and early Middle Pennsylvanian strata is in the ferrous state, occuring in abundant siderite (Cecil et al., 1985). Conversely, a significant amount of iron occurs in oxidized forms in red beds in the Dunkard. The comparatively minor amounts of reduced iron in the Dunkard occurs in relatively small amounts in siderite, predominantly in the Waynesburg Formation and the lower half of the Washington Formation. Additional reduced iron occurs in pyrite, mostly associated with coal beds. In contrast, siderite is very abundant in Early and early Middle Pennsylvanian strata whereas pyrite occurs only in trace amounts. The major contrast in the mineralogical distribution of iron among the Early through early Middle Pennsylvanian strata and Dunkard Group strata is indicative of major differences in sedimentary geochemistry.

The original major coal resources in Early and early Middle Pennsylvanian strata also contrast with the extremely meager coal resources of the Dunkard Group (upper barren measures of Rogers, 1858). Furthermore, kaolinite is the predominant clay mineral, by far, in Early and Middle Pennsylvanian underclay paleosols, including the Allegheny Formation. These paleosols were subjected to subaerial exposure, weathering, and pedogenesis. The kaolinite-rich underclays are diagnostic indicators of intense chemical weathering during stage-scale low-stands and pedogenesis. The acidic and reducing pedogenic conditions were sufficient to cause the reduction and removal of iron, either by illuviation and precipitation in a B_S soil horizon (Spodic horizon), or by dissolution and outflow through fluvial systems on colloidial organic matter, or as dissolved ferrous iron. In contrast, Dunkard paleosols are commonly calcareous, and they generally do not exhibit characteristics of intense chemical leaching.

The major differences in the occurrence and concentration of syngenetic minerals in the southern coal fields of West Virginia and the Dunkard are indicative of relative extremes in syndepositional Eh and pH conditions. Surficial waters during deposition of nonmarine strata in the southern coal fields generally must have been acidic with a pH perhaps as low as 4.5 analogous to rivers draining the regions of ombrogenous peat deposits in the perhumid region of equatorial Indonesia (Cecil et al, 2003b). Such low pH values preclude the formation of calcite. During deposition of the Early and early Middle Pennsylvanian strata, the Eh was at least mildly reducing with values as low as -0.4V to -0.6V, all that is required to reduce iron under acidic conditions (see Garrels and Christ, 1965, p. 234). Given the ubiquitous calcite in Dunkard strata, the general pH condition during deposition of the Dunkard must have been alkaline, $\geq$ pH 8. The Eh probably was never less than 0.2V, except possibly during deposition of the sparse organic-rich deposits.

Uniquely different water chemistries are required to explain the differences in the lithologic characteristics among Early Pennsylvanian rocks in southern West Virginia and the much younger Dunkard Group. The different water chemistries were almost assuredly caused by differences in paleoclimate (Cecil et al., 1985; Cecil, 1990). The Early Pennsylvanian was dominated by a humid to perhumid climate whereas the Late Pennsylvanian and early Permian were controlled by climate extremes (Cecil, 1990) that may have ranged from semiarid during the formation of calcic paleosols to humid during peat formation. Acidic black water rivers, analogous to modern streams in equatorial Indonesia, with a pH as low as 4.5 (Cecil et al., 2003b), dominated fluvial systems and estuaries during the Early and early Middle Pennsylvanian. Concentrations of dissolved solutes in fluvial systems were very low, but the Eh-pH of continental hydrologic systems were sufficient to reduce and transport both dissolved and collodial iron to depositional environments where the iron was precipitated in the ferrous state as siderite. In addition, the humid climate resulted in intense chemical weathering that ultimately resulted in oligotrophic nutrient conditions that included all surface drainages, surficial waters and paleosols.

Figure 2. Alternating red and gray-green strata in the Dunkard alluvial plain. Guard rail (bottom) and light pole (left side of photo) provide scales. Road cut is on US 50 bypass near Parkersburg, WV.

In contrast to the acidic waters of the Early Pennsylvanian, low gradient muddy alkaline rivers, saturated by dissolved solutes such as calcium carbonate, criss-crossed the Dunkard landscape. Oxidizing conditions in the well-drained alluvial plain gave way to reducing conditions whenever and wherever waterlogging or flooding occurred during pluvial climate states, particularly in the lacustrine basin center. Alternating red and gray-green strata in the upper alluvial plain **(**Figure 2**)** are herein attributed to fluctuations in oxidizing and reducing conditions driven by stage-scale allocyclic changes in precipitation. Oxidizing conditions prevailed during dry periods when water tables were relatively low and vegetation on the alluvial plain landscape was sparse. Reduction of iron on the alluvial plain occurred during relatively humid periods when soil moisture was high, terrestrial organic matter was abundant, and the water table was at or above the surface in the lacustrine basin.

Intra- and Extra-Basinal Weathering

Climate-driven physical and chemical weathering are always a factor in pedogenesis and sediment production. The nature of intra-basinal weathering during Dunkard deposition is indicated by physical and chemical properties of paleosols, and the textural and mineralogical properties of fluvial sediments. If any sediment was derived from outside the basin, then weathering in extra-basinal provenance regions is also indicated by the textural and mineralogical properties of fluvial sediments. If there were differences in weathering in either of these environments, they are not apparent in field observations. Repeated occurrences of wacke sandstones as well as an increase in the occurrence of petrocalcic paleo-Vertisols indicate that there was a period-scale decrease in the degree and intensity of chemical weathering up section. The general absence of kaolinitic paleosols in Dunkard strata relative to the abundance of Pennsylvanian kaolinitic paleosols also indicates greatly diminished weathering while Dunkard strata were being deposited relative to these earlier times when coal-resource-bearing Pennsylvanian strata were being deposited.

DUNKARD GROUP DEPOSITIONAL ENVIRONMENTS

Deposition of strata within the Dunkard Group has been attributed to alluvial plain and lacustrine depositional environments (e.g., Beerbower, 1961, 1964, 1969; Berryhill and Swanson, 1962; Berryhill, 1963; Martin, 1998). In the present study, the alluvial plain environment is interpreted as a low gradient alluvial fan where autocyclic fan drainages discharged from the south-southeast (Martin, 1998) into a lacustrine basin along the synclinal axis of the present-day basin center. Laterally fluctuating facies between the autocyclic alluvial plain sedimentation and allocyclic lacustrine conditions is indicated by intertonguing of laterally extensive beds of lacustrine origin (coal beds, limestones, and fan-delta lacustrine siliciclastic rocks) with strata of alluvial plain origin. Some limestone units have been traced southward from the lacustrine basin center into coeval petrocalcic paleo-Vertisols that formed on the alluvial plain. The lateral extent of the limestone and the coeval petrocalcic-Vertisols is indicative of a time of low siliciclastic sediment input in response to a dry-subhumid to semiarid paleoclimate.

The diminishing lateral extent of coal beds and limestones up-section (see cross section by Fedorko, included with the CDROM of this guidebook) suggests that lacustrine conditions became more restricted spatially through time. Figure 3 presents a schematic cross section depicting alluvial fan and lacustrine basin depositional settings.

Base level fluctuations are recorded in Dunkard Group strata, especially in the lithostratigraphy in the basin center. Well developed regional-scale basin-center paleosols are indicative of well-drained soils and a subsurface water table when base level migrated beyond the basin. In contrast, nonmarine limestones, coal beds, and fan-deltas, are indicative of relatively high intra-basinal base levels and a variety of lacustrine conditions. The common stratigraphic succession of paleosol, limestone and/or coal, dark shale, conformably overlain by flat-bottom sandstone is indicative of rising water level and the onset of lacustrine conditions.

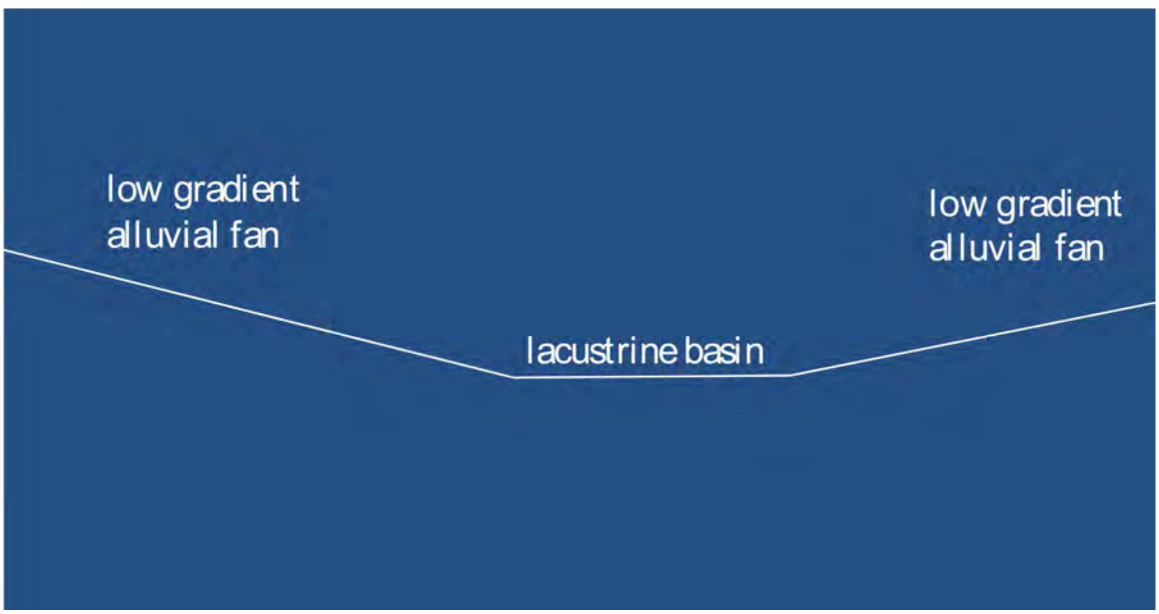


Figure 3. Schematic cross section illustrating a low gradient alluvial fan and a lacustrine basin.

Mineral Paleosols

An extensive variety of paleosols occurs in the Dunkard Group, ranging from Inceptisols to petrocalcic-paleo-Vertisols. Inceptisols are common in alluvial plain sequences where soil horizonation is not developed, but root traces indicate incipient soil formation. Inceptisols are common in aggrading alluvial plain sequences. Hydromorphic Histosols include coal beds and certain dark shales. Plant fossils in the dark shales are indicative of waterlogged conditions and a clastic swamp (see the description of Stop 4, Rosby's Rock). The frequency of occurrence and thickness of hydromorphic paleosols decrease up section. Soils underlying coal beds (underclays) analogous to the humid climate paleosols that underlie the thick, mineable, Pennsylvanian coal beds are poorly developed at best or not developed at all in the Dunkard. The paucity of well developed underclays is indicative of the relatively dry climates that prevailed during Dunkard deposition.

Vertisols with petrocalcic horizons and nodules are common, especially in the upper half of the Dunkard Group. The seat rock (underclay) beneath the Waynesburg A coal bed (Waynesburg Formation) is apparently the oldest petrocalcic-Vertisol in the Dunkard. The regional distribution of the petrocalcic-Vertisol under the Waynesburg A coal is indicative of landscape topography and a paleosol catena where a paleosol developed on topographic highs and lacustrine carbonate developed in topographic lows, as is the case below the Waynesburg A coal at the Roberts Ridge Road stop (Stop 2). Nonetheless, petrocalcic-Vertisols are very common in younger strata, especially in the Greene Formation (Figure 4).

Coal Beds

The Dunkard Group is known for its lack of coal resources (the upper barren measures of Rogers, 1858) (see cross section by Fedorko, on the CDROM of this guidebook). With the exception of the Waynesburg coal, which marks the base of the Dunkard, only the Washington coal complex has been mined locally, but it never represented a significant coal resource. Other Dunkard coal beds are generally thin and rather discontinuous although coal bed horizons can be traced regionally (see Fedorko and Skema, 2011). In addition, where coal does occur, generally it is relatively high in ash yield and sulfur content (see Eble et al., 2011). All the characteristics of Dunkard Group coal beds are suggestive of both topogenous (ground and surface water peat hydrology) and eutrophic (high dissolved solutes) conditions of peat formation (see Cecil et al., 1985, p. 202). In addition, coal underclay paleosols generally are thin and poorly developed relative to underclay paleosols that underlie Pennsylvanian coal beds that contain or did contain significant coal resources prior to mining. In short, conditions of peat formation during deposition of the Dunkard Group were marginal at best (see Cecil et al., 1985 for a more comprehensive treatment of factors that control peat formation). Coal beds are indicative of waterlogged peat swamps whereas the overlying dark gray shales are interpreted as lacustrine prodelta deposits.

Limestones

Limestones in the Dunkard Group (see Fedorko and Skema, 2011, for the stratigraphy of Dunkard limestones) appear to be of lacustrine origin. The predominant primary lithologies common to many of the Dunkard Group limestones are ostracode-peloidal wackestones and packstones (Isabel Montanez, 2011, personal communication). Limestones, most common in the Waynesburg and Washington formations, are composed of multiple benches that contain

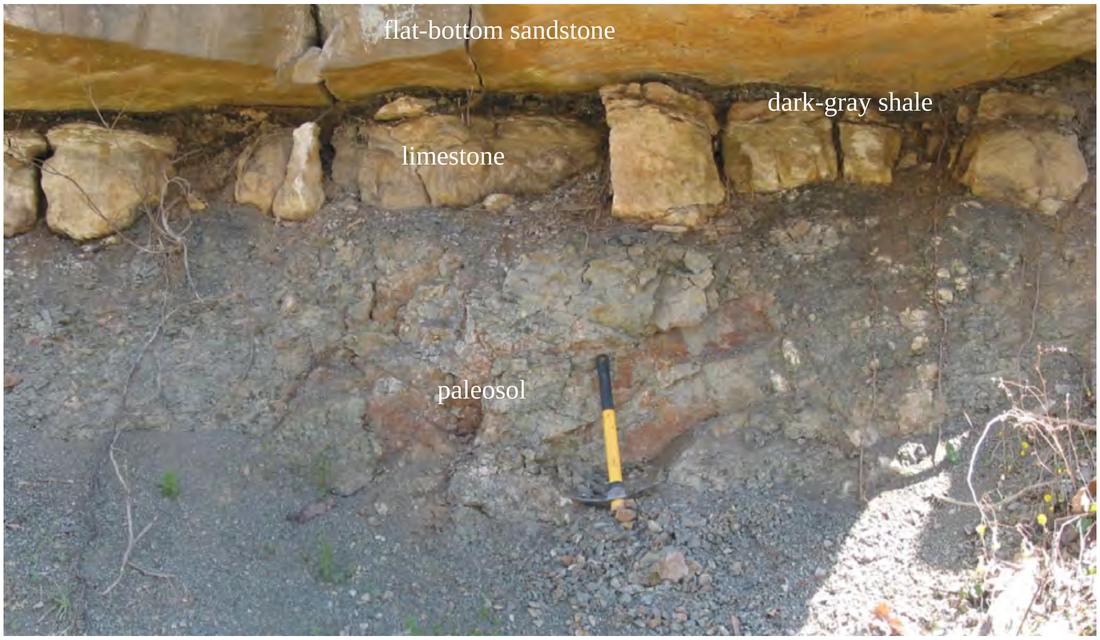


Figure 4. Petrocalcic Vertisol (unit with mattock for scale) in the Greene Formation overlain by the blocky Middle Rockport lacustrine limestone (buff), thin dark gray shale, and flat-bottom sandstone (top of photo) respectively. The dark gray shale occupies the stratigraphic position coal. Coal is not present at this locality. Mattock handle in the center of the photo is 40 cm (1.3 ft). Outcrop is on Greathouse Hill Road, Wetzel Co., WV.

subaerial macroscopic features such as subaerial crusts, micro-karsting, and brecciation induced by pedogenic processes similar to most Pennsylvanian nonmarine limestones including the Middle Pennsylvanian Upper Freeport limestone (Cecil, et al., 1985). The complex of limestone beds with subaerial exposure features suggests repeated rise and fall of water within carbonate-rich lakes. In contrast, limestones in the Greene Formation sometimes consist of a single bed without well developed subaerial features, and may contain plant fossils, such as tree ferns, typical of relatively abundant soil moisture (perhaps growing on the lake margins). Again, though, the Greene Formation is more or less confined to the lacustrine basin center and, therefore, potentially less well drained and wetter relative to the alluvial plain.

The carbonate-rich lakes were desiccated periodically either by draining, or by evaporation during relatively dry intervals. Evaporation should have led to increased salinity, but, except for some dolomitic beds, field evidence for increased salinity has not been reported. If drainage was the cause, then the nonmarine carbonate systems must have been somewhat analogous to the fresh water marls in the Florida Everglades, where a small drop in sea level in Florida Bay would result in drainage of the Everglades. Subsequently, the entire fresh-water lacustrine-palustrine Everglades system would disappear and subaerial exposure of the fresh water marls would lead to subaerial crusts and brecciation similar to those features in Pennsylvanian nonmarine limestones. Such small-scale sea-level fluctuations might also account for the multiple benches of limestone within individual named limestone units such as the Benwood in the Monongahela Group or the Washington in the Dunkard. A distal connection to the sea, and hence to sea level influences, perhaps through a breach in a sill represented by the Cincinnati arch, would also explain the apparent time-stratigraphic correlation of marine and nonmarine limestones (Cecil et al., 2003a). The Gulf of Carpentaria in Northern Australia represents an alternative to the Florida Everglades for carbonate deposition. Eustatic cycles appear to have been an allocyclic control on the Pleistocene/ Holocene stratigraphy of bottom sediments in the Gulf of Carpentaria basin (Edgar et al., 2003). Low stand exposure of the floor of the Gulf of Carpentaria basin contributed to the development of calcic-paleosols (Chivas et al., 2001). Subsequent ponding of drainage by sea level rise and lake formation could have led to lacustrine carbonate deposition.

Although sea level fluctuations could account for carbonate deposition and subsequent subaerial exposure, it does not appear that sea level fluctuations can account for high base level and deeper water conditions that led to fluvial-deltaic lacustrine deposition. Thus, unless definitive data become available that document a connection to sea level fluctuations, climate drying appears to be the cause of lake desiccation. Climate drying is also indicated by coeval alluvial plain paleosols containing petro-calcic features.

Sandstones and Shales

Sandstones and shales in the alluvial plain environment are the result of autocyclic fluvial processes (Beerbower, 1969). However, allocyclic changes in paleoclimate likely controlled stratigraphic variation in sediment supply as well as oxidation-reduction states in the alluvial plain environment.

Sediments were supplied to open-water lacustrine systems where fan deltas are indicative of fluvial deltaic progradation during lake level high stands. The dark gray shales that commonly overlie coal beds are interpreted herein as prodelta deposits. Flat-bottom sandstones, conformably overlying the dark gray shale facies, are interpreted as distributary mouth bars derived from fluvial progradation. Sandstones with an erosional base commonly incise the upper part of the flat-bottom sands. Incision, as indicated by these latter sandstones, is attributed to fluvial progradation analogous to the well-known fluvial incision of mouth bars by prograding distributaries in the Mississippi River delta (e.g., Tye and Coleman, 1989).

Recurring red bed units that extend from the alluvial plain into the basin center may have been the result of prograding alluvial fans during drier intervals when lake levels were low.

MODERN ANALOGUES

Two modern low-gradient alluvial fan complexes appear to approximate the dry and humid end-member climatic conditions that prevailed during deposition of the Dunkard Group. The Okavango Fan at the northern edge of the Kalahari Desert in Botswana (Milzow et al., 2009), southern Africa, appears to represent the dry climate end member condition during Dunkard deposition when dry climate paleosols extended from the upper alluvial plain into the basin center (Figures 5A and B). Standing water rarely if ever develops in the basin at the toe of the Okavango Fan complex. During the rainy season, water is delivered to the fan by rivers from headwaters outside the basin where it is mostly absorbed into soils or lost to evapotranspiration before reaching the basin.

In contrast to the Okavango fan, the alluvial fan complex along the eastern edge of the Pantanal in southern Brazil (Figures 6A and B) (Assine and Soares, 2004; Assine, 2005) is a likely candidate for the more humid climatic Dunkard conditions, when the alluvial plain was vegetated and lacustrine conditions developed in the basin center. In the modern Pantanal, the

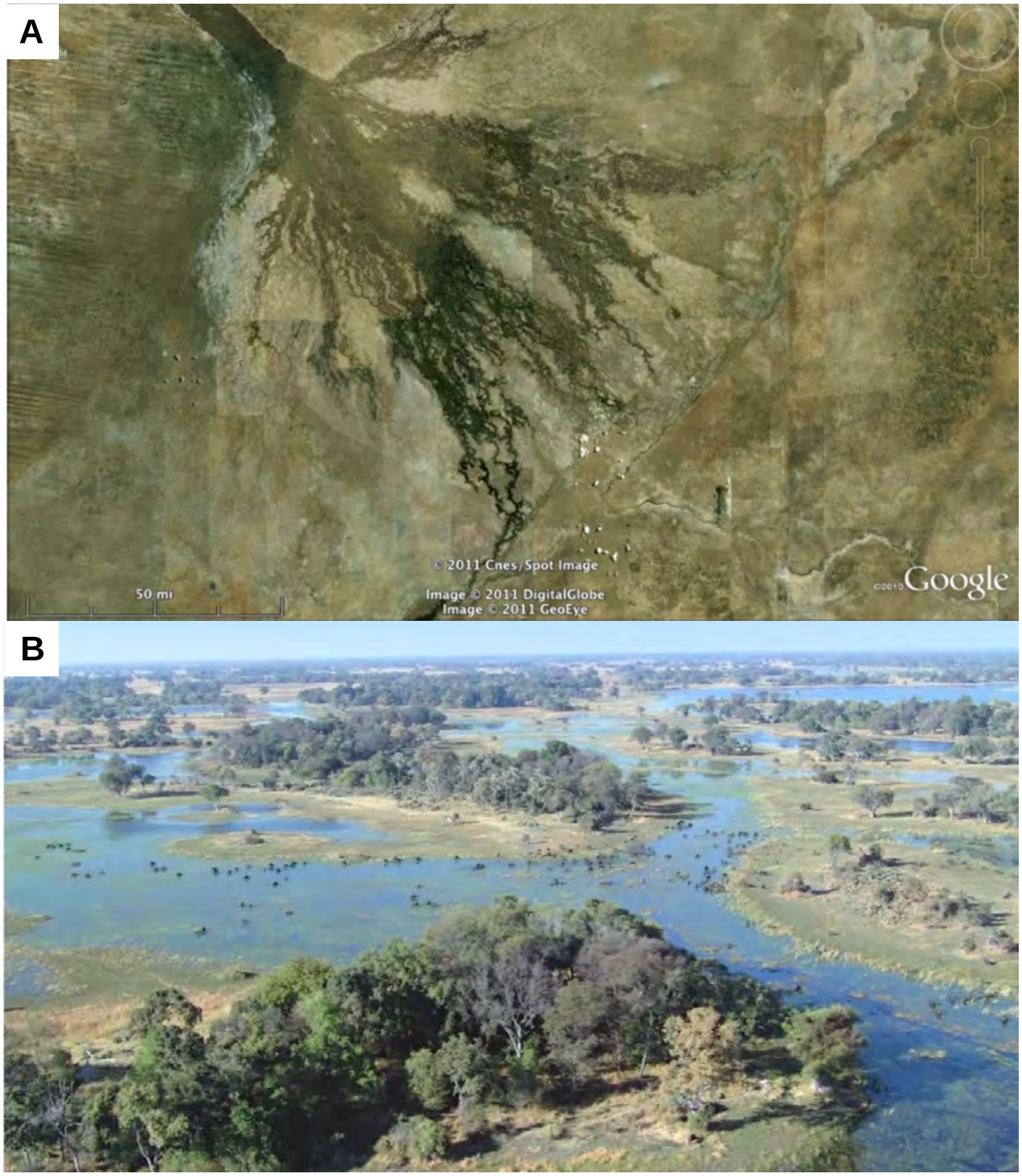
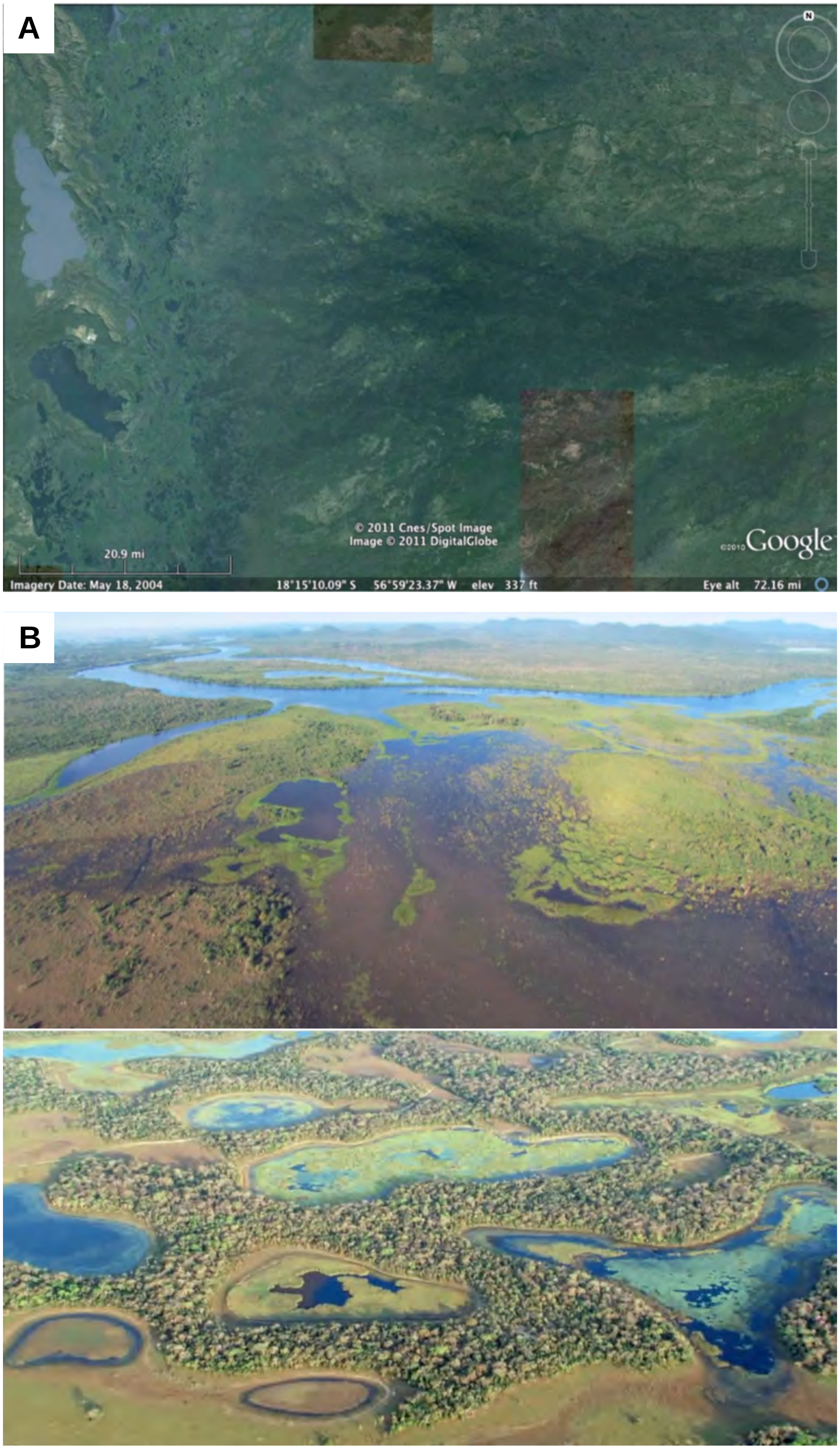


Figure 5. Photos of Okavango Fan in southern Africa. A—Okavango Fan as a dry climate analogue for dry periods on the Dunkard alluvial fan. Satellite view of the Okavango fan during the rainy season (source, Google Earth). Active fluvial drainages are indicated by anastomosing dark green colors. Scale – 2.54 cm = ~32 km (1 in = ~20 mi). B—Low altitude view of anastomosing streams and lakes on the vegetated Okavango fan surface during the rainy season (view is from *Planet Earth series).* Trees and large mammals (dark dots in streams) are indicative of scale. Roots of aquatic plants in streams are probable analogues of root traces in Dunkard fluvial mudstones and sandstones. Streams are dry during the dry season.

basin center is flooded during the rainy season, but is then desiccated to randomly distributed lakes during the dry season (Assine and Silva, 2009). The lack of flooding during the dry season may be mostly the result of drainage of the extremely low-gradient basin center rather than evapotranspiration. The lakes in the basin center appear to represent environments where either peat or limestone could form whenever the necessary water levels and chemistry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conditions are met. Either organic or carbonate sedimentation in scattered lakes in the Pantanal basin center may be analogous to the discontinuous coal beds or limestones in the Dunkard.

The extent of stage-scale climate change in the Okavango and Pantanal regions is not known. Neither is the effect of climate and climate change as a control on sedimentation and stratigraphy. An extensive coring program in these two regions could certainly help elucidate autocyclic and allocyclic controls on sedimentation and stratigraphy in low-gradient fan systems. The results of modern analogue studies would almost assuredly help unravel the genetic controls on ancient alluvial systems such as the Dunkard Group.

The semiarid region of the Gulf of Carpentaria basin, Northern Australia, may represent a good analogue for Pennsylvanian deposition during the Missourian dry interval when mixed marine and nonmarine strata were deposited. Given the elevation of the Gulf basin, it can only be a partial analogue for Dunkard deposition because of marine flooding during high stands. Low stand paleosols overlain by ancient lake sediments beneath modern marine sediments in the Gulf are prototypes for paleosol types and lacustrine limestones (Edgar, et al., 2003; Chivas et al., 2001). Drainage gradients in the predominant fluvial source areas to the south of the Gulf are less than 10.6 m per km (< 2 ft per mi).

SUMMARY AND CONCLUSIONS

Determination of the relative importance of autocyclic and allocyclic processes provides a comprehensive analytical method by which the origin of alluvial-lacustrine depositional systems can be determined. Although Beerbower (1964) attempted to use the Mississippi River delta as a modern analogue for Dunkard stratigraphy, it now appears that modern fan-basin systems such as the Okavango and Pantanal are much better analogues, particularly in terms of stream gradients on the fan and in basin centers, as well as climate end-member conditions.

It also appears that autocyclic processes dominated alluvial plain sedimentation whereas allocyclic climate changes, ranging from humid to dry subhumid and perhaps even semiarid dominated the lacustrine systems . Alternating red and green strata are indicative of sediments that were either oxidizing and relatively dry (red) or waterlogged (green) by both autocyclic alluvial plain processes or by allocyclic changes in paleoclimate.

In contrast to the alluvial plain, allocyclic processes predominantly controlled basin center deposition and stratigraphy. However, stage-scale climate cycles (cyclothems), so apparent during the Pennsylvanian (Cecil, 1990), became increasingly less extreme and less distinct during Dunkard deposition. As in the upper alluvial plain, stage-scale climates ranged from humid to dry-subhumid or even semiarid (Figure 7). Humid periods led to a variety of waterlogged and base level conditions that determined peat formation, limestone deposition, or lacustrine deltaic sedimentation. Dry stages resulted in basin exposure and basin-scale calcic

Figure 6 (opposite page). Photos of the Pantanal fan in southern Brazil. A—Satellite image of the Pantanal fan and basin (area with lakes on the left-side of photo) as a humid climate analogue for humid periods on the Dunkard alluvial fan and basin (source, Google Earth). Low gradient active fluvial drainages on the fan are indicated by westward flowing anastomosing streams (darkest green colors) from the right center toward the center of the photo. Scale – 2.54 cm = ~32 km (1 in = ~20 mi). Lakes ~32 km (~20 mi) in length are visible on the left side of the photo. B—Two low altitude views of the Pantanal basin surface during the dry season (views are from the *Planet Earth series).* The region is flooded and under water during the rainy season. Trees and dirt road in the upper left corner of the bottom photo are indicative of scale. Vegetated lakes and ponds are potential analogues for the discontinuous Dunkard coal beds.

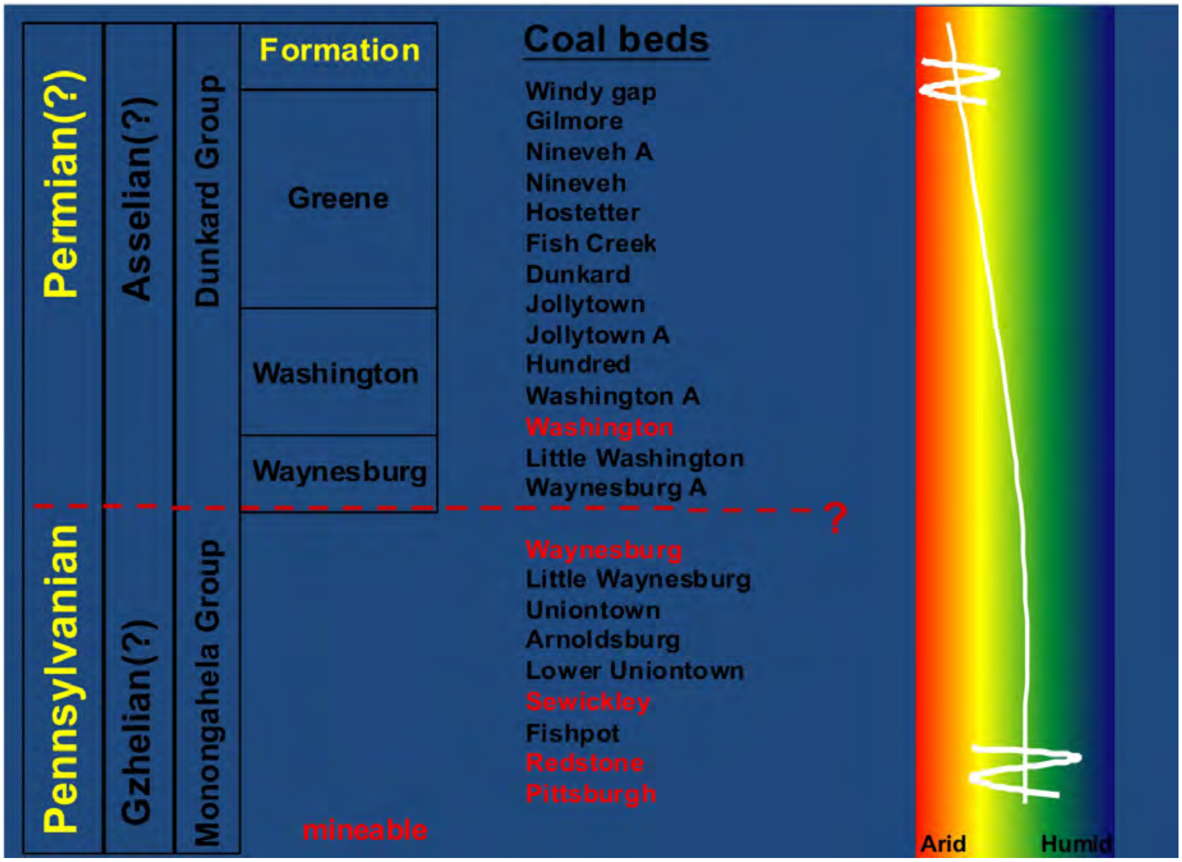

Figure 8. Stratigraphy of Monongahela Group and Dunkard Group coal beds. The diagram on the right side of the figure illustrates the period-scale climate transition and stage-scale climate cycles.

paleosols. These stage-scale paleoclimates became increasingly dampened, as they were part of a period-scale climate transition (Figure 7).

It is more difficult to explain the period-scale climate drying transition that began during deposition of the Late Pennsylvanian Monongahela Group and progressed during deposition of approximately 366 m (1,200 ft) of preserved Dunkard strata (Figure 7). Period-scale climate drying observed in Dunkard strata may have been caused by the assembly of Pangea (Parrish, 1993). If the Dunkard basin was positioned in the Pangean interior, it was far removed from a moisture source. A temporal increase in rainout, as prevailing equatorial winds moved from coastal regions toward the continent interior, would have caused period-scale climate drying (e.g. Tabor and Poulsen, 2008). Another factor in the slow drying trend may have been the rise of the central Pangean mountains and the slow development of a rain shadow over the Dunkard basin and beyond. Regardless of the cause(s), a time-stratigraphic drying trend, as shown in Figure 7, is indicated by the following: a) a stratigraphic increase in syngenetic calcite, b) a decrease in coal bed occurrence and thickness, c) a decrease in coal quality, and d) an increase in the occurrence of petrocalcic-paleosols.

The period-scale trend in climate drying represented in the approximately 366 m (1,200 ft) of Dunkard strata probably lasted for at least 10 million years and at least some, if not most, of the Dunkard is Permian (Asselian to Kungurian; see Lucas, 2011). If so, then Dunkard deposition was coeval with the transition between the humid Late Carboniferous and the arid late Permian (Kungurian in Russia) (summarized by Wardlaw et al., 2004).

ACKNOWLEDGEMENTS

This paper is dedicated to the memory of Prof. James R. (Dick) Beerbower whose pioneering work on the origin of the Dunkard led to his formulation and declaration of autocyclic and allocyclic models of sedimentation and stratigraphy.

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COMPOSITIONAL CHARACTERISTICS OF DUNKARD GROUP COAL BEDS: PALYNOLOGY, COAL PETROGRAPHY AND GEOCHEMISTRY

Cortland F. Eble, William C. Grady, and Bascombe M. Blake

OVERVIEW

The "Dunkard Series" was named by I.C. White in 1891 for exposures along Dunkard Creek in SW Pennsylvania. His original description is as follows:

> "The uppermost beds are found at the headwaters of Dunkard Creek, a large stream which heads near the West Virginia – Pennsylvania line, on the eastern slope of the watershed separating the Ohio and Monongahela River drainage system, and flowing eastward puts into the Monongahela two miles above Greensboro, Greene County, Pennsylvania, and four miles north from the West Virginia line. This stream flows over Permo-Carboniferous rocks from its source to the point at which it leaves the West Virginia line at Mount Morris, Pennsylvania, a distance of more than thirty miles, furnishing very fine exposures of these rocks along its banks and bluffs; hence, the geographical name (Dunkard Creek), which I have given the series." (White, 1891, p.20)

Although originally designated "Dunkard Creek Series" by White (1891, 1899), the term "Dunkard Series" was used in subsequent publications by White (1903, 1910), and other workers (Grimsley, 1907, 1910; Hennen, 1909, 1911, 1912; Hennen and Reger, 1913). The "Dunkard Series" is referred to the "Dunkard Group" in more modern literature, and includes all strata that occur above the top of the Waynesburg coal bed in the Appalachian basin.

INTRODUCTION

This study reports on the palynology (spores and pollen), petrology (maceral composition) and geochemistry (ash yields and total sulfur contents) of Dunkard Group coal beds. Specifically, this paper outlines the types of plants (as determined from palynology) that contributed to the formation of Dunkard coal beds, and provides commentary on the nature of the swamps that later were buried and converted to coal. A comparison of Dunkard coals, with the more economic coal beds of the underlying Monongahela Group is also provided. Finally, problems associated with using coal palynology to determine whether the Dunkard Group is Late Pennsylvanian, Early Permian, or both are also discussed.

SAMPLES

Palynologic data were derived from the analysis of 17 samples collected from a series of cores from test drilling in advance of long-wall mining operations in northern WV and SW

Eble, C. F., Grady, W. C., and Blake, B. M., 2011, Compositional characteristics of Dunkard Group coal beds: Palynology, coal petrography and geochemistry, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin: Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 46-59.

Pennsylvania by the Consolidation Coal Co. (Consol). Petrographic and geochemical data are based on results from these cores, and also from collections made during coal bed mapping and resource assessment projects conducted by the West Virginia Geologic & Economic Survey (WVGES, www.wvgs.wvnet.edu). A summary of sample preparation and analytical techniques can be found in Eble and others (2003). Please refer to stratigraphic sections of Fedorko and Skema (2011) for the temporal placement of individual coal beds.

PALEOBOTANICAL CONSIDERATIONS

An overview of Late Pennsylvanian/Early Permian plant reconstructions, with an example of their corresponding spore/pollen types, is shown in Figure 1.

Marattialean Ferns

Based on spore records, tree ferns were a dominant component of nearly all Late Pennsylvanian/Early Permian Dunkard swamps in the Appalachian basin (Cross, 1952; Clendening, 1960, 1962, 1966, 1970, 1975; Clendening and Gillespie, 1963; Habib, 1968; Eble,

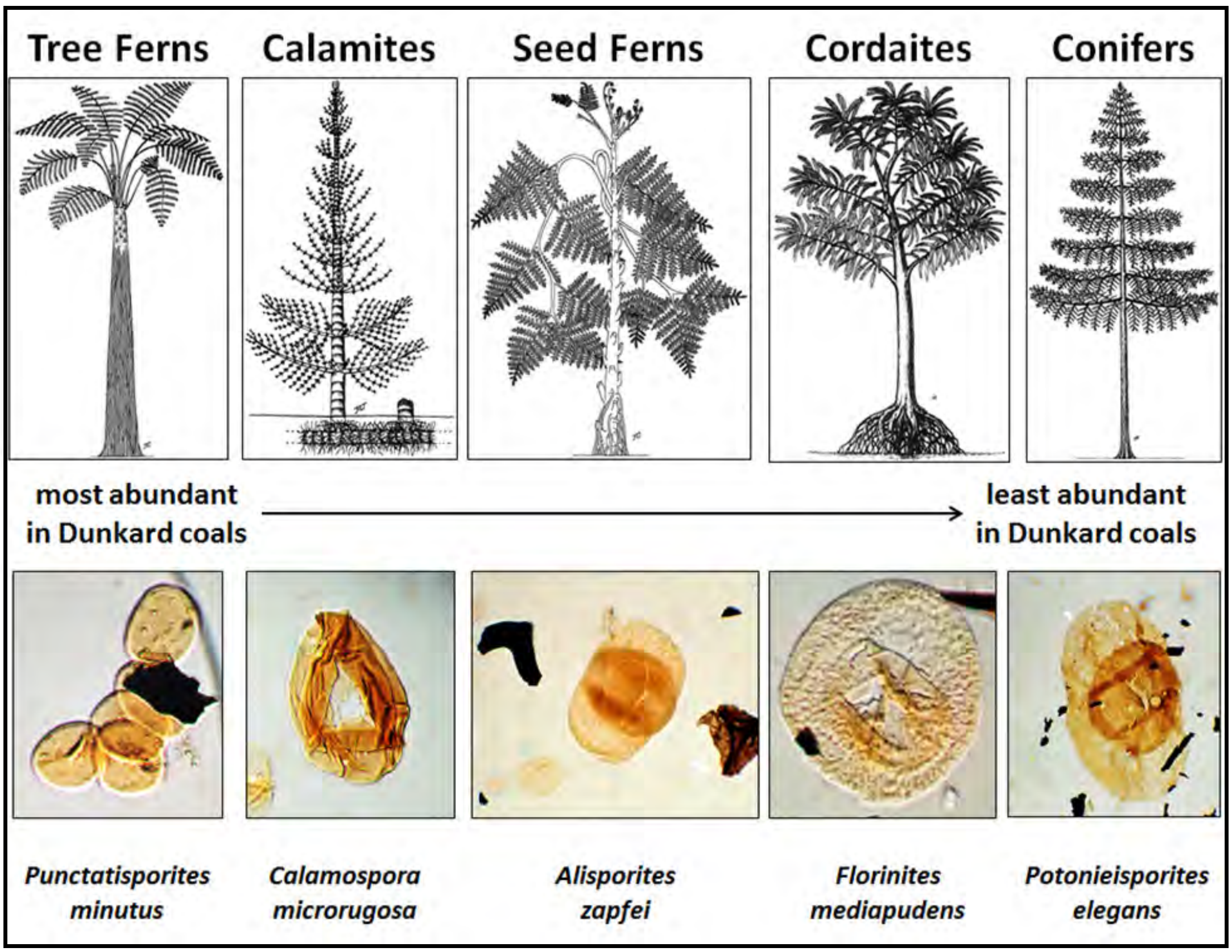

Figure 1 - Reconstructions of the major plant types, and their representative spores/pollen, in Dunkard coal beds. Plant reconstructions are based on drawings by Jerry Jenkins (Gillespie et al., 1978). Spore and pollen photograph magnifications are 850 to 1000X.

1985; Grady and Eble, 1989; Eble et al., 2003, 2006). Marattialean ferns in the Late Pennsylvanian and Early Permian were fairly large trees, some probably more than 5 m (16 ft) in height (Lesnikowska, 1989). The stem portions of the plant are ascribed to the genus *Psaronius*, which by convention, is applied to the entire plant. Sterile foliage remains (mainly adpressions) are assigned to the form genus *Pecopteris*.

Because of its size, *Psaronius* was an important contributor to peat biomass. However, its construction was significantly different from a modern day angiosperm (e.g., oak, maple), or gymnosperm (e.g., pine, cedar) tree. *Psaronius* achieved the tree habit primarily from an adventitious root-mantle that surrounded and structurally supported a relatively narrow vascular cylinder (Figure 2). As such, *Psaronius* possessed very little actual "wood", even though it had the stature of a small to medium-sized tree (DiMichele and Phillips, 1994).

Psaronius had a massive reproductive output (DiMichele and Phillips, 2002), producing small spores assigned to the following taxa in Late Pennsylvanian/Early Permian coals: *Thymospora* spp., *Punctatisporites minutus*, *Punctatosporites minutus*, *Laevigatosporites minimus*, *Spinosporites exiguus,* and *Apiculatisporites saetiger*. In his study of Dunkard Group

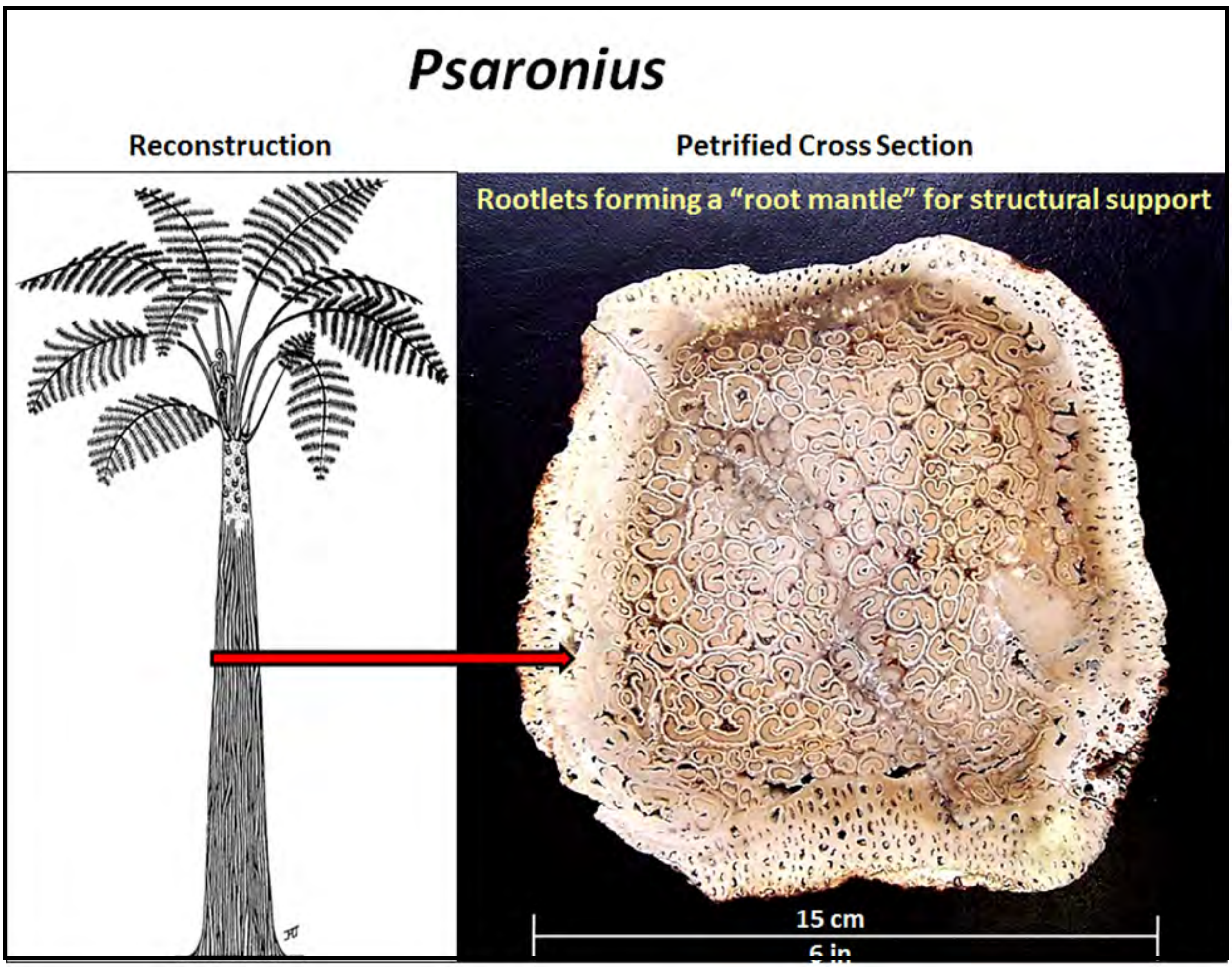

Figure 2 - A reconstruction of *Psaronius*, drawn by Jerry Jenkins (Gillespie et al., 1978), and a cross section of a *Psaronius* trunk. In cross section, the abundant rootlets, which formed an adventitious root mantle for structural support, can be clearly seen.

coal bed palynofloras, Clendening (1960, 1962) noted the overwhelming dominance of very small species of *Laevigatosporites*, which were later reassigned to *Fabasporites* (Clendening, 1968). *Fabasporites* is all but indistinguishable from small forms of *Punctatisporites minutus*; the only difference in taxonomic assignment is whether or not a trilete suture can be observed (*Fabasporites* is alete, whereas *P. minutus* is trilete).

Thymospora is represented by three species in Late Pennsylvanian/Early Permian coals, *Thymospora thiessenii, T. psuedothiessenii*, and *T. obscura*. *Thymospora thiessenii* was first identified in thin sections of the Pittsburgh coal by Thiessen and Staud (1923), and designated as the "Pittsburgh spore", because of its abundance. Cross (1952) later noted a similar abundance of *Thymospora thiessenii* in chemical macerations of the Pittsburgh coal. In a study of the Sewickley coal bed (Eble et al., 2003), a direct relationship was observed between bed thickness, ash yields, and *Thymospora* abundance. Thicker (>100 cm), lower ash (average 11.5%, dry basis) coal had whole-bed average *Thymospora* percentages between 20% and 65%. *Thymospora* was also found to increase in abundance from the base of the coal bed to the top. In contrast, thinner (< 100 cm [<39 in]), higher ash (average 22.5%, dry basis) coal contained much lower percentages of *Thymospora*. These results indicate that the parent plant of *Thymospora*, *Scolecopteris vallumii*, probably was part of a more established flora in the Sewickley paleoswamp, a suggestion made previously by Phillips and others, (1985) and Mahaffy (1988) for the Herrin and Springfield coal beds of the Eastern Interior (Illinois) basin (Late–Middle Pennsylvanian, Westphalian D). Noting that *Thymospora* was more abundant away from paleochannels, it was hypothesized that *Scolecopteris vallumii* may have favored areas of swamps that remained undisturbed for extended periods of time.

Thymospora is present in coals throughout the Dunkard section, but rarely attains any appreciable abundance (Clendening, 1960; this paper). This may be (likely is?) a reflection of the thin, discontinuous nature of Dunkard coals. The edaphic conditions needed for the proliferation of *Scolecopteris vallumii* probably never became established in the local swamps that characterize the Dunkard.

Small Ferns

Small ferns, represented by *Anachoropteris, Botryopteris, Zygopteris* and others, were a very diverse group of groundcover plants that, collectively, are indicative of exposed peat substrates. They are represented in the palynological record by the following genera, which are sphaerotriangular in shape and trilete: *Granulatisporites, Deltoidospora, Lophotriletes and Acanthotriletes*. Other dispersed spore taxa occur as well. Though diverse in occurrence, small ferns were minimal contributors to peat biomass (DiMichele and Phillips, 1994).

Sphenopsids (Calamites)

Calamites was a small to medium-sized, woody tree. Late Pennsylvanian/Early Permian forms may have achieved heights of 3 to 5 m (10 to 16 ft). Anatomically-preserved wood is assigned to the genera *Arthropitys, Calamodendron* and *Arthroxylon*. Foliage is referred to *Annularia* or *Asteroxylon*. *Calamospora* and larger species of *Laevigatosporites* (e.g., *L. minor, L. vulgaris*) are the most common dispersed spore taxa in Late Pennsylvanian/Early Permian coal beds. Calamite spore abundance patterns in Monongahela Group coal beds (Grady and

Eble, 1989; Eble et al., 2003, 2006) indicate increased percentages in, and around, areas of swamp disturbance (clastic partings, high ash coal layers).

Pteridosperms (Seed Ferns)

The principle seed ferns in Late Pennsylvanian/Permian strata were *Medullosa* and *Sutcliffia*, small to medium-size monoaxial trees with large, expansive fronds (DiMichele and Phillips, 1994; DiMichele et al., 2006). Most of the "fern" fossils that are found and collected are actually seed fern foliage; common genera include *Neuropteris* and *Alethopteris*. Seed fern pollen, *Schopfipollenites* (produced by *Medullosa*) and *Punctatisporites kankakeensis* (produced by *Sutcliffia*) is rare in Late Pennsylvanian/Early Permian coal beds in the Appalachian basin, suggesting that seed ferns were not swamp-centered vegetation. It is more likely that they inhabited clastic lowland environments adjacent to swamps, perhaps a result of higher nutrient requirements, which mineral (compared to peat) substrates would provide (Pfefferkorn and Thompson, 1982).

Cordaites

Cordaites were small, woody trees with large strap-shaped leaves. Reconstructions usually resemble modern mangroves, though recent studies are ambiguous about the habitat of these plants (Raymond et al., 2001). When found preserved, cordaite wood is referred to either *Mesoxylon* or *Pennsylvanioxylon*, with the latter being more common in Late Pennsylvanian/ Early Permian strata. *Artisia*, a common fossil, is a distinctly septate pith cast of cordaite stems. *Florinites* is the most common cordaite pollen found in Dunkard coal beds, though it rarely exceeds one to two percent of most assemblages. Foliage is referred to *Cordaites*.

Lycopods

In Dunkard coals, *Crassispora kosankei*, the dispersed spore of *Sigillaria*, is the only Late Pennsylvanian/Early Permian representative of the large lycopod trees (e.g., *Lepidodendron, Lepidophloios*) that were dominant components of Early and Middle Pennsylvanian swamps. *Crassispora* is rarely observed in Dunkard coal beds, and tends to be associated with inorganic partings and high ash coal layers in underlying Monongahela Group coals (Eble et al., 2003, 2006).

"Other Gymnosperms"

Pollen from coniferous plants, which are believed to have inhabited more moisture-stressed areas of the Late Pennsylvanian/Early Permian landscape (DiMichele et al., 2008; Falcon-Lang et al., 2009; Plotnick et al., 2009) are extremely rare in Dunkard coal palynofloras. The most common representative is *Pityosporites*, a genus that closely resembles modern pine pollen. Bisaccate-striate pollen (e.g., *Hamiapollenites, Striatopodocarpites, Protohaploxypinus*), are essentially non-existent in Dunkard coals. The lack of bisaccate-striate forms may be explained, in part, by the fact that only swamp environments (coal) are reported in areas where conifers were most likely sparse or absent. Most coniferous foliage is referred to the form genus *Walchia*.

RESULTS

Geochemistry and Coal Petrology

Geochemical and petrographic analyses of Dunkard Group coal beds are shown in Figures 3 and 4. Ash yields range from 19.5% to 32.6% (average = 27.8%, dry basis); total sulfur contents range from 3.5% to 7.2% (average = 5.1%, dry basis). Petrographically, Dunkard coal beds are dominated by vitrinite (range 79.7% – 93%, average = 83.1%, mineral matter free basis [mmf]), which forms principally from wood and wood-like tissues (Teichmüller, 1989). Inertinite macerals, which form from wildfire (Scott, 1989; Scott and Jones, 1994) and/or intense aerobic degradation of botanical material (Hower et al., 2011) average 13%, and range from 6.2% to 16.2% (mmf) in Dunkard coals. Liptinite macerals, which represent hydrogen-rich plant parts including spores and pollen, resins and cuticles, are very low in abundance in Dunkard coal beds (range 0.8% to 5%, average = 2.1%, mmf).

Dunkard coals are higher in vitrinite content, lower in inertinite and lower in liptinite than older Pennsylvanian coals lower in the strata. A statistical average of 1,157 coal samples, representing the entire range of Pennsylvanian coals in West Virginia from the WVGES Coal Quality Database, indicates an average vitrinite content of 75.5% (mmf), an average inertinite content of 19.3% (mmf), and an average liptinite content averaging 5.1% mmf.

Palynology

The distribution of spores and pollen in Dunkard coal beds is shown in Figure 5. All of

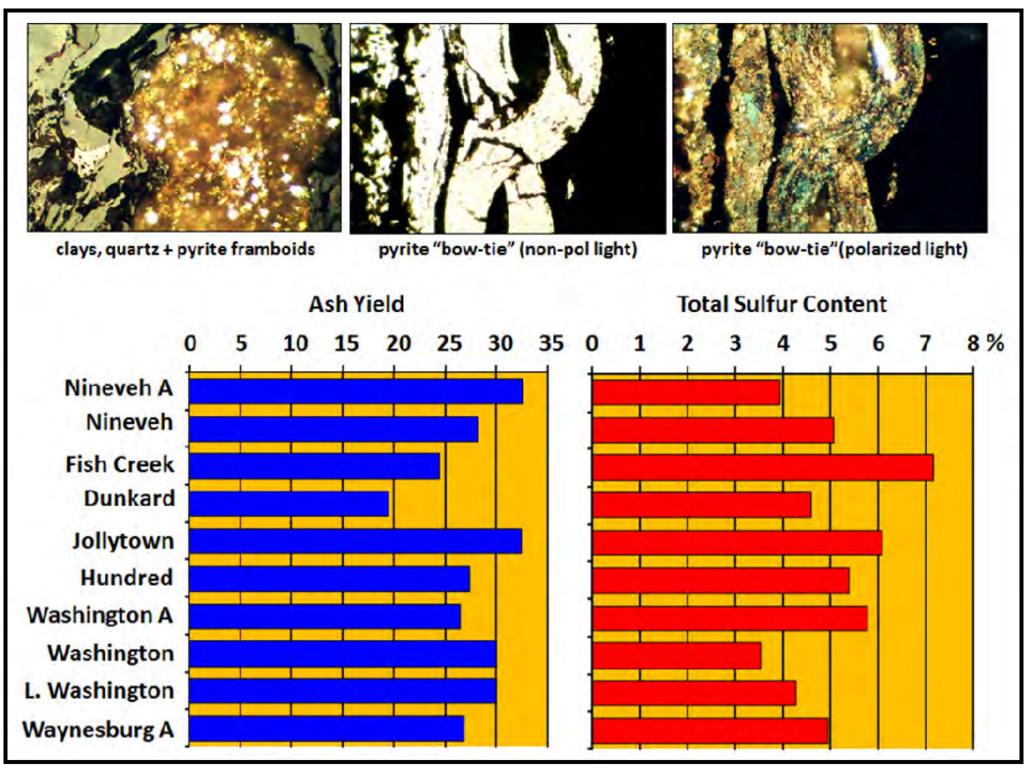

Figure 3 - Average ash yield and total sulfur contents of Dunkard coal beds. Photographs of common minerals found in Dunkard coals are from polished surfaces, with reflected light (magnification 960X).

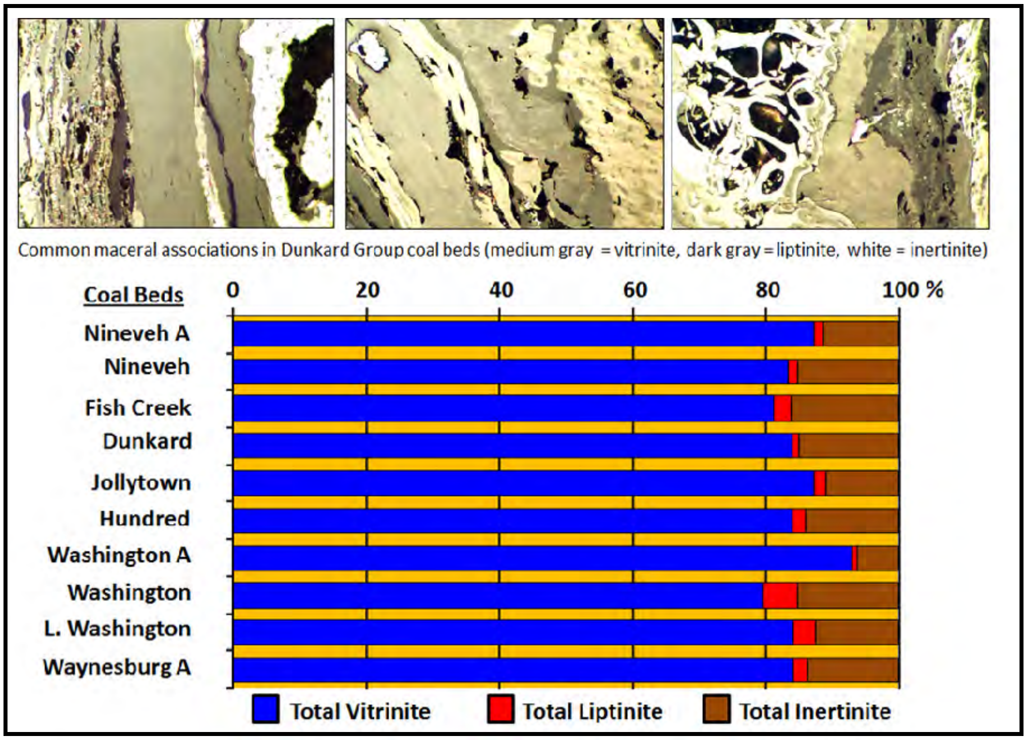

Figure 4 - Average distribution of vitrinite, liptinite and inertinite (coal macerals) in Dunkard coal beds. Maceral photographs are from polished surfaces, with reflected light (magnification 960X).

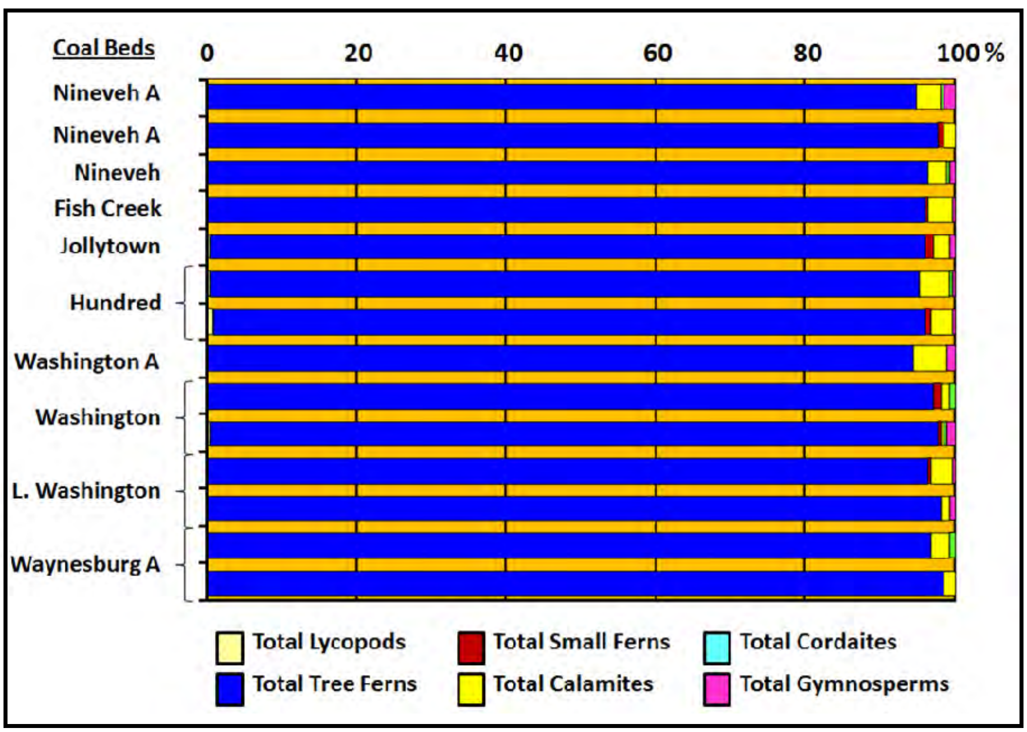

Figure 5 - Distribution of spores and pollen, grouped according to plant affinity, in selected Dunkard coal samples.

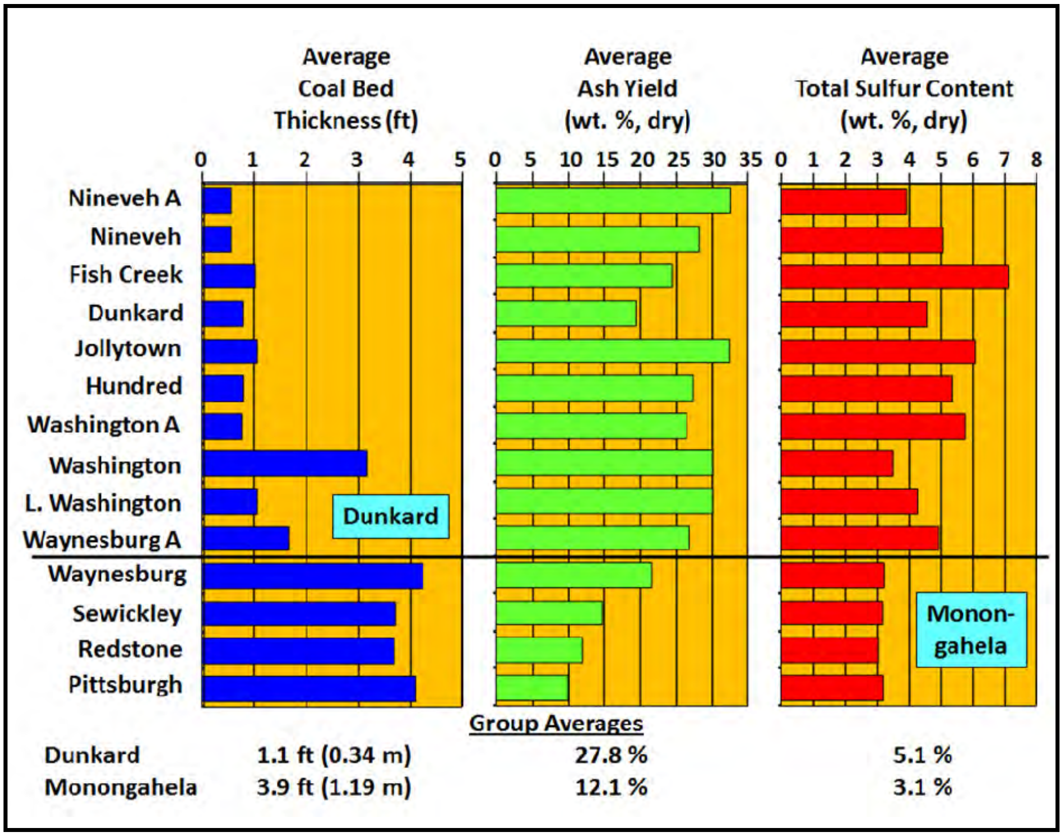

Figure 6 - Comparison of coal bed thickness, ash yields and total sulfur contents for Dunkard and Monongahela Group coal beds.

the coals are dominated by tree fern spores, which on average, comprise 96.3% of the overall pollen/spore floras (range 94.4% to 98.4%). *Punctatisporites minutus* is the most abundant tree fern spore in Dunkard coals (average = 82.8%), followed by *Laevigatosporites minutus* (average = 12.0%). Contributions from other plant groups are minor by comparison.

Comparison with Monongahela Group Coal Beds

The Washington coal represents the stratigraphically-youngest coal in the northern Appalachian basin that attains both mineable thickness (average thickness = 1 m [3.2 ft]), and exhibits lateral continuity on a regional scale. In this respect, the Washington coal resembles mineable coals of the underlying Monongahela Group (Pittsburgh, Redstone, Sewickley and Waynesburg, Figure 6), which average 1.2 m (3.9 ft) in thickness. However, all of the coal beds above the level of the Washington coal typically are thin (average thickness = 0.2 m [0.8 ft]), and discontinuous.

Palynologically and petrographically, Dunkard coal beds are essentially indistinguishable from Monongahela Group coals. Coal beds of both groups are high in vitrinite, with correspondingly low percentages of inertinite and liptinite. Coal palynology indicates paleoswamp floras strongly dominated by tree ferns. Geochemically, Dunkard Group coals contain more than twice the amount of ash (average = 26.8%) of Monongahela Group coal beds (average = 12.1%). Total sulfur contents of Dunkard coal beds (average = 5.1%) are nearly twice that of Monongahela Group coals (average = 3.1%, Figure 6).

Dunkard Group Coal Beds – Late Pennsylvanian or Early Permian?

Palynologic identification of Permian age strata relies principally on the recognition of common (5 – 10%) to abundant (>10%) amounts of "bisaccate-striate" pollen, which were produced by conifers. Common genera include *Protohaploxypinus, Striatopodocarpites, Striatoabietites, Hamiapollenites* and *Vittatina*. These (and additional) forms have been reported from the U.S. by Wilson (1962), Jizba (1962), Shaffer (1964), Gupta (1977), Winston (1983) and Willard (1992). The nearly complete lack of bisaccate-striate pollen in Dunkard Group strata led Clendening (1970, 1975) and Clendening and Gillespie (1972) to conclude that the entire Dunkard Group was Late Pennsylvanian, and not Early Permian, in age. However, it must be pointed out that most of Clendening's samples were from coal and carbonaceous shale, lithologies that are indicative of wet, "lowland" environments. However, coniferous plants, some of which would be producing bisaccate-striate pollen, are generally regarded as plants of seasonally dry environments, habitats not particularly suitable for the accumulation of organic material. As such, the lack of characteristic Permian conifer pollen reported in Clendening's (1970, 1972, 1975) work may simply be the result of sample set biased towards, and reflective of, a lowland flora. Coniferous plants may have been, and probably were, present, but not in the lowland areas that were sampled.

In addition, a majority (all?) of the Dunkard Group may have been deposited during a time of global polar ice fluctuations (Fielding et al., 2008), and there is evidence in the Dunkard of fluctuating depositional conditions driven by factors external to the basin (allocyclic controls, Beerbower, 1961). As such, it is quite possible that if conifers were present regionally, they would have been in the Dunkard basin at different times in geological history than wetland plants, alternating with them as basinal climate fluctuated on a time scale of orbital-forcing mechanisms.

Comparison with an Early Permian coal from Texas

An example of the effect depositional environment can have on palynofloras can be demonstrated from a Texas coal that was recently analyzed. The coal was recovered from the Rendova Loretta #1 discovery well, drilled in the Kerr basin in west-central Texas (Figure 7). Fusilinids sampled from strata that bracket the coal are indicative of an Early Permian (Wolfcampian) age (Hollingsworth, 1960).

The coal, which was cored principally to determine if it had any potential as a coal bed methane (CBM) resource, is thick (1.75 m [5.74 ft]), and high in ash (average = 36.6%). Petrographically, the coal is high in vitrinite (average = 86%, mmf), with correspondingly low to moderate amounts of inertinite (average = 10%, mmf) and liptinite (average = 4%) (Barker et al., 2003).

Palynologic analysis of closely-spaced (~6.4 cm [~2.5 in]) bench samples (Figure 7) indicate that the coal is dominated by tree fern spore taxa (average = 83.3%), especially *Thymospora thiessenii* (average = 33.8%) and *Punctatisporites minutus* (average = 40.6). Calamite spores were the second most abundant plant group (average = 12.4%). Even though the coal is of Early Permian age, based on fusilinid evidence, bisaccate-striate conifer pollen is virtually absent. Indeed, the coal palynology is very similar, if not identical, to Late Pennsylvanian Monongahela Group (Late Pennsylvanian) coal spore floras from the Appalachian basin.

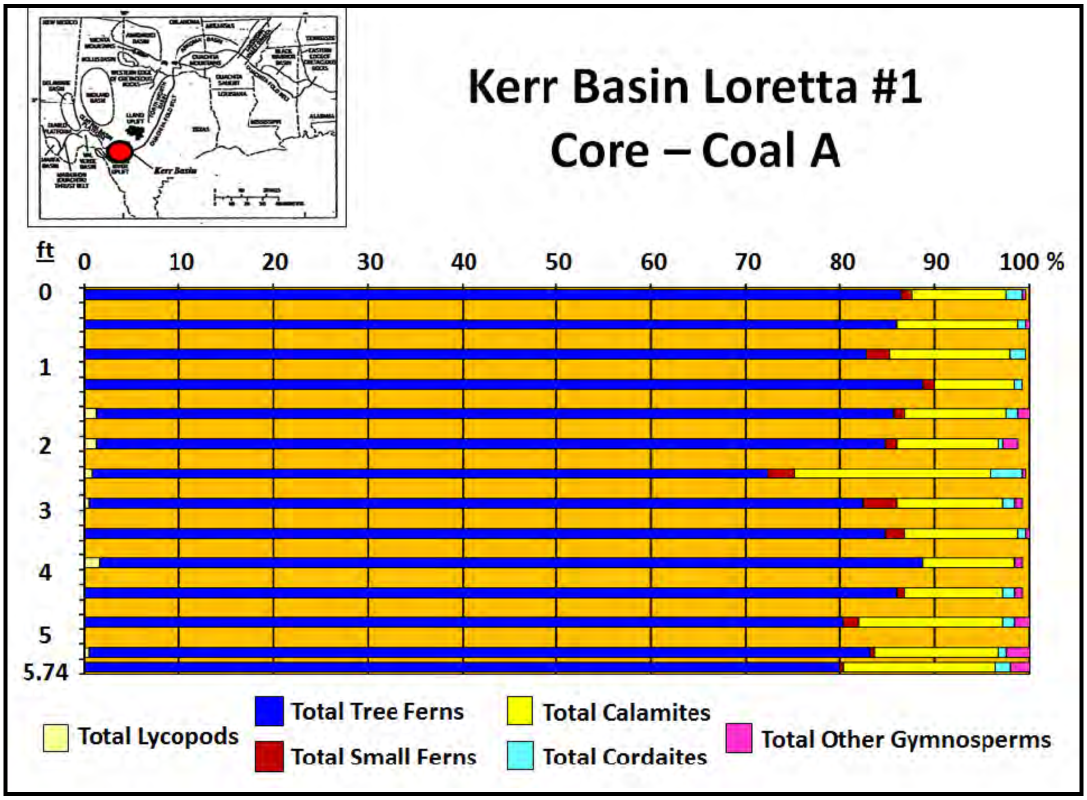

Figure 7 - Distribution of spores and pollen, grouped according to plant affinity, in an Early Permian (Wolfcampian) coal bed from the Kerr basin in east-central Texas. Although Permian in age, the coal is nonetheless dominated by tree fern and calamite spores.

The important point here is that floral elements that inhabited lowland and swamp environments during the Late Pennsylvanian and Early Permian were more similar than dissimilar. As such, coal palynology serves as a poor proxy to identify and differentiate Late Pennsylvanian from Early Permian strata, simply because of strong ecological biases. Future work should be directed at sampling environments that have better potential of recording the presence of floras that were able to develop in more water-stressed areas.

SUMMARY

In the Dunkard Group, the Washington coal bed, located near the base of the group, is the last coal in the entire Pennsylvanian/Permian section of the Appalachian basin that attains mineable thickness (average = 1 m [3.2 ft]) across a relatively broad area. As such, the Washington coal is the stratigraphically-youngest coal in the Appalachian basin that is considered to be a "coal resource" (West Virginia Geological & Economic Survey coal resource data, www.wvgs.wvnet.edu). All of the coal beds above the Washington are thin (average = 0.2 m [0.8 ft]), and many are laterally discontinuous.

Geochemically, Dunkard Group coal beds are high in ash yield (average = 27.8%) and total sulfur content (average = 5.1%). Petrographically, Dunkard coals are high in vitrinite (average = 84.8%, mmf), with correspondingly low to moderate amounts of liptinite (average = 2.1%, mmf) and inertinite (average = 13.1%, mmf). Palynologically, Dunkard coal beds are all dominated by tree fern spore taxa (average = 96.3%), especially *Punctatisporites minutus* (average = 82.8%). Calamite spores are the second most abundant plant group (average = 2.4%), with others (lycopods, small ferns, cordaites and other gymnosperms) having very minor representation.

Coal beds of the underlying Monongahela Group (Pittsburgh, Redstone, Sewickley and Waynesburg) are thicker (average = 1.2 m [3.9 ft]), and lower in ash (average = 12.1%) and sulfur (average = 3.1%) than their Dunkard counterparts. Palynologically and petrographically, Dunkard Group coal beds are indistinguishable from coals of the underlying Monongahela Group.

Collectively, Dunkard swamps were all planar and topogenous, their formation being controlled by topography and moisture availability. The lifespans of swamps above the level of the Washington coal are difficult to assess. Thin, discontinuous coal beds can result from short-lived conditions favorable to organic accumulation, or they may result from high-decay rates during marginally favorable conditions (considering the creation of accommodation space to be constant). Alteration of the peat was controlled largely by anaerobic, rather than aerobic, processes. Anaerobic degradation contributed to coals with high vitrinite contents, and low to moderate inertinite contents. Anaerobic conditions also promoted the formation of abundant pyrite, the dominant source of sulfur in the coal. It is also likely that some portion of the ash in the coal was authigenic in nature, resulting from the decay of organic material. A progressive loss of organic material through anaerobic destruction effectively concentrates the inorganic portion of the original plant material, which helps explain the high ash yields of Dunkard coals (Cecil et al., 1982).

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PENNSYLVANIAN-PERMIAN VEGETATIONAL CHANGES IN TROPICAL EURAMERICA

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Spencer G. Lucas, Richard Lupia, W. John Nelson, and Neil J. Tabor

INTRODUCTION

Vegetational changes across the Pennsylvanian-Permian boundary are recorded in several largely terrestrial basins across the Euramerican portions of equatorial Pangea. For the purposes of this paper, these include the Bursum-Abo Formation transition and its equivalents in several small basins in New Mexico, the Halgaito Formation of southeastern Utah, Markley Formation of the eastern shelf of the Midland Basin in north-central Texas, Council Grove Group of northern Oklahoma and southern Kansas, and Dunkard Group of the central Appalachian Basin. This transition also is recorded in numerous basins in Europe, reviewed by Roscher and Schneider (2006), based on paleoclimate indicators preserved in those regions. Collectively, these deposits form a west-to-east transect across the Pangean paleotropics and thus provide a paleogeographic setting for examination of both temporal and spatial changes in vegetation across the Pennsylvanian-Permian boundary (Figures 1 and 2).

The Pennsylvanian-Permian transition records the change from wetland vegetation as the predominant assemblages found in the plant fossil record, to seasonally dry vegetation. This has often been called the “Paleophytic-Mesophytic” transition, a concept that is flawed

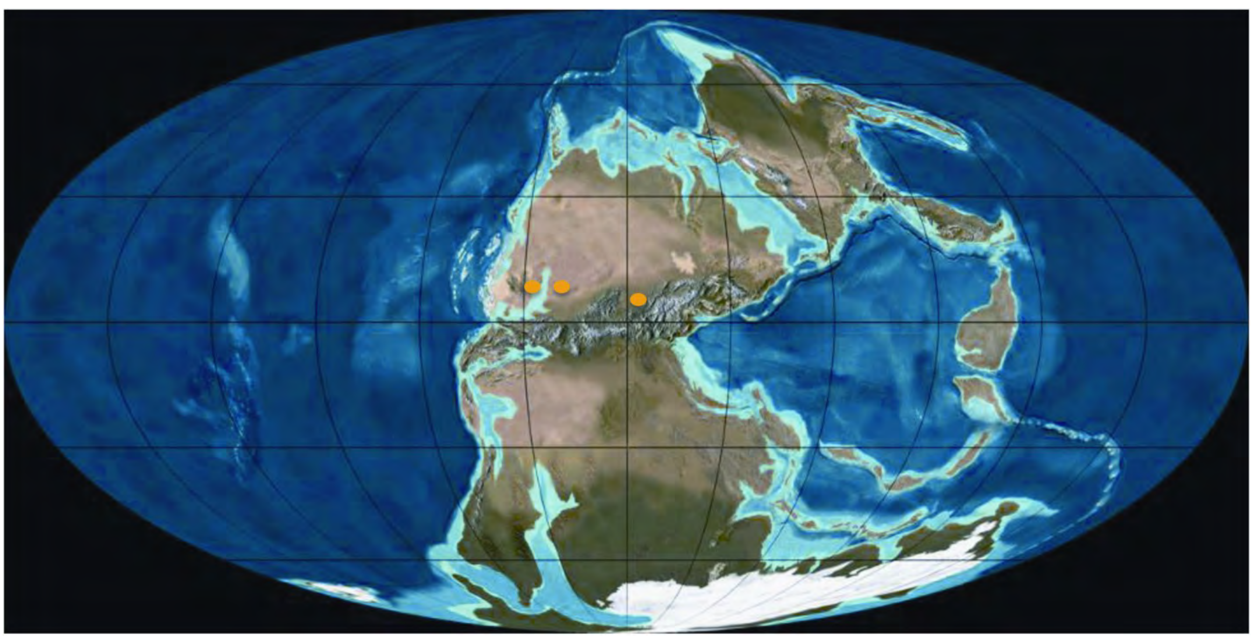

Figure 1. Continental configuration at the Pennsylvanian-Permian boundary. Yellow ovals indicate the principal areas discussed herein: Left – New Mexico and Utah, Center – Texas and Oklahoma, Right – Central Appalachians/Dunkard. Map courtesy of Ron Blakey, Northern Arizona University.

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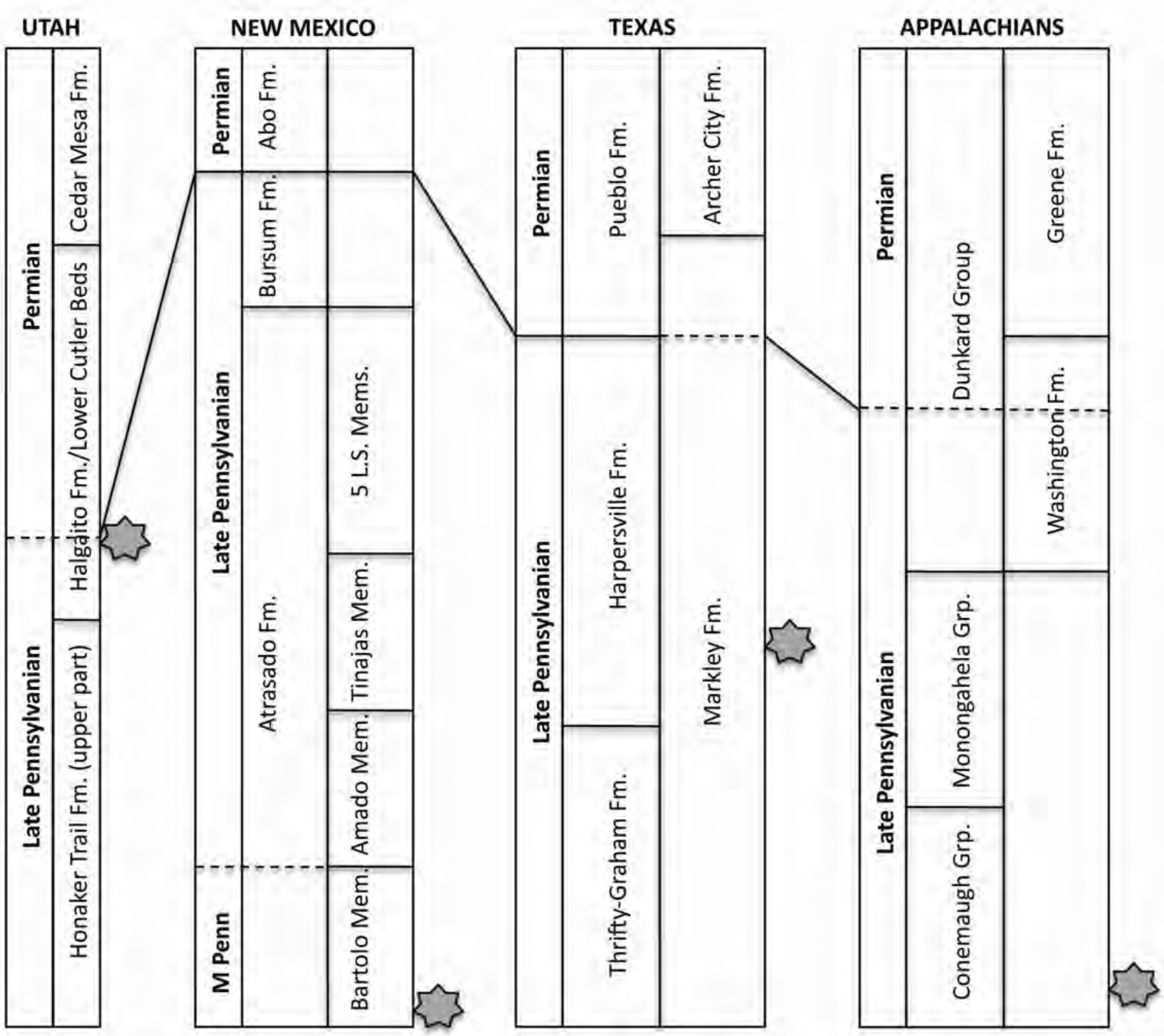

Figure 2. Stratigraphic sections for the areas discussed herein, positioned West-to-East along the Pangean paleoequator. Stars indicate the position of the first appearance of seasonally dry flora in the local stratigraphic section. Solid lines are accepted temporal boundaries or boundaries of formations or groups. Dotted lines are inferred temporal boundaries.

conceptually but does reflect a pattern of vegetational change at a certain coarse level of resolution (DiMichele et al., 2008). Interpreted simplistically, this change is asynchronous across the Pangean tropics. In detail, however, it is clear that "dry" floras appear as early as the Middle Pennsylvanian in central and western Pangea (Scott et al., 2010; Falcon-Lang et al., 2009; Plotnick et al., 2009; Galtier et al., 1992; Dolby et al., 2011), and that coals are effectively gone from the stratigraphic record of far western Pangea after the end of the Atokan (Bolsovian, mid-Moscovian) (Lucas et al., 2009a). In contrast, organic facies persist, even if much reduced in thickness, quality and abundance, together with associated wetland plants, into the Early Permian in central and east-central Pangea, at the same time as deposits dominated by plants of seasonally dry environments are increasing in frequency (e.g., Martin, 1998; Kerp and Fichter, 1985).

When the complexity of this transition is embraced, it is found that its nature takes on different qualities depending on the spatial and temporal scale at which it is resolved and examined. On the temporal scale of glacial-interglacial cycles, floras consisting of species tolerant of seasonal moisture deficits are found intercalated with floras of humid environments, resulting in the alternation of taxonomically distinctive assemblages, at least on the northern side of the Appalachian-Variscan central Pangean mountain range (Mapes and Gastaldo, 1986; Broutin et al., 1990; DiMichele and Aronson, 1992 – the history and dynamics of this mountain belt were complex, such that the mountain range was not contemporaneously present across the Pangean continent, as is frequently shown in paleogeographic reconstructions – see discussion in Roscher and Schneider, 2006). In some instances, this floristic alternation can occur between different beds of the same outcrop (DiMichele et al., 2005a; Falcon-Lang et al., 2009). The changing floras reflect climate contrasts that have been interpreted to change in concert with glacial-interglacial cyclicity (Falcon-Lang, 2004; DiMichele et al., 2010c; Falcon-Lang and DiMichele, 2010). Such vegetational cyclicity is interpreted in this paper to reflect shifts between cool-wet (glacial) and warm-dry (interglacial) (Cecil and Dulong, 2003) tropical climate states. Furthermore, the cyclicity observed for vegetation, and inferred for climate, occurs within depositional packages interpreted to result from sea-level changes, recorded in the classic Pennsylvanian cyclothems. Also seen are climate trends on longer time scales of 2-6 million years (Darrah, 1969; Cecil, 1990; Falcon-Lang et al., 2011a; Allen et al., 2011) linked to extended periods of drier climate, higher mean sea-level, and reduced polar ice volume (Fielding et al., 2008; Rygel et al., 2008; Eros et al., in press). At the longest temporal scale is the long-recognized trend toward drier tropics manifested in the Pennsylvanian-Permian transition in the western and central parts of equatorial Pangea (Remy, 1975 for discussion from a paleobotanical perspective). This longest-term trend is not monotonically unidirectional, but is an average of an oscillatory trend ultimately leading to a reduction in the number and length of intervals of significantly wet tropical climate. Within this longer-term trend it might be said that climate fluctuations continue, but that, on average, the wet periods become less wet and the dry periods become more dry. Cecil et al. (2011) suggest that this trend is due either to the final assembly of Pangea or to the rise of the central Pangean mountains. The trend happens, inexplicably, at a time when evidence has been adduced for the return of large amounts of ice at the beginning of the Permian (e.g., Montañez et al., 2007; Fielding et al., 2008).

BASES OF INFERENCE/RATIONALE

The environmental preferences of fossil plants can be resolved in many environmental dimensions. Very commonly, however such preferences are broadly construed as favoring habitats with year-round high soil-moisture content or environments where there are seasonal deficits of moisture. These preferences are determined from plant architecture, growth habit, and environmental correlates of occurrence, including the sedimentary environments in which the fossils are found. Such inferences come from the literature and from the personal experiences of the authors, which are both published and unpublished. The extent to which such patterns are reflective of regional climates is contextual, dependent on broader spatial and temporal patterns of floristic occurrence, the sedimentological context of the floras, and the extent of the depositional systems of which they are a part. Consequently, the focus herein is more on the plants and less on the accessory information that accompanies the paleobotanical samples. That said, there are clear preservational biases in the fossil record of plants. Paleosols tell us that vegetation was present during the times of drier climate, but we have a limited understanding of that vegetation, at least at present, despite a considerable number of publications that discuss it (for recently published papers see, for example, Feldman et al., 2005; Falcon-Lang et al., 2009; Plotnick et al., 2009; Scott et al., 2010; DiMichele et al., 2010c; Falcon-Lang et al., 2011a; Dolby et al., 2011). The dry soil plants do become well represented in the Permian fossil record of the equatorial region, reflecting, perhaps, changed sedimentary systems and changed paleobotanical search images (Kerp, 1996).

To the extent possible, we consider information from paleosols relevant to rainfall and water table, sedimentary patterns, geochemical proxies for various aspects of climate, climate modeling, near-field studies of the ice record, and other sources of information (e.g., Cecil et al., 2003; Feldman et al., 2005; Montañez et al., 2007; Tabor and Montañez, 2004; Poulsen et al., 2007; Tabor and Poulsen, 2008; Tabor et al., 2002, 2008; DiMichele et al., 2010c; Horton et al., 2010).

In a glacial world, there also is very likely to be a close, mechanistic relationship among tropical climate, sediment transport patterns, global sea-level dynamics, polar ice-volume, and the composition of the atmosphere. This is framed broadly, though not entirely, by considering rock stratigraphic sequences in terms of a combined climate-sea level-tectonics framework (Cecil and Dulong, 2003; Poulsen et al., 2007; Elrick and Nelson, 2010; Allen et al., 2011; Eros et al., in press). Here we are operating under the presumption that sea-level lowstand, particularly from its midpoint into the earliest phases of sea-level rise, is associated with the wettest equatorial climates. If peats/coals are present in a cycle, this is the most likely time for them to have formed. The driest equatorial climates occur at sea-level highstand and continue, diminishing, into the phases of sea-level fall.

It is during these stages of the cycle that limestones, evaporites, and paleosols indicative of seasonal climates are most likely to form. This model (Cecil et al., 2003; Poulsen et al., 2007; Eros et al., in press) has been slow to emerge for several reasons. Perhaps foremost is incumbency. Earlier models, such as that of Bohacs and Suter (1997), which call upon sea-level rise to initiate and sustain cratonic-basin blanket-peat formation, treat climate as a constant (e.g., p. 1618, where they note that "only accommodation varies significantly and most other variables remain constant (flora, climate, environment, etc.)"). This, in effect, removes climate as a variable in their sequence stratigraphic model, without specifying the state of that invariant climate. Such models, therefore, attribute all changes in lithological sequences to sea-level changes. When considered within a "total evidence" framework, however, these models fall short. For example, were sea-level rise the driver of peat formation, it is then necessary to explain why, during the late Paleozoic, peat beds did not form in the western parts of the Pangean equatorial region while, at the same time, extensive peats were forming in the Midcontinent and eastern coal basins (e.g., Cecil et al., 2003; Bishop et al., 2010). Furthermore, modern sea-level rise has not created vast peat-forming swamplands along coastlines worldwide.

The prevailing understanding that the modern tropics are wettest during interglacials and dry during glacial phases of glacial-interglacial cycles is not as much an iron-clad truth as assumed, and is changing as greater study of historical Amazonian climate patterns emerges. The Amazonian lake record is consistent with an equatorial band of the Amazon being wettest at or near glacial maximum, possibly caused by constraint of the intertropical convergence zone (ITCZ) (e.g., Bush et al., 2002). In addition, the great majority of modern Amazonian interglacial rainfall reflects recycling of water by the tropical rainforest (e.g., Bush and Flenley, 2007), something that may be entirely dependent on the presence of angiosperms and their unique water transport and evapotranspirative capacities (Boyce and Lee, 2010; Boyce et al., 2010). The latter point means that pre-angiospermous systems, such as those of the late Paleozoic, may not have been able to create “rainforest” levels of moisture over large areas during inter-glacial episodes (Remy, 1975). Without the powerhouse evapo-transpiration created by the angiosperms, the implication is that tropical wetness must have been generated by a largely physical mechanism (e.g., the restriction of the ITCZ by polar ice and attendant atmospheric conditions at the ice front – Cecil et al., 2003).

SPATIAL PATTERNS WITHIN THE PALEOEQUATORIAL REGION

The main geomorphological trend that is most easily observed and tracked in space and time is the presence or absence, or relative abundance, of coal and the subjacent “underclay” paleosols (Cecil et al, 2003; Cecil et al., 2011). Well developed “underclay” paleosols can often be traced in somewhat elevated areas well beyond the occurrence of the superjacent coal beds where soils with aquic to udic soil moisture regimes inhibited or even excluded the development of peraquic conditions that are conducive to the development of histosols (peat precursors to coal) on the land surface (Cecil et al., 2011). The physical, chemical, and mineralogical properties of the underclay paleosols beneath coal beds clearly indicate low stand sea level conditions and pedogenesis under a humid to perhumid paleoclimate. Thus, coal represents the “wettest” (glacial) climate phase of a glacial-interglacial cycle when rainfall was sufficient to perpetually maintain the water table at the surface.

Also relatively readily observed as indicators of climate are seasonally-dry floral elements. It is possible to assess: (a) their “on-average” abundance or the commonness with which they are encountered in a particular geological unit, such as a formation, (b) their degree of mixing with the more hygromorphic (wetland) floral elements, and (c) the time, or times, at which they first appear and/or become the most commonly encountered vegetational elements. With regard to this latter point, “dominance” is a term we will use with caution and constrain to relative abundance within a single fossil assemblage, which is considered herein a proxy for a natural plant community. The term “dominance” when expanded to larger temporal or spatial scales implies the existence of a single vegetation type or species pool in equilibrium with, or responding to, a monotonic climate. Rather, we see two or more distinctive, climatically characteristic species pools reflective of different mean climate states (sensu Cecil et al., 2011). The presence of these different species pools in sedimentary basins is a response to changes in soil moisture. Soil moisture changes, allowing for the ever-present effects of local edaphics, may reflect changes in regional climate brought about by glacial-interglacial cyclicity. As mentioned previously, we consider climatic cyclicity within a single glacial-interglacial cycle to be wet/cool during sea level lowstands (maximum ice volume) and seasonally dry during sea level highstands (minimum ice volume). These glacial-interglacial patterns, on the 10^5-year time scale, will be superimposed on the longer term changes in climate, on the 10^6, 2 to 6 million year scale (Allen et al., 2011), when mean climate state was overall warmer-drier or cooler-wetter, reflective of changes in mean global ice volume and sea-level. And patterns on both these time scales will be encapsulated within the long term Pennsylvanian-Permian climate change, on the 10^7-year time scale.

At the 10^7-year, Pennsylvanian-Permian time scale, there clearly is a trend toward overall drier climate beginning in the late Middle Pennsylvanian and continuing through the Early Permian (Remy, 1975; Roscher and Schneider, 2006; Tabor and Poulsen, 2008), perhaps attributable to the final assembly of the mega-continent Pangea, the northward drift of the continental landmass, and changes in atmospheric and oceanic circulation (Cecil, 1990; Parrish, 1993; Roscher and Schneider, 2006; Tabor et al., 2008). During glacial-interglacial oscillations, this means that the wetter phases became less wet and the drier phases became even drier. This pattern, ultimately, permits the survival in basinal lowlands of plants more tolerant of moisture seasonality, at all phases of the cycle. As basinal landscapes “dry out” over time, the wetland specialist plants become less common and often are spatially concentrated, occurring together as habitat specialists in reduced diversity assemblages in landscape “wet spots” (DiMichele et al., 2006).

Westernmost Pangea: New Mexico and Southeastern Utah

During the late Paleozoic, present day New Mexico was located in the western part of equatorial Pangea as an archipelago of basement cored uplifts surrounded by shallow marine and, in some cases, locally terrestrial basin floors of the Ancestral Rocky Mountain basin and range (Figure 1). This region of the supercontinent appears to have been persistently more seasonally dry than more central parts of the megacontinent throughout the later Pennsylvanian and into the Early Permian, with a distinct monsoonal pattern of precipitation, evidently sourced from Panthalassa rather than Tethys (Parrish, 1993; Tabor and Montañez, 2004; Tabor and Poulsen, 2008). Glacial-interglacial scale lithological cyclicity is more obscure in this region than in the central Pangean coal basins due in large part to the near absence of coal/underclay as a low-stand marker bed. Cyclicity can be detected in some of the marine (Elrick and Scott, 2010) and terrestrial Pennsylvanian sections, and in mixed marine and terrestrial Lower Permian rocks (Mack et al., 2010). It also can be detected as the higher-level, stage-scale fluctuations in climate state (sensu Cecil et al., 2011), e.g., a much drier Missourian (Kasimovian) than either the earlier Desmoinesian (Moscovian) or the later Virgilian (Falcon-Lang et al., 2011a), and in the same overall drying trend that characterizes the Pennsylvanian-Permian transition in general (Remy, 1975; Cecil et al., 1985; Cecil, 1990).

Our studies in New Mexico reveal the shift from wetter-to-drier climates earlier than any other area we have examined in detail. In the central part of the state, Socorro County, a nearly complete lithological transition, with associated flora and both marine and terrestrial fauna is preserved from the Middle Pennsylvanian to the Early Permian (Lucas et al., 2009a; Krainer and Lucas, 2009). The Pennsylvanian-Permian transition is well exposed in both this central region, and to the north in areas near Albuquerque (Lucas and Krainer, 2004; Krainer and Lucas, 2004; Lucas and Zeigler, 2004), and still farther north in areas to the west of Sierra Nacimiento, east of Cuba (Lucas et al., 2010a, b; Krainer and Lucas, 2010; DiMichele et al., 2010a). Seasonally-dry floral elements begin to appear in the Middle to Late Pennsylvanian in each of these areas (depending on the age of the oldest exposed terrestrial rocks) and are a part of nearly every plant assemblage collected thereafter, becoming outright the overwhelmingly most common vegetation type either in the later part of the Late Pennsylvanian, in the Bursum Formation or in the wholly Lower Permian Abo Formation (e.g., Mamay and Mapes, 1992; Utting et al., 2004; Tidwell and Ash, 2004; DiMichele et al., 2004; DiMichele et al., 2007; Lucas et al., in press a). Geological ages throughout New Mexico are among the best constrained of any of the Pennsylvanian-Permian transition regions considered in this paper, all accurately calibrated with conodonts and/or fusulinids from adjacent and intercalated marine strata (e.g., Lucas et al., 2009a).

Floras from these areas require some explanation because of their mode of preservation. In the Atokan (Bolsovian/middle Moscovian) Sandia Formation, coal beds are represented only by a few isolated lenses, geographically very localized (e.g., Thompson, 1942; Armstrong et al., 1979; Kues and Giles, 2004; Krainer et al., 2011) and not thick enough or laterally extensive enough to be of economic significance. They are associated with wetland floras (Lucas et al., 2009b), but are not even remotely comparable in extent to coal-forming paleoenvironments of similar age in the eastern USA and in Western Europe. Paleosol deposits, so called “fire clays”, associated with the coal beds have been mined locally for brick and pottery (Lucas et al., 2009b), and indicate periods of humid to perhumid climates and intense weathering.

The Desmoinesian (Asturian/late Moscovian) in New Mexico reflects a time of elevated sea levels relative to the earlier Pennsylvanian, as across much of the Ancestral Rocky Mountain zone, with local tectonics creating an essentially non-marine basin floor setting in an area east of Socorro, central New Mexico, that persisted through the late Missourian (Lucas et al., 2009a). In this area, glacial-interglacial-scale oscillation of wetland and seasonally-dry floras can be detected in parts of the section. The wetland floras, although rare, are authochthonous to parautochthonous and completely dominated by typically Pennsylvanian wetland plants, such as pteridosperms and *Psaronius* tree ferns. These assemblages appear to be preserved in lowstand coastal-plain deposits, based on their sequence position (lowstand), and sedimentary characteristics (thin, widespread, autochthonous, rooted siltstone deposits). They are intercalated with shales that appear to represent marine high-stand to falling-stage shallow shelf deposits, containing brackish-to-marine fauna and allochthonous plant remains (Figure 3). These latter floras contrast sharply with those of the lowstand, wetter phases, and contain predominantly the xeromorphic/seasonally dry plants, such as *Sphenopteris germanica*, *Charliea*, and walchian conifers, with admixtures of wetland elements, mainly pteridosperms (DiMichele et al., 2010b). The Missourian of this area is particularly dry (Lerner et al. 2009; Falcon-Lang et al, 2011a) (Figure 4), as it appears to be across the North American part of the Pangean tropics (Cecil et al., 1985; Cecil, 1990; Martino and Blake, 2001; Martino, 2004; DiMichele et al., 2010b; Cecil et al., 2011). Floras preserved in Socorro County vary from completely dominated by seasonally-dry substrate species, including conifers, callipterids, *Charliea*, *Taeniopteris* and *Sphenopteris germanica* (Falcon-Lang et al., 2011a), to mixed, with wetland elements forming a background to primarily seasonally-dry substrate taxa (Lerner et al., 2009). A coeval flora of Missourian age is known from 100 km (62 mi) to the north, near Albuquerque, from the Kinney Brick Company quarry in the Manzanita Mountains (Mamay and Mapes, 1992). This deposit was formerly thought to be of middle Virgilian age, is dominated by wetland plants, but has a well represented and diverse array of seasonally dry-substrate taxa, including *Sphenopteris germanica* (*Sphenopteridium manzanitanum*, Mamay, 1992), *Charliea manzanitana* (Mamay, 1990), *Dicranophyllum readii* (Mamay, 1981), several forms of conifer and pteridosperms typically found in drier substrate settings (Lucas et al., in press a).

At the Pennsylvanian-Permian transition, including the Bursum (and its equivalents) and lower Abo formations (Lucas and Krainer, 2004; Krainer et al., 2009; Krainer and Lucas, 2010; Mack et al., 2010), floras are regionally variable and show different mixtures of wetland and seasonally dry elements. In Socorro County, Bursum floras are preserved in coastal plain environments (Figure 5A) that are not substantially different from those of the Desmoinesian and Missourian in the same area. These floras are, however, dominated by dryland species,

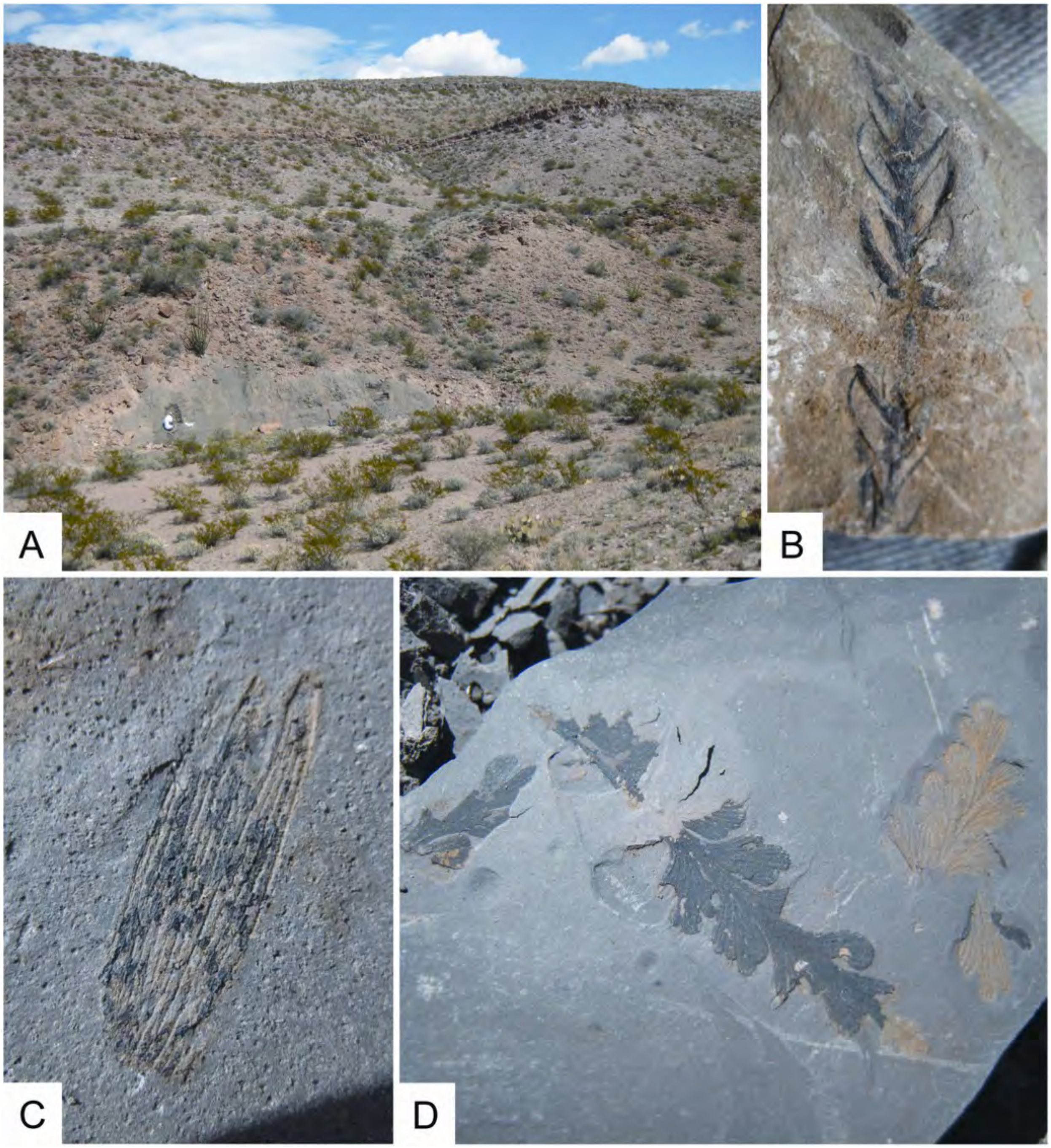

Figure 3. Desmoinesian (Middle Pennsylvanian) age exposure and plants from the Bartolo and Tinajas members of the Atrasado Formation, Socorro County, New Mexico. A—Bartolo Member – two lower shales separated by calcareous sandstone, each contain allochthonous, seasonally dry flora. Crest of ridge marks the position of the Amado Member, immediately above the Desmoinesian-Missourian boundary. B—Walchian conifer twig with leaves (USNM 543964; USNM loc 43527). C—*Charliea* pinnule (USNM 543967; USNM loc 43469). D—*Sphenopteris germanica* pinnae (USNM 508828; USNM loc 43469). All specimens approximately 3 X magnification.

including walchian conifers and cordaitaleans. The similar sedimentary environments permit an isotaphonomic comparison (sensu Behrensmeyer and Hook, 1992) and suggest that the wettest intervals of climate cycles were significantly drier than during earlier time periods in the Desmoinesian through Virgilian. In the Lucero Uplift, 80 km (50 mi) to the NNW of the

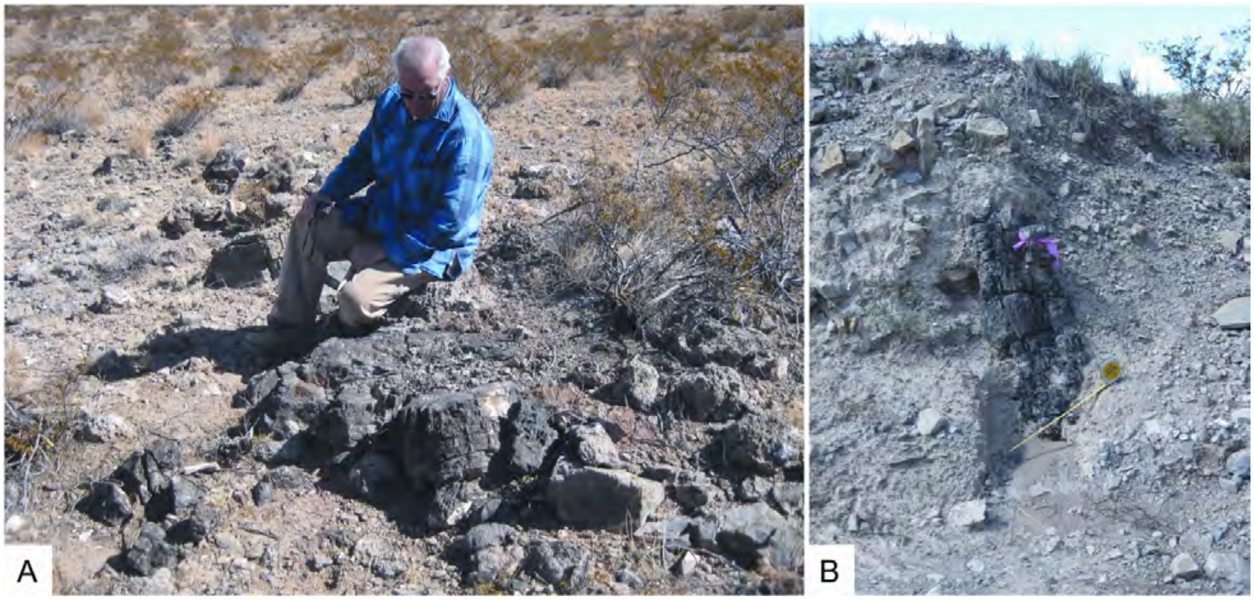

Figure 4. Missourian (Late Pennsylvanian) gypsum-carbonate sabkha deposit, Tinajas Member of the Atrasado Formation, Socorro County, New Mexico. The bases of coniferopsid (conifers or cordaitaleans) trees are preserved within this deposit, rooted in micritic limestone and buried by gypsum and carbonate (see Falcon-Lang et al, 2011a). A—Large tree base with attached roots, extending into the basal micrite, exposed by erosion of the gypsum-carbonate bed. B—Tree trunk extending upward into the gypsum-carbonate bed (scale = 50 cm).

Socorro area, latest Pennsylvanian floras from the Red Tanks Member of the Bursum Formation are also preserved in lacustrine sediments within a coastal plain setting. Conditions were rarely humid enough to engender peat/coal formation (Krainer and Lucas, 2004). The plants are predominantly wetland taxa but with a large admixture of conifers, callipterids and taeniopterids (Carrizo Arroyo flora, see Tidwell and Ash, 2004), all elements of seasonally dry vegetation. Such a mixed flora suggests that the wetland species were confined to wet places on an otherwise seasonally dry landscape (as suggested for the regional climate by Krainer and Lucas, 2004). The northern-most Late Pennsylvanian plant-bearing strata (El Cobre Canyon Formation, a Bursum equivalent, and Lower Permian Arroyo del Agua Formation) (Lucas et al., 2010b), were deposited in braided river systems (Krainer and Lucas, 2010). The floras are numerically dominated by wetland plants, like those from the approximately equivalent aged Red Tanks Member in the Lucero Uplift, but, as there, these wetland elements appear to have been drawn from stream banks or from locally colonized areas within the shifting braidplains. Most of the assemblages contain uncommon, but consistently present, plants from substrates that were likely seasonally dry, such as walchian conifers, *Taeniopteris* and cordaitaleans (DiMichele et al., 2010a).

In all regions of New Mexico, the Early Permian rocks record the long-term transition to a seasonally drier climate mean at all stages of climate cycles (Mack et al., 1991; Krainer and Lucas, 2010). Depending on facies, the flora of these Early Permian rocks varies considerably. A major facies change takes place as a consequence of a pulse of the Ancestral Rocky Mountain uplift beginning in the early Wolfcampian (latest Pennsylvanian). This produces the synorogenic Abo Formation, which formed on a vast coastal plain, represented by channel-form fluvial deposits and overbank siltstones and mudstones (Figures 5B, 6). Abo Formation exposures are intercalated with marine deposits of the Hueco Group in southern New Mexico,

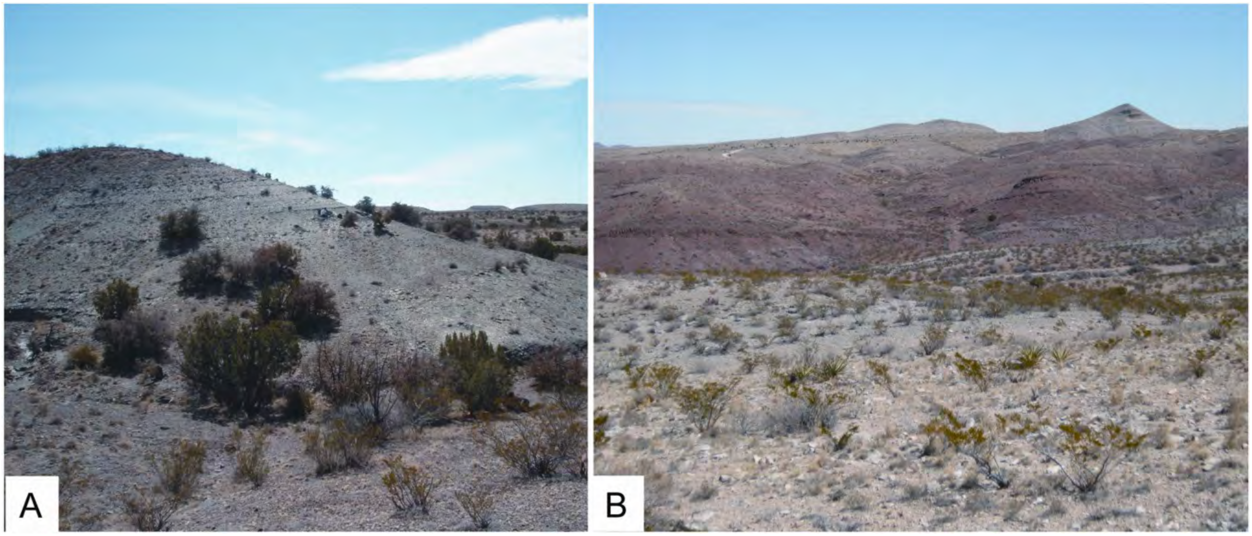

Figure 5. Wolfcampian (latest Pennsylvanian) age Bursum Formation and Early Permian age Abo Formation, Socorro County, New Mexico. A—Bursum Formation plant-bearing shales, perhaps deposited in a floodplain lake. Most of the Bursum consists of paleosols and limestones, and is without plant fossils. B—Succession in the Socorro County area: Atrasado Formation (white, limestone in foreground), Bursum Formation (gray, low slope and hills in middle foreground), Abo Formation (red hills), Yeso Formation (yellow-gray hills on skyline). Pennsylvanian-Permian boundary lies at the Bursum-Abo contact just above the center of the photograph.

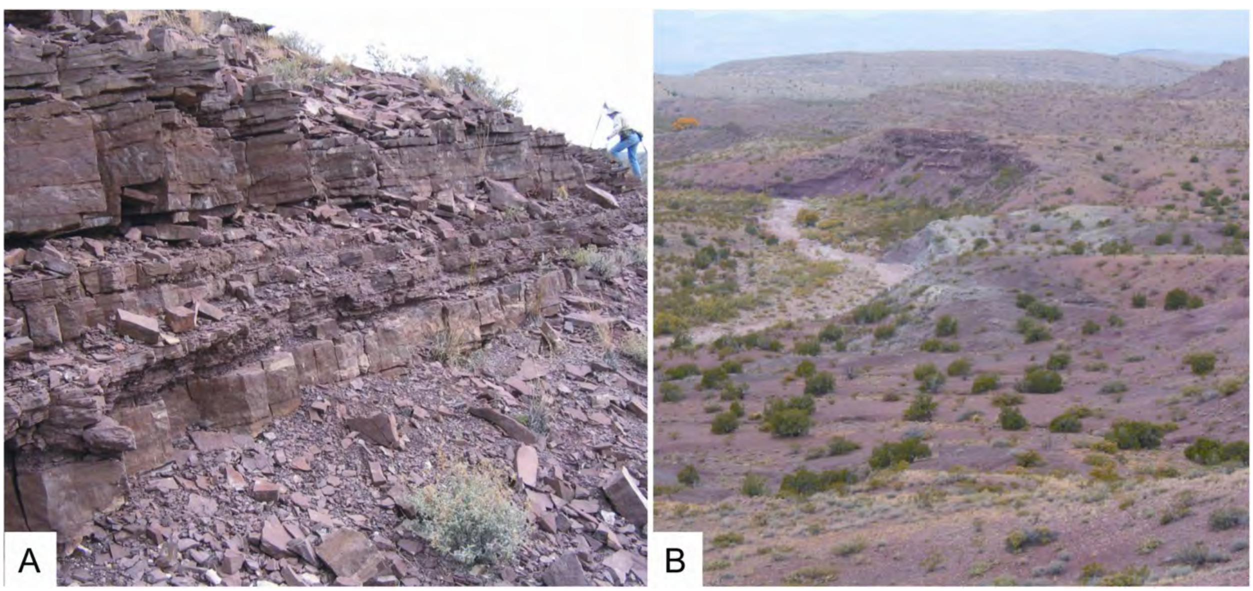

Figure 6. Early Permian age Abo Formation, Socorro County, New Mexico. A—Siltstone channel fill characteristic of deposits in which plants are commonly found. B—Lower and Middle portions of the Abo formation shown in stream cutbank. Siltstones with strong pedogenic overprint characterize the lower portions of the Abo Formation. Thin, sheet siltstones and fine grained sandstones are generally more common in the middle and upper part of the Abo Formation.

(Lucas et al., 1995), but otherwise the Abo is entirely terrestrial, with thick calcic Vertisols and stream deposits, with loessites further to the north (Mack et al., 1991; Kessler et al., 2001; Mack et al., 2003; Mack, 2007). The flora from these beds comes primarily from siltstones of the channel facies and is of remarkably low diversity (DiMichele et al., 2007), consisting of patches of various walchian conifers and the peltasperm *Supaia*, with a small, and generally rare,

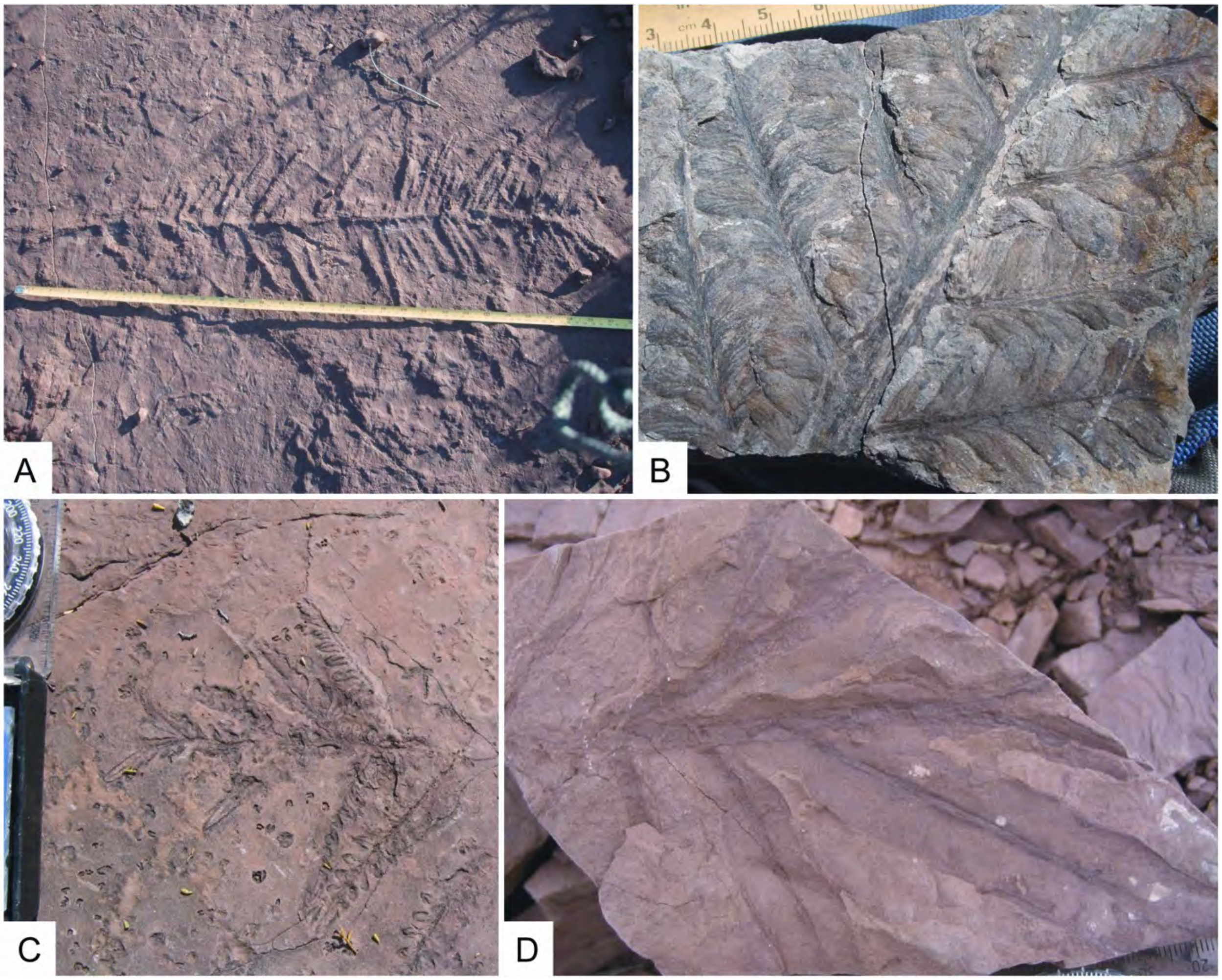

Figure 7. Plant fossils typical of seasonally dry flora, from the Bursum and Abo formations, Socorro County, New Mexico. A—Walchian conifer branch (field photo; specimen not collected). B—*Rhachiphyllum* sp., callipterid peltasperm frond segment (USNM 543963; USNM loc. 43460). C—*Autunia conferta*, callipterid peltasperm frond segment (tip) (USNM 543955; USNM loc. 42251). D—*Supaia anomala*, supaioid peltasperm leaf fragment (USNM 543956; USNM loc 42255).

admixture of callipterids and other plants of seasonally dry habitats (Figure 7). This is particularly true for the central and southern areas of the state. In the northern areas, such as Arroyo del Agua, the flora is more diverse than in the typical Abo redbeds, but of lower diversity than the Late Pennsylvanian floras of the same area. The most common elements include conifers, callipterid peltasperms, and *Taeniopteris*, all seasonally dry plants. However, there also are the common wetland elements, calamitaleans, pteridosperms, and tree ferns, that probably grew within the floodplains of braided stream systems (DiMichele and Chaney, 2005).

Approximately 500 km (311 mi) to the NW of Socorro County, and perhaps at the margin of the paleotropical, equatorial belt (Dubiel et al., 1996), are deposits at the Pennsylvanian-Permian boundary in southeastern Utah (Figure 8). The sedimentary systems in this region reflect greater aridity at all times than those in New Mexico (Soreghan et al., 2002; Tabor et al., 2008), with regional climate varying from arid (aeolian dune sands, loessites and Calsisols) to subhumid (rivers and flood plains with Calcisols, colonized by plants). Keeping in mind that this area is paleolatitudinally north of the equatorial region, there appears to be an overall trend

Figure 8. Setting Hen Butte, Valley of the Gods, southeastern Utah. Stratigraphic succession. Camera angle causes the apparent overlap in the units. Late Pennsylvanian Honaker Trail Formation. Pennsylvanian-Permian Halgaito Formation. Early Permian Cedar Mesa Formation.

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from semiarid during the early Late Pennsylvanian to seasonally wetter conditions in the Early Permian (Soreghan et al., 2002). Like the Abo Formation, in New Mexico, the Early Permian deposits in southeastern Utah are significantly more "inland" than older rocks of the Late Pennsylvanian.

In a simplified stratigraphy from this area, the Middle and Late Pennsylvanian strata are assigned mostly to the Honaker Trail Formation, the uppermost part of which is represented by a cyclic series of paleosols, channel or braidplain sandstones, non-marine limestones and shales possibly reflective of a strong global-scale, allocyclic eustatic signal (Hite and Buckner, 1981; Soreghan et al., 2002). We did not find any plant macrofossils, other than a few unidentifiable axes, in these latest Pennsylvanian rocks, though a flora composed entirely of Pennsylvanian wetland species was reported from clastic rocks in the Honaker Trail Formation, 150 km (93 mi) to the north, near Moab, Utah (Tidwell, 1988). Palynological analyses of clastic and evaporitic rocks from the Paradox Formation of SE Utah, which is of Desmoinesian age (Middle Pennsylvanian), were carried out by Rueger (1996). Substantial climate cycling at this time is detectable from these analyses, as it was in similar age rocks from New Mexico. Overall, the SE Utah palynological assemblages are dominated, on a percent-abundance basis, by striate bisaccate monosaccate grains, generally considered suggestive of seasonally dry flora. However, these grains cycled in abundance through halites and their clastic interbeds. Typical wetland-flora palynomorphs were found to be rare but present (including *Lycospora*), and were negatively correlated with striate bisaccate grains.

The Honaker Trail Formation is overlain by the Halgaito Formation, which in the Valley of the Gods, in SE Utah (~ 16 km [~10 mi] north of Mexican Hat, Utah), consists of eolian sandstones and loessites, non-marine limestones, paleosols indicative of seasonally dry climates, and water-lain sandstones of various geometries, from thin bedded and flat-bottomed to bar forms of moderately large channels (Soreghan et al., 2002; Tabor et al., 2008). The Halgaito is thought to cross the Pennsylvanian-Permian boundary (e.g., Condon, 1997), and contains well preserved plant fossils. It is overlain by eolian sands and evaporites of the Cedar Mesa Formation (Loope, 1984, 1985, 1988).

The Halgaito Formation in the Valley of the Gods and Monument Valley contains both plants and non-marine aquatic and terrestrial vertebrate fossils. The vertebrates are found mainly in carbonate facies and conglomeratic deposits in small channels (Vaughn, 1962; Scott and Sumida, 2004), and are represented by taxa typically found in the Early Permian from elsewhere in the southwestern and eastern United States. Plants have been known from this region for many years (Vaughn, 1962, 1973; Mamay and Breed, 1970), though no thorough paleobotanical study of them has been undertaken. Specifically, in the Valley of the Gods, these plants are confined, for the most part, to various facies of a widely traceable interval consisting of channel cut-and-fill features. The flora was predominantly calamitalean sphenopsids and tree ferns, mainly preserved as trunks of relatively large size (10-15 cm [4-6 in] in diameter) that had been only locally transported. As with deposits of similar age in New Mexico, the flora also contained a background of plants typical of seasonally dry substrate environments, including walchian conifers. This, again, leads us to conclude that the preservation of parautochthonous, well preserved, typically wetland plants within channel facies strongly suggests a riparian flora growing along and perhaps within these channels where soil moisture was high for much of the year. However, the background of drier-site plants suggests that the landscape outside of the riparian zone was populated by plants tolerant of seasonal drought; this is consistent with the record of the paleosols, as well, which indicate only subhumid climates at the wettest (Soreghan et al., 2002). The most unusual thing about these channel and paleosol deposits is their presence within what otherwise appears to be a succession of arid to hyper-arid deposits.

Vegetational temporal patterns across the entire western Pangean region parallel those in areas farther to the east. For example, xeromorphic plants typical of seasonally dry environments have been reported from the Middle Pennsylvanian in the Illinois coal basin (Falcon-Lang et al., 2009; Plotnick et al., 2009), and recently a callipterid has been reported from the Middle Pennsylvanian in this basin, as well (Pšenička et al., 2011). Conifer pollen has been found to dominate assemblages from the late Middle Pennsylvanian in Atlantic Canada (Dolby et al., 2011). These patterns indicate that seasonally dry, subhumid climates alternated on glacial-interglacial scales (Milankovitch) with intervals of humid climate that characterized times of peat formation. At a still larger scale, the early Late Pennsylvanian (Missourian/ Kasimovian) has been characterized as a time of relatively high sea-level and significantly greater climatic dryness across the western and west central Pangean tropics, including Nevada (Bishop et al., 2010), New Mexico (Falcon-Lang et al., 2011a), Utah (Soreghan et al., 2002), and the Applachian Basin (Cecil et al., 1985; Cecil, 1990; Joeckel, 1995; Martino and Blake, 2001; Martino, 2004; Greb et al., 2008). Wetter conditions appear to have returned in the later Late Pennsylvanian (Virgilian/Gzhelian), indicated by the mixed floras and even a rare coal in New Mexico (Krainer and Lucas, 2004), and major coal beds in the Appalachians (e.g., Cecil et al., 1985; Cecil, 1990; Ruppert et al., 1999) with associated typically wetland Pennsylvanian floras (Blake et al., 2002). By the Permian, those floras that are known from the western Pangean regions are uniformly composed of species typical of seasonally dry habitats, with local patches of wetland plants surviving in those parts of landscapes that had semi-permanent, probably ground-water-based, soil moisture (see papers cited above, plus Lucas et al., in press b).

Moving East: North-central Texas

The Eastern Shelf of the Midland Basin of north-central Texas preserves one of the "classic" Permian red beds successions in North America (Figure 9). This area has been the source of numerous studies, mainly in Early Permian vertebrate paleontology, which brought

Figure 9. North-central Texas, Early Permian (Kungurian) red beds deposit, formed in ox-bow lake. Deposits such as these have yielded both flora and vertebrate fauna.

the terrestrial geology of the area into focus (e.g., Romer, 1935; Olson, 1952, Sander, 1989; Nelson et al., 2001; Montañez et al., 2007). Various aspects of paleobotany also have been investigated, mainly systematics (e.g., White, 1912; Mamay, 1967, 1968, 1976, 1986), but also stratigraphy (Read and Mamay, 1964). The Pennsylvanian portion of the section has received less paleontological acclaim, but has been studied in terms of its geology (Moore and Plummer, 1922; Brown, 1967; Feray, 1967; Galloway and Brown, 1973; Hentz, 1988; Tabor and Montañez, 2004) and paleobotany (e.g., Gupta, 1977; DiMichele et al., 2005a). This area was in the southwestern portion of equatorial Pangea during the Pennsylvanian and Permian. As such, the area was subject to atmospheric circulation patterns similar to those affecting areas farther to the paleo-west, in New Mexico and Utah. During the Virgilian (Gzhelian) the wetter parts of climate cycles, presumably driven by glacial-interglacial cyclicity, were usually wet enough for coals to form, though these were generally of low quality, high in ash, thin, and limited in areal extent (e.g., Gennett and Ravn, 1993). These coaly facies become thinner, less widely distributed and ultimately disappear from the stratigraphic succession near the Pennsylvanian-Permian boundary (Hentz, 1988).

The Pennsylvanian-Permian boundary on the Eastern Shelf as defined by conodonts, occurs at about the level of the Saddle Creek Limestone Member of the Harpersville Formation, just below its contact with the Pueblo Formation (Wardlaw, 2005). Farther to the north, where we have studied the paleobotanical succession, the geological section is predominantly terrestrial, with different formational names, but not enough marine limestones to place the Pennsylvanian-Permian boundary accurately with conodont (or any other marine invertebrate) biostratigraphy. However, it can be projected to be in the upper part of the Markley Formation (Hentz, 1988).

Fossil plants are preserved across the Pennsylvanian-Permian boundary, occurring in a variety of facies that preserve both wetland assemblages and those from seasonally dry or better drained substrates. There is very little mixing of these assemblages, based on study of the facies distribution of macrofossils, which appear to have occurred at different stages of glacial-interglacial cycles (DiMichele et al., 2005a). The fossil assemblages are preserved in abandoned channel deposits on landscapes that are represented primarily by paleosols (clastic deposits with a strong pedogenic overprint – see Tabor and Montañez, 2004), or channel-form

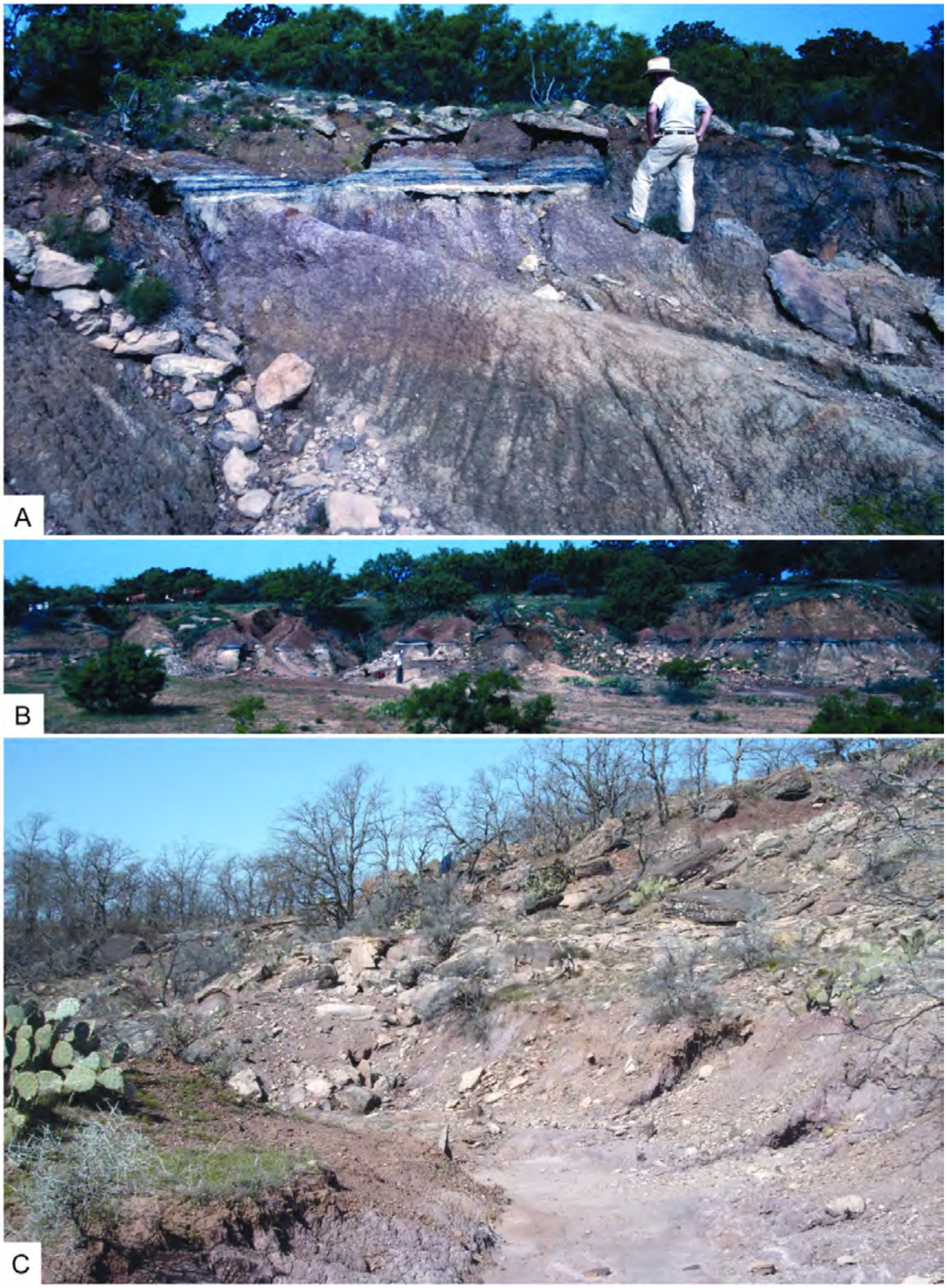

Figure 10. North-central Texas Pennsylvanian and Early Permian age outcrops of the Markley Formation. A—Typical succession of beds from the latest Pennsylvanian. From base: paleosol, quartz-kaolinite siltstone (containing seasonally dry flora), organic shale/coal bed (containing wetland swamp flora), gray siltstones (containing wet floodplain flora), sandstone (represented by float blocks, rarely containing seasonally dry flora). B—Same location as in (A), viewed from a distance. Note the prominent white, quartz-kaolinite siltstone and overlying organic shale bed in the center of the outcrop. C—Earliest Permian plant bearing outcrop, consisting of floodplain siltstone and channel-form sandstone deposits.

sandstones. Plant-bearing deposits occur mainly in association with various coal bed horizons, across which coals/organic shales are discontinuous and variable in thickness and organic content.

A typical outcrop in the lower part of the Markley Formation (Figure 10 A, B) has a distinctive lithological sequence, which is strongly associated with plant biofacies patterns. It must be stressed that this pattern has been observed at multiple outcrops and throughout the Pennsylvanian portion of the Markley Formation, through several cycles, marked by distinct coal-bed horizons. Closer to the Pennsylvanian-Permian boundary, coaly facies are no longer present in outcrops, but the rest of the facies patterns and plant compositional aspects remain the same. These patterns are described by DiMichele et al. (2005a), and from base to top of a typical outcrop they are as follows:

1. A basal paleosol, evidencing wet but well drained conditions, sometimes containing roots that cannot be attributed to specific plant groups. Paleosols may be several meters thick and, throughout the section, evidence a drying trend in the later Pennsylvanian and into the earliest Permian (Tabor and Montañez, 2004).
2. A kaolinite-quartz siltstone bed that may vary in thickness from a few centimeters to over a meter. This bed generally rests unconformably on the upper surface of the paleosol. It contains a flora composed of *Sphenopteris germanica* and walchian conifers, with rare elements including *Charliea*-like pinnules and various pteridosperms. Typically wetland plants have not been identified in this facies. The flora is characteristic of Pennsylvanian seasonally dry climates.
3. An organic-clastic bed that may vary from an organic-rich, finely laminated clay shale to a normally bright-banded coal bed. Where plants are found in this facies, the most common and abundant are typically *Macroneuropteris scheuchzeri*, *Pecopteris* of various species, and *Sigillaria brardii*. This low diversity assemblage represents the typical Pennsylvanian clastic-to-peat swamp flora of flooded substrates with high water tables throughout most of the year. Such beds may vary from less than a meter in thickness to several meters.
4. A sequence of gray to buff siltstone deposited in shallow scours with erosional bases. The siltstones commonly contain a weak pedogenic overprint. Where pedogenesis has not proceeded too far, plant fossils often are preserved in the channel scours. The flora preserved in the fill of any one small scour can vary greatly in species richness, but the overall flora of any given outcrop, consisting of several such subdeposits, is generally the most diverse of any of the plant-bearing facies, consisting of greater than 30 species. These species are principally tree ferns and pteridosperms, typical of Pennsylvanian wetland floras.
5. Sandstone, often many meters in thickness (difficult to measure because of talus formation and vegetational cover) with an erosional basal contact with the floodplain shales may, on occasion, contain plant fossils. Such sands probably are at the base of the subsequent cycle but, because of the nature of weathering in the area, they tend to support hillslopes and occur at the tops of exposed sections. Only rarely have plants been identified in these rocks. Where they have occurred, the plants are cordaitaleans and conifers, both xeromorphic, coniferalean taxa that have been identified as elements of seasonally dry floras. Cordaitaleans are an extremely diverse group, and appear to occur from coastal environments

(Raymond, 1988; Falcon-Lang, 2005; Raymond et al., 2010) into remote interior areas of tropical latitudes (Falcon-Lang, 2000). However, the appearance of conifers and cordaitaleans in this channel facies is consistent with the appearance of such elements, typical of seasonally dry floras, throughout the Middle and Late Pennsylvanian in coal-bearing rock sequences of the Western Interior basin (Feldmann et al., 2005), Eastern Interior (Illinois) basin (Falcon-Lang et al., 2009; Plotnick et al., 2009), Central Appalachian basin (Martino and Blake, 2001) and the Maritimes area of Canada (Dolby et al., 2011).

The over-riding paleobotanical pattern observed in the Texas deposits is the co-occurrence, at the outcrop scale, of wetland flora and seasonally dry flora (Figure 11), throughout the Virgilian and into the earliest parts of the Wolfcampian. Both vegetation types occupied the coastal plain environment, but they are principally unmixed, occurring in lithologically distinct beds, in stereotypical patterns of succession. Thus, the degree to which these floras overlapped in space and time continues to be an open question. They were certainly likely to have been present contemporaneously in the equatorial region, but did they coexist on lowland landscapes in close spatial proximity, during the latest Pennsylvanian and earliest Permian, as we have surmised from assemblages in New Mexico and Utah? In the case of north-central Texas, the evidence is most parsimoniously interpreted, in our opinion, as indicating they did not share the lowlands contemporaneously. In this area, the wetland plants do not appear to have been confined to channels or channel belts, surrounded by seasonally dry flora. Rather, the confinement of the Texas floras to distinctly different lithofacies, which themselves are of types closely tied to broader indicators of regional climate and sea-level or local base-level, argues for differentiation in space and time.

In light of the climate-rock framework and models described in the introductory portion of this paper, we interpret the floristic and lithofacies patterns in north-central Texas as follows. Coal and organic shale beds reflect the wettest portions of the climate cycles and likely formed during lowstand to late lowstand on a flat lying coastal plain, developing under moist subhumid to humid paleoclimate, differentially in lower areas where standing water was most likely to accumulate for long periods, including channel belts, formed during the earlier phases of sea-level fall. Overlying these organic beds are floodplain deposits, which frequently are in gradational contact with the organic facies. We interpret these as the early phases of sea-level rise, during which climate was still moist subhumid, favorable to development of a wetland flora. These sections contain few or no limestones. So, these low areas of the landscape may have been embayed or served as areas of active drainage during highstand, surrounded by interfluves. From highstand, through falling stages and into early lowstand, paleosols formed on these floodplain sediments, resulting in deep pedogenesis. In addition, shallow channels developed across the landscapes, in which sandstone channel bars were preserved. The climate during these phases, based on the plants, appears to have been seasonal, but still likely dry subhumid. The kaolinite beds have flow features associated with them, and the contained flora is, in most instances fragmentary and evidences at least local transport. We interpret these deposits as representing sediments washed into the channels during early lowstand as rainfall is increasing on the landscape, under the transition from dry to moist subhumid conditions.

The Pennsylvanian-Permian transitional interval in north-central Texas is marked by a strong floristic discontinuity (DiMichele et al., 2010c). Coal beds become progressively thinner and localized in development prior to disappearing entirely in the latest Pennsylvanian,

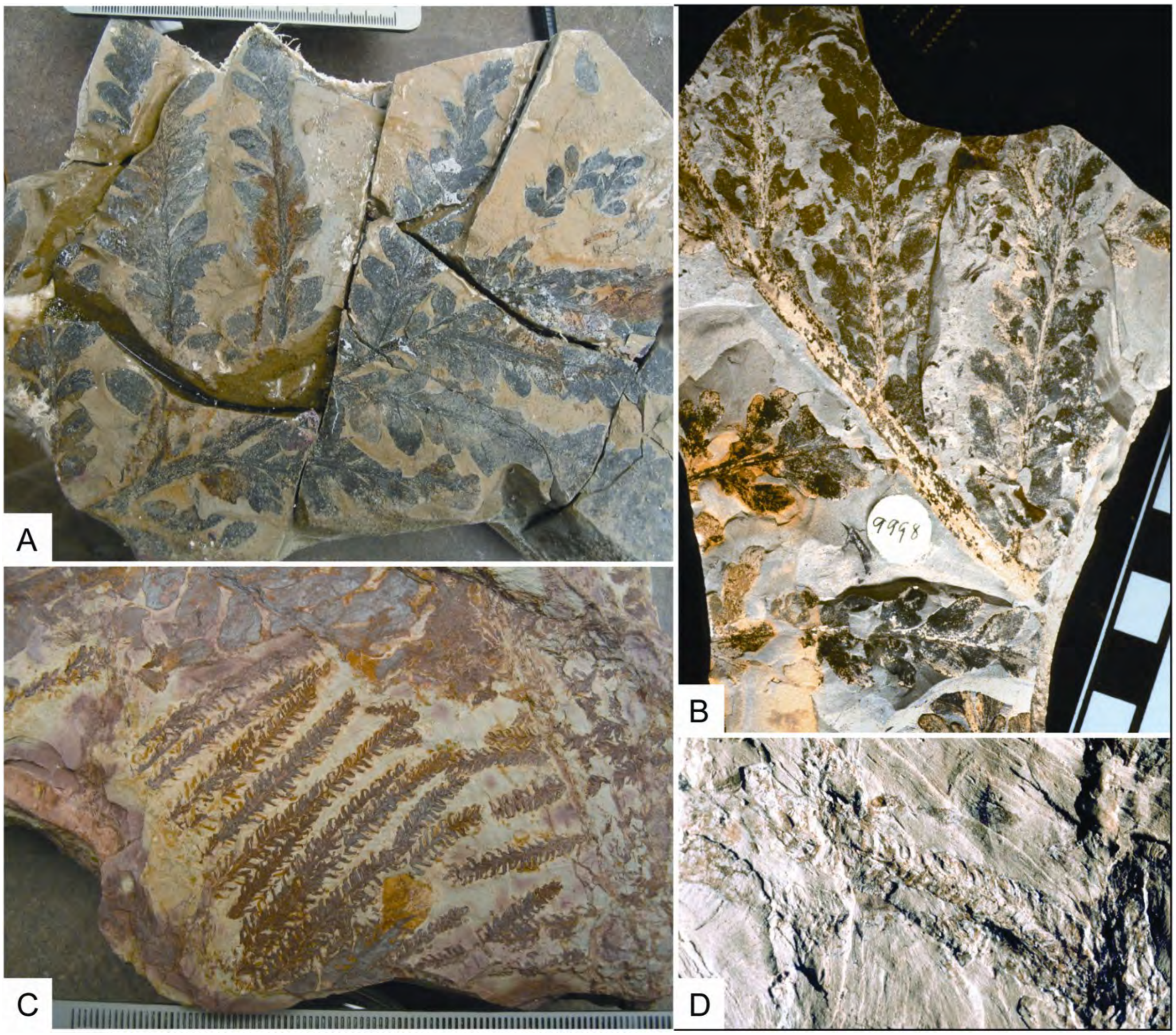

Figure 11. North-central Texas, seasonally dry flora from the Pennsylvanian-Permian Markley and Early Permian Archer City formations. A. *Rhachiphyllum* cf. *schenkii*, callipterid peltasperm frond segment, Early Permian, floodplain siltstone (USNM 536518; USNM loc 40037). B. *Sphenopteris germanica*, quartz-kaolinite siltstone bed (USNM 543965; USGS loc 9998). C. Walchian conifer, Early Permian ox-bow lake deposit (USNM 543966; USNM loc 40027). D. Walchian conifer, quartz-kaolinite siltstone bed (USNM 543968; USNM loc 39998).

suggesting a diminishment of moisture in the wettest phases of glacial-interglacial cycles. In the Permian, there is a change to much simplified lithological patterns on outcrop, and organic beds are lacking (Figure 10 C). Channel sediments, where plants have been identified, are dominated entirely by a flora that resembles that typical of the kaolinite-quartz beds of the Pennsylvanian outcrops, that is: *Sphenopteris germanica*, walchian conifers, and various pteridosperms, but with an admixture of callipterids (such as *Autunia conferta*) and other seasonally dry elements such as cordaitaleans and *Taeniopteris*. This kind of flora remains the most commonly encountered throughout the Archer City Formation, which lies immediately above the Markley and, together with the upper 10 % or so of Markley thickness, is primarily Asselian and Sakmarian in age (Montañez et al., 2007), a time interpreted to be one of major glaciation in the southern hemisphere (Fielding et al., 2008). Montañez et al., (2007) demonstrated strong vegetational tracking of climate changes and dynamics associated with this and subsequent intervals of inferred glaciation and the intervening non-glacial periods. However, we hasten to point out that significant unconformities indicative of prolonged sea level fall associated with long-term, abundant continental ice have not been identified in this region or on the North American craton in general.

Palynology of the north-central Texas section through the Pennsylvanian-Permian boundary (Gupta, 1977) suggests a more gradual turnover than that found in the macrofossils (leading Gupta to place the palynological boundary well above that for either fusilinids or conodonts, at the time his paper was written). These palynological data suggest that the wetter and drier elements of the flora were in reasonably close proximity on the Eastern Shelf, even though this is not supported by the macrofossils, which do not show the patterns of wet-flora/ dry-flora co-occurrence seen in the deposits from the more westerly regions of the equatorial belt that suggest spatial adjacency.

In summary, the vegetational transition in north-central Texas, based on macrofossils, appears to be detectable on the temporal scales of both glacial-interglacial oscillations and on the longer scale of the general Pennsylvanian-Permian drying trend. However, the changeover to a more consistently sampled seasonally dry flora is superficially later in Texas, and to the north in Oklahoma (see below), than in New Mexico and Utah, appearing not to have occurred until very near to the Pennsylvanian-Permian boundary. This may, in no small part, reflect a taphonomic sampling bias: wetland deposits, and their associated flora, which may have represented a much smaller temporal fraction of lowland basin occupancy than that of seasonally dry flora (Falcon-Lang et al., 2009; Falcon-Lang and DiMichele, 2010), are more abundant and more easily located by collectors than the seasonally dry flora (often represented either by paleosols or by scattered deposits with poorly preserved fossils), therefore giving the appearance of greater concentration of wetlands on the landscape in time and space than is warranted by fact. Floral samples in the overlying Archer City Formation of Texas, which lacks indicators of climates as wet as or wetter than moist subhumid, are rare and widely distributed in time and space. These floras are typically of the seasonally dry type, which may, in fact, be those of the wettest climate phases. The Markley Formation actually may record only the fact that the wetter parts of glacial-interglacial cycles, such as they were in the latest Pennsylvanian, were simply wetter and more capable of generating widespread wetland landscapes than were similarly much drier wettest phases in equivalent age rocks of New Mexico and Utah. If this is the case, the longer term drying trend in these two areas is less conspicuously offset than would appear, though it is still present.

Northern Oklahoma and Southern Kansas

Understanding of Late Pennsylvanian and Early Permian paleobotany of north-central Oklahoma and adjacent south-central Kansas derives primarily from palynology. This transition occurs in rocks of the latest Pennsylvanian through earliest Permian Wabunsee, Admire, Council Grove, and Chase groups. Macrofossils, although recorded, have not been collected or studied systematically or in their stratigraphic context. In Oklahoma, foliage has been found in cores from the Wood Siding Formation (Wabunsee Group, latest Virgilian) and the Winfield Limestone (Chase Group, Early Permian) (Lupia, 2010), and also has been reported in outcrop in the Doyle Shale (Chase Group) (Chaplin 2010).

The Council Grove and Chase groups, of latest Pennsylvanian through Early Permian age, contain alternating shales and limestones associated with epicontinental sea-level fluctuations, set against a pattern of overall drying through the Early Permian (West et al., 2010). The Pennsylvanian-Permian boundary is placed in the Red Eagle Limestone of the Council Grove Group on the basis of conodonts (Sawin et al., 2006). Abundant evidence of cyclicity consistent with glacial-interglacial drivers has been documented across the Penn-Permian boundary by Olszewski and Patzkowsky (2003), who conclude that eustatic lows were intervals of dry climates whereas eustatic highs were intervals of humid climates, but this may reflect terminological differences from the interpretations presented here. They also note that thin, but persistent coals occur above and below the boundary. Mazullo et al. (2007) also concluded that depositional sequences in these rocks reflect glacial-interglacial drivers, based on examination of isotopic evidence from brachiopods, which they inferred to track changes in ice-volume rather than local ocean temperature.

Previously, two comprehensive palynological sampling programs have studied the Pennsylvanian-Permian boundary interval in this area. Clendening (1970, 1975) investigated Kansas palynology for comparison with his Dunkard studies. Wilson and Rashid (1971) likewise sampled Virgilian through Lower Permian sediments to establish the boundary. In both cases, they concluded that the vegetational change occurred far above the conodont-based boundary assignment in the Red Eagle Limestone, placing it within the lower Chase Group or at its top, respectively. Although lacking in the resolution necessary to detect oscillations, both suggest that Early Permian vegetation was substantially similar to the underlying Late Pennsylvanian. More recently this interval has been reinvestigated (Lupia, 2010), specifically in the context of the conodont-based boundary, and has affirmed these prior large-scale findings as well as observed smaller-scale oscillations in palynological content across the boundary, without a sharp vegetational discontinuity. Rather, wetter-drier phase oscillations are superimposed on a longer-term, coarser scale “drying” trend. The abundance of striate bisaccate pollen, indicative of taxa from seasonally dry habitats (e.g., conifers and peltasperms), is negatively correlated with the abundance of spores representative of wetland taxa (e.g., ferns, horsetails). In the course of these analyses, considerable lateral variation in lithologies was found, suggesting differences in local conditions, not unlike the regional variations found in the Dunkard Group (see below; Cecil et al., 2011; Fedorko and Skema, 2011). One core showed a Virgilian section that was overall quite dark grey and organic-rich, whereas another, from less than 32 km (20 mi) away, showed dominantly red bed shales in equivalent Virgilian strata assuming correct correlation.

The Central Appalachian Basin – Dunkard Group

The Dunkard Group, of the central Appalachian Basin, was located in the western central portion of Pangea during the Pennsylvanian-Permian transition. In this position, it was well inland and backed up against the early phases of tectonic uplift of the Appalachians, which was the source of most Dunkard sediment (Martin, 1998). The occurrence of linguloids in the Washington coal zone (Cross and Schemel, 1956; Berryhill, 1963) suggests that the basin may have had a drainage to the open ocean, but the lack of unequivocally marine fossils in any Dunkard strata strongly suggests that marine waters never actually entered the basin. In fact, the last certain occurrence of marine conditions in the Applachians is the early Virgilian (Gzhelian) Ames Limestone, which has a date securely established by conodonts (Barrick et al., 2008), and lies approximately 100 meters below the base of the Dunkard Group. Dunkard strata are dominantly fine-grained, clay to silt-sized clastic sediments with high mica and low feldspar content, perhaps suggesting a distant source or one poor in such minerals. Common to abundant non-marine limestones, vertic to calcic-vertic, deeply developed paleosols, and evidence of fluctuating water tables indicate a prevalence of subhumid climates, moist and dry, with some periods of greater aridity (Cecil et al., 2011). Coals, indicators of humid climate, are thickest in the lower parts of the Dunkard, where the Waynesburg and Washington coals have been mined commercially. However, above the Washington coal zone, coals tend to be thin and often discontinuous (Fedorko and Skema, 2011). These coals are high in ash and sulfur content (Eble et al., 2011), consistent with their formation under climatic conditions that favored topogenous peat swamps, barely within the climatic and sedimentological window favorable for peat formation (Cecil et al., 1985; Cecil et al., 2011).

Dunkard sedimentation has been described as "cyclic" (Beerbower, 1961, see also Fedorko and Skema, 2011). These patterns are most detectable in the northern portions of the Dunkard Group where facies diversity is highest. On the southern and western margins of the basin, the section consists mainly of interbedded channel sandstones/siltstones and paleosols (Martin, 1998; Fedorko and Skema, 2011). In the deeper portions of the basin, the lithofacies included in this cyclicity indicate the oscillation of regional climate extremes: at the wettest end are the coal beds, indicative of widespread, high-soil-moisture conditions, probably under moist subhumid to humid conditions. At the other extreme are non-marine limestones, indicative of high evapotranspiration and a range of conditions varying from dry subhumid to semi-arid to occasionally arid. Also alternating are conditions that favored widespread vertic paleosol development, often with calcareous deposits, reflective of a strongly seasonal distribution of moisture under subhumid climatic conditions, and deposits of sandstone, siltstone, and mudstone, the grain sizes and geometries of which indicate deposition in standing water bodies of considerable areal extent (Martin, 1998; Cecil et al., 2011; Fedorko and Skema, 2011).

It can be concluded from these observations that the Dunkard Basin was subject to cyclic variations in water table and atmospheric moisture delivery on several different temporal scales. Even at its "wettest" (Waynesburg and Washington coals) there were still intervals of moisture fluctuation and drought of variable duration, sometimes enough to interrupt peat formation and return the region temporarily to high levels of clastic transport (see Cecil and Dulong, 2003, for a discussion of the relationship between climate and sediment transport). Even at its "driest", the region appears to have been wet enough to support fluvial siliciclastic sediment transport, reflected in features such as frequent clastic partings within limestone, high clastic content of some limestones (making them almost calcareous shales) and the occasional preservation of plants in the limestones themselves, indicative of wet-substrate conditions (specimens of the tree fern foliage *Pecopteris* have been found within the Windy Gap Limestone of the upper Greene Formation).

The Dunkard flora is well established and is characterized overwhelmingly by assemblages of typically Late Pennsylvanian (late Virgilian: Gzhelian) character (see Gillespie and Pfefferkorn, 1979; Blake et al., 2002), typical of wetland habitats. This was clear even from the illustrated macroflora of Fontaine and White (1880), the original description of the flora, from Gillespie et al. (1975), who reinterpreted the flora to some degree, and from the commentaries of White (1936) and Darrah (1969, 1975). Palynological analyses (Clendening, 1972, 1974, 1975; Clendening and Gillespie, 1972) present a similar picture, though they are focused primarily on those rocks representative of the wettest portions of climate cycles (Eble

Figure 12. Northern West Virginia, Brown's Bridge locality from which the original Fontaine and White (1899) callipterids were collected. Site of railroad grade. Modern exposure is heavily vegetated.

et al., 2011), the coals and coal roof shales. In our own collecting, we have found wetland species, particularly pecopterid ferns, in nearly every facies (except clastic paleosols), including limestones, sandstones, the latter representing both high -water deltaic deposits (flat-bottomed sands) and erosionally based channel-form deposits representing low stand or falling stages of base-level, floodplain mudstones, and organic-rich swamp deposits.

The exception to the wetland Dunkard flora is the occurrence of callipterid peltasperms and conifers in some floodplain mudstones and clastic partings in limestones, occurring sporadically (at current levels of stratigraphic resolution) from the Washington coal zone, in the Washington Formation, to the level of the Nineveh coal in the middle Greene Formation (DiMichele et al., 2011) (Figures 12 & 13). These are exceptional occurrences, of which perhaps as many as 10 have been reported (see Darrah, 1975). There appear to be few other taxa in the callipterid assemblages. Most significantly, Darrah (1969, 1975) reports conifers, including what he identified as *Lebachia*, from at least one site. No illustration was provided, and the location of Darrah's collections is unknown to us. In the David White collections made in 1902, and those of Aureal Cross, made in the late 1940s, other taxa, including the pteridosperm *Odontopteris*, calamitalean stems and *Annularia* foliage, are rare. However, most of the collections are small and it is not certain that the non-callipterid material was collected from the same beds as the callipterids (based on examination of the matrix and on what can be gleaned from surviving field notes and notes in collections).

The spatio-temporal distinctiveness of the wetland and callipterid floras most likely reflects environmental control. In keeping with the more widely known and documented occurrences of callipterids (e.g., Kerp and Fichter, 1985; Read and Mamay, 1964) and of peltaspermous seed plants in general (DiMichele et al., 2005b), these plants are likely representative of seasonality of moisture distribution with periods of soil moisture deficit, probably in subhumid climatic regimes. Numerous calcic paleosols, throughout the Dunkard, document climatic intervals with a probable ustic soil moisture regime indicative of an intense dry season for most of the year (Cecil, et al, 2011). And for most such soils, the surface vegetation is not known – it might have been a callipterid-conifer assemblage.

Poor exposure of Dunkard strata (Figure 12) limits our understanding of the environmental context of both the common wetland and rare seasonally dry floras, but much more so the latter. Wetland floras are represented by many more collections and collecting sites than are seasonally dry floras, which provides a "statistically" richer picture of the context of the wetland vegetation, despite the generally limited outcrop exposure. However, the cyclicity of Dunkard lithotypes, and environments they represent (Beerbower, 1961), suggest that the callipterids may (1) characterize one part of any given climate-deposystem cycle, and (2) that there was an increase in climate contrasts in the Dunkard beginning near the time of deposition

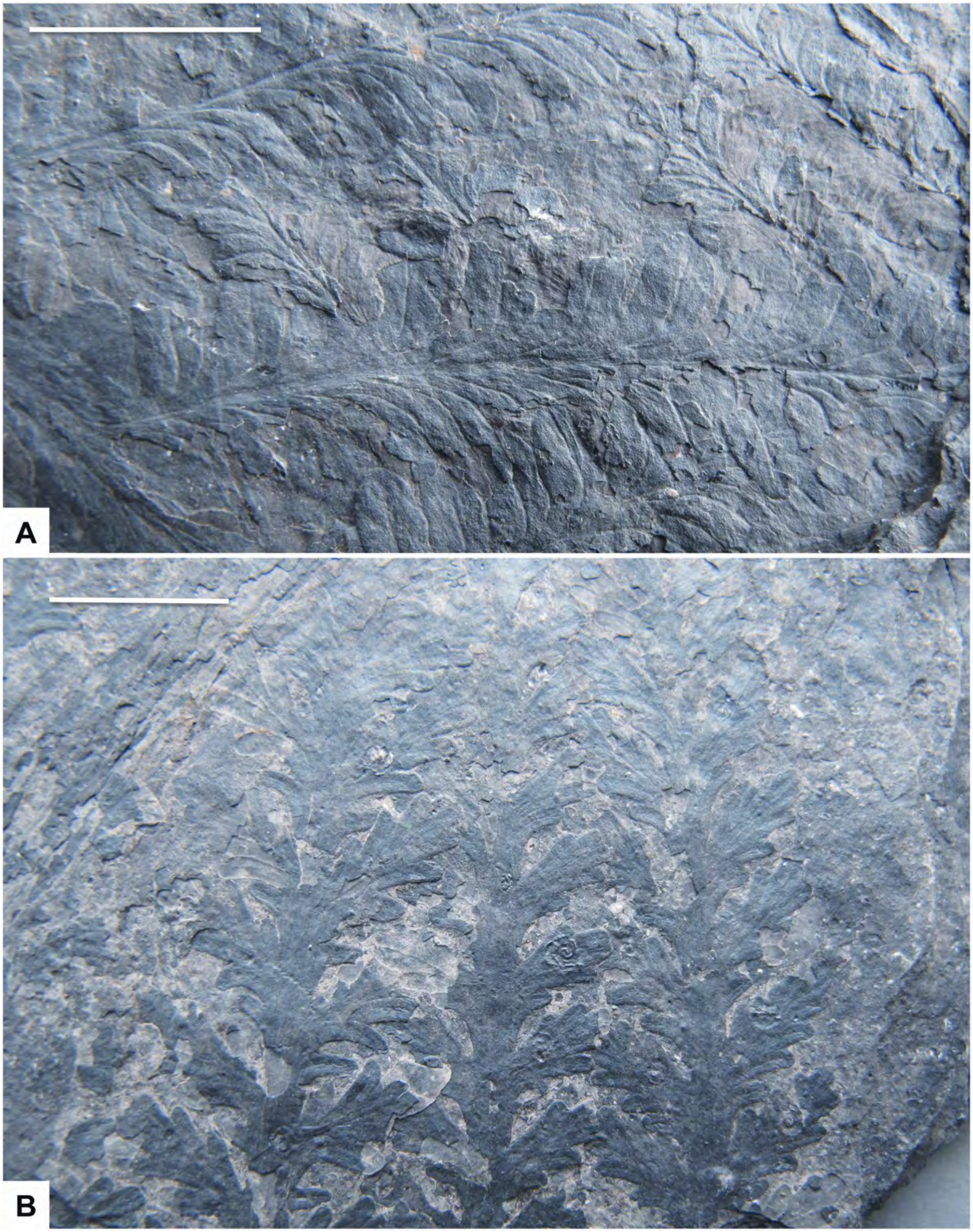

Figure 13. Callipterids from Brown's Bridge locality, West Virginia. Collected by David White, 1902. These species, typical of seasonally dry habitats, were one of the major reasons that Fontaine and White inferred a Permian age for the Dunkard Group. A—*Autunia conferta* (USNM 543957; USGS loc 2926), 4 X magnification. B—*Lodevia oxydata* (USNM 543958; USGS loc 2926), 3X magnification.

of the Waynesburg A coal bed, compared to the oldest portions of the Dunkard and the underlying Monongahela Formation. This climate contrast continued at least through deposition of the lower to middle Greene Formation. Evidence of this increasing contrast between the wetter and drier portions of climate-deposystem cycles is seen as early as the time of formation of the paleosol seat-earth that lies beneath the Waynesburg A coal bed, which is locally a thick, well developed, calcic Vertisol, evidencing strong moisture seasonality with periods of high evapotranspiration. This paleosol is succeeded by a return to humid conditions with the formation of the peat that formed the Waynesburg A coal bed. Paleosols of this type have been found, intermittently, well up into the Greene Formation (such as the exposures on Great House Hill Road, near Wylieville, West Virginia, which is stratigraphically near the position of the Lower Rockport Limestone: Fedorko and Skema, 2011).

There is no definitive marine or radioisotopic evidence by which the Pennsylvanian-Permian boundary can be located within the Dunkard. There are, however, strongly suggestive paleontological and lithological data suggesting that it may lie close to the level of the Washington coal. Such evidence include the occurrence of brackish fauna at that horizon (Berryhill, 1963), consistent with a latest Pennsylvanian marine high stand (Davydov et al., 2010) at that time, non-marine ostracode data (Tibert et al., in press; Tibert, 2011) and tetrapod vertebrate data (Lucas, 2011), both of which suggest a Permian age, possibly beginning around the time of deposition of the Washington coal complex. Thus, the interval within which callipterids appear is, indeed, not an unreasonable candidate for placement of a systemic boundary. Such a placement would be consistent with the early thinking of David White (1904), based on plant fossils (though he later joined I.C. White in considering the entire Dunkard Group to be Permian – White, 1936). That upward diminishment of coaly, or organic-rich, facies throughout the Greene Formation is a similar pattern to that of the other American basins we have discussed. In the Dunkard, the organic facies persist longer than in the basins from the more western regions of the Pangean equatorial belt. This pattern suggests that the wetter end of climate cycles remained moist subhumid to humid longer in a progressive west-to-east direction. As a consequence, a wetland flora remains the most commonly encountered throughout the Dunkard section, even into the Early Permian.

DISCUSSION

There are clear trends in the spatial and temporal patterns of vegetational change across the Pennsylvanian-Permian boundary of the Pangean equatorial region in North America. Wetland species, characteristic of humid and moist subhumid climates, are the most commonly encountered in the terrestrial fossil record of most of the Pennsylvanian, a pattern that has long been recognized. Taxa typical of seasonally dry environments, those not typically associated with organic-rich deposits or in association with physical indicators of humid climates, are characteristically the most commonly encountered plants of Permian age rocks. This pattern appears to be time transgressive. It begins first in the western equatorial regions of Pangea and occurs progressively later in time along the paleoequator to the east, at least on the northern and western side of the Central Pangean mountain belt. As in the Dunkard Group, strata in Western Europe, such as the Lower Rotliegends, which transgress the Pennsylvanian-Permian boundary, have numerous organic-rich deposits, often well into what is interpreted as the Lower Permian, depending on the particular basin and its tectonic setting (see Roscher and Schneider, 2006). These may preserve intercalated wetland and seasonally dry floras, with elements of seasonally dry floras appearing in the Pennsylvanian and, conversely, wetland elements persisting into the Permian (e.g., Remy, 1975; Broutin, 1977; Wagner and Martinez-García, 1982; Wagner, 1983; Kerp and Fichter, 1985; Jerzykiewicz, 1987; Kerp et al., 1990; Popa, 1999; Steyer et al., 2000; Cassinis and Ronchi, 2002; Wagner and Mayoral, 2007; Boyarina, 2010).

Underlying the average vegetational trend is the concept of a progressive "drying" from the Pennsylvanian into the Permian, which is correct at a coarse resolution (see summary in Remy, 1975). However, as resolution is increased, it can be seen that this general drying trend contains a great deal of finer scale climate variation, including that likely to be orbitally forced on the temporal scales encompassed by Milankovitch cyclicity (e.g., Heckel, 2008; Connolly and Stanton, 1992; Rasbury et al., 1998; "stage scale" of Cecil et al., 2011), consistent with glacial-interglacial cycles of the 10^5-year scale. And, on longer time frames, such as that evaluated by Allen et al. (2011), these Milankovich-scale fluctuations in climate can be seen to be superimposed on broader fluctuations in moisture that range in duration from 2 to 6 million years, corresponding broadly to fluctuations in southern hemisphere ice volume (Fielding et al., 2008).

At the glacial-interglacial scale of 10^5 years, increased contrasts can be identified between the intervals of peat formation (humid climate) and those of paleosol formation below the peat beds (subhumid climate) (Cecil et al., 2003; DiMichele et al., 2010c). This began at least by the Middle Pennsylvanian across much of the Pangean tropics west of the Central Pangean mountain belt, which may have simply been highlands at that time, in what is now North America (Roscher and Schneider, 2006). It is reflected in the nature of the fossil-plant assemblages found in association with coals vs. those deposits formed during the drier inter-coal time intervals (Galtier et al., 1992; Falcon-Lang et al., 2009; Plotnick et al., 2009; Scott et al., 2010; Falcon-Lang and DiMichele, 2010). Climate contrasts on the glacial-interglacial scale become even more accentuated during the drier of these intervals, such as the Missourian (early Late Pennsylvanian), where excellently preserved "Permian"-type floras may be found and lithologies record strong contrasts between wetter and drier parts of climate-sealevel-deposystem cycles (Cridland and Morris, 1963; Darrah, 1975; Martino and Blake, 2001; Martino, 2004; Martino and Greb, 2009; Falcon-Lang et al., 2011a; Allen et al., 2011; Lucas et al., in press a).

The physical and biological aspects of these inferred glacial-interglacial oscillations in climate are manifested differently among Pangean depositional basins, depending on regional climatic means, and this affects our understanding of the floras in those respective areas. For example, in western Pangea (New Mexico and Utah), even at those times of wettest climate, peat formation was at best sporadic and generally did not occur. The regional climate system was shifted to the drier end of the "wet-dry" spectrum. Cyclicities can be recognized in the western basins at multiple temporal scales, and they follow the same basic patterns as in more easterly parts of the Pangean tropics: indicators of sea-level change can be identified, there are fluctuations in terrestrial floristic composition associated with these changes, and longer-term changes can be identified associated with changing polar ice volume. The "apparency" of vegetational change, that is our ability to discern such change from the plant fossil record, is, however, strongly affected by regional climate differentiation. The lack of coal in western Pangea removes a focal point for both recognition of and preservation of wetland vegetation in autochthonous and parautochthonous accumulations. As a consequence, the field search images for locating plant-fossil-bearing deposits are very different in a central Pangean coal basin than in areas of western Pangea, where coaly rocks are rare to absent. In the west, we often find plants, after considerable searching and excavation, in rock units that preserve mainly allochthonous material and that formed during portions of the sea-level and climate cycle, such as deltaic deposits formed at highstand or falling stages of sea level, that are considerably different from those represented by coals and the sedimentary rocks immediately associated with them (such as floor and roof shales with which they intergrade). Strikingly, a search for like kinds of lithotypes in coal basins often reveals the same kinds of non-wetland plants as are found in western Pangea, and often in allochthonous assemblages (White, 1912; Arnold, 1941; Cridland and Morris, 1963; Feldmann et al., 2005; Plotnick et al., 2009; Falcon-Lang et al., 2009; Pšenička et al., 2011).

Major changes in the volume of polar ice occur on temporal scales that are considerably longer than those driven by orbital cyclicity (Allen et al., 2011). Glacial-interglacial cycles are superimposed on variations in mean ice volume, and thus on the change in climate state they cause (Cecil et al., 2011). The larger scale changes do not appear to occur rhythmically, as if driven by various combinations of orbital forcing factors, but this is presently uncertain. Such changes appear to have occurred near the Atokan-Desmoinesian (Bolsovian-Asturian, mid-Moscovian) boundary and the Desmoinesian-Missourian (approximately the Moscovian-Kasimomvian) boundary, and to have resulted in threshold-like changes in climate state at those boundaries.

At the Atokan-Desmoinesian boundary, there appears to have been an overall shift toward greater dryness at all phases of glacial-interglacial cycles. This resulted in a shift from an oscillation between perhumid climates during peat formation and moist subhumid climates during the intervening periods, to humid peat-forming climates and dry subhumid climates between periods of peat formation (Cecil, 1990). The result was a change from raised, ombrotrophic peat swamps to planar, minerotrophic swamps (e.g., Cecil et al., 1985; Greb et al., 2002; Eble et al., 2001, 2003), an increase in sulfate and carbonate minerals in the system (Cecil et al., 1985), changed sedimentary patterns (Cecil, 1990; Bertier et al., 2008), and changes in the dominant vegetation of wetland intervals (Peppers, 1996; Cleal, 2007).

Major vegetational change also took place at the Desmoinesian-Missourian (Westphalian-Stephanian in traditional useage, approximately Moscovian-Kasimovian) boundary (Phillips et al., 1974; DiMichele and Phillips, 1996). This vegetational change is closely correlated with, and was seemingly caused by, climatic warming and drying that began earlier in the Middle Pennsylvanian and culminated near the Desmoinesian-Missourian boundary (Cecil, 1990; Phillips and Peppers, 1984), resulting in great diminishment of polar ice (Fielding et al., 2008; Allen et al., 2011) and a high-stand of sea level (Rygel et al., 2008; Heckel, 2008). One of the driving forces of this warming may have been CO_2 (Cleal et al., 1999), which, though not shown definitively for the Desmoinesian-Missourian boundary, has been demonstrated for similar changes in Permian ice volume by Birgenheier et al. (2010), which may serve as a model for this older time interval. Similarly, model studies of late Paleozoic atmospheric composition, ice volume and vegetational feedbacks implicate fluctuations in CO_2 as a major controlling variable (Horton and Poulsen, 2009; Horton et al., 2010). Another exacerbating factor may have been the rise of the Central Pangean Variscan mountain range across Central Europe, which both exposed earlier-deposited coals to erosion and reduced dramatically the area of basinal depocenters (Cleal and Thomas, 2005; Cleal et al., 2009). The Missourian/ Kasimovian warm-dry period transitioned to a seemingly wetter Virgilian/Gzhelian, based on patterns of coal thickness and areal distribution in the central Pangean coal basins (Cecil, 1990), and from patterns seen in parts of the western US, but this phase in particular, as has been discussed in this paper, was spatially quite heterogeneous. The changes at the Desmoinesian-Missourian boundary are instructive when viewed in comparison with Pennsylvanian-Permian boundary. In both cases there is a trend toward drying and perhaps warming through time. However, in the Missourian, this warming and drying appears to be correlated with a significant diminishment in polar ice volume (Fielding et al., 2008) and a distinct global rise in mean sea-level (Rygel et al., 2008). In stark contrast, there is evidence of a major reappearance of ice during the Early Permian, the first of several Permian ice intervals (Fielding et al., 2008; Montañez et al., 2007), though evidence of widespread, long-term sea-level fall has not been demonstrated. Thus, the changes at these two time periods do not appear to be replicates of the same basic kind of global event or set of drivers, and the vegetational consequences certainly were different (DiMichele et al., 2009).

The longest-term climatic trend considered here is that characterizing the general drying beginning in the later Middle Pennsylvanian and continuing throughout the Permian. Difficulties in picking a lithological boundary between the Carboniferous and Permian reflect the gradual, directionally oscillatory, lithological changes that occurred in much of the world between the two Periods – there is no “natural” break, in most instances (see discussion in Wardlaw et al., 2004). This lack of a natural lithological break (lacuna) argues against a global lowstand of sea level, which is consistent with the paleobotanical changes, as described herein. Whereas there certainly is an “on average” difference between a Pennsylvanian tropical flora and one from the Permian, there is always the chance of finding a “Permian”-type assemblage in the Pennsylvanian and vice versa. This has been a point of confusion for many years for paleobotanically-based stratigraphic frameworks. Consider, for example, the remarks of W.C. Darrah in the “Discussion” following Remy’s paper (Remy, 1975), regarding the Garnett, Kansas Missourian flora, which Darrah insisted had to be Permian, or the remarks of Remy (1975) himself, who could not accept explanations of high-frequency climate changes to explain the intercalation of wetland and seasonally dry floras within Pennsylvanian and Permian sections, despite arguing forcefully for strong climatic controls on plant distribution. Such disagreements have continued to the present (e.g., Wagner and Lyons, 1997; Falcon-Lang et al., 2011b). Certainly, the recognition of “xeromorphic” or seasonally dry floral elements as facies fossils, representative of drier climatic settings, has a long history (Elias, 1936, 1970; Arnold 1941).

Long-term drying is the spatio-temporal scale that is most frequently discussed when referring to Pennsylvanian-Permian climatic changes (see, for example, Remy, 1975, who recognizes changes at multiple scales, but focuses strongly on the long-term patterns). It is often discussed, despite clear awareness of the patterns at more resolved temporal scales, as if there is a monotonic trend toward drier and warmer climates, accompanied by a simple vegetational transition. In so doing, an unintended consequence is lack of acknowledgement of the tremendous complexity of this interval, even in an area confined to the low latitudes, encompassing climate and sea level changes, changes in continental positions, the rise and subsequent erosion of mountain ranges, atmospheric circulation patterns and composition, and the interaction of all of these factors. The effects of this multiplicity of variables, many changing in concert at one or more scales of resolution, on the distribution of terrestrial vegetation should be expected to be enormous, and there is no reason to believe that the tropics were not populated by several distinct, barely overlapping species pools, including both plants and animals, that migrated in and out of lowland, basinal settings where the likelihood of preservation was greatest.

This raises the issue of "apparency" in vegetational sampling, and what may be described as a taphonomic megabias (sensu Behrensmeyer and Hook, 1992; Behrensmeyer et al., 2000), that is a large scale, persistent bias in the basic structure of the fossil record. In this case, we use "apparency" to refer to what is most likely to be discovered in the course of normal field activities when prospecting for and collecting plant fossils. For example, there is unambiguous evidence of seasonally dry vegetation living in basinal lowlands beginning at least by the early Middle Pennsylvanian (Atokan/Bolsovian/early Moscovian), from every major terrestrial basin across North America and some in Europe (discussed, with citations, in the preceding paragraphs), that is, not simply in allochthonous assemblages transported into the basins from local "uplands". Yet, that vegetation is poorly represented and any random sampling of likely host rocks will yield almost invariably a wetland flora. This reflects the strong differences in the preservational conditions during the wetter times of climate-sea level-deposystem cycles in comparison to those when seasonally dry climates prevailed – paleosols may be our best evidence of the plants that once occupied the relatively dry landscapes. Yet, given estimated times of accumulation of peat vs. the development of thick, calcic vertic paleosols, it is not unreasonable to believe that seasonally dry floras were the dominant vegetation of basinal lowlands for extended periods of time during the late Middle and Late Pennsylvanian (DiMichele et al., 2010c), perhaps even occupying these areas for longer than the wetland vegetation that is such an iconic representation of the "coal age" (Falcon-Lang and DiMichele, 2010; Dolby et al., 2011). Conversely, there is ample evidence of the survival of wetland floras well into the Permian, even into the classic Kungurian-age red beds of north-central Texas (DiMichele et al., 2006; Chaney and DiMichele, 2007), demonstrating continued presence on the landscapes, even in the drier parts of equatorial western Pangea.

Asynchonous patterns of vegetational change appear, unsurprisingly, to characterize the Permian itself, as much as the transition from the Pennsylvanian to the Permian. Recent papers on early occurrences of such plants as gigantopterids (Ricardi-Branco, 2008; Booi et al, 2009a) and comiods (Booi et al., 2009b), and certain groups of conifers (Looy, 2007) extend the impression that these plants evolved in tropical envionments and spread out through time, into western Pangea (e.g., Chaney et al., 2009) and still later into such extra-tropical areas as Angara (Mamay et al., 2009). Similarly, early occurrences of typically Mesozoic plants (e.g., DiMichele et al., 2001; Kerp et al., 2006) indicate that there was a tremendous amount of evolutionary and ecological dynamics in the terrestrial landscape that either escapes detection entirely, appears only in brief glimpses as conditions in preservational basins create exceptional windows of opportunity, or appears millions of years after the actual evolutionary origin of the plants and their ecological associations. Thus, the rare occurrence of exotic plant fossils may provide more information about the dynamics of climate than do the abundant wetland floras. Caution is called for when interpreting the fossil record without understanding, or even just an awareness, of the local and regional climatic and sedimentological context of a collection.

CONCLUDING REMARKS

The characterization of changes in terrestrial vegetation across the Pennsylvanian-Permian boundary is literally a problem of "seeing the forest for the trees." One's understanding of it depends on the scale or scales of space and time at which one resolves the data empirically. It also depends on the geographic region with which one is most familiar. Smaller scale, glacial-interglacial oscillations can be detected in most basins along the paleo-equator. However, the patterns and timing of the longer-scale trends – the prominence and areal extent of aridity, say, are spatially patterned and largely time transgressive from west-to-east along the equatorial region. Additionally, given the approximately 10 My of the Pennsylvanian-Permian transition, plant evolution also must be factored into paleoecological analyses. The ultimate conclusion that we offer is not unique, but is similar to that of many who have looked at the Carboniferous-Permian transition, lithologically, faunistically, or floristically: the end points are distinct, but between these different terminal conditions lies a long transition interval, probably as long as either of the better characterized periods. The rationale for a "Dyassic" Period, long ago discussed as a time interval of gradational change between the last coal beds of the Carboniferous and gypsum deposits of the Permian, is clear.

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HARPER'S GEOLOGICAL DICTIONARY

AUTOCHTHONOUS - [Gr. Autos - self + chthon - earth, land] Your own personal piece of property, belonging only to you and no one else.

THE ENIGMATIC DUNKARD MACROFLORA

Bascombe M. Blake, Jr. and William H.Gillespie

INTRODUCTION

During the Carboniferous and Early Permian, various continental land masses consolidated along the paleoequator to form Pangea. The weight of the growing imbricated thrust sheets depressed the crust, forming a series of foreland basins within the collision zone between present day Africa and South America (Gondwanaland) with a combined North America and Europe (Laurussia). Additionally, during the Late Paleozoic the Earth was experiencing a global ice age (Fielding et al., 2008). Milankovitch orbital cycles drove cyclic changes in continental ice volumes and produced the cyclothemic (transgression/regression) sediment packages so prominent during this time (Wanless and Shepard, 1936; Heckel et al., 2007). The resultant tropical basins extended eastward from the North American Midcontinent region, through the Appalachian region, the Canadian Maritimes, the British Isles, western and central Europe and into China (Schopf, 1975). These paralic basins are collectively known as the Pennsylvanian paleoequatorial belt (Wagner, 1993), although they began forming in the Late Mississippian and continued into the Early Permian (Blake et al., 2009). The long-term climate stability resulted in the establishment of the well-studied paleoflora found in these low-lying coastal basins known as "The Coal Flora," the Euramerican Floral Realm (Wagner and Winkler Prins, 1991, 1994) or the Amerosinian Floral Realm (Chaloner and Lacy, 1973).

The youngest strata in the Appalachian Basin crop out in southwestern Pennsylvania, northern West Virginia and southeastern Ohio. Early workers (e.g. H.D. Rogers, 1858; W.B. Rogers, 1884) mirrored western European usage and referred to these strata as the Upper Barren Measures due to the lack on minable coals. I.C. White (1891) named these rocks the Dunkard Creek Series for exposures along the Dunkard Creek in northern West Virginia and southwestern Pennsylvania. He (White, 1903) later shortened the name to the Dunkard Series. Modern usage has replaced the incorrect temporal term "Series" with the correct lithostratigraphic term "Group."

The Dunkard Group (Figure 1) comprises a clastic succession with minor non-marine carbonates and mostly discontinuous muck puddle coaly beds. The Dunkard Group is an erosional remnant preserved within the Pittsburgh-Huntington Synclinorium, a feature frequently and incorrectly termed "The Dunkard Basin." These strata occur hundreds of kilometers from their nearest correlatives in the Anthracite basins of eastern Pennsylvania and various basins to the west, a separation that has hindered development of regional syntheses. Please refer to Fedorko and Skema (2011) for specific discussions on the stratigraphy of the Dunkard Group. Interested readers are referred to Martin (1998) for a recent treatment of the Dunkard Group in the Appalachian Basin.

This paper reviews plant macrofloral assemblages reported from the Dunkard Group and discusses why they do not provide an unequivocal answer to the age of the Dunkard Group. Specifically, this paper discusses the chronostratigraphic confusion that results because

Blake, Jr., B. M. and Gillespie, W. H., 2011, The enigmatic Dunkard macroflora, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 103-143.

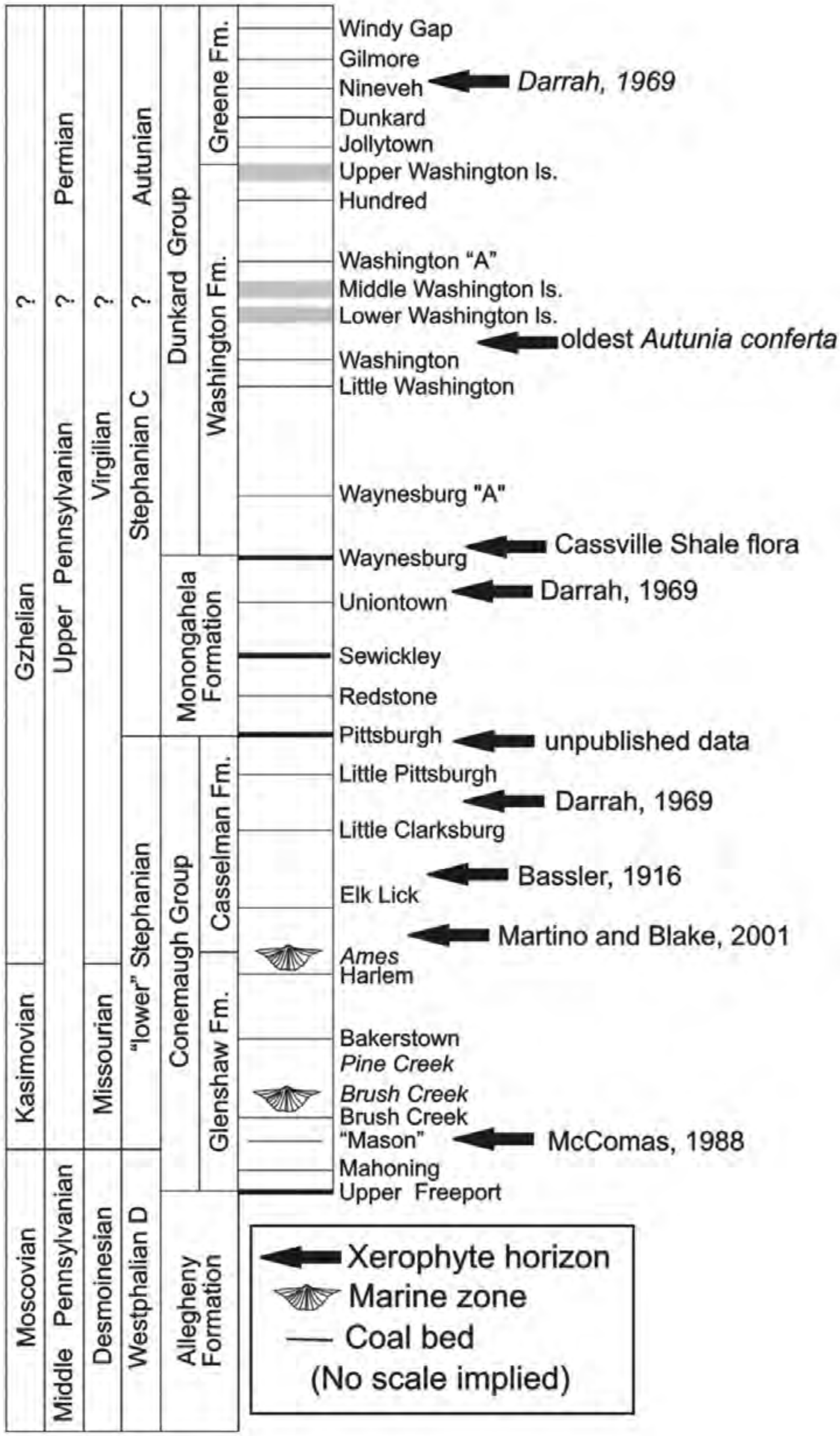

Figure 1. Stratigraphic column showing important Upper Pennsylvanian/Lower Permian stratigraphic units. The stratigraphic level of xerophyte (dryland)

Dunkard plant macrofossil assemblages varied in response to changing paleoecologic conditions. Additionally, Dunkard plant macrofossils are compared with slightly older assemblages reported from the Upper Pennsylvanian Conemaugh Group and Monongahela Formation in the northern Appalachian Basin. Finally, the question as to whether the age of the Dunkard Group is Late Pennsylvanian, Early Permian, or both is discussed.

Biostratigraphy

Traditional biostratigraphic techniques have allowed paleontologists to subdivide determine which chronostratigraphic information various fossils can provide. The ideal index species are easy to identify, geologically short-lived, generally common in collections, and cosmopolitan in occurrence. Species restricted to specific facies tend not to make ideal index fossils. The temporal range of a species is also important. Long-lived forms, such as *Lingula*, do not provide the short-term temporal resolution required to subdivide the geologic time scale into increasingly finer increments. First and last occurrences of species are frequently cited in biostratigraphic studies. Knowing the preserved geologic range of a taxon can be valuable, but can we ever be sure? The Signor-Lipps Effect (Signor and Lipps, 1982) suggests that the incompleteness of the fossil record precludes finding either the first or last member of a taxon, even if it was preserved. This effect is even more problematic when dealing with facies-constrained taxa such as plants that evolved in areas remote from the receiving basin or survive for long time periods in remote refugia (e.g. *Metasequoia*). If these species reappear in the rock record, it may be difficult to determine if they are a “Lazarus” species returned from the dead (see Jablonski, 1986) or an “Elvis” species reshaped by convergent evolution (see Erwin and Droser, 1993). This problem is compounded in paleobotany, which is forced frequently to deal with dispersed organs, a problem that can mask huge morphological differences within a single kind of plant (e.g. *Stigmaria ficoides*, the rooting organ of arborescent lycopsids). Employing new tools, such as sequence stratigraphy, which can have great impact on paleoclimate and facies architecture, can shed new light on old problems.

THE DUNKARD MACROFLORA

The Dunkard macroflora has been well-studied beginning with the publication of Fontaine and White’s (1880) seminal monograph suggesting a Permian age for the Dunkard. Later important contributors to Dunkard paleobotany include David White, Aureal T. Cross, William C. Darrah, and John, Clendening, plus those made by the authors of this paper. The debate over the position of the Carboniferous-Permian boundary in the Appalachian Basin culminated in a symposium “The Age of the Dunkard,” held in Morgantown, West Virginia in 1972. Many presentations were given and a symposium volume was published (Barlow and Burkhammer, 1975), but no consensus was reached. The paleobotanical component of this debate remains unresolved because the Dunkard macroflora is a continuation of the underlying Conemaugh and Monongahela macrofloras which are clearly middle to late Stephanian (see discussions in Read and Mamay, 1964; Darrah, 1969, 1975; Gillespie et al., 1975; Blake et al., 2002). Complicating the issue further, the Dunkard contains several elements characteristic of older formations (discussed below) and a few new introductions suggestive of a Permian age (discussed below), notably several species of callipterids. For a discussion of the callipterids (Plate 11, Figures 1-2, 5-7) of the Dunkard, see DiMichele et al. (2011). Also refer to Eble et al. (2011) for a

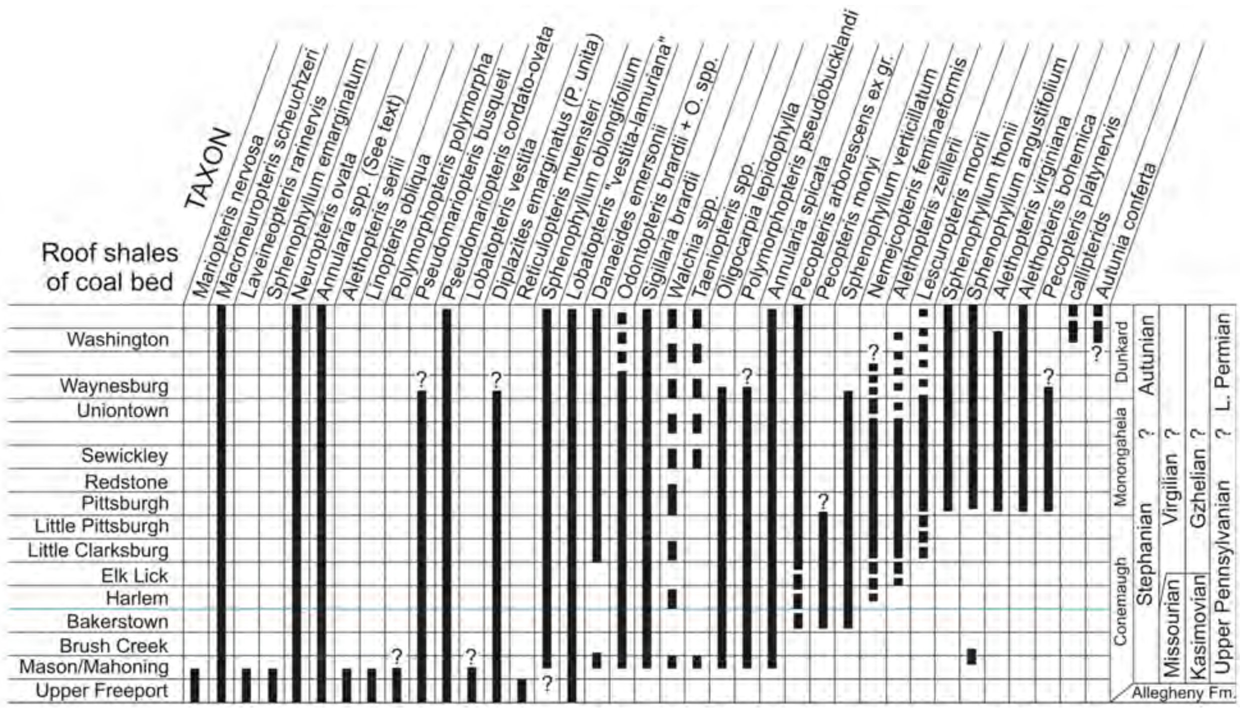

Figure 2. Seriation showing ranges of common and important Upper Pennsylvanian/ Lower Permian plant macrofossils for the northern Appalachian Basin. Chart is adapted from Blake et al. (2002). The horizontal rows represent the roof shale interval above the named coal bed. Note that while the Allegheny-Conemaugh contact is the top of the Upper Freeport coal bed, the Middle-Upper Pennsylvanian chronostratigraphic boundary occurs in the Mason/Mahoning interval between the Upper Freeport and Brush Creek coals.

discussion of the mire-centric plants of the Dunkard. Herein is a brief description of the most common Conemaugh, Monongahela and Dunkard (Upper Pennsylvanian/Early Permian) plant fossils likely to be encountered during the field trip. Refer to Figure 2 for species ranges. The following descriptions are excerpted and adapted from Gillespie et al. (1978).

Lycopsids

The large tree-like lycopsids that dominated the coal flora during the Early and Middle Carboniferous suffered a major extirpation near the end of the Middle Pennsylvanian due to climate changes (Phillips et al., 1974; Kosanke and Cecil, 1996; Peppers, 1996, 1997). There was one notable survivor, *Sigillaria brardii* (Plate 12, Figure 3) which was a moderately-size tree but did not live exclusively in the wet, poorly-drained environments as did the earlier lycopsids. As such, *Sigillaria brardii* rarely dominated wetland habitats and is relatively unimportant in Upper Pennsylvanian/Lower Permian fossil plant collections. Lycopsid leaves, assigned to the form genus *Lepidophylloides*, are long and thin, similar to blades of grass. Size of the leaves varies with position on the tree, becoming shorter in the upper portions as any individual tree approached the end of its determinate life cycle. Lycopsid leaf adpressions cannot reliably be separated into natural species. Lycopsid rooting organs are nearly identical in all species and are assigned to the genus *Stigmaria*, which can be thought of as an underground branch network. *Stigmaria* extends from the base of the tree and rapidly dichotomizes into two or four large axes that continue to branch as they grow through the substrate. *Stigmaria* axes are commonly preserved as casts marked by circular pits where rootlets exited. Ribbon-like adpressions of *Stigmaria* rootlets can be common in poorly-drained paleosols.

Sphenopsids (*Calamites* and *Sphenophyllum*)

There are two main taxonomic orders of the sphenopsids: the Equisetales, which include the tree forms attributed to *Calamites,* and the Sphenophyllales, which include the ground cover form *Sphenophyllum*. As a group the sphenopsids had jointed stems, underground rhizomes with adventitious roots, upright axes with monopodial branches bearing whorls of leaves and terminal cones. All sphenopsids have distinct nodes (attachment points of leaves and branches) and internodes, which are marked by distinct, regular-spaced, longitudinal ribs and furrows. *Calamites* species were generally small tree-like plants that rarely reached 10 m (33 ft), although some large, highly woody species are known. Some species could sprout new growth from buried trunks, an adaption to life on shifting substrates. See fig. 1 in Eble et al. (2011) for a reconstruction of a calamite plant.

Calamites stems were filled with easily decayed pith or were hollow, except at the nodes, where a thin layer of tissue (a diaphragm) was present. When the plants died and fell, the hollow cavity was filled with sediment which hardened into rock and was preserved. The surrounding woody tissues are rarely preserved except as a thin layer of coal. The resulting pith cast shows the jointed arrangement of the stem, a pattern of longitudinally aligned ribs and furrows in the internodal regions, and sometimes circular scars indicating the positions of branches at the nodes.

There are two common types of *Calamites* foliage: *Asterophyllites* and *Annularia*, which are considered “form genera” as many sphenopsids had polymorphic leaves. All leaves have single veins in these genera. *Asterophyllites* leaflets are narrow and “grass-like” and the whorls are not fused together. The whorl is concave upwards around the stem and the leaflet tips often overlap the base of the next whorl. The whorl was normally flattened during preservation. *Asterophyllites equisetiformis* and *A. longifolium* are the two common Dunkard species.

The second calamitean foliage form genus, *Annularia,* has whorls of lanceolate leaves. The leaves are fused into a ring at the node to which they are attached which makes them stiff and preservable in a flat plane; thus the entire whorl is seen. The leaves, and therefore whorls, vary in overall size according to the position on the plant. Four species of *Annularia* are common in Upper Pennsylvanian/Early Permian collections; *Annularia carinata*, *A. mucronata*, *A. stellata,* and *A. sphenophylloides*. *Annularia carinata* has oval whorls of spatulate leaves of unequal length with mucronate tips. In *Annularia mucronata* (Plate 3, Figure 2) and *Annularia sphenophylloides* (Plate 5, Figure 3) leaves are equal in length, spatulate and mucronate. There is some question whether these forms are separate species. *Annularia stellata* has lanceolate, equal-length leaves. As discussed by DiMichele et al., (2010a), populations of *Annularia carinata* often are considered mixtures of *Annularia sphenophylloides* (smaller whorls) and *Annularia stellata* (larger whorls). As they noted, the latter two species have symmetrical whorls formed by leaves of approximately equal length. Because of the co-mingling of species in published collections, the ranges shown on Figure 2 are combined under *Annularia spp.*

Sphenophyllum is characterized by jointed stems with swollen nodes, longitudinally ribbed internodes, whorls of multi-veined, wedge-shaped leaves, and terminal cones. The stems are generally much less than a centimeter in diameter. *Sphenophyllum* is envisioned as a scrambling or climbing understory vine-like plant. Some species have spine-like attachment hooks. Wedge-shaped leaves occur in whorls at nodes. The leaves are polymorphic, varying in shape and degree of incision along the plant leading to over splitting. Depending on the species, one or two veins enter the base of the leaf and then divide repeatedly ending in a terminal "tooth." There are a number of species reported from Upper Pennsylvanian/Lower Permian collections. *Sphenophyllum angustifolium* (Plate 7, Figure 3) internodes are unribbed. Whorls contain 4 to 6, 2.5 to 4 cm (1 to 1.6 in) long leaves and average about 1 cm (0.4 in) wide at the tips. Each leaflet is split about two-thirds of its length; tips are bluntly toothed.

Sphenophyllum oblongifolium (Plate 6, Figure 7; Plate 12, Figure 5) is the most common of the *Sphenophyllum* species and specimens exceeding about 1 m (several ft) in length are common. Whorls contain 6 leaves that are arranged in 3 pairs; the lower pair deflexed at about 90 degrees to the others. This characteristic makes this species easy to identify. The first occurrence of this species has been considered an index to the beginning of the Stephanian Substage (Wagner and Lyons, 1997; Wagner and Álvarez-Vázquez, 2010).

Sphenophyllum thoni (Plate 12, Figure 3) is the largest species of the group found in the Upper Pennsylvanian/Lower Permian of the Appalachian Basin. Six similar-sized leaves comprise the whorls of *Sphenophyllum thoni* and are considerably larger than other members of the genus. Individual leaves have rounded tips which may be lobed or fringed. The range of this species in the Appalachian Basin is poorly known with specimens being reported from the base of the Conemaugh (Wagner and Lyons, 1997), shales below and above the Pittsburgh coal bed and Cassville Shale (roof shale of the Waynesburg coal).

Marattialean Tree Ferns

As discussed by Eble et al. (2011), marattialean tree ferns were a major component of the Late Pennsylvanian/Early Permian paleoflora in the northern Appalachian Basin. Marattialean tree fern foliage is assigned mainly to the form genus *Pecopteris* and is generally very common in Dunkard macrofloral collections. Pinnule shape can exhibit considerable variation within specific tree ferns depending on position on the frond (e.g. Darrah, 1969) which has led to excessive division into species in the literature, especially in early reports. Accurate identification requires large collections or large frond fragments. There are probably only six or seven natural species of tree ferns in the Dunkard notwithstanding the plethora of names. Common pecopterids include *Pecopteris arborescens, P. candolleana, P. cyathea, P. hemitelioides, P. lamuriana (=Lobatopteris tenuivervis), P. polymorpha,* and *P. unita* (*Diplazites* cf. *emarginatus*). Tree ferns are subdivided into natural groups based on fructifications.

Pecopteris lamuriana is the Stephanian evolutionary end product of an apparent sequence of morphotypes that represents a gradational series of forms. The series began with *Pecopteris miltoni* in the Kanawha Formation (Duckmantian; Westphalian B), proceeded through *Pecopteris micromiltoni* (Bolsovian; Westphalian C) and *Pecopteris vestita* in the Middle Pennsylvanian Allegheny Formation (Asturian; Westphalian D), terminating with *P. lamuriana* in the early Stephanian. Minor taxonomic differences and a postulated substantial unconformity between the Middle and Upper Pennsylvanian in North America (see discussion in Wagner and Álvarez-Vázquez, 2010; Falcon-Lang et al., 2011), led Wagner and Lyons (1997) to suggest that specimens of *P. lamuriana* identified in the North American Upper Pennsylvanian were, in fact, an Elvis taxon, meaning a distinct species that morphologically “imitates” the older form. To support their hypothesis, they resurrected, without adequate diagnosis, Fontaine and White’s (1880) name *Pecopteris tenuinervis* (Plate 10, Figure 3) to replace *P. lamuriana* in the Appalachian region. This complex can be difficult to identify based on fragmentary specimens. Its pinnules are extremely polymorphic due to a complex series of lobing that occurs during leaf development. This species occurs commonly in late Middle Pennsylvanian through Upper Pennsylvanian/Lower Permian strata from mid-Allegheny to the lower Dunkard.

A second genus of probable tree fern leaves is *Danaeides* which has only one known species, *D. emersonii* (Plate 8, Figures 1-3; Plate 9, Figure 1, Plate 12, Figure 2). The fronds were quadripinnate and up to 1 m (3.3 ft) long. Sterile pinnules are easily confused with certain other pecopterids. In general, they are attached by the entire base, oblong, blunt-tipped, have straight-to-convex sides, and are basally constricted. The veins are often obscured in the thick leaflets, but sometimes they are about all that can be seen. Lateral veins arise from a well-developed midvein and arch to the margin, sometimes singly, but usually forking once just after leaving the midvein. In well-preserved specimens, veins entering directly from the stem are evident. Fertile pinnules have rows of linear sporangia arranged on the lower surface along the veins from the midvein to the leaflet margin. In most instances, the sporangia are fused to such a degree that single sporangia cannot be distinguished. A fertile leaflet greatly resembles a sowbug. *Danaeide*s occurs throughout the Upper Pennsylvanian of North America, although it does not become common until about the horizon of the Pittsburgh coal bed.

Ferns (Filicales)

Small, ground cover ferns are not common in Upper Pennsylvanian/Lower Permian plant fossil collections probably due to preservation biases. Fern foliage is not as robust as tree ferns and the plants were smaller, producing less preservable biomass. In addition, as largely ground cover and possibly vines, these plants were less likely to be naturally “sampled” and have their leaves transported into depositional settings. These plants exhibit a variety of leaf shapes and are assigned to several form genera, notably *Pecopteris* and *Sphenopteris* (Brousmiche, 1983). These plants were the understory ground cover and perhaps were early colonizers of newly disturbed substrates similar to those that created the “fern spike” that appears immediately above the post-Cretaceous impact clay layer. Compared to tree ferns, these plants were relatively small and tend to occur as fragmentary fossils. A few widespread forms are important in biostratigraphic studies in this interval.

One of the more commonly encountered, and highly distinctive, that belongs to this group is *Pecopteris feminaeformis*, which has been restudied and named *Nemejcopteris feminaeformis* (Plate 5, Figures 1-2). It has stiff-appearing, toothed leaflets with a thick midvein from which straight, unbranched lateral veins depart and run to a tooth on the margin. Because of a strongly -toothed margin (Plate 5, Figure 2), *N. feminaeformis* is easy to identify and is likely not a typical tree fern species. It is often found in a very fragmentary condition in Upper Pennsylvanian strata from the base upward into the Dunkard Group. Darrah (1969) reports finding two very poorly preserved specimens in the Lower Kittanning interval in Pennsylvania. Wagner and Lyons (1997) suggest that this report is untenable because the stratigraphic position is too low. These workers consider *N. feminaeformis* to be strictly a Late Pennsylvanian/Early Permian species.

Seed Ferns (pteridosperms)

The seed ferns (pteridosperms) are a group of extinct vines, shrubs, and small trees. They had large fern-like, compound leaves (fronds) and reproduced by seeds instead of spores. Most of the famous "fern fossils" so common in earlier Carboniferous "Coal Flora" plant collections are actually seed ferns. However, these nearly obligate wetland plants were in decline due to climate change during this time and are less common in Upper Pennsylvanian/Lower Permian assemblages of the Appalachian Basin than in older deposits.

Seed fern fronds could be quite large, exceeding 3 m (10 ft) long and 1.5 m (5 ft) wide (Laveine, 1986). The pinnule shapes of the various form-genera are quite different and are used to subdivide the seed ferns into more or less natural groups. Common wetland seed fern form genera in Dunkard collections include: *Alethopteris*, *Lescuropteris*, *Linopteris*, *Macroneuropteris*, *Neuropteris*, *Odontopteris* and *Pseudomariopteris*. The seasonal-drought adapted callipterids, assigned to the Peltaspermales (Kerp, 1988), also appear to be a type of pteridosperm, or an evolutionary derivation from that group; the genera *Autunia*, *Lodevia*, and perhaps *Rhachiphyllum*, have been identified in Dunkard strata (see DiMichele et al., 2011).

Alethopteris pinnules are thick and were probably leathery in life. They have a convex adaxial surface and are attached by their entire bases, which are usually decurrent and confluent. Pro-minent midveins extend nearly to or to the tip of the pinnule. Numerous lateral veins emerge from the midvein at nearly right angles and fork variously to the pinnule margin. Numerous veins also pass directly from the rachis (axis on which the pinnules are borne) into the decurrent portion of the pinnule base. Common species include *Alethopteris bohemica* (Plate 13, Figures 7-8), *A. virginiana*, and *A. zellerii*. *Alethopteris virginiana* (Plate 3, Figure 3; Plate 9, Figure 2-3) is common in the main parting of the Waynesburg coal at Cassville, West Virginia, the classic Fontaine and White (1880) locality.

Neuropteris pinnules are tongue-shaped, usually blunt-tipped, with cordate (heart-shaped) bases. Pinnules are narrowly attached to the rachis. Midveins seldom extend past the middle of the pinnule. Lateral veins are numerous, dividing variously while arching to the pinnule margin. The rachis of some species, particularly *Neuropteris* (*Macroneuropteris*) *scheuchzeri*, has broadly triangular spurs which appear as "rose thorns" but are the points of attachment of the pinnules that have fallen away. *Neuropteris ovata* (Plate 4, Figure 1; Plate 5, Figure 5-6) has variable-sized pinnules characterized by a nearly straight base, rounded acroscopic corner, and a basiscopic corner characteristically prolonged into an auricle or ear. The midvein is poorly marked, with numerous lateral veins dividing four times while arching to the margin. The terminal pinnule is relatively long, broad, and asymmetrical. Intrafrond variations have lead to excessive division of *Neuropteris ovata* into additional species. Stalked fimbriate cyclopterid pinnules (Plate 4, Figure 1; Plate 5, Figure 6) occur near the base of the fronds (*Cyclopteris fimbriata*). Early reports confused this species with the older *Neuropteris heterophylla*. *N. ovata* first appears at the base of the Westphalian D (Asturian) substage and, in North America, ranges upward into the Permian (DiMichele et a;., 2010a; see discussions in Barlow and Burkhammer, 1975).

Neuropteris (Marconeuropteris) scheuchzeri (Plate 6, Figures 1-6) had very large fronds. The pinnules are large and polymorphic, but intergrading specimens connect the different morphologies. Pinnules are generally found dispersed. Deeply lobed (polymorphic) forms occur commonly in Monongahela and Dunkard strata. The pinnules are typically tongue-shaped with blunt tips, cordate bases, and may have two paired, circular cyclopterid pinnules

(*Cyclopteris orbicularis* when dispersed) pinnules at the base. A midvein extends nearly to the tip of the pinnule. Thin lateral veins emerge at an acute angle and divide three to four times, usually four, with the first division coming immediately after departure, and the fourth no more than half way to the margin, which is met at a wide angle. Stiff, abaxial pinnule hairs are characteristic of this species (Plate 6, Figure 6), but they can be lost when specimens are removed from the entombing matrix. The hairs frequently lie cross lateral veins, aiding identification. The polymorphic nature of the pinnules and frequent absence of the abaxial hairs due to preservation factors, have resulted in division of this species into several segregate forms in the literature. Cuticle studies have shown that there are several unrelated plants that have large pinnules similar in shape to *N. scheuchzeri*: th presence of abaxial hairs is diagnostic. *Neuropteris scheuchzeri* first appears in the upper part of the Pottsville Group and extends into the Permian (Figure 2) in North America (DiMichele et al., 2010a).

A somewhat uncommon but interesting form genus is *Linopteris.* Pinnules in this genus are neuropterid in appearance, but the veins anastomose and form networks. This characteristic becomes common in many Permian forms. It is sometimes necessary to use a hand lens to see the pattern. This genus occurs in the late Middle and Upper Pennsylvanian, being most common in Allegheny and lower Conemaugh strata.

Odontopteris (Plate 5, Figure4; Plate 12, Figure 6) pinnules are smooth margined, tongue-to-sickle-shaped, and are attached by their entire base. Several veins pass directly, or after dividing one or more times, from the rachis to the margin. *Odontopteris* is an Upper Pennsylvanian and Permian genus that seems to have evolved from *Neuropteris* through the breakup of large leaflets into smaller segments. Some of the polymorphic leaflets of *Neuropteris* found in the Upper Pennsylvanian often resemble this genus.

Lescuropteris (Plate 3, Figure 1; Plate 7, Figures 1-2) resembles *Odontopteris* and is characterized by broadly attached, scythe-shaped (fulcate) leaflets in which the lower veins emerge directly from the rachis and the upper ones emerge alternately from a thin midvein. Lateral veins may approach each other, and on rare occasions, may anatomose (unite). The main rachis bears half-round leaflets between the lateral rachises, termed intercalary pinnules. This plant has been found from the roof shales of the Pittsburgh coal bed upward into the Dunkard, but it is rare. Several recorded occurrences are mis-identified tips of *Odontopteris brardii*. *Lescuropteris moorei* is the only species in the Appalachian region.

Pseudomariopteris (Plate 13, Figure 3) has triangular-to-elongate-oval leaflets with a distinct midvein, but the lateral veins are rarely visible because the pinnules were thick. This seed fern is a vine (Krings and Kerp, 2000; Krings et al., 2003). *Pseudomarioteris* first occurs in the late Middle Pennsylvanian, but is by and largely an Upper Pennsylvanian species and can be common in Dunkard Group collections. The common species is *Pseudomariopteris cordato-ovata*.

Coniferophytes

Two groups of woody, coniferophyte gymnosperms were present in the Late Pennsylvanian/Early Permian: the cordaitaleans and the conifers. See discussions in Eble et al. (2011). Cordaitaleans were trees that existed from the Upper Mississippian into the Triassic. The form genus *Cordaites*, originally used as the generic name for strap-shaped, paralleled-veined leaf adpressions, is now applied to the entire plant. Cordaitalean trees were apparently adapted to live on dry ground as well as in swamps (Falcon-Lang and Scott, 2000; Falcon-Lang and Bashford, 2004). Some had a mangrove-like (stilt) root system and probably lived in a near-shore (lake or marine) environment similar to today's tropical mangroves. Cordaites leaves are long and strap-shaped with fine veins, parallel to the leaf margin and running the length of the leaf. Because of their size, complete specimens are uncommon although cordaitalean leaves can dominate collections in some facies. Leaves are difficult to separate without cuticle studies.

Conifers are gymnospermous trees most of which have needle-like foliage and seeds borne in cones. *Walchia* (Plate 11, Figures 3-4; Plate 13, Figures 1-2) is the artificial classification used for poorly preserved Upper Pennsylvanian and Lower Permian coniferous foliage throughout the world. Since conifers lived in the drylands surrounding the Euramerican wetlands, they have a poor fossil record through the Carboniferous. Coniferous pollen has been reported from the Upper Mississippian (Scott and Chaloner, 1983) and macrofossils have been reported from the Middle Pennsylvanian in Great Britain (Duckmantian; Bashkirian) (Scott, 1974) and Oklahoma (Asturian; Moscovian) (Rothwell, 1982). They occur sporadically throughout the Upper Pennsylvanian/Lower Permian of the northern Appalachian (Figure 1) and are common elements of Late Pennsylvanian, especially Missourian, deposits in the Midcontinent and western portions of the Pangean equatorial region (Elias, 1936; Cridland and Morris, 1963; Cridland et al., 1963; Scott, 1974; Scott and Chaloner, 1983; Winston, 1983; Leisman et al., 1988; Tidwell, 1988; Lyons and Darrah, 1989). They become relatively abundant in the Permian. Unbranched and poorly preserved specimens can easily be confused with leaf-bearing lycopsid twigs.

Conifers occur sporadically in Upper Pennsylvanian (Stephanian) strata in the Appalachian region, being reported from Ohio (McComas, 1988), Pennsylvania (Darrah, 1969), and West Virginia (Lyons and Darrah, 1989; Martino and Blake, 2001).

DISCUSSION

The age of the Dunkard Group has been controversial since publication of Fontaine and White’s (1880) seminal monograph on the Dunkard macroflora. Since then, the Dunkard has been considered wholly Middle Pennsylvanian (Bode, 1975), wholly Upper Pennsylvanian (Stevenson, 1907; Clendening, 1972, 1974, 1975; Clendening and Gillespie, 1972; Gillespie et al., 1975), both Upper Pennsylvanian and Lower Permian with a boundary at the first occurrence of *Autunia conferta* (Figure 1) (Darrah, 1937; Berryhill et al., 1971; Cross; 1975; Havlena, 1975) or perhaps entirely Permian (Fontaine and White, 1880; White, 1931, 1936). The first occurrence of *Callipteris* (*Autunia*) *conferta* was adopted by early International Carboniferous congresses as the terrestrial index fossil to the base of the Permian, a conclusion since complicated by the report of *C. conferta* from the Stephanian C of Europe (Havlena 1970) and the Virgilian of Kansas (Leisman et al., 1988) where it is associated with undoubted Pennsylvanian marine fossils. In Euramerican basins, the first appearance of *C. conferta* broadly corresponds to the sporadic occurrence of a Permianesque floral assemblage (Blake and Gillespie, 2011) that actually first appears earlier in the Upper Pennsylvanian (Figures 1 and 2). Wagner (1984) considered that the sporadic appearance of *Autunia conferta* above the Washington coal bed suggested a correlation with a level high in the European Stephanian C (Late Pennsylvanian) or perhaps even Lower Autunian (Early Permian). Wagner (1984) correlated these strata with his *Autunia conferta* biozone which was consistent with Read and Mamay’s (1964) opinion. An international symposium was held in Morgantown, West Virginia in 1972 to address the question of the position of the base of the Permian in the Appalachian Basin (Barlow and Burkhammer, 1975).

A complicating factor in North American basins is the common occurrence of *Neuropteris ovate*, which was long considered an index species of the Westphalian D (now Asturian) in western European basins (see discussions in Bode, 1975, and Darrah, 1969, 1975). Several associated Asturian species common in late Middle Pennsylvanian collections from North America, notably *Macroneuropteris scheuchzeri*, are also common in some Dunkard collections (Figure 2). In the Dunkard, *Autunia conferta* is relatively rare while *Neuropteris ovata* and *Macroneuropteris scheuchzeri* are two of the more common taxa in many collections. Using plant macrofossils, it can be argued that the Dunkard may be both late Middle Pennsylvanian and Early Permian. Currently the base of the Permian is defined on purely marine criteria (Heckel et al., 2007) which are absent in the terrestrial Dunkard.

Three facies-dependant, paleofloral associations have been described from the Late Pennsylvanian/Early Permian transition (Figure 3) (Darrah, 1969, 1975; DiMichele and Hook, 1992; Blake et al., 2002; DiMichele et al., 2001, 2010b; Blake and Gillespie, 2011). Havlena (1961) recognized that these assemblages were separate communities with different moisture requirements living contemporaneously on a single landscape. A hygrophilous (poorly-drained substrate) paleoflora (the Coal Flora) dominated wetland habitats throughout the Lower and Middle Pennsylvanian (Westphalian), changing little in overall structure until the extinction of the mire-centered lycopsids (Phillips et al., 1985; Peppers, 1996) during the Middle-Upper Pennsylvanian (Westphalian-Stephanian) transition. An impoverished remnant of this paleoflora extended, in wetland habitats, through the Upper Pennsylvanian into the Permian (Darrah, 1969, 1975; Gillespie et al., 1975; DiMichele et al., 2001, 2010; Blake et al., 2002), but it never matched its Westphalian antecedent in diversity or areal extent. In the middle Westphalian, a mesophyte association developed consisting of cordaitaleans, tree ferns, pteridosperms, and sphenopsids, but rarely lycopsids (Darrah, 1969). This allochthonous mesophyte association (flözfern of Havlena, 1961) lived in better-drained areas adjacent to the poorly-drained lowlands occupied by the autochthonous or parautochthonous (sensu Gastaldo et al., 1996a, b) wetland coal floral association (flöznah flora of Havlena, 1961). Havlena's (1961). Flözfern assemblages may represent polymorphous assemblages of taxa from different growth sites within the basin, possibly combined with extrabasinal or basin-margin species (Gastaldo, 1996a, b; Gastaldo et al., 2009).

A contemporaneous xerophytic dryland (well-drained) paleoflora existed in better-drained areas outside the peat-forming regions from at least the earliest Mississippian (Figure 3) (Havlena, 1961; DiMichele and Hook, 1992) and periodically moved into the now better-drained wetlands during the drier parts of glacial-interglacial cycles. As can be expected, records of this "extrabasinal" paleofloral association are rare in the rock record. As briefly summarized by DiMichele and Hook (1992), these paleofloras are represented by Late Mississippian (Namurian) to Early Permian collections from the central and western United States (e.g. Cridland and Morris, 1963; Tidwell, 1988; Leary, 1976; Leary and Pfefferkorn, 1977; Winston, 1983; Leisman et al., 1988; Rothwell and Mapes, 1988). In the Upper Pennsylvanian of the Appalachian Basin, this Permianesque association consisted of various taeniopterids, walchian conifers, and *Plagiozamites*, with pteridosperms such as *Lescuropteris*, *Odontopteris* and various callipterids (Cridland and Morris, 1963; Cridland et al, 1963; Moore et al., 1936; Darrah, 1969; Blake et al., 2002).

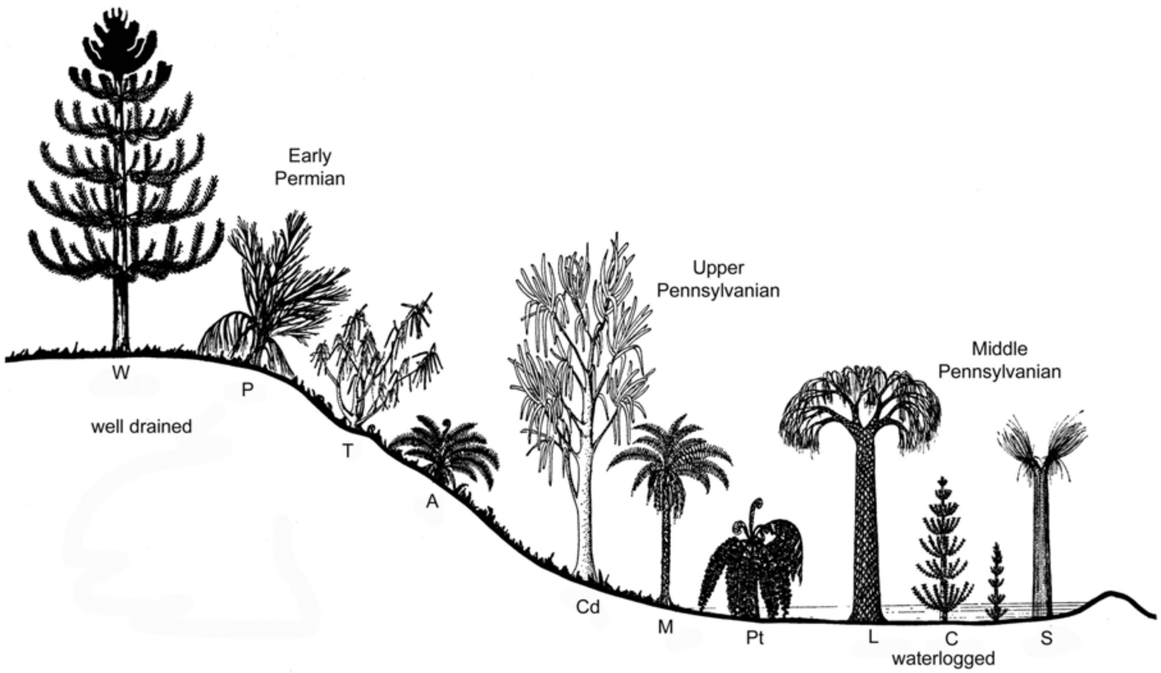

Figure 3. Sketch the shows the distribution of plant types on a hypothetical landscape during Dunkard time. During the Late Paleozoic Ice Age, changes in moisture regimes and sea level were driven by the waxing and waning of southern hemisphere continental glaciers. The above configuration shows the distribution of assemblages during wet periods. During dry climate intervals, the Permianesque (Early Permian) dryland flora migrated down into the now better-drained valleys. Apparent age of the plant associations as discussed in text are shown at top of the figure. W-walchian conifer; P-*Plagiozamites*; T-*Taeniopteris*; A-*Callipteris* (*Autunia*) *conferta*; Cd-cordaitian gymnosperm; M-Marattialean tree fern; Pt-pteridosperm; L-lycospora-bearing lycopsid; C-*Calamites*; S-*Sigillaria*. Figure adapted from Cridland and Morris, 1963.

Paleoecology

On a continental scale, climate is the main factor controlling the geographic distribution of plants (Thompson et al., 2000). Local criteria that can affect the distribution of plants include competition between species, climatic or edaphic conditions, and the presence of physical barriers preventing migration into new areas of appropriate habitat. During the Late Paleozoic Ice Age, Milankovitch-driven cyclic sea-level changes (cyclothems) caused major fluctuations in global climate (Tandon and Gibling, 1994; Flint et al., 1995; Hampson et al., 1999; Poulsen et al., 2007; Blake et al., 2009), which impacted the composition of Upper Mississippian (Blake et al., 2009) and Pennsylvanian tropical vegetation (Falcon-Lang, 2004; Driese and Ober, 2005; Feldman et al., 2005; Falcon-Lang et al., 2009; Falcon-Lang and DiMichele, 2010). A tropical rain forest developed that was dominated by lycopsids, pteridosperms, and tree ferns, the typical Carboniferous Coal Flora (DiMichele and Phillips, 1994; DiMichele et al., 2001; Falcon -Lang and DiMichele, 2010; Falcon-Lang et al., 2011). The near absence of dryland macrofloral elements within the tropical wetlands that dominated the prePangean (=Euramerican) paleoequatorial belt for millions of years cannot be taken as proof of absence. To the contrary, the lack of a record should be read as one of showing the success and ecological stability of the non-mire Coal Flora during long periods of ecosystem stasis (Pfefferkorn et al., 2000). Changes in plant communities frequently result from disruptions in the environment, such as climatic change, rather than by replacement by new forms (DiMichele and Hook 1992). This is especially true for ecological specialists such as the mire-centered, *Lycopspora*-bearing lycophytes and associated plants that dominated tropical everwet and peat-forming environments from the Late Mississippian through Middle Pennsylvanian. Increased aridity towards the end of the Middle Pennsylvanian in the Appalachian region, and the rest of North America, resulted, at least in part, from the formation of a shadow zone to the west of the rising Appalachian Mountains, which cut off the moisture-bearing easterlies as well as the northward movement of Euramerica (later Pangea) through and out of the wet tropics (Cecil, 1990; Kosanke and Cecil, 1996).

Ecologic partitioning of the Euramerian Floral Realm occurred as Late Pennsylvanian orogenies, coupled with the onset of the third stage of the Late Paleozoic Ice age, periodically changed sea levels and disrupted atmospheric circulation patterns, impacting moisture availability beginning in late the Westphalian. Climatic disruptions became widespread in the Upper Pennsylvanian (Stephanian) (Rothwell, 1982; Kerp, 1996; Blake et al., 2002). Semiarid climates began with deposition of the Conemaugh Group in the Appalachian region (Cecil, 1990), as indicated by the presence of calcareous paleovertisols and the occasional introduction of dryland plants into the former wetlands. The onset of semiarid climates is conspicuously marked by the extinction of the mire-centered lycophytes in North America, although they survived into the Permian of China (Phillips et al., 1985).

It is important to recognize that frequent sea-level and paleoclimatic changes (Elias, 1936; Moore et al., 1936; Cridland and Morris, 1963; Cridland et al., 1963; Leary, 1976; Leary and Pfefferkorn, 1977; Pfefferkorn, 1980; Knight, 1974, 1983; Broutin et al., 1990; DiMichele and Aronson, 1992; Blake et al., 2002; Falcon-Lang and Miller, 2007; DiMichele et al., 2008), create problems for macrofloral-based biostratigraphic studies utilizing the traditional tools discussed above that do not consider the impact of paleogeography or paleoclimate on the distribution of macrofloral assemblages. As discussed by many workers (e.g. Blake, 1997; Blake et al., 2002, 2009), it is important that facies-restricted assemblages be compared only with other similar facies-restricted assemblages. In other words, it is important to compare oranges with oranges - wet floras with wet floras, dry floras with dry floras. Studies that try to compare dryland assemblages with wetland assemblages, such has constantly been done with the Upper Pennsylvanian/Lower Permian macrofloras of the Appalachian Basin, will likely yield cryptic results. As mentioned previously, dryland assemblages have been described in the literature because of low preservation potential in areas surrounding the wetlands (e.g. Leary and Pfefferkorn, 1977; Falcon-Lang, 2004; Feldman et al., 2005; DiMichele and Gastaldo, 2008; Falcon-Lang et al., 2009; Falcon-Lang and DiMichele, 2010; DiMichele et al., 2010b).

Conifers are an excellent example of problems associated with using traditional biostratigraphic techniques. Since the common occurrence of conifers is frequently considered evidence of a Permian age, should the first appearance of conifers indicate the onset of the Permian? Conifers first occur in Appalachian Basin in the 7–11 Mine flora near the base of the Conemaugh (Figure 1) (McComas, 1988; Wagner and Lyons, 1997). They are known from a few isolated occurrences from across Euramerica beginning in the Middle Pennsylvanian (Scott and Chaloner, 1983; Lyons and Darrah, 1989; Scott et al., 2010) and their pollen occurs earlier in the late Mississippian (Stephenson et al., 2008; DiMichele et al., 2010a). However, conifers do not become common in the lowland tropical floras until near the Pennsylvanian–Permian boundary (DiMichele et al., 2010b). Conifers were only able to replace the previously dominant wetland assemblage when the drying trend in Euramerica that began in the late Middle Pennsylvanian changed conditions enough to allow the conifers to replace the previously entrenched wetland floras. Other Permianesque macroflora, such as *Taeniopteris, Walchia, Dichophyllum, Plagiozamites,* or *Callipteris* (traditional sense), occur sporadically throughout the Upper Pennsylvanian/Lower Permian of the Appalachians and the North American Midcontinent (Elias, 1936; Cridland and Morris, 1963; Cridland et al., 1963; Leisman et al., 1988). Using these plants, perhaps we cannot draw the proverbial line in the sand, but rather come to realize that we are dealing with a transitional sequence and macrophytes cannot provide the definitive answer.

SUMMARY

The Dunkard macroflora is comprised of three paleofloral assemblages that occupied separate habitats on the Late Pennsylvanian/Early Permian landscape, constrained largely by the availability of water. One paleofloral assemblage consisted of a depauperate relic of the wetland Coal Flora that dominated tropical Euramerica for millions of years. A marattialean tree fern dominated assemblage replaced the Coal Flora as the climate dried towards the Middle -Upper Pennsylvanian boundary and wetland habitats became scarce. The third assemblage comprised xerophytes that occupied moisture-stressed (drylands) areas surrounding the wetter lowlands and, episodically, moved into the former wetlands when conditions allowed. These dryland plants, such as the conifers and callipterids, evolved outside the depositional basins, a factor that tended to obscure their early evolution as they were rarely preserved in the fossil record. Numerous studies utilizing traditional biostratigrahic tools have provided unsatisfactory answers on the age of this paleoflora. In the Appalachian Basin, many workers placed utmost importance on the first occurrences of Permianesque forms such as *Taeniopteris jejunata* and *Walchia*. These xerophytes, first reported from near the base of the older Conemaugh Group, and the callipterids, which appeared later, lived contemporaneously in dryland settings with the relic late Westphalian-like wetland floras, and replaced the “normal” lowland Stephanian flora during drier interglacial periods. As a result, Dunkard paleofloras provide a mixed and confusing biostratigraphic signal. In fact, after 130 years of research, what do plant fossils say about the age of the Dunkard? Darrah (1975) makes the following observations that are still valid, with minor changes, today:

1. While there may be minor sequence-stratigraphic breaks, there is no floral discontinuity from near the base of the Conemaugh through the highest Dunkard
2. The Upper Pennsylvanian/Lower Permian flora is obviously transitional
3. *Callipteris*, notably *C. conferta*, arbitrarily accepted by early International Congresses for Carboniferous Stratigraphy as an index of Permian age, has been found in the Dunkard from the Washington coal and higher
4. A number of Permianesque forms such as *Taenioptereis, Walchia* and *Plagiozamites* first appear near the base of the Conemaugh and are found sporadically throughout the section in the Appalachian Basin
5. A number of survivors of the antecedent Coal Flora persist throughout the entire Upper Pennsylvanian/Lower Permian sequence in the Appalachian Basin
6. Numbers 1 through 5 hold true for correlative strata in Kansas, Texas, and New Mexico. In addition, similar floras occur in western European basins in the Czech Republic,

France, Germany, and Spain

Recognizing the climatic changes associated with glacioeustatic sea level changes provides a framework for explaining the distribution of the enigmatic Dunkard macroflora, but doesn't answer the question as to the Age of the Dunkard. The base of the Permian is defined on marine criteria, which are missing from the Dunkard. As a result, the best we can say on the basis of plants is that the Conemaugh, Monongahela, and Dunkard sequence of the Appalachian Basin is transitional and, based on known macroflora, any level selected to serve as the systemic boundary can only be considered an approximation.

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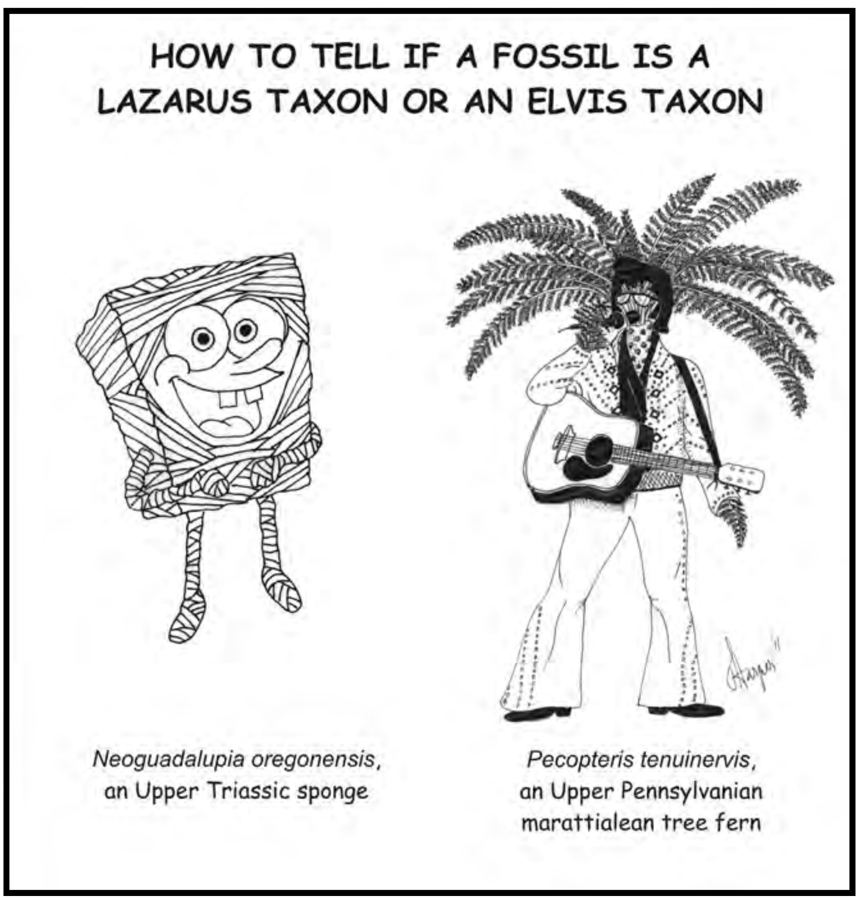


Plate Captions

These plates have been previously published in Blake et al. (2002), although a few of the images have been moved and plates slightly changed. The plate numbers are not the same as in the original report. These plates are provided herein as reference material for the field trip. Anyone desiring to cite these figures should refer to the original publication. Scale bar on all figures equals 1 cm, unless noted. Unless noted, specimens are curated at the West Virginia Geological and Economic Survey (WVGES), Morgantown, WV and are part of the WVGES' paleobotanical collection collected by the authors..

Plate 1.

Figure 1. ***Dicksonites pluckeneti*** **(Schlotheim) Sterzel. Upper Pennsylvanian Greene Formation (Dunkard Group). Between the Upper Rockport Limestone and the Gilmore coal bed. 0.2 km (0.1 mi) from Wood County line, Wirt County, WV. PSS 420/15.**

Figure 2. ***Sphenopteris durbanensis*** **Corsini. Upper Pennsylvanian Washington Formation (Dunkard Group). Roof shale of Waynesburg coal bed, along Route 100, near Pursglove, Monongalia County, WV. PSS321/155.**

Figure 3. ***Dicksonites pluckeneti*** **(Schlotheim) Sterzel. Upper Pennsylvanian Greene Formation (Dunkard Group). Between the Upper Rockport Limestone and the Gilmore coal bed. 0.2 km (0.1 mi) from Wood County line, Wirt County, WV. PSS 420/15.**

Plate 2.

Figure 1. ***Lescuropteris moorei*** **(Lesquereux) Schimpter. Harmon Creek surface mine along Route 18 between Burgettstown and Florence, about 15 miles west of Pittsburgh, PA. Roof shale of the Pittsburgh coal bed. Collected in 1947 by Anthony Miklausen, NE corner of Burgettstown 7.5' quadrangle (1:24000), Findley Township, SW Pennsylvania. Specimen in Carnegie Museum, Pittsburgh PA. 10 cm. scale bar.**

Plate 3.

Figure 1. ***Sphenopteris minutisecta*** **Fontaine and White. Roof shale of Upper Pennsylvanian Waynesburg coal bed, Washington Formation, Dunkard Group. Dippel and Dippel surface mine west of Osage, Monongalia County, WV. PSS-328/3.**

Figure 2. ***Annularia mucronata*** **Shenk. Upper Pennsylvanian Washington Formation (Dunkard Group). Roof shale of Waynesburg coal bed, 1.5 km (0.9 mi) south of Mount Morris, PA., Monongalia County, WV. WG-80.**

Figure 3. ***Alethopteris virginiana*** **Fontaine and White. Upper Pennsylvanian Washington Formation (Dunkard Group). Roof shale of Waynesburg coal bed, road cut along WV Route 7 at Cassville, WV. WG-8212. This is the Cassville Shale locality of Fontaine and White (1880).**

Figure 4. ***Polymorphopteris pseudobucklandii*** **(Andra) Wagner. Upper Pennsylvanian Monongahela Formation. Siderite nodules in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Plate 4.

Figure 1. ***Neuropteris ovata*** **Hoffmann (=*Neuropteris frimbiata* Lesquereux). Stalked frimbriate cyclopterid pinnules from below main frond division. Upper Pennsylvanian Dunkard Group. Cassville Shale (roof shale of Waynesburg coal bed) between Pursglove and WV Route 100, Osage 7.5' quadrangle, Monongalia County, WV.**

Figure 2. ***Callipteridium cf. gigas*** **Gutbier. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV. WGF-6/83.**

Figure 3. ***Sphenophyllum latifolium*** **Fontaine and White. Upper Pennsylvanian Washington Formation (Dunkard Group). Roof shale of Waynesburg coal bed in surface mine between Route 100 and Pursglove, WV, Osage 7.5' quadrangle (1:24000), Monongalia County, WV. PSS-321.**

Figure 4. ***Sphenophyllum*** **cf.** ***verticillatum*** **Schlotheim. Upper Pennsylvanian Glenshaw Formation (Conemaugh Group). Roof shale of Bakerstown coal bed in Amerikohl surface mine along Route 92, Nestorville 7.5' quadrangle, Barbour County, WV. WG-100/70.**

Plate 5

Figure 1. ***Nemejcopteris (Pecopteris) feminaeformis*** **(Schlotheim) Barthel. Upper Pennsylvanian Conemaugh Group (Casselman Formation). Roof shale of Little Pittsburgh coal bed, surface mine near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 2. ***Nemejcopteris (Pecopteris) feminaeformis*** **(Schlotheim) Barthel. Upper Pennsylvanian Conemaugh Group (Casselman Formation). Roof shale of Little Pittsburgh coal bed, surface mine near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV. Close up of Figure 1 above.**

Figure 3. ***Annularia sphenophylloides*** **(Zenker) Gutbier. Middle Pennsylvanian Kanawha Formation. Shales 1.5 m (5 ft) above Stockton coal bed, road cut along Quick-Sanderson road at railroad tunnel curve, near Puncheon Camp Branch, Kanawha County, WV. PSS-232.**

Figure 4. ***Odontopteris osmundaeformis*** **(Schlotheim) Zeiller. Upper Pennsylvanian Monongahela Formation. Red, fissile shale in third bench, road cut along I-77 near Kanawha-Putnam county line. Sayre School section, Sissonville 7.5' quadrangle, Kanawha County, WV. PSS-59**

Figure 5. ***Neuropteris ovata*** **Hoffmann. Upper Pennsylvanian Dunkard Group. Cassville Shale (roof of Waynesburg coal bed), classic Fontaine and White (1880) locality, near Cassville, Monongalia County, WV.**

Figure 6. ***Neuropteris ovata*** **Hoffmann. Close up of frimbriate cyclopterid pinnule, Cassville Shale (roof shale of Waynesburg coal bed), Anchor Coal Co. surface mine at Osage, Osage 7.5' quadrangle, Monongalia County, WV**

Plate 6

Figure 1. ***Macroneuropteris scheuchzeri*** **(Hoffmann) Cleal, Shute, and Zodrow. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV**

Figure 2. ***Macroneuropteris scheuchzeri*** **(Hoffmann) Cleal, Shute, and Zodrow. Upper Pennsylvanian Monongahela Formation. Example of polymorphism common in Upper Pennsylvanian examples of this species. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 3. ***Macroneuropteris scheuchzeri*** **(Hoffmann) Cleal, Shute, and Zodrow. Upper Pennsylvanian Monongahela Formation. Example of polymorphism common in Upper Pennsylvanian examples of this species. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 4. ***Macroneuropteris scheuchzeri*** **(Hoffmann) Cleal, Shute, and Zodrow. Upper Pennsylvanian Monongahela Formation. Example of polymorphism common in Upper Pennsylvanian examples of this species. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 5. ***Macroneuropteris scheuchzeri*** **(Hoffmann) Cleal, Shute, and Zodrow. Middle Pennsylvanian**

Kanawha Formation. Roof shale 1.5 m (5 ft) above Stockton coal bed, road cut along Quick-Sanderson road at railroad tunnel curve, near Puncheon Camp Branch, Kanawha County, WV. PSS-232.

Figure 6. ***Macroneuropteris scheuchzeri*** **(Hoffmann) Cleal, Shute, and Zodrow. Close up photograph to show diagnostic hairs. Middle Pennsylvanian Kanawha Formation. Roof shale 1.5 m (5 ft) above Stockton coal bed, road cut along Quick-Sanderson road at railroad tunnel curve, near Puncheon Camp Branch, Kanawha County, WV. PSS-232.**

Figure 7. ***Sphenophyllum oblongifolium*** **(Germer and Kaulfuss) Unger. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Plate 7.

Figure 1. ***Lescuropteris moorei*** **(Lesquereux) Schimper. Holotype. Originally figured as Plate XIX, Figures 1 and 1a in Lesquereux (1858). Upper Pennsylvanian Monongahela Formation. Roof shale, Pittsburgh coal bed, Irwin Station, Allegheny County, PA. Examples of *L. moorei* also have been collected from the roof shale of the Pittsburgh coal bed at St. Clairsville, Belmont County, OH, near Pittsburgh, PA, and Morgantown, WV. Specimen of *Emplectopteris triangularis* Halle from China in author's (WHG) collection is seemly identical to *Lescuropteris moorei.* HBM 5989 (Harvard Botanical Museum).**

Figure 2. ***Lescuropteris moorei*** **(Lesquereux) Schimper. Upper Pennsylvanian Monongahela Formation. Roof shale, Pittsburgh coal bed, Irwin Station, Allegheny County. Same specimen as Pl. XXV, Figure 1.**

Figure 3. ***Sphenophyllum angustifolium*** **(Germer) Göppert. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Plate 8

Figure 1. ***Danaeides emersonii*** **Lesquereux. Fertile and sterile foliage on same specimen. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 2. ***Danaeides emersonii*** **Lesquereux. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 3. ***Danaeides emersonii*** **Lesquereux. Upper Pennsylvanian Monongahela Formation. Notice the short terminal pinnules. Nodular siderite layer in roof shales of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker** County, WV.

Plate 9

Figure 1. ***Danaeides emersonii*** **Lesquereux. Upper Pennsylvanian Monongahela Formation. Sterile foliage. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 2. ***Alethopteris virginiana*** **Fontaine and White. (=*A. leonensis* Wagner) Upper Pennsylvanian Dunkard Group. Upper clay parting in Waynesburg coal bed, classic Fontaine and White (1880) locality, near Cassville, Monongalia County, WV.**

Figure 3. ***Alethopteris virginiana*** **Fontaine and White. (=*A. leonensis* Wagner) Upper Pennsylvanian Dunkard Group. Upper clay parting in Waynesburg coal bed, classic Fontaine and White (1880) locality, near Cassville, Monongalia County, WV.**

Figure 4. ***Dolerotheca pennsylvanica*** **Dawson ex White. Dorsal (top) view. This is a campanulate synangium usually associated with *Alethopteris zeillerii* Ragot. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV**

Figure 5. ***Dolerotheca pennsylvanica*** **Dawson ex White. Ventral (lower) view. This is a campanulate synangium usually associated with *Alethopteris zeillerii* Ragot. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV**

Figure 6. ***Neuropteris odontopteroides*** **Fontaine adn White. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shales of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Plate 10

Figure 1. ***Polymorphopteris subelegans*** **(Potonié) Wagner. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shales of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 2. ***Pecopteris platynervis*** **Fontaine and White. Upper Pennsylvanian Conemaugh Group (Glenshaw Formation). Roof shales of Bakerstown coal bed in surface mine, along Big Cove Run, near Moatsville, Nestorville 7.5' quadrangle, Barbour County, WV.**

Figure 3. ***Lobatopteris vestita*** **(D. White) Wagner. Middle Pennsylvanian Allegheny Formation. Roof shale of Middle Kittanning coal bed, road cut on Gladesville Road, 1 km (0.6 mi) west of WV Route 73, Monongalia County, WV. PSS-144.**

Figure 4. ***Polymorphopteris pseudobucklandii*** **(Andrae) Wagner. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV**

Plate 11

Figure 1. ***Rhachiphyllum shenkii*** **(Heyer) Kerp. Upper Pennsylvanian Dunkard Group. Shale 1.2 m (4 ft) above Lower Washington Limestone, 1 km (0.6 mi) east of Waynesburg, Greene County, PA. Originally figured (Pl. I, A) as *Callipteris* cf. *conferta* by Gillespie *et al*. (1975).**

Figure 2. ***Rhachiphyllum shenkii*** **(Heyer) Kerp. Juvenile form. Upper Pennsylvanian Dunkard Group. Shale 1.2 m (4 ft) above Lower Washington Limestone, 1 km (0.6 mi) east of Waynesburg, Greene County, PA. .**

Figure 3. ***Walchia piniformis*** **(Schlotheim) Florin. Upper Pennsylvanian Conemaugh Group (Glenshaw Formation). Shale between Upper Freeport and Brush Creek coal beds (Mahoning/Mason interval) on hilltop southwest of bridge 2.5 km (1.6 mi) east of County Route 17 off WV Route 26, north of Albright, Preston County, WV. PSS-353. Same specimen as Pl. XXXIV, Figure 2.**

Figure 4. ***Walchia piniformis*** **(Schlotheim) Florin. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV. Same specimen as Pl. XXXIV, Figure 1.**

Figure 5. ***Autunia conferta*** **(Sternberg) Kerp. Upper Pennsylvanian Dunkard Group. Red shale in road cut below sandstone, 0.8 km (0.5 mi) west of Liberty Post Office, along WV Route 34, Putnam County, WV. PSS-358.**

Figure 6. ***Autunia conferta*** **(Sternberg) Kerp. Upper Pennsylvanian Dunkard Group. Top of Lower Washington Limestone, 1 km (0.6 mi) east of Waynesburg, Greene County, PA.**

Figure 7. ***Autunia conferta*** **(Sternberg) Kerp. Upper Pennsylvanian Dunkard Group. Red shale in road cut below sandstone, 0.8 km (0.5 mi) west of Liberty Post Office,** along WV Route 34, **Putnam County, WV. PSS-358.**

Plate 12

Figure 1. ***Sphenophyllum thoni*** **Mahr. Upper Pennsylvanian Dunkard Group. Cassville Shale (roof shales of Waynesburg coal bed) in abandoned surface mine between Pursglove and WV Route 100, Monongalia County, WV. PSS-321. OUPH-12,637.**

Figure 2. ***Danaeides emersonii*** **Lesquereux. Fertile pinnules. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV. Same specimen as Pl. XXVI, Figure 2. OUPH-12,635. Scale bar 0.5 cm.**

Figure 3. ***Sigillaria brardii*** **Brongniart. Upper Pennsylvanian Dunkard Group. Shale bed in Lower Washington Limestone at limestone quarry, near Washington, Washington County, PA. Loaned for display in panorama in Carnegie Museum, Pittsburgh PA. OUPH-12,638.**

Figure 4. ***Odontopteris brardii*** **Brongniart. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shales of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV. OUPH-12,639.**

Figure 5. ***Sphenophyllum oblongifolium*** **(Germer and Kaulfuss) Unger. Upper Pennsylvanian Dunkard Group. Cassville Shale (roof shales of Waynesburg coal bed) in surface mine 3.3 km (2 mi) north of Osage, 550 m (1,805 ft) west of Bethel Church, and 120 m (394 ft) north of secondary road, Monongalia County, WV. PSS-315. OUPH-12,640.**

Figure 6. ***Odontopteris brardii*** **Brongniart. Aphlebia-like basal leaflets sometimes confused with pinnatified basal cyclopterid-like leaflets of** ***Neuropteris ovata*** **Hoffmann. Compare with Plate XXI, Figure 3. Note broad attachment. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shales of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV. OUPH-12,641.**

Figure 7. ***Mariopteris nervosa*** **(Brongniart) Zeiller. Middle Pennsylvanian Allegheny Formation. Roof shale of Middle Kittanning coal bed, road cut on Gladesville Road, 1 km (0.6 mi) west of WV Route 73, Monongalia County, WV. PSS-144. OUPH-12,677.**

Plate 13

Figure 1. ***Walchia piniformis*** **(Schlotheim) Florin. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shales of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 2. ***Walchia piniformis*** **(Schlotheim) Florin. Upper Pennsylvanian Conemaugh Group (Glenshaw Formation). Shale between Upper Freeport and Brush Creek coal beds (Mahoning/Mason interval) on hilltop southwest of bridge 2.5 km (1.6 mi) east of County Route 17 off WV Route 26, north of Albright, Preston County, WV. PSS-353.**

Figure 3. ***Pseudomariopteris cordato-ovata*** **(Weiss) Gillespie** ***et al*****. Upper Pennsylvanian Conemaugh Group (Glenshaw Formation). Roof shale of Bakerstown coal bed in surface mine, along Big Cove Run, near Moatsville, Nestorville 7.5' quadrangle, Barbour County, WV.**

Figure 4. ***Alloiopteris angustissima*** **(Sternberg) H. Potonié. Lower Pennsylvanian Kanawha Formation. Roof shale of Douglas coal bed in surface mile between Crane Creek and Mile Branch, Iaeger 7.5' quadrangle, McDowell County, WV.**

Figure 5. ***Pecopteris monyi*** **Zeiller. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV**

Figure 6. ***Pecopteris monyi*** **Zeiller. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV. Same specimen as Pl. XXXIV, Figure 5.**

Figure 7. ***Alethopteris bohemica*** **Franke. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV.**

Figure 8. ***Alethopteris bohemica*** **Franke. Upper Pennsylvanian Monongahela Formation. Nodular siderite layer in roof shale of Pittsburgh coal bed, near Fairfax Stone State Park, Davis 7.5' quadrangle, Tucker County, WV. Same specimen as Pl. XXXIV.**

Plate 1

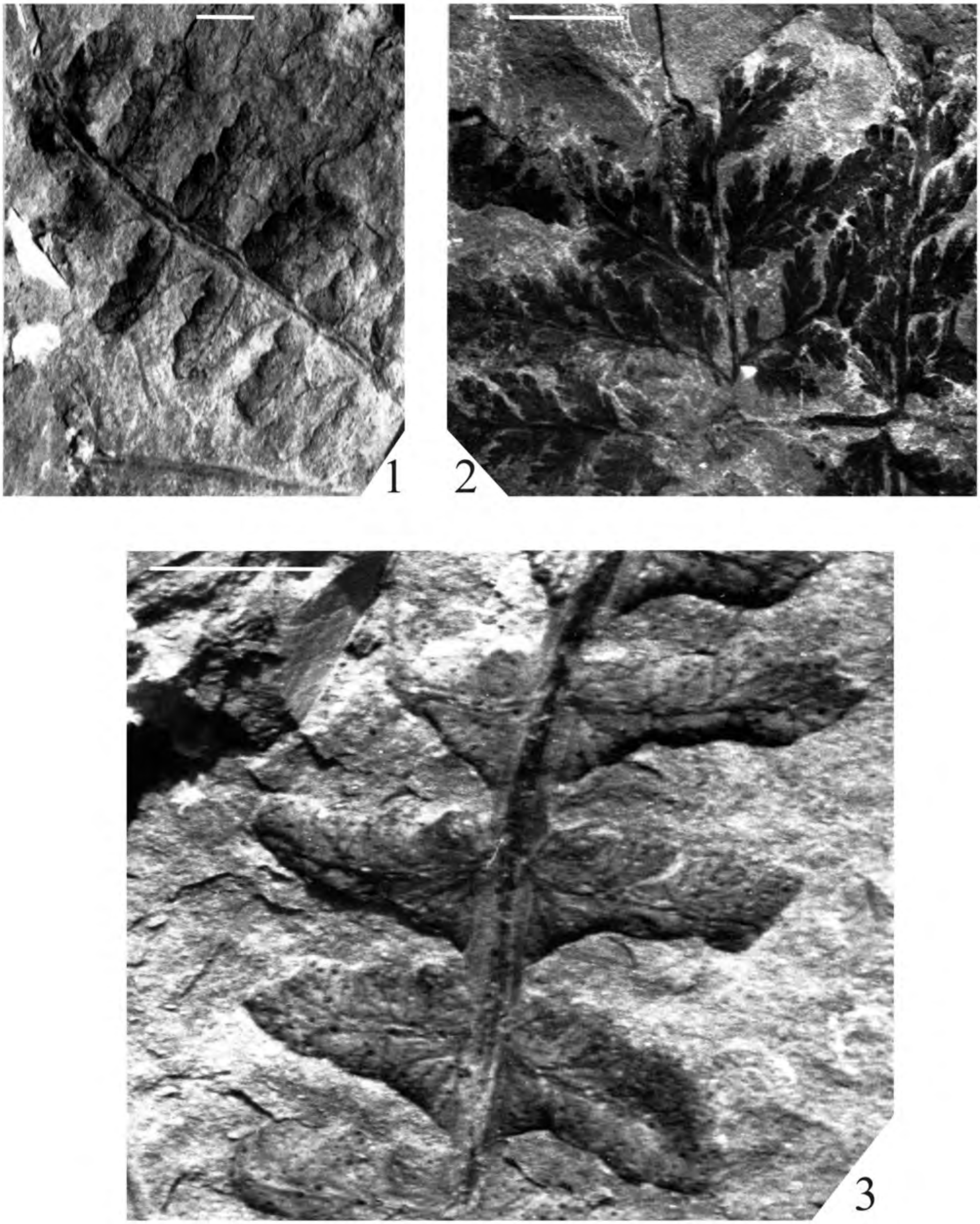
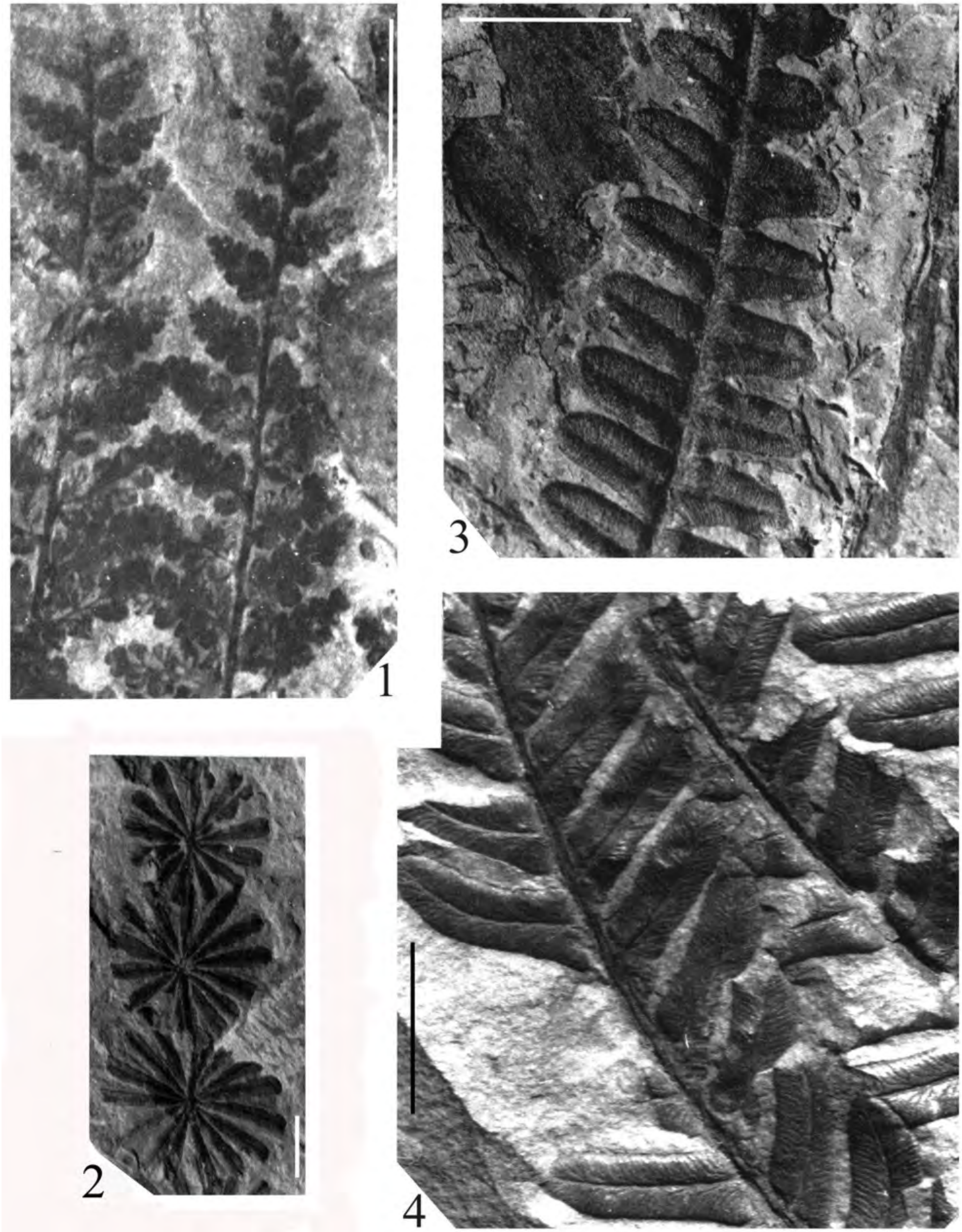


Plate 2

Plate 3



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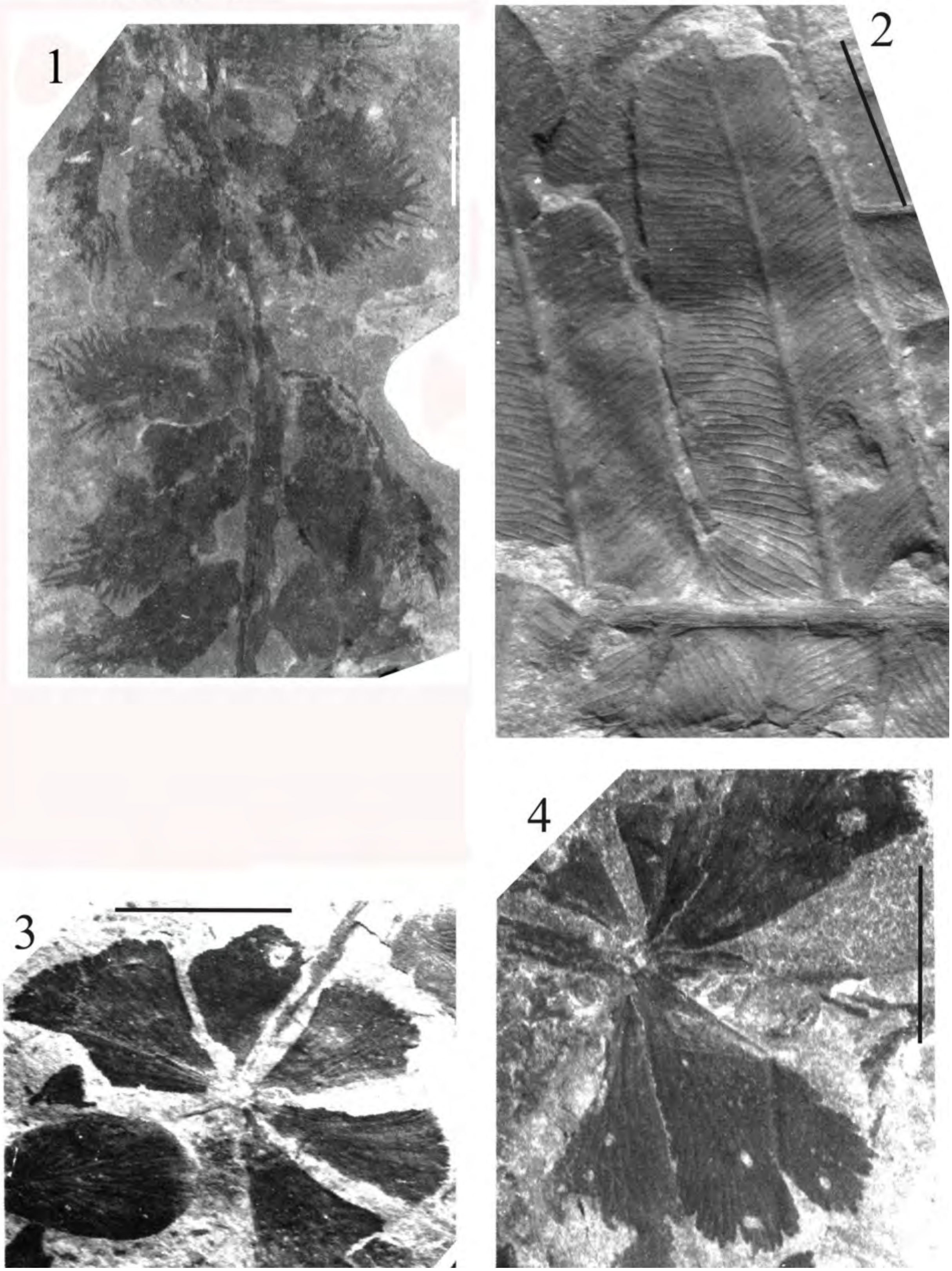

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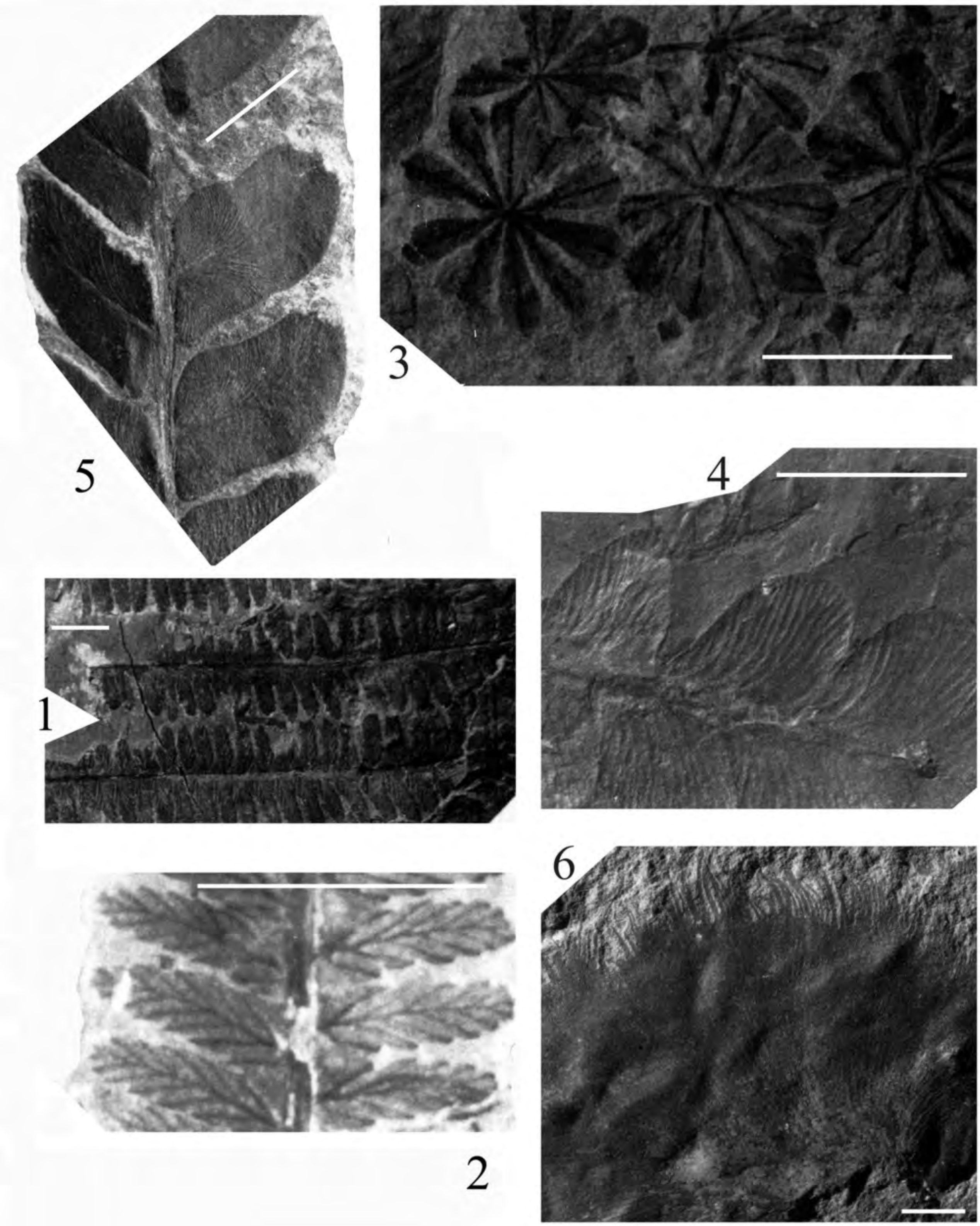

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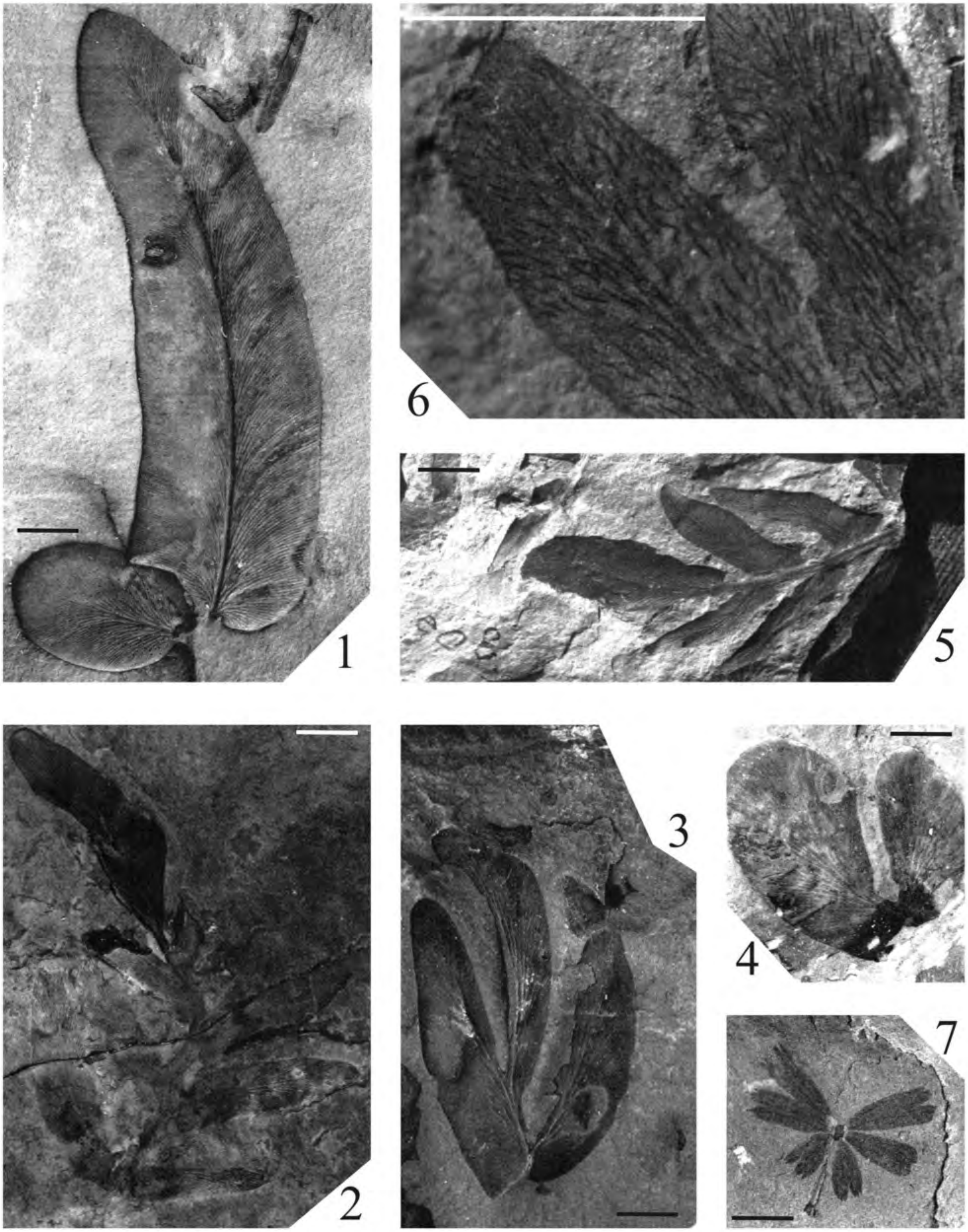

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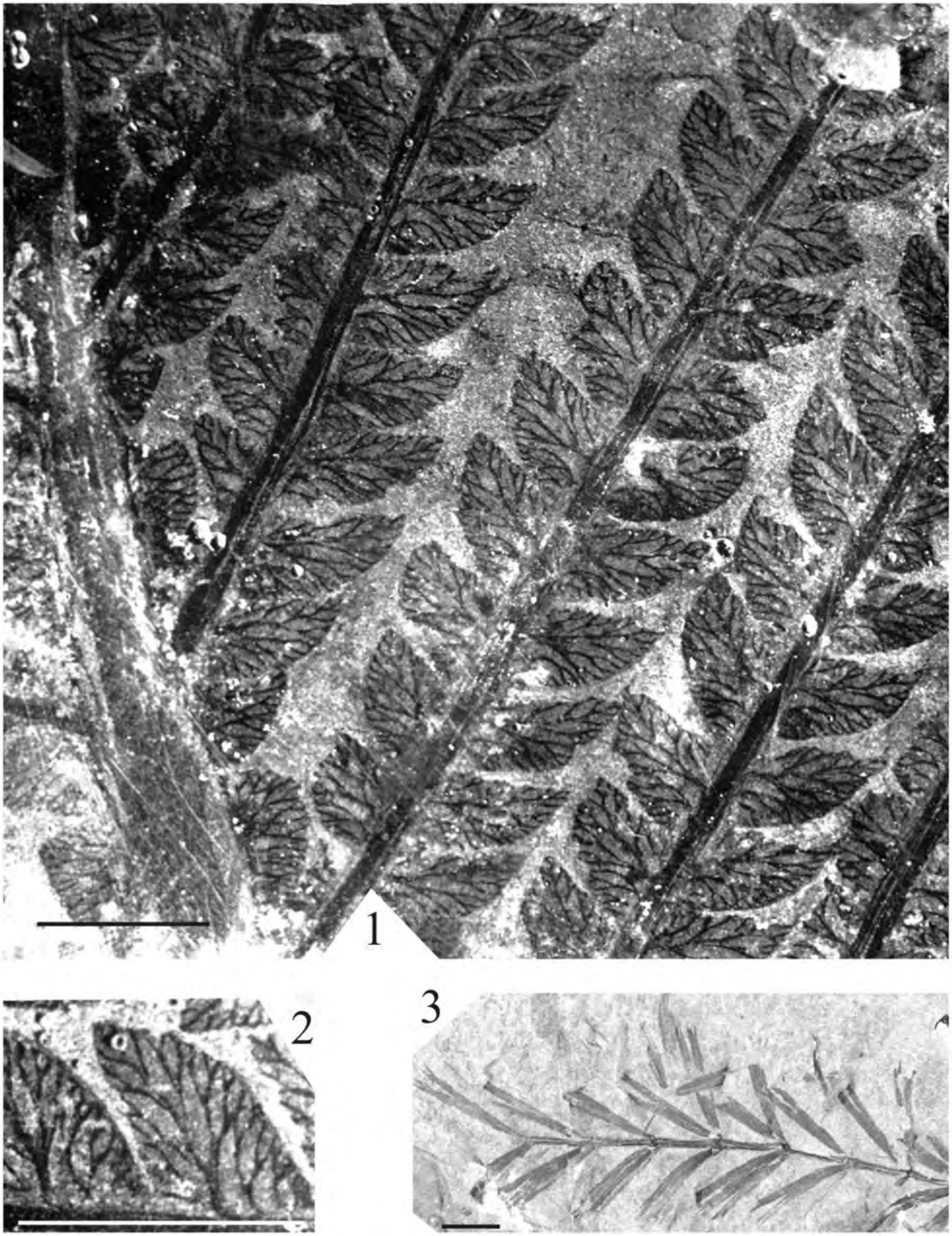

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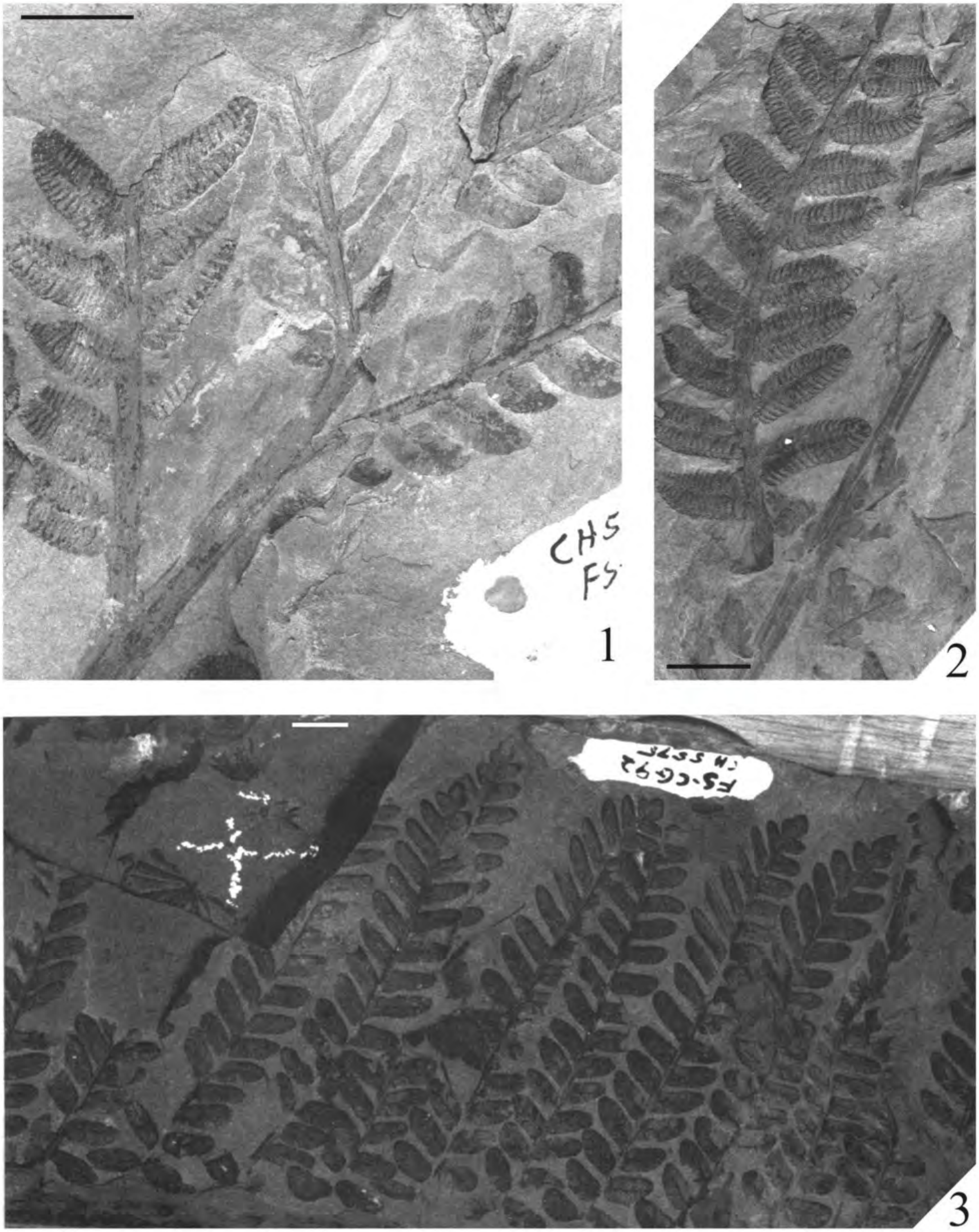

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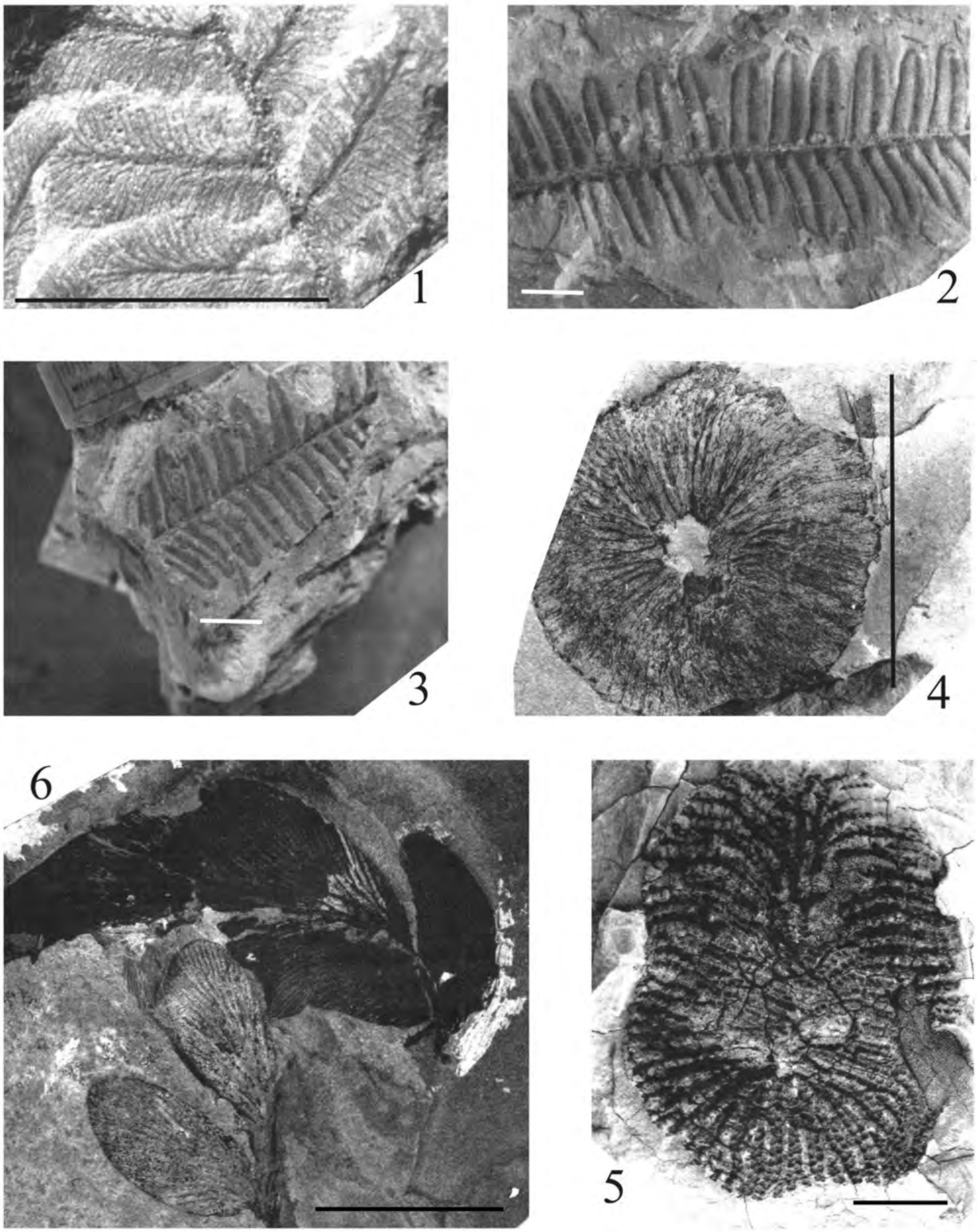

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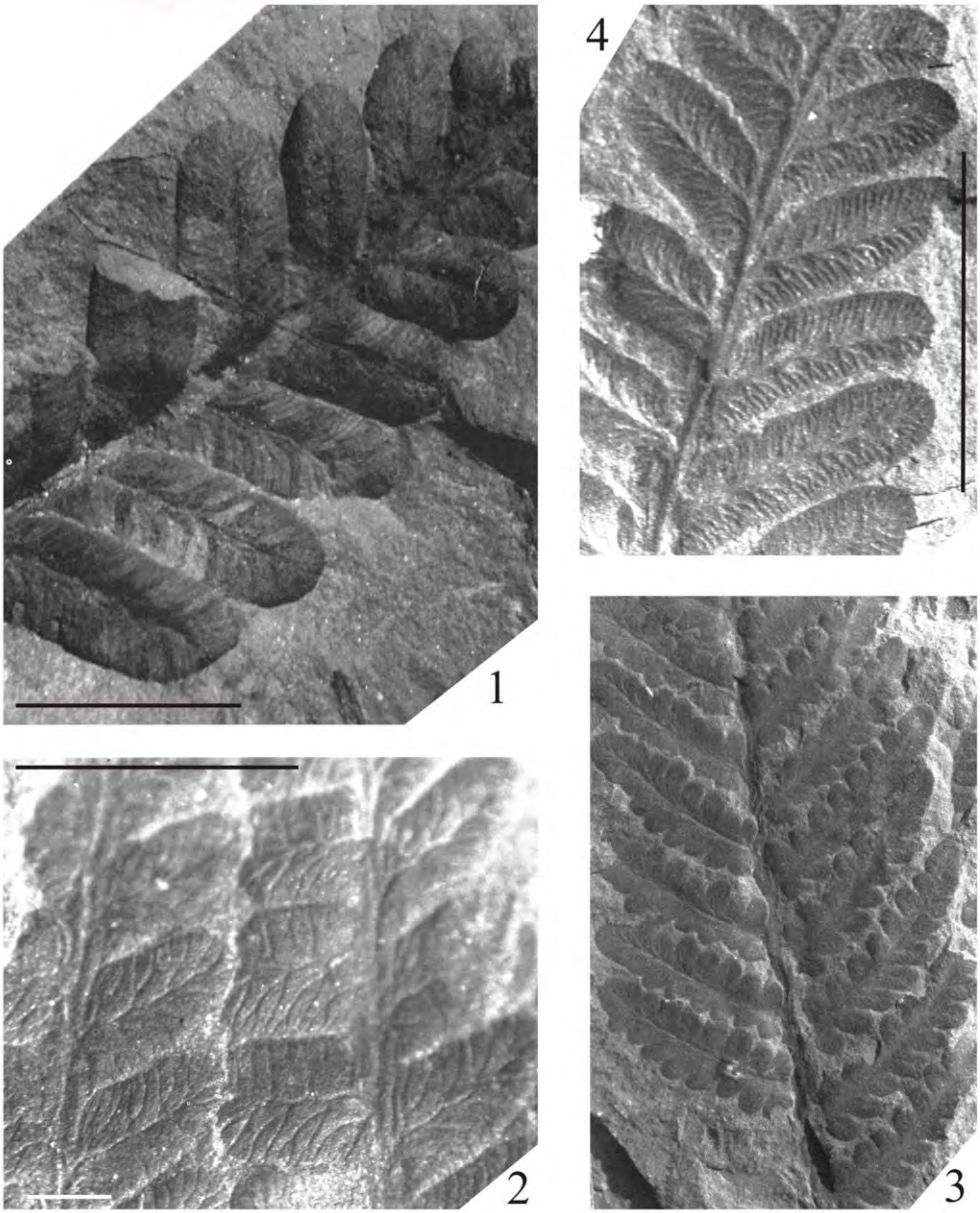

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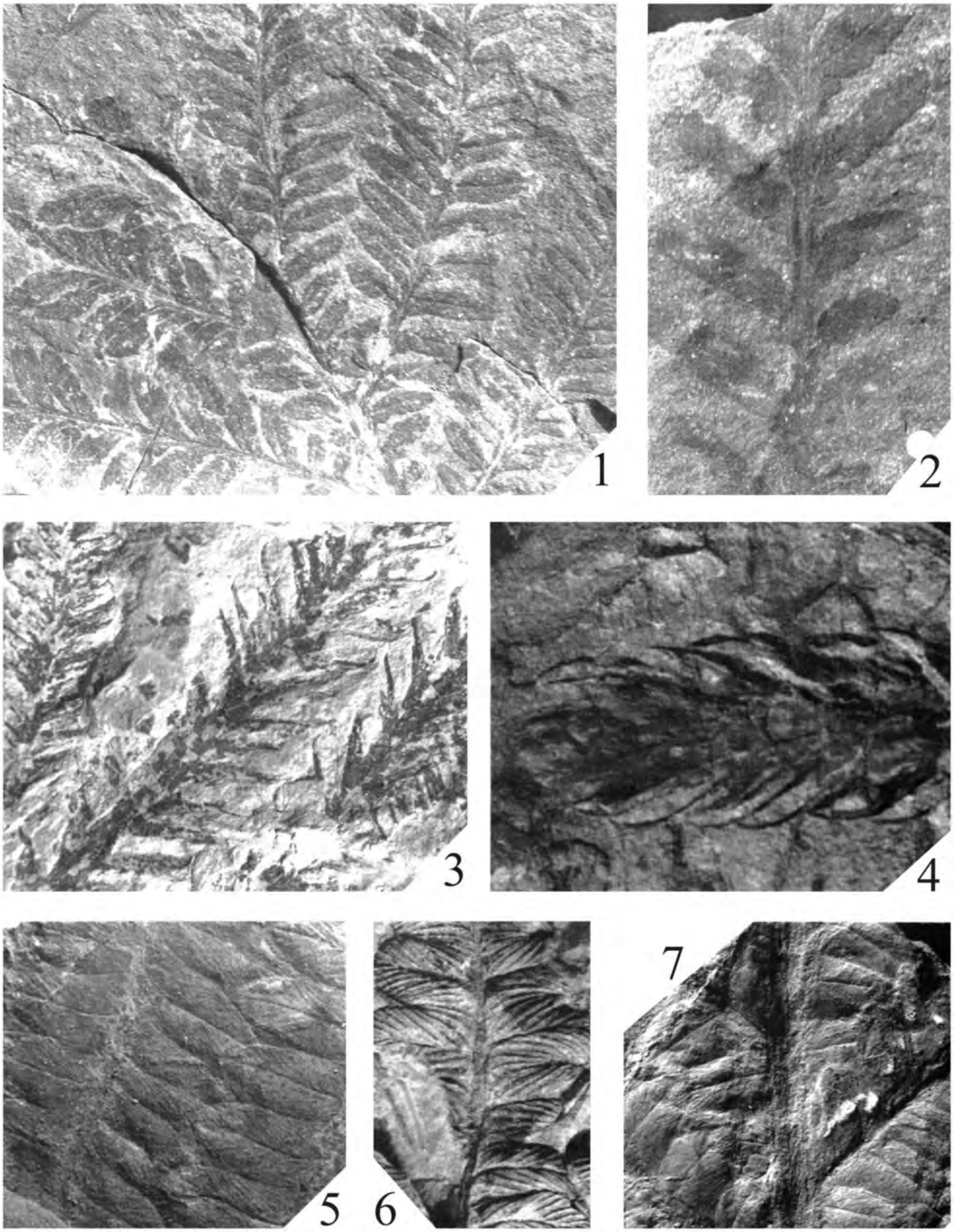

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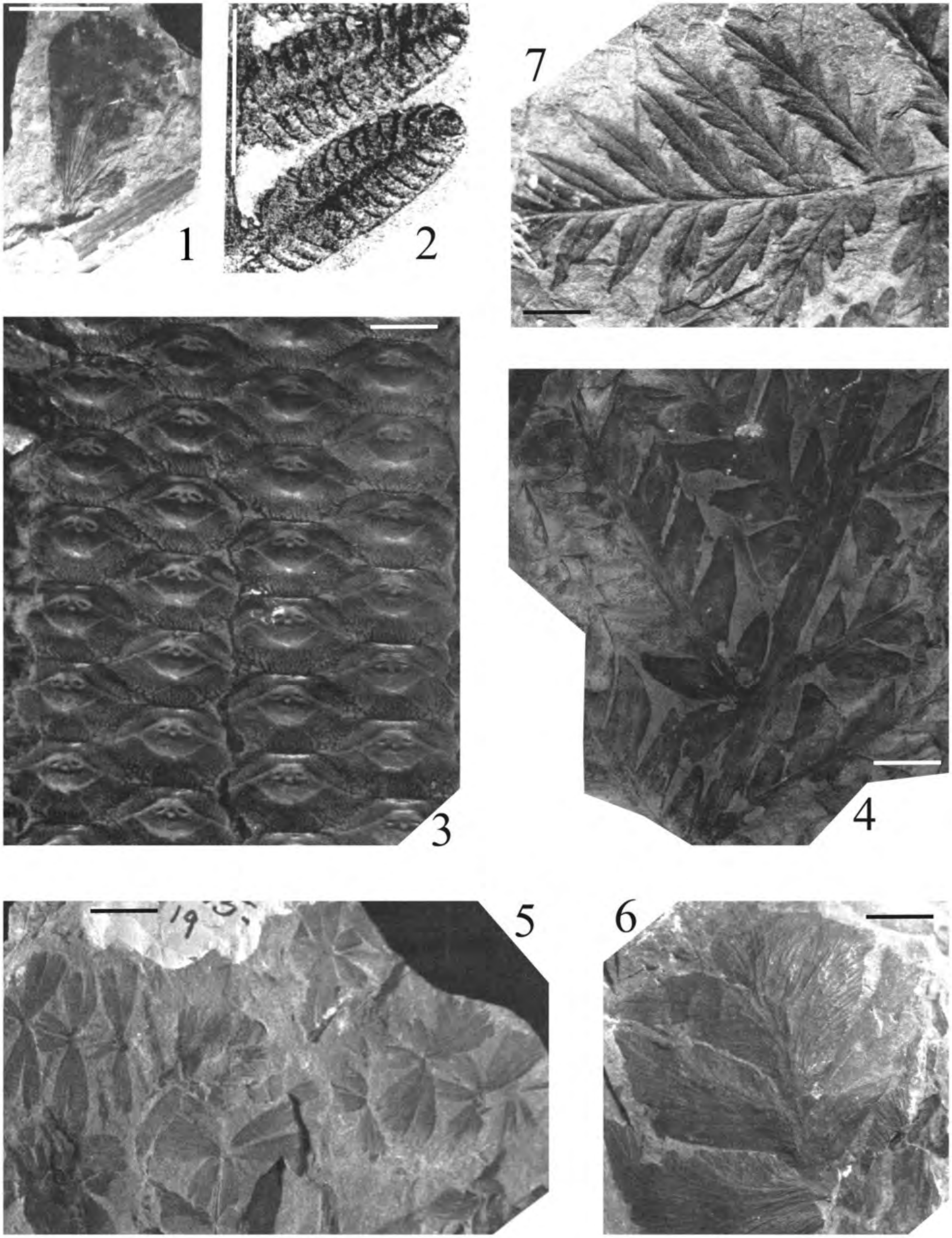

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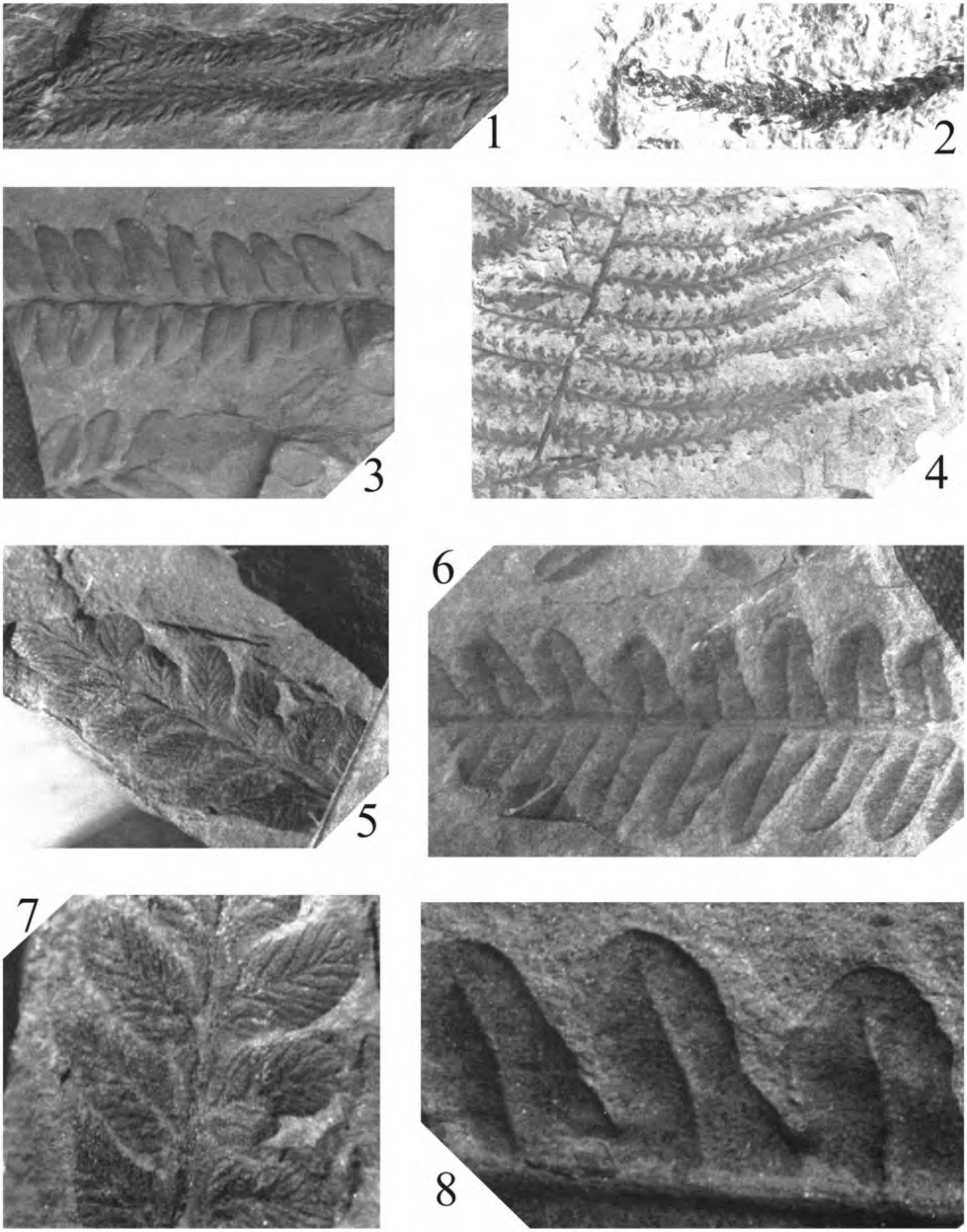

CALLIPTERIDS OF THE DUNKARD GROUP OF THE APPALACHIAN BASIN: THEIR IDENTITY AND PALEOENVIRONMENTAL SIGNIFICANCE

William A. DiMichele, Bascombe M. Blake, Jr., Blaine Cecil,
Nick Fedorko, Hans Kerp, and Viktoras Skema

INTRODUCTION

The Dunkard Group, which contains the youngest sedimentary rocks in the Appalachian Basin, is a >340 m (1,116 ft)-thick erosional remnant sited hundreds of kilometers from its nearest correlative. With the exception of rare *Lingula* in the Washington coal zone (Cross and Schemel, 1956), the Dunkard is considered non-marine, therefore lacking definitive index fossils. The geological age of these strata has been a matter of intense debate since the publication of Fontaine and White's (1880) monograph on the Dunkard Group's macrofloras. The age of the Dunkard has turned strongly on the presence of callipterids, a wholly extinct group of peltaspermous plants, in the lower third of the Group, and especially in the 10-20 m (33-66 ft) of strata above the Washington coal in the Washington Formation. Aside from these plants, which globally became abundant and widely distributed beginning near the Pennsylvanian-Permian boundary and through much of the Permian (Kerp, 1988), the Dunkard flora is overwhelmingly characterized by typically Pennsylvanian plant species (Fontaine and White, 1880; Clendening, 1972, 1974, 1975; Clendening and Gillespie, 1972; Gillespie et al., 1975; Cross et al., 1996; Blake et al., 2002; Blake and Gillespie, 2011a, b). Other groups of organisms have been brought to bear on the question of the age of the Dunkard (see summaries in Barlow and Burkhammer, 1975), but none seem to have been emphasized like the callipterids to point both paleobotanists and others toward a Permian interpretation.

Many paleobotanists have collected plant fossils from Dunkard strata. To our knowledge, however, the only illustrations of the callipterids are the original engravings of Fontaine and White (1880, Plate XI, Figures 1-4) and photographs in Darrah (1975, Figures 1-3) and Gillespie et al. (1975, Plate I A and Plate VIII A; refigured in Blake et al., 2002, Plate XXXV, Figures 1-2, 6) and Gillespie and Pfefferkorn (1979, Plate 3, figures 6-8; refigured in Blake et al., 2002, Plate XXXV, Figures 5, 7). Herein we describe and illustrate callipterids from two, previously unillustrated collections. These were made by David White in the summer of 1902, during which time he visited I.C. White in Morgantown, West Virginia, and collected regionally in the Dunkard (though he never published an illustrated flora, see White, 1904, 1936), and from the collections of Aureal T. Cross, now housed at the Field Museum of Natural History, who also worked extensively in the Dunkard in the late 1940s and early 1950s (Cross et al., 1950; Cross, 1954, 1958). Both White and Cross expanded greatly the known stratigraphic range of callipterid occurrence, White high into the Greene Formation.

Four species of callipterid characterize known Dunkard collections. Two are attributable to the genus *Autunia*, a seed plant belonging to the Peltaspermales (Kerp, 1988), a likely descendant of the pteridosperms so common to Pennsylvanian wetlands. These include *A.*

DiMichele, W. A., Blake, Jr., B. M., Cecil, Blaine, Fedorko, Nick, Kerp, Hans, and Skema, Viktoras., 2011, Callipterids of the Dunkard Group of the Appalachian basin: Their identity and paleoenvironmental significance, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 144-167.

conferta (a larger and a smaller form) at three of the localities considered here and *A. naumannii*, from one locality. In addition specimens attributable to *Lodevia oxydata* (Goeppert) Kerp and Haubold (*Sphenopteris coreacea* Fontaine and White, identified by Darrah, 1969 as *Callipteris lyratifolia*) occur at one locality. Gillespie et al. (1975) illustrated (Plate VIII a) a specimen that appears much like *Rhachiphyllum schenkii* Kerp.

GEOLOGICAL CONTEXT

The Dunkard Group has been divided traditionally into the Washington Formation, encompassing those rocks between the top of the Waynesburg coal and the top of the Upper Washington Limestone, and the Greene Formation, for the rock sequence above the Upper Washington Limestone (Berryhill, 1963). The U.S. Geological Survey recognizes three formations, dividing the rocks below the Greene into a lower Waynesburg Fm, defined as encompassing those rocks between the bottom of the Waynesburg coal and the bottom of the Washington coal, and an upper Washington Formation, encompassing the Washington coal to the top of the Upper Washington Limestone. This nomenclature was based on mapping done in Washington County, Pennsylvania (Berryhill and Swanson, 1962; Berryhill et al., 1971). The Greene Formation is thicker than the Waynesburg and Washington Formations combined. Dunkard rocks crop out in Pennsylvania, West Virginia, Ohio and Maryland and reach their maximum thickness in SW Pennsylvania, in Greene County, and adjacent parts of Wetzel County, West Virginia where the Dunkard exceeds 340 m (1,116 ft). In the northern part of the Dunkard Basin, in the adjacent areas of northern West Virginia, NE Ohio, and SW Pennsylvania, the Dunkard consists of cyclic successions of coal, limestone, shale and sandstone (Beerbower, 1961). The clastic strata are primarily buff and gray in color, but red beds are present higher in the section, beginning near the top of the Washington Formation, and, and occurring in conjunction with decreased abundances of coal. To the S and SW the Dunkard section becomes increasingly red and is mostly a sequence of stacked paleosols interbedded with channel-form sandstone bodies (Berryhill, 1963; Martin, 1998; Fedorko and Skema, 2011a, b; Cecil et al., 2011a, b).

The Little Washington coal horizon marks the base of the Washington coal complex (Skema et al., 2011), which straddles the Waynesburg-Washington Formation contact. This coal complex is one of the most persistent units in the section and the Little Washington coal at its base, which has a distinctive underclay paleosol (Hennen, 1911), can be traced through much of the basin (Fedorko and Skema, 2011a, b; Skema et al., 2011). The Washington coal complex consists of two to four separate coals separated by shale and or sandstone. In the northwestern part of the basin (Belmont County, Ohio and the western edge of Marshall County, WV) the coal complex consists of two benches separated by a gray shale wedge up to 2 m (7 ft) or more thick. In this clastic parting, brackish water fauna have been found, including linguloid brachiopods and myalinid pelecypods (Cross and Schemel, 1956; Berryhill, 1963). This brackish-water fauna is the highest occurrence of any evidence of marine influence in the Appalachian Basin and may correlate with a global marine highstand pulse identified by Davydov et al. (2010) just below the Pennsylvanian-Permian boundary.

Throughout the Dunkard, if plant fossils are encountered they are almost always typical Pennsylvanian wetland flora, dominated by the tree fern *Pecopteris*, but including Middle Pennsylvanian holdovers, such as the pteridosperms *Neuropteris ovata* and *Macroneuropteris scheuchzeri*, in addition to a wide variety of other plants (Gillespie et al., 1975; Blake et al.,

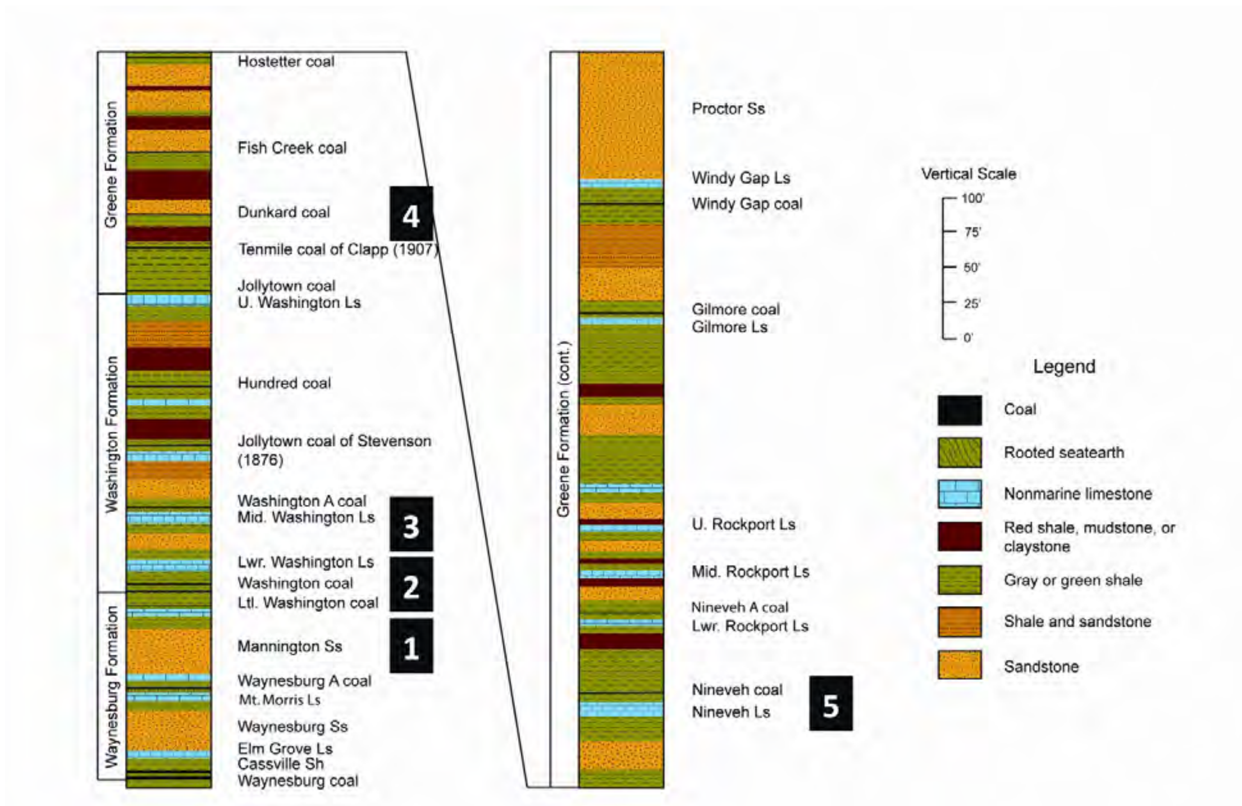

Figure 1. Dunkard Stratigraphy, after Fedorko (this guidebook). Numbers indicate horizons from which callipterids have been positively identified: 1— Between the Waynesburg A and Washington coals; 2— Within or above the Washington coal zone; 3—Within or proximate to the Lower and Middle Washington limestones; 4— Proximate to/above the Dunkard coal; 5—Above the Nineveh coal.

2002). This is true regardless of the facies sampled (Blake and Gillespie, this guidebook). *Pecopteris* foliage has been found in all lithologies sampled from the bottom of sands deposited in channels, to buff and red mudstones presumably deposited in alluvial floodplain or delta plain (alluvial plain) environments, in association with coals or not, and rarely even within limestones.

Dunkard callipterids, on the other hand, have been reported to occur mainly in shales within the Washington Formation (in the more traditional two formation scheme, as used by the West Virginia Geological and Economic Survey), beginning at the approximate horizon of the Washington coal complex through the Lower and perhaps Middle Washington Limestone. These fossils occur in a variety of lithologies, from dark, micaceous finely laminated shales, to buff and red mudstones with poor fissility. At one location, on Robert's Ridge Road near Moundsville, West Virginia, specimens were found in a shale parting within the uppermost portion of a limestone sequence, either the Lower or Middle Washington LS (not differentiable at the site). The stratigraphy of this part of the Dunkard section is complex and the strata are highly variable laterally. The callipterids appear to occur at a number of different stratigraphic horizons within the above mentioned interval. Darrah (1975, in the captions to his Figures 1 and 3 on page 95) provides the only report of callipterids from below the level of the Washington coal. He attributes the collections to James Barlow of the West Virginia Geological Survey, purportedly collected from between the Waynesburg A and Washington

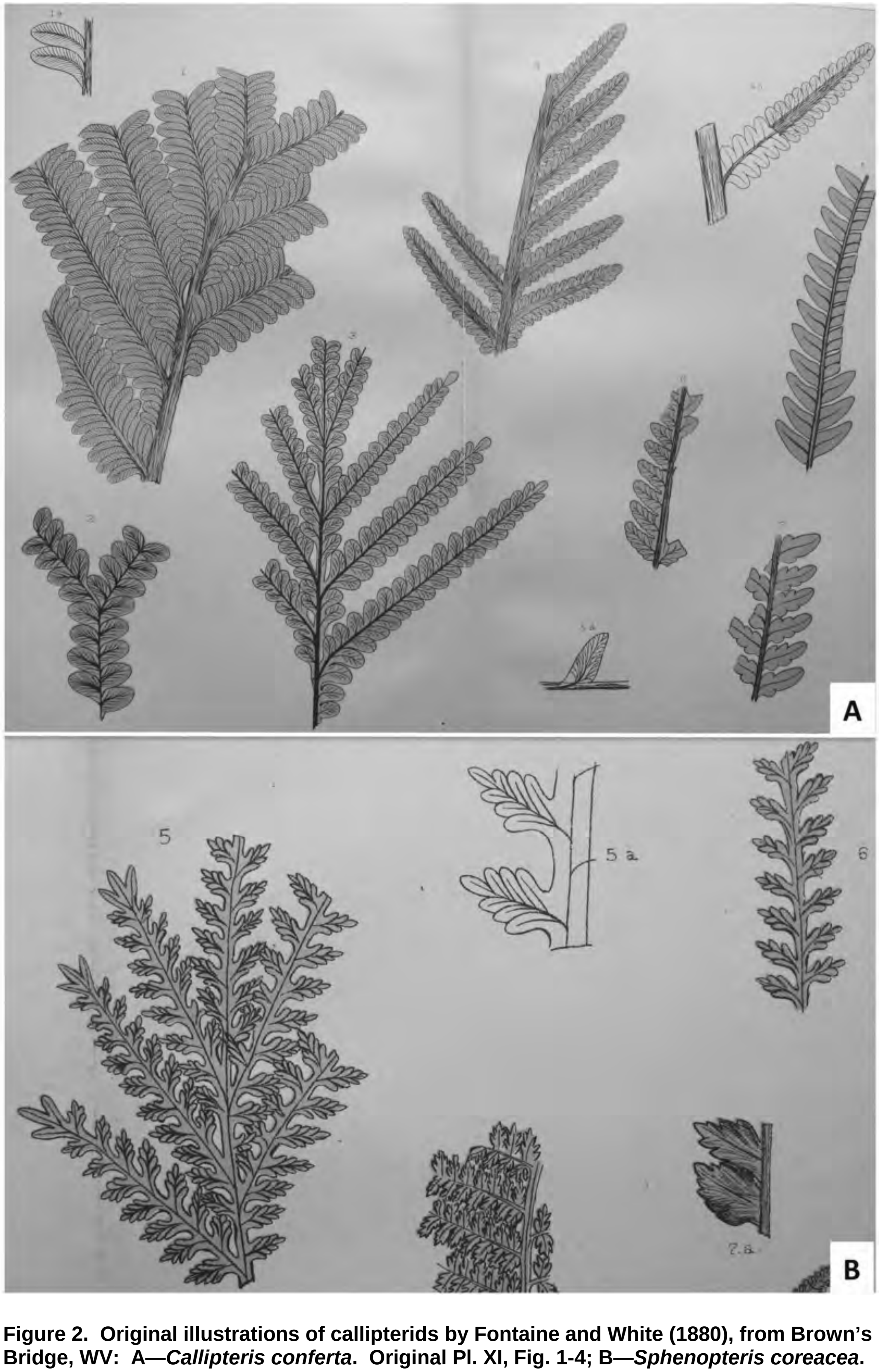

Figure 2. Original illustrations of callipterids by Fontaine and White (1880), from Brown's Bridge, WV: A—*Callipteris conferta*. Original Pl. XI, Fig. 1-4; B—*Sphenopteris coreacea*. Original Pl. V, Fig. 5.

coals near Williamstown, in Wood County, West Virginia. However, coals are very poorly developed in the Dunkard Group in Wood County, and Darrah provides no detailed information on the location of the collecting locality or on the ultimate disposition of the samples, so, at present, we treat this report with reservation. In no instances have callipterids been documented in the immediate roof shales of any coal bed or coaly facies.

There are reports of other kinds of plants uncommon in Pennsylvanian floras, but common in the Permian, such as *Taeniopteris*, *Plagiozamites*, and conifers, from various horizons in the Dunkard Group, including the Cassville Shale, at the base of the Dunkard Group (Fontaine and White, 1880; D. White, 1904, 1936; Darrah, 1975). Such occurrences led some to conclude that the entire Dunkard Group is of Permian age, including David White at the end of his professional career (D. White, 1936). However, callipterids have not been reported from such a low stratigraphic position.

STRATIGRAPHIC AND LITHOLOGICAL OCCURRENCES OF CALLIPTERIDS

We report and illustrate callipterids from three sites investigated by David White in 1902: "Browns Bridge" (also known as Browns Mills or Worley), West Virginia, "Railroad Cut West of Littleton", West Virginia, and "Pleasant Hill Gap", Pennsylvania, and one collected by Aureal Cross, on August 27, 1948: "Winklers Mill", Ohio. We also note and comment on other illustrated callipterids from the Dunkard, including those of Darrah (1975), Gillespie et al. (1975), and Gillespie and Pfefferkorn (1979). The stratigraphic position of these callipterid beds, if securely known, is shown in Figure 1.

Brown's Bridge

Brown's Bridge was the original callipterid discovery site of Fontaine and White (1880) and their illustrated specimens come from that site. To our knowledge, this is the only site from which callipterids have been collected more than once (four times in fact). The location lies on the south side of a bend in Dunkard Creek (NW quadrant of the Blacksville 15' Quadrangle map/NE quadrant of the Blacksville 7.5' Quadrangle map, opposite the town of Worley) along present-day West Virginia Route 7, probably close to the grade of a former railroad at that location.

Fontaine and White's specimens are lost, unfortunately, with a single exception, discussed below. In addition, the bed from which they were collected has never been accurately reidentified or described in the literature or in extant field notes of collectors, so neither the exact stratigraphic nor paleoenvironmental context is known. Their original illustrations are reproduced here as Figure 2. The exception to the lost callipterid specimens is a single specimen US National Museum of Natural History collections from the Brown's Bridge locality, identified as *Callipteris conferta* by Leo Lesquereux (USNM Specimen Number 27141 – Figure 3). This specimen presumably was supplied to Lesquereux for identification by Fontaine and White or perhaps by R.D. Lacoe (see note in Darrah, 1969, p. 153). It is possible that this specimen was part of a suite of typical Dunkard specimens donated by Fontaine and White to the well known private collector R.D. Lacoe, which later came to the U.S. National Museum when Lacoe donated his entire collection to that institution in the late 1800s. Fontaine and White (1880, p. 54) describe their plants as coming from a "calcareous iron ore" that occurs in the roof of the Washington coal (page 54). The specimen identified by Lesquereux (USNM

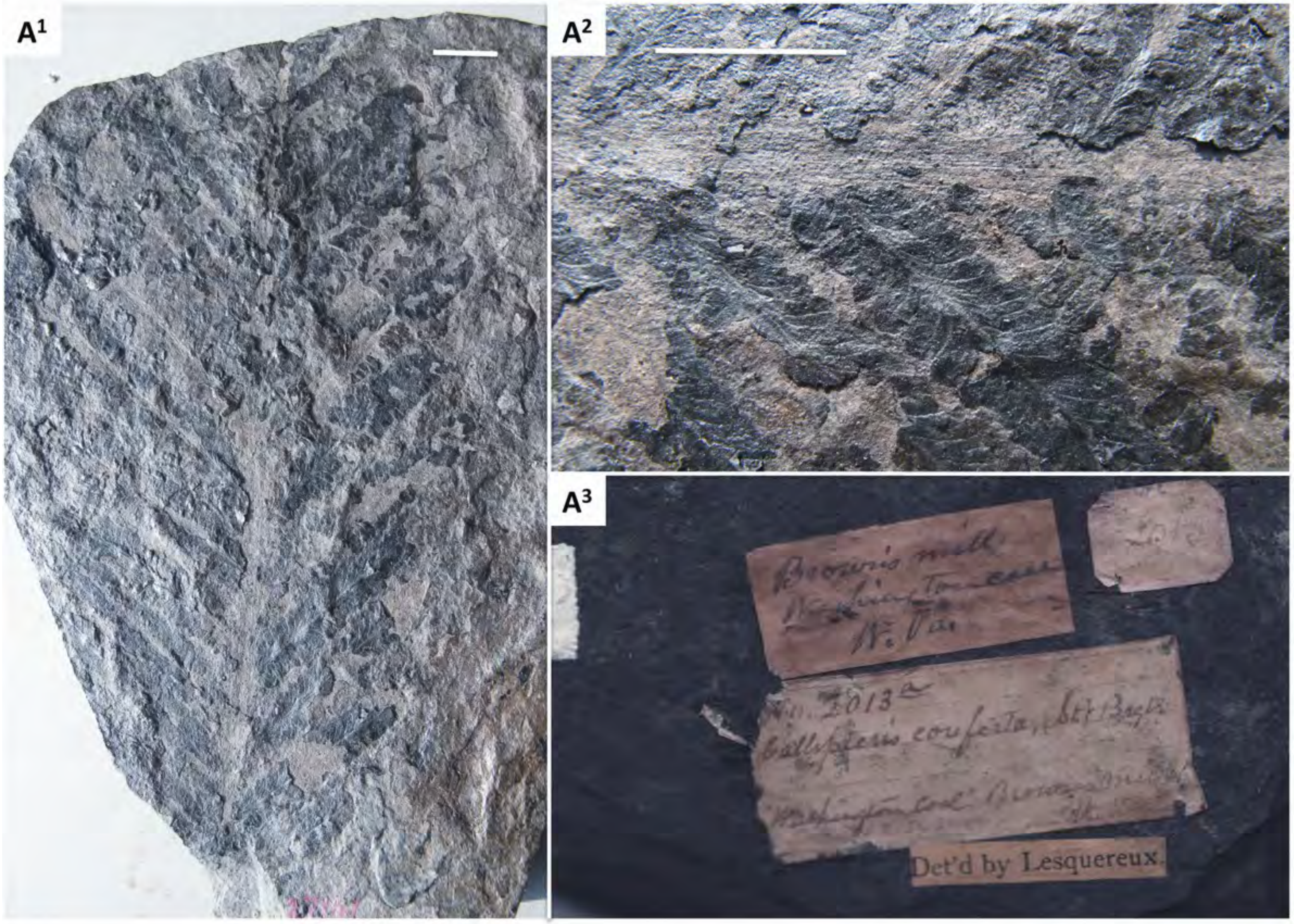

Figure 3. Brown's Mill (Brown's Bridge), WV. *Autunia conferta* from original collecting locality of Fontaine and White (1880). USNM specimen number 27141: A1—Specimen. Typical small-pinnule form of *Autunia conferta*; A2—Close up of pinnae to show the blunt apices of the pinnules and upswept midvein; A3—Reverse side of specimen showing original Lesquereux identification label (as *Callipteris conferta*) and locality identifier label. 2013a is the original Lacoe specimen number. Scale bars = 1 cm.

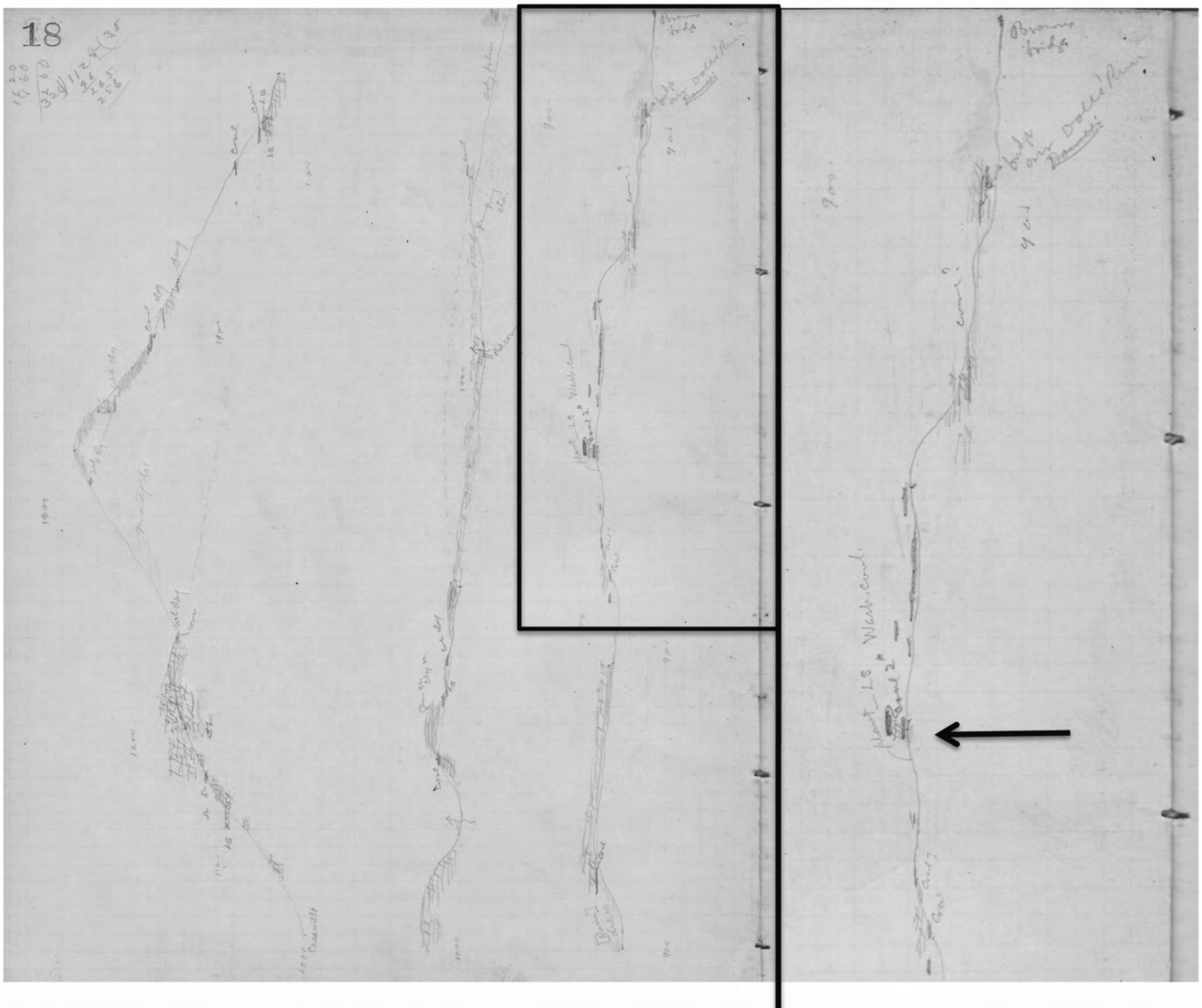

Figure 4. David White's 1902 field notes for Brown's Bridge, WV. Area in black box on left side of figure is enlarged on right side. "Brown's Bridge" noted in lower right hand corner of notebook page. Notation of plants, but not callipterids, at arrow, placing a plant-bearing bed ambiguously in the vicinity of the Washington coal or an overlying limestone.

Specimen Number 27141) is in a heavy, thin bedded, dark gray shale that, from its weight, must indeed contain considerable siderite, but is not what one would describe as an ironstone.

David White visited this site in 1902, during extended fieldwork in the Dunkard, and collected plants from a very dark shale, described below (see White, 1936). Darrah (1975) reports collecting callipterids from this site in the 1930s, from a brown iron stone layer above the Washington coal, which seems similar to the beds from which Fontaine and White collected. Gillespie et al. (1975, p. 229) also report finding callipterids at this site more than once (one time by opening up the site with dynamite), but the fossils were taxonomically indeterminate due to poor preservation in a coarse, sandy matrix, which is quite distinct from the lithology collected by Fontaine and White. The diversity of lithologies from which callipterids have been identified at this site strongly suggests that they were widespread and occurred across the landscape for an extended period perhaps both before and after deposition of the Washington coal.

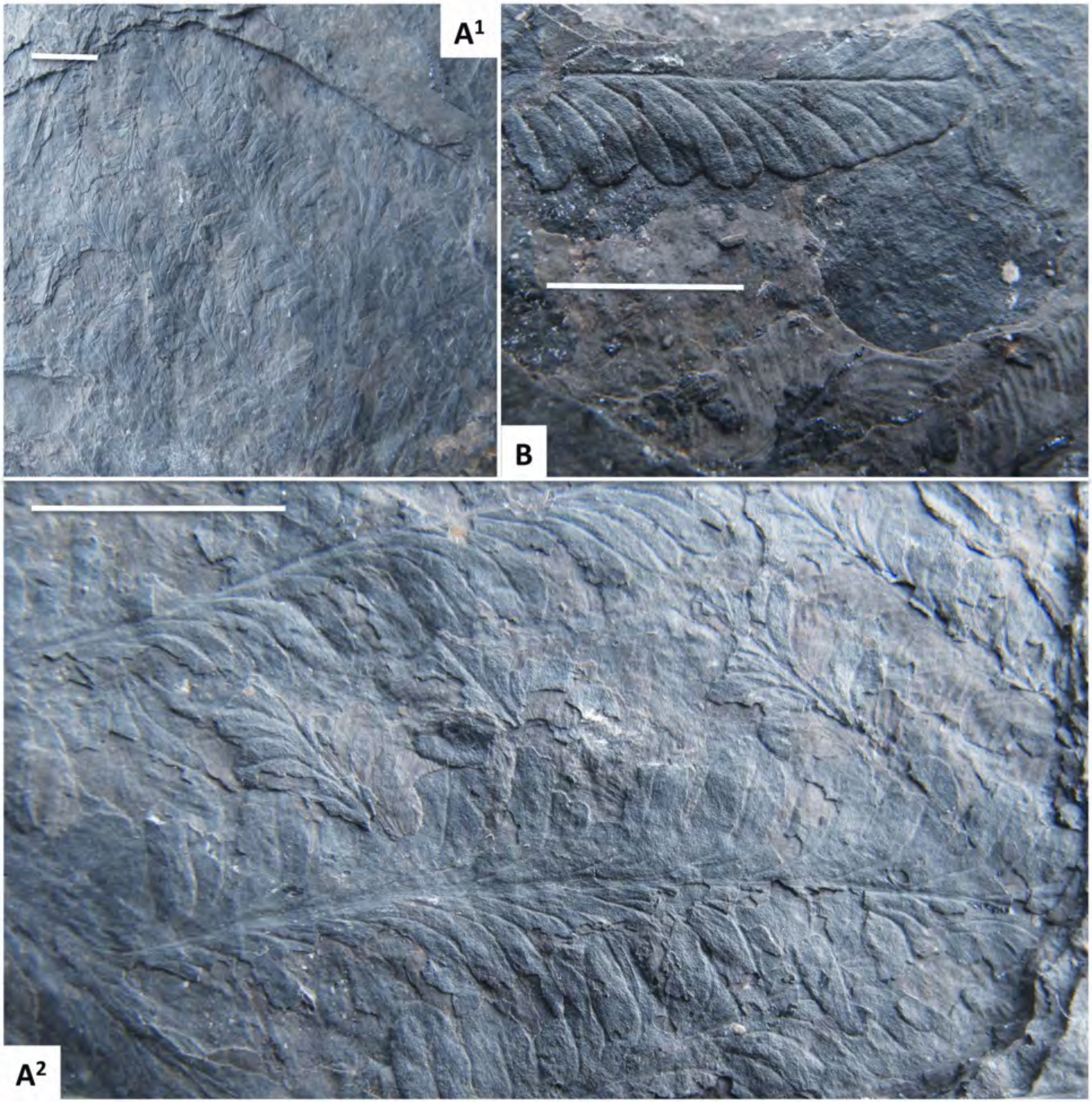

Figure 5. ***Autunia conferta*** **from David White's Brown's Bridge, WV, collecting location. USGS locality number 2926: A1—Fragmentary frond with several pinnae attached to a rachis; A2—3X enlargement of A1 illustrating the shape of the pinnules, which is the same as that of the Lacoe/Lesquereux specimen from the original Fontaine and White (1880) collecting site. Note the thickness of the pinnules, such that the steeply ascending lateral veins are mostly obscure; B—Specimen illustrating the shape of the pinna tip, pinnules gradually fused acroscopically and the steeply ascending lateral veins. Scale bars = 1 cm.**

The lithological matrix of the 1902 collections of David White (USGS Locality Number 2926) is not identical to the single specimen of Fontaine and White that resides in the Smithsonian collections, mainly in being less sideritic and somewhat more distinctly laminated, but both are very dark shales with considerable siderite content, so they may represent the same bed. Darrah recollected from the general site in 1932, but does not illustrate it or provide a geological interpretive section of the site; he described the plant bed as a thin, dense,

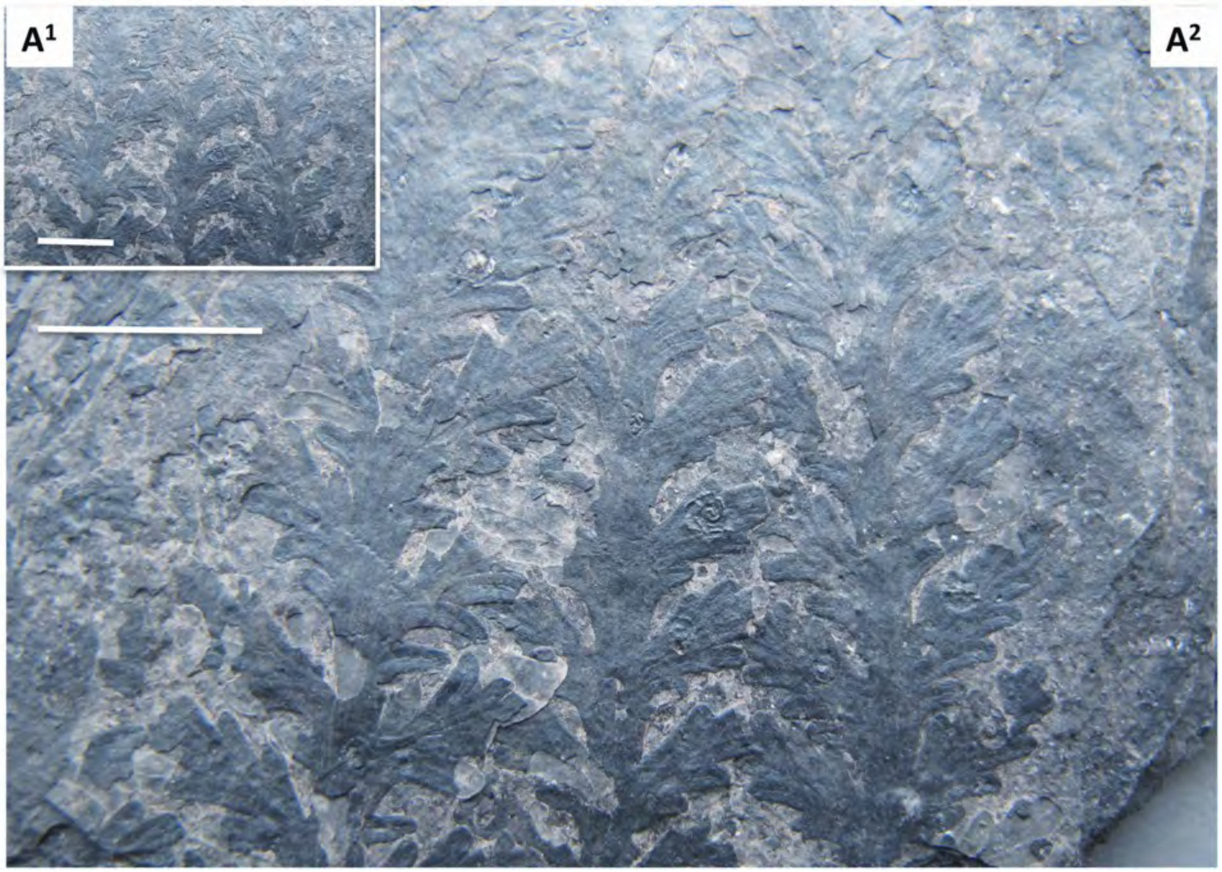

Figure 6. *Lodevia oxydata* from David White's Brown's Bridge, WV, collecting location. USGS locality number 2926. Note spirorbids attached to the foliage: A1—Three pinnae with typical flabellate, lobed pinnules; A2—3X enlargement of A1 illustrating pinnule shape, steeply ascending lateral venation, and thick pinnule aspect. Scale bars = 1 cm.

arenaceous shale, which broadly fits the other known collections, though he does not mention the conspicuously dark color.

As to the physical position of the callipterid bearing beds, David White, whose 1902 field notes are the only known record of this locality by someone who collected callipterids there, barely mentions the site. In the White notes, page 18, upper right corner (Figure 4), the collecting location is shown without comment. He notes a questionable coal near the grade of Brown's Bridge and "plants" below a 5-cm (2-in) thick coal and thin limestone, 12 m (40 ft) above the base of the exposure. It is this thin coal that he identified as the Washington. The callipterid-bearing shale White collected contains non-marine ostracodes, snails, and spirorbids (Tibert et al., in press), and thus may have come from a clastic interbed facies of the Lower Washington Limestone. We recently identified three discontinuous, thin benches of limestone at this location from above what may be a bench of the Washington coal, above the level of the railroad grade in about the position specified in White's field notes. Plants were present, but not callipterids. However, we also have noted many discontinuous siderite layers below the railroad grade at the site and there is evidence of in situ coal below the grade also, which may be another bench of the Washington coal. If the callipterid beds were at this level, they are no longer accessible without major excavation. Modern cover at the location makes it difficult to sort out these relationships.

The Brown's Bridge callipterids are of two types. One we tentatively identify as a small pinnuled form of *Autunia conferta* (Figure 5), consistent with the identification of Fontaine and White (1880), who called these specimens *Callipteris conferta* (F & W Plate XI, Figures 1-4, Figure 2 of this paper). The other we identify as *Lodevia oxydata* (Goeppert) Kerp and Haubold (1988) (Figure 6), originally identified as *Sphenopteris coreacea* Fontaine and White (F & W Plate 5, Figure 5, Figure 2 of this paper). Notably, they compared the species to *Sphenopteris oxydata* Goeppert and *S. lyratifolia* Weiss. Darrah (1969) identified this material as *Callipteris lyratifolia*.

Pleasant Hill Gap

Pleasant Hill Gap is located between New Freeport and Jollytown, Pennsylvania, in the southeast corner of the New Freeport 7.5' Quadrangle map. Here White found callipterids in association with the Nineveh coal bed, high up into the Greene Formation (White 1902, Dunkard Field Notes, page 27) (Figure 7), though he does not indicate the position of the callipterid-bearing rocks in relationship to that coal (i.e., roof shale or some distance above the

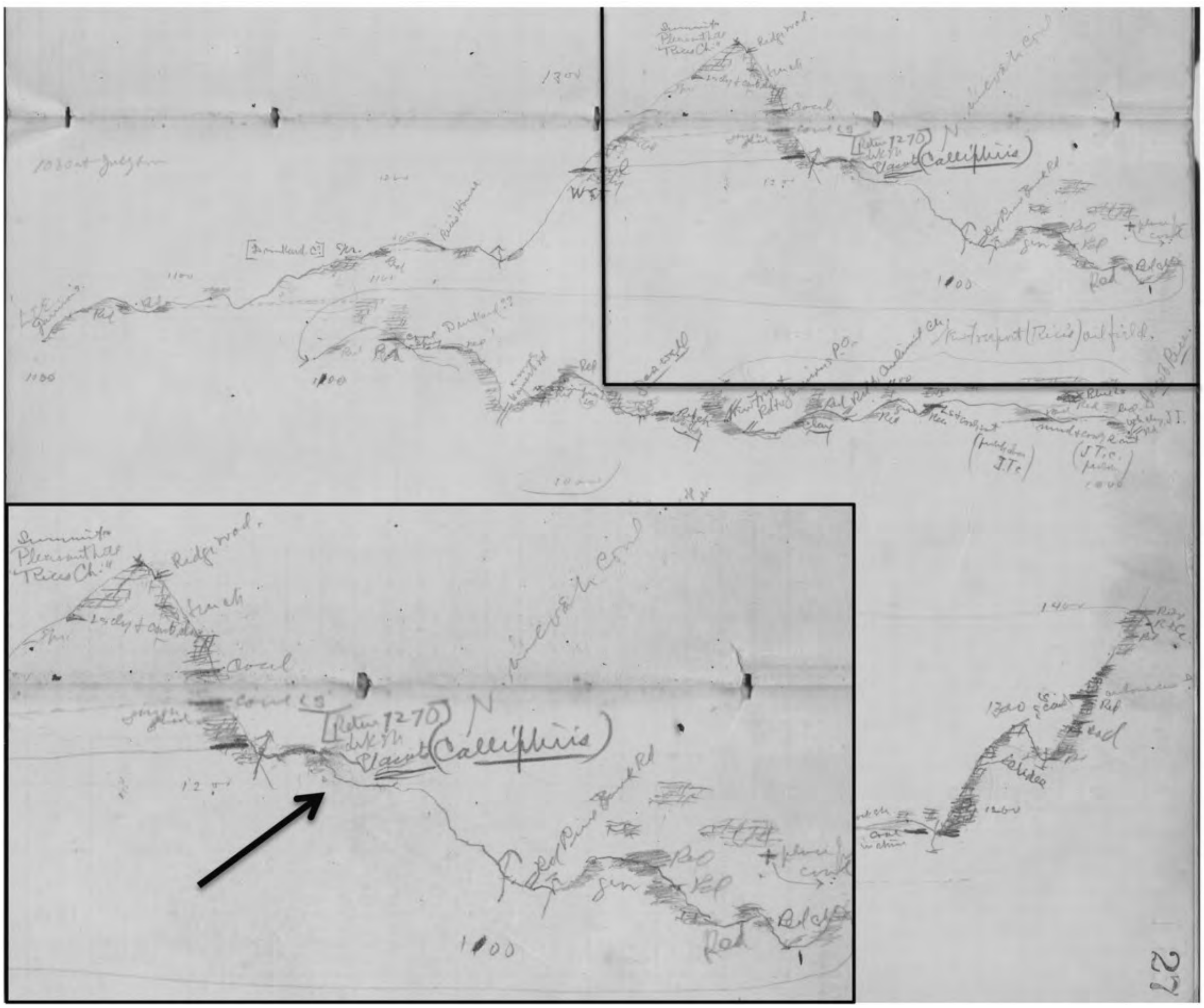

Figure 7. David White's 1902 field notes for Pleasant Hill Gap, PA, USGS Locality Number 2915. Page is rotated 90 degrees clockwise. Area in black box in upper right of image is enlarged in lower left. The occurrence of *Callipteris* is noted in proximity to the Nineveh coal (at arrow).

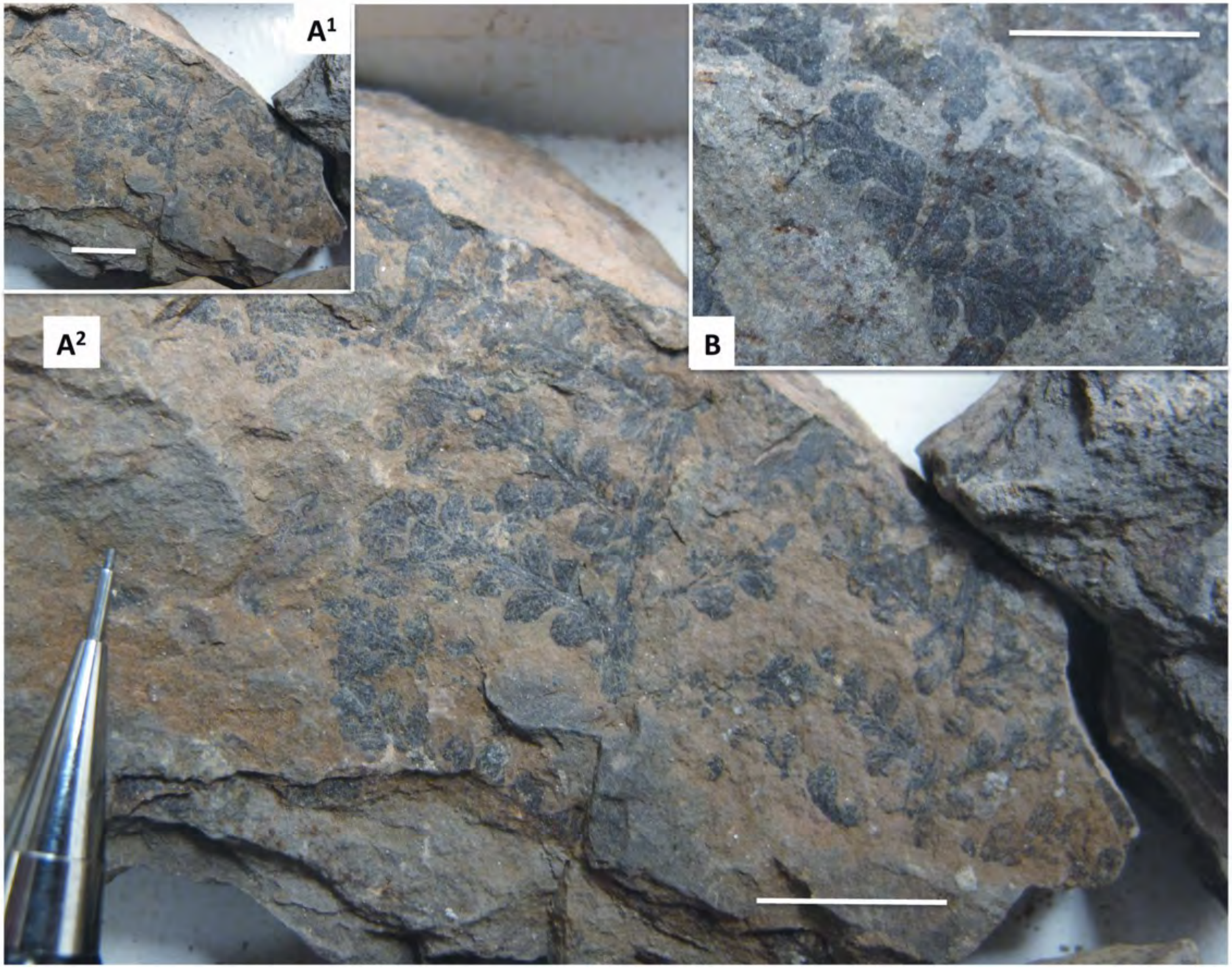

Figure 8. ***Autunia naumannii*** **from David White's Pleasant Hill Gap, Pennsylvania collecting locality, USGS Locality Number 2915. A1. Specimen showing the general aspect and relatively delicate nature of the frond. A2. 3X enlargement of specimen illustrated in A1. Note weakly lobed nature of the pinnule margins and steeply ascending venation. B. Specimen illustrating the rounded to weakly lobed pinnule shape and steeply ascending venation. Scale bars = 1 cm.**

coal). This is USGS Locality Number 2915, also known as "Rice's Gap" and noted by White as occurring on the west side of the gap, 2.4 km (1.5 mi) east of New Freeport. The 1904 road configuration (refer to the Rogersville 15' Quadrangle map, southwest quadrant) appears to be only slightly different from today, and a church is located at the gap as today. However, the area is now heavily vegetated and we were not able to locate the exact beds from which White collected. Although poorly exposed, the section does not appear to contain continuous beds of limestone and the fossil collection was made from a brown siltstone. Darrah (1969) reported relocating this site after discussions with David White in the 1930s. Darrah reports the fossiliferous beds in claystones above the Nineveh coal, with reservation due to the inability to identify any limestones in the section in this section.

We determine the callipterids from this site to be *Autunia naumannii* (Figure 8). Darrah (1969, p. 17) reported additionally finding *Autunia conferta* at this locality.

West of Littleton

The collecting locality West of Littleton is described by David White in his 1902 field notes, page 28 (August 22nd 1902) (Figure. 9). A note in the collection drawer at the museum

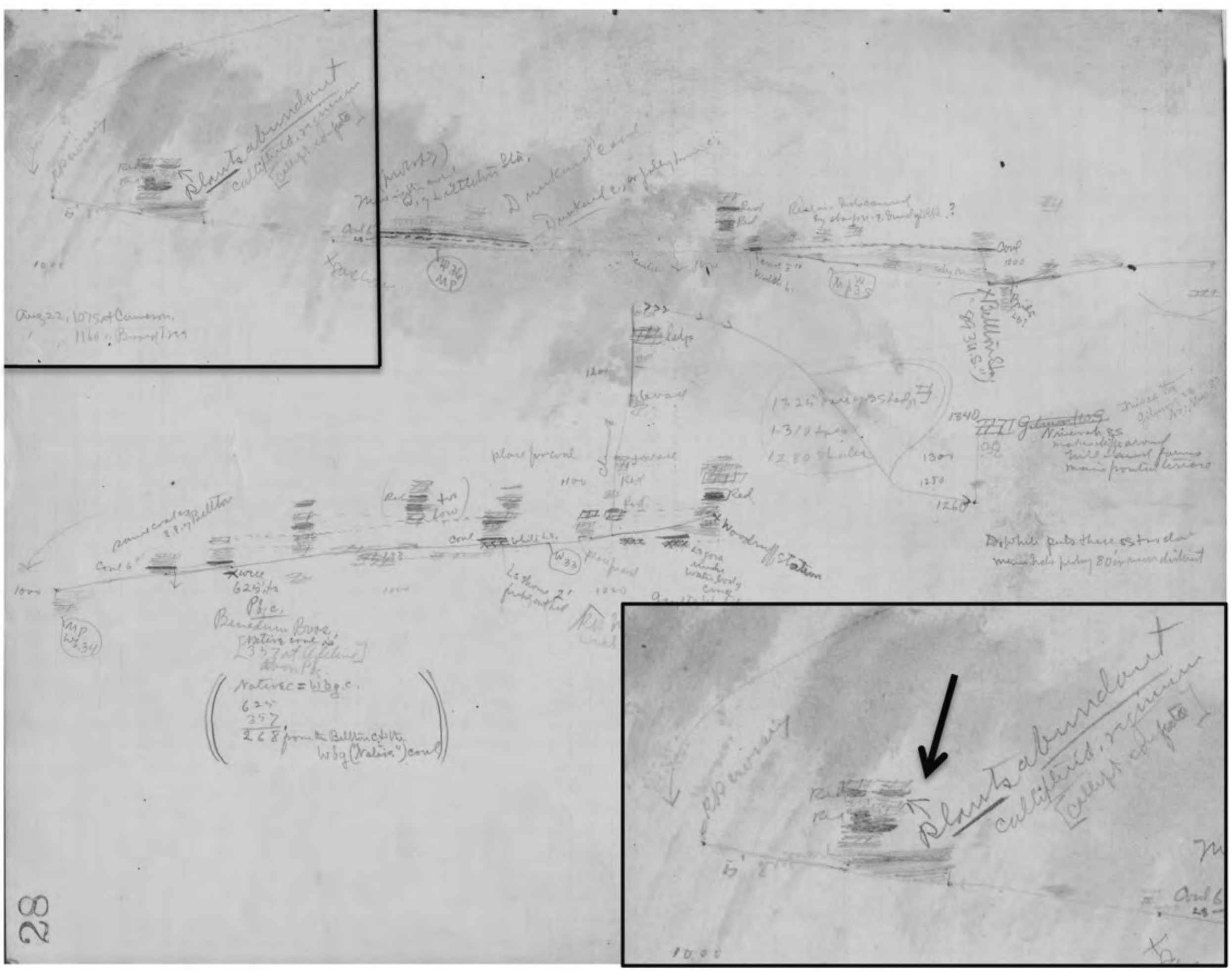

Figure 9. David White's 1902 field notes for Railroad Cut West of Littleton, WV. Page is rotated 90 degrees counter clockwise. Area in black box in upper left of image is enlarged in lower right. Note (at arrow) the

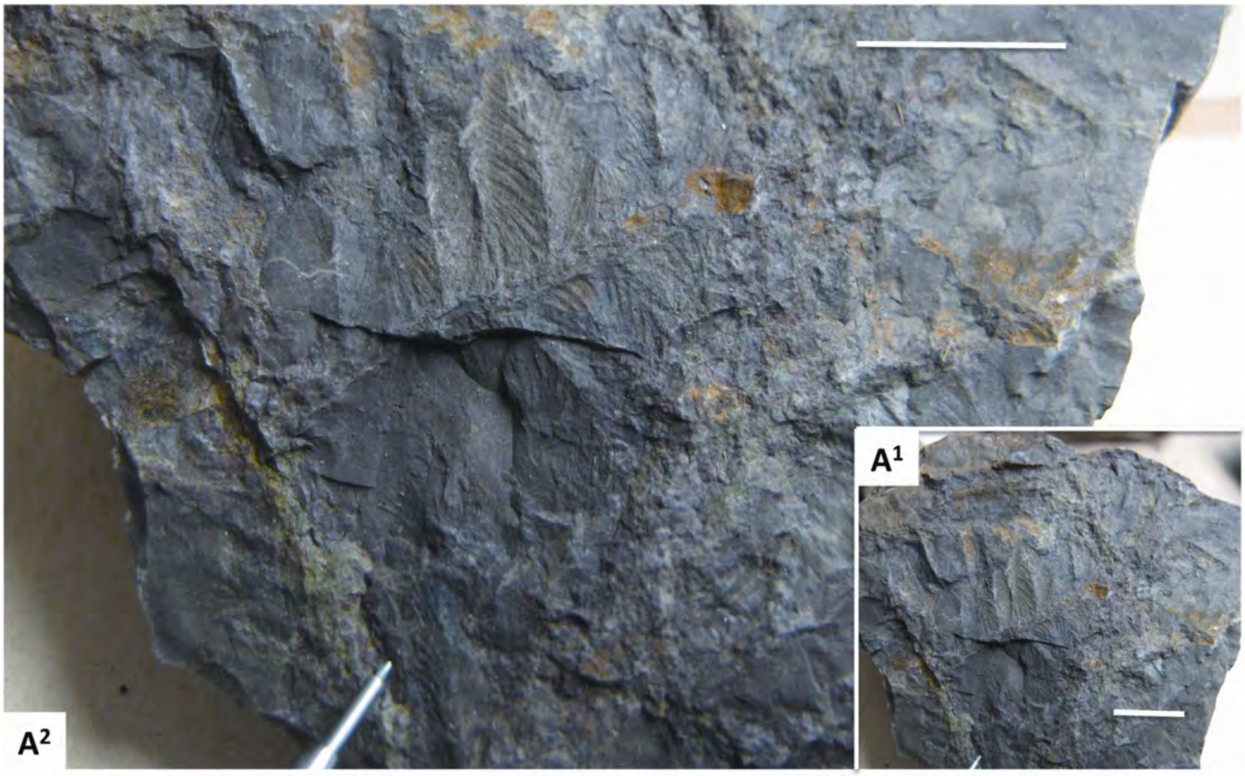

Figure 10. *Autunia conferta* from David White's Railroad Cut West of Littleton, West Virginia, collecting locality USGS Locality Number 2909x. A1. The single callipterid specimen collected from this site, the flora from which is dominated by wetland plants. This plant may have been collected from a different bed, given the distinct locality number (2909x rather than 2909). A2. 3X enlargement of specimen illustrated in A1, showing the bluntly acute pinnule apices, nearly centrally located pinnule midvein, and steeply ascending but convexly arched lateral veins. These pinnules do not appear to be as thick as those from Brown's Bridge. They resemble the larger form of *Autunia conferta* reported elsewhere in the Dunkard and in the Permian of North America (see text).

(written on the back of an envelope from the "Hotel Creed, Cameron, West Virginia") describes the collection locality more accurately as occurring at the west end of the first railroad cut west of Littleton. The fossils occur in red shales above what White identified as the "Dunkard (or perhaps the Jollytown)" coal, a bed 6" thick, in the Greene Formation. He notes the presence of "*Callipteris reginum*", which either he or someone else has annotated in brackets as "[Callipteris conferta]". There appear to be other annotations in the notebook referring to "D. White" in the third person, taking exception with a stratigraphic determination. However, those describing this West of Littleton site seem to be in White's own handwriting.

The West of Littleton plant collection (USGS Locality Number 2909) consists mainly of typically wetland plants, including tree fern foliage and possibly *Callipteridium,* in red shales. However, there is a single specimen, segregated from the others, labeled USGS Locality Number 2909x, identified on a note as *Callipteris*, with an illegible species name (David White's handwriting is often nearly illegible). We determine the plant from this site to be the typically larger form of *Autunia conferta* (Figure 10), although the lateral veins are somewhat more flexuous than is typical for most specimens of that species.

Rinehart Tunnel Station

On the last field day (August 23rd) of his 1902 summer field season White (field notes page 34) stopped near a newly excavated railroad tunnel on the West Virginia Shortline Railroad. He called the site "Tunnel Sta." (?Tunnel Station) located, according to his description, "…in SE corner of Wetzel County in Centerpoint 15' quad and Folsom 7 1/2 ' quad. It is between Rinehart and Folsom." Again, the fossils seem to come from the Greene Formation, located approximiately 373 m (1,225 ft) above the Pittsburgh coal and about 91-122 m (300-400 ft) below the top of the hill.

White explicitly notes "*Callipteris* in plenty at the tunnel in the **** [illegible] dump and in dump at the w. end of switch below tunnel sta." We were able to relocate and reach the eastern portal the tunnel, but could find no evidence of waste rock dumps from tunnel excavation, the presumed source of the callipterid fossils. It is probable that the excavated rock was hauled off and perhaps was used locally as fill in the construction of the railroad bed. Unfortunately, no collections appear to have been made from this site, at least as far as we can tell from an examination of the USGS locality records or from a survey of the Smithsonian collections.

Winkler's Mill

Winkler's Mill is a site collected by Aureal Cross on August 27, 1948. It is in the NW quarter of New Martinsville 15' Quadrangle, approximately due west of New Martinsville WV. Cross et al. (1950) describe the general area, though not this specific site, in Stop 12 of the Day 1 road log and as Geological Sections 8C, 8D, and 8E on Bares Run, which is the drainage to the immediate east of Winklers Mill. The area exposes the highest Greene Formation rocks in Ohio. A fossiliferous horizon containing callipterids is not noted in the geological sections or road log of that field guide. Only the plant containing beds noted in Section 8E are not from shales immediately above coaly horizons, making these the most likely candidates for the bed from which the callipterids were obtained. However, the descriptions of the beds note that they are gray-green to purple in color (purple is often used, however, for red beds with hematitic inclusions, so red coloration is not ruled out by this description).

The Winklers Mill collection is presently held by the Field Museum of Natural History in Chicago, Illinois, and bears A.T. Cross' original locality number C-339(2) and B-3667. We identify the callipterid in this collection as the larger form of *Autunia conferta* (Figure 11). The flora of that collection contains other, more typically wetland elements, in the same matrix as *A. conferta*, including an *Alethopteris* similar to *A. zeilleri* and *Annularia carinata*.

Other collections

Darrah (1975) illustrates three callipterid specimens, all of which he identified as *Callipteris conferta*. We are in agreement with this identification for his Figures 1 and 2, which appear to be the small pinnuled form of *Autunia conferta* (Figures 12 A and B). Darrah's Figure 3 is much more fragmentary and appears to be lobed. It may be the small form of *A. conferta*, but the lobes are not separated adequately to be described as pinnules. Ultimately, the specimen, as photographed, is too small for a positive identification. The specimens illustrated in Darrah's Figure 1 (Figure 12B) and Figure 3 were collected by J.A. Barlow on Williams Run

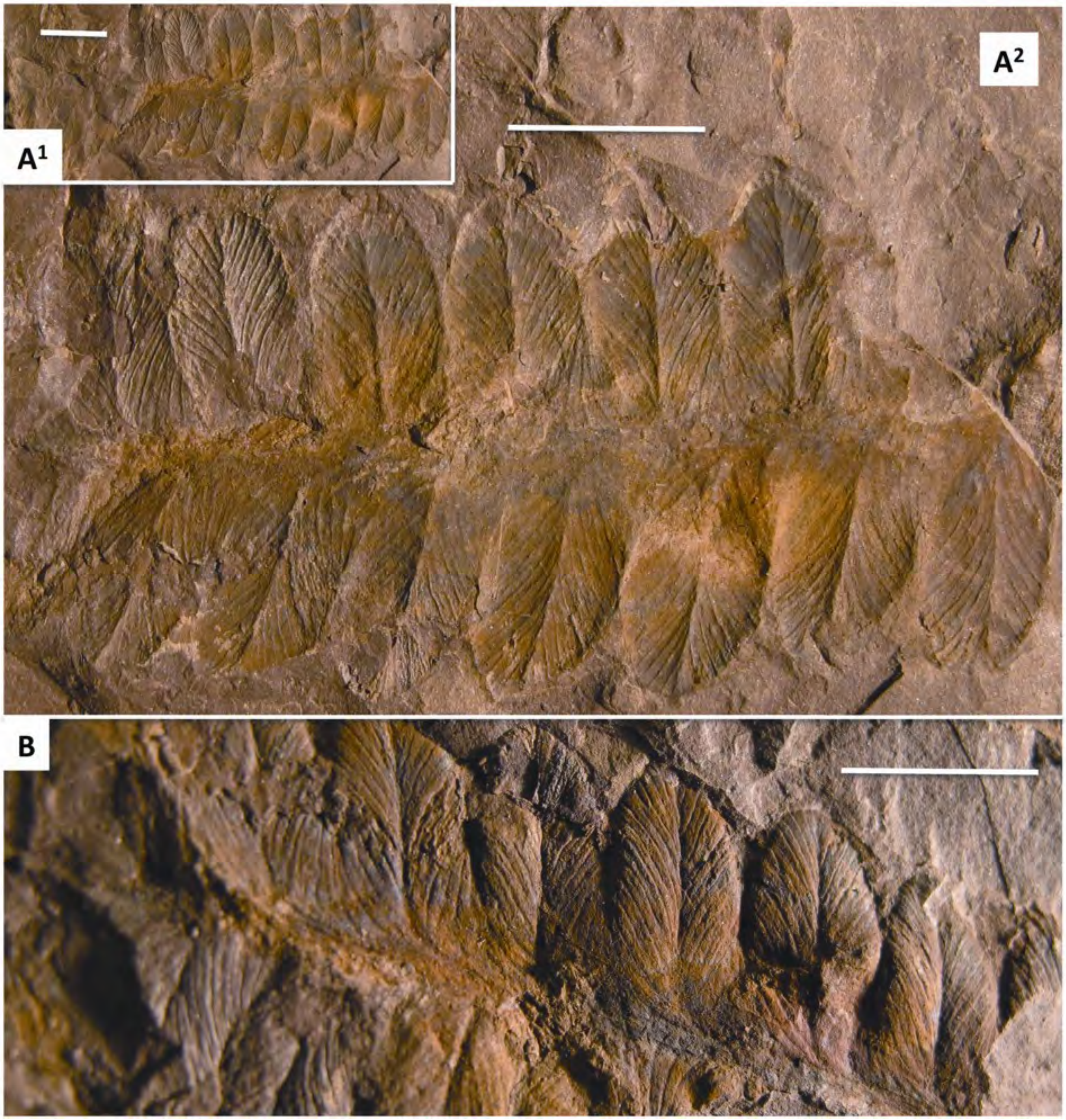

Figure 11. ***Autunia conferta*** **from Aureal Cross' Winkler's Mill, Ohio collecting locality, A.T. Cross collection numbers C-339(2) and B-3667. A1. Specimen showing general aspect of a single pinna. A2. 3X enlargement of specimen illustrated in A1 showing the bluntly acute pinnule apices, nearly centrally located pinnule midvein, and steeply ascending but convexly arched lateral veins. These pinnules, as with those illustrated in Figure 10, do not appear to be as thick as those from Brown's Bridge and resemble the larger form of** ***Autunia conferta*** **reported elsewhere in the Dunkard and in the Permian of North America (see text). B. Specimen illustrating the pinnule shape and venation. Because this is a "positive" specimen, it likely preserves an impression of the underside of the pinna and pinnules. Note the prominence of the veins, as if they were well marked on the lower surface of the leaf, which was pressed into the sediment matrix, leaving a "footprint". Scale bars = 1 cm.**

near Williamstown, West Virginia. The specimen illustrated in Darrah's Figure 2 (Figure 12A) comes from a ravine north of Wolfdale in Canton Township, northwest of Washington, Pennsylvania.

Gillespie et al. (1975) illustrate two callipterid specimens (Figure 13). Both were collected from 1.2 m (4 ft) above the Lower Washington Limestone, approximately 1.6 km (1 mi) east of Waynesburg, Pennsylvania. The first (Gillespie et al., Plate I A, 2/3 X) (Figure 13A) is the larger pinnuled form of *Autunia conferta*, which Gillespie et al. identify as "possibly *C. conferta*". The second specimen (Gillespie et al., Plate VIII A, no scale) (Figure 13B) is also identified as "possibly *C. conferta*", but appears to be *Rhachiphyllum schenkii* Kerp (Kerp, 1988). Pinnules are large, elongate with lobed margins, are strongly decurrent and have distinct lobes on the basiscopic margins near the point of lamina contact with the midrib. The pinnules appear to be thin and non-vaulted, of relatively flat aspect with the midvein and lateral veins flush with the surface of the lamina; details of venation cannot be observed in the photograph of the specimen.

Darrah (1969, p. 17) notes the occurrence of *Autunia* (*Callipteris*) *conferta* at a minimum of 8 sites in Pennsylvania and West Virginia, in addition to Brown's Bridge and Pleasant Hill Gap of David White. Most of these sites he describes as north and northwest of Waynesburg, in both red shales and dark gray sandstones above the Lower Washington Limestone.

Gillespie and Pfefferkorn (1979) report and illustrate several specimens that they attribute to *Callipteris conferta* from red shale, probably from the Washington coal interval, exposed near Liberty, Kanawha County, West Virginia. This locality is in a basin margin setting, over 161 km (100 mi) south of the classic Dunkard area being covered by this field trip, and well outside the area where coals and limestones are best developed in the Dunkard Group. As a consequence these specimens are extremely important indicators of the occurrence of callipterids in areas where soil moisture was likely seasonal in distribution.

DISCUSSION

A diverse group of callipterids has been identified in the Dunkard Group. This diversity was recognized very early in the study of the Dunkard flora, and various names have been applied to these fossils. Our present list includes the following: *Autunia conferta*, both a large and a small form, *Autunia naumannii*, *Lodevia oxydata*, and *Rhachiphyllum schenckii*. The large and small pinnuled forms of *Autunia conferta* may indeed be different species. The small pinnuled form bears a close resemblance to the more traditional European morphotypes attributed to *A. conferta* (see Kerp, 1988). The larger pinnuled form, a very nice specimen of which Gillespie et al. (1975) illustrate in their Plate I A, is quite like large pinnuled callipterids from Kungurian red beds of North-Central Texas (Chaney and DiMichele, 2007) in terms of pinnule shape (ovoid), moderate-to-large size, and the tip shape (varying from sharply to bluntly pointed). There are other names that have been applied to some of the Dunkard callipterid specimens, but nearly all these names seem to swirl around the "*Callipteris conferta*"-like specimens from Brown's Bridge, the commentaries related mainly to the matter of *C. conferta* as a traditional marker species for the bottom of the Permian, thus the import of a correct species identification at the time most of the papers were written. Gillespie et al. (1975, p. 229) note that David White (1933) described the Brown's Bridge fossils by a number of names (*Callipteris lyratifolia*, *C. diabolica*, and *C. currettiensis*) but not as *Callipteris conferta*. They also note that Hans Bode (1958) reports examining the Brown's Bridge material first hand

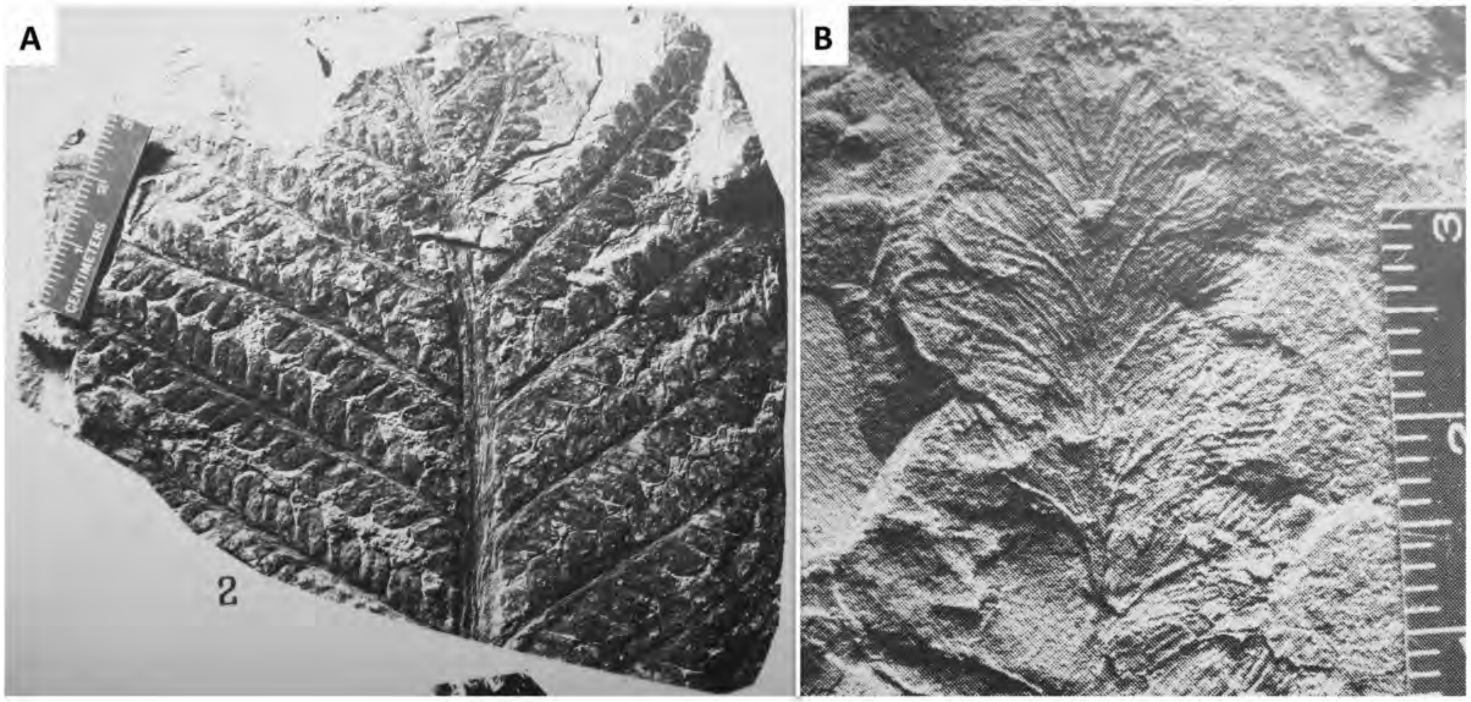

Figure 12. Callipterid specimens originally illustrated by Darrah (1975). A. *Autunia conferta* of the large pinnule form. This is Darrah's original Figure 2. B. *Autunia conferta* of the large pinnule form. A portion of Darrah's original Figure 1. Note the bluntly acuminate pinnule apices and the steeply ascending, but convexly arching lateral veins, and fused pinnules at the pinna apex.

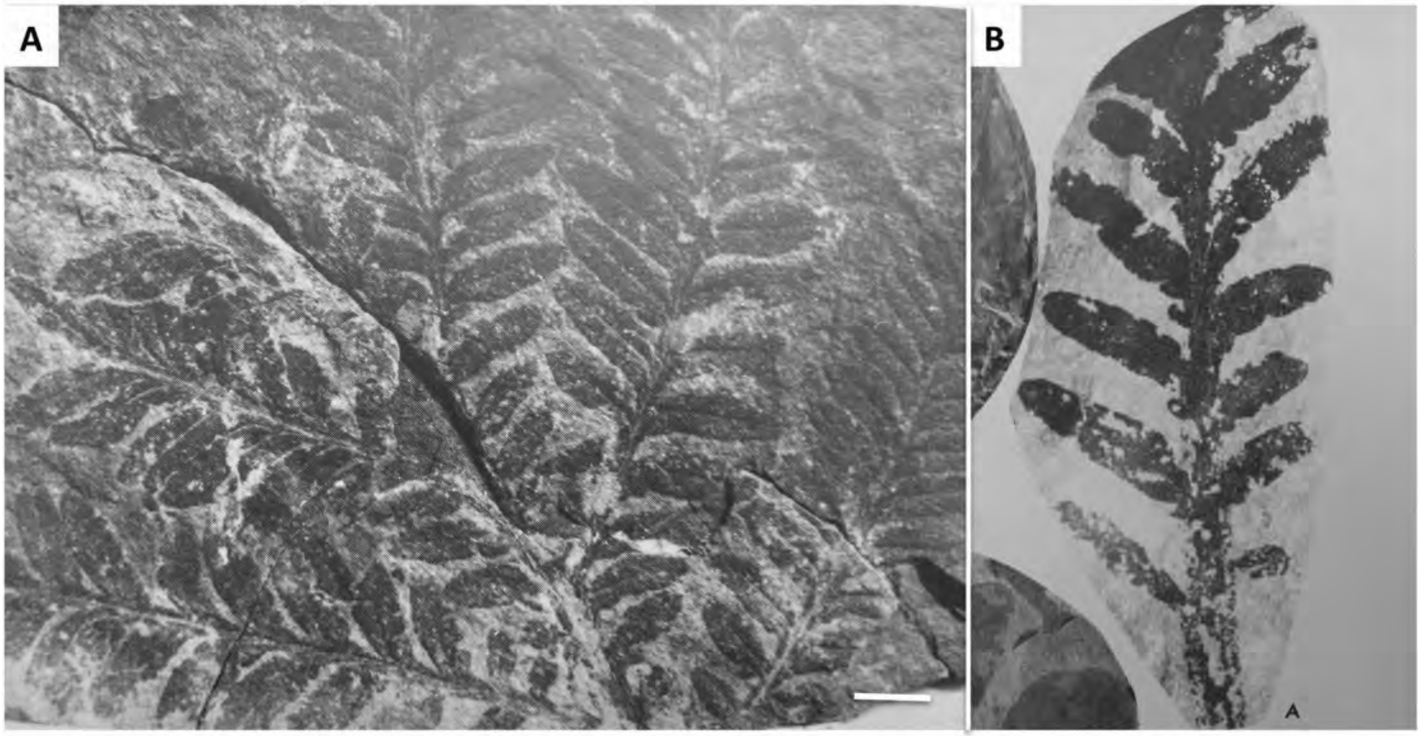

Figure 13. Callipterid specimens originally illustrated by Gillespie et al. (1975). A. *Autunia conferta*. Gillespie et al.'s original Plate IA figure. This is the most extreme example of *Autunia conferta* with large pinnules. In this case the pinnules are distinctly obovate with tapered tips. Intercallary pinnules, characteristic of callipterid fronds, are clearly visible on the main rachis of the specimen, above the first left-side pinna from the base. Scale bar = 1 cm. B. Possible *Rhachiphyllum schenkii*. Gillespie et al.'s original Plate VIIIA, which was identified as "Possible *Callipteris conferta*". Note flat aspect to the pinnules; they are not vaulted and the midvein is not sunken in the lamina surface. Pinnule margins are wavy and the pinnules are distinctly lobed at the base. The terminal pinnule, as far as it can be seen, is free. No scale was given for the original illustration.

in 1956 while the collection was on loan to Aureal Cross at West Virginia University. He also doubted the *C. conferta* identification and attributed the lot to *C. lyratifolia*, noting that they were rather poorly preserved – beauty being in the eye of the beholder.

Callipterids appear, as has been noted before by every author who has looked closely at the matter, in the middle of the Washington Formation (in the two formation scheme) and continue to occur upward, sporadically, into the Greene Formation. Darrah (1969, p. 47) noted that *Callipteris conferta*, *C. lyratifolia*, and *C. naumanni* occur at several horizons in the Dunkard and generally are abundant where found. However, the number of localities at which they have been found are rather few and scattered compared to occurrences of typically wetland plants. Darrah interprets the coriaceous leaves of the Dunkard callipterids to indicate "a facies flora of drier environments. " He notes that he "never found *Callipteris* in fine grained gray or black shales. … but in the upper Dunkard section it always occurs in coarse micaceous sandstone of yellow brown to chocolate brown color." He also notes finding *Lebachia* in association with callipterids above the Nineveh coal (?presumably at the Pleasant Gap locality), which is in keeping with an overall interpretation of these plants as indicative of seasonally dry conditions.

Darrah (1969, p. 154) states: "The occurrence of *Callipteris* is much more sporadic than I had long ago assumed. … When *Callipteris* does occur, however, it is abundant almost to the exclusion of other forms, at many localities." He notes that it occurs from the level of the Washington coal upward into the upper Dunkard.

Interpreting the callipterid growth environment from the patterns of occurrence and the environments of deposition, such as they are known, is not particularly revealing. Callipterids have been found in various kinds of lithologies, from limey zones (such as that at Brown's Mills and on Robert's Ridge Road) to siltstones (Pleasant Gap) and red claystones (Winkler's Mill, East of Littleton). In Darrah's words (1969, p. 68), "…the *Callipteris* Permian flora...appears to be indicative of physiologically xerophytic conditions. It is a biofacies usually associated with a peculiar lithofacies." However, that lithofacies has not been characterized well enough for any generalities to be drawn. Perhaps the most telling occurrences are those that clearly are associated with non-marine limestones, such as David White's Brown's Mills collection or our more recent collection from Robert's Ridge Road. In each of these instances the callipterids occur in close association with invertebrates (the ostracodes and non-marine snails of the Brown's Mills shales) or non-marine limestones themselves (the occurrence of callipterids in a shale parting at the top of the Lower Washington Limestone on Robert's Ridge Road). These occurrences suggest environments of high evapotranspiration, at least seasonally, that led to relatively high concentrations of dissolved solids in lacustrine environments. It is noteworthy biofacies and lithofacies similar to those of the Dunkard occur in the Appalachian section in the older Missourian (Kasimovian) Conemaugh Formation and, to a lesser extent, in the Virgilian (Gzhelian) Monongahela Formation, and conifers have been reported from these rocks (Blake, 1992; Darrah, 1969; Martino and Blake, 2001; Blake et al., 2002). However, no callipterids are known. Callipterids have been reported in association with sabkha facies in Missourian (Kasimovian) age rocks of New Mexico (Falcon-Lang et al., 2011) and the earliest known callipterid has been reported from late Desmoinesian (Westphalian/Moscovian) age rocks of the Illinois Basin (Pšenička et al., 2011), so these plants were "out there" on the landscape someplace much earlier than they appear in the Dunkard Group.

The sense of callipterid rarity may be exaggerated by the limited amount of outcrop, particularly of the non-resistant shales, in the presently humid to moist subhumid, highly vegetated northern Appalachian landscape. Were the Appalachian region to be exposed under a semi-arid to arid climate, such as that of the Permian of Texas or New Mexico, the controversy over this matter might be less conspicuous because callipterid-bearing beds might be better exposed, found to be more numerous, and be better understood depositionally/ sedimentologically. Callipterid-bearing beds seem to be unassociated with the conspicuous coaly facies of the Dunkard Group. Rather they occur in various kinds of shales, even clastic beds within limestones. Most plant sampling, in the Greene Formation, in particular, comes from coaly layers, which certainly tend to produce wetland plants of a type generally considered to be a "Pennsylvanian" flora. Of course, the real issue with plants is fidelity to climatic conditions (e.g., see many references to this in Roscher and Schneider, 2006; DiMichele et al., 2008). There seems to be little controversy surrounding the suggestion that whatever the occurrence of callipterids in the Dunkard may signify about its age, they most certainly indicate subhumid paleoclimates, demonstrated by the return of red Vertisols in the Greene Formation. Within the scope of the Pennsylvanian, such climatic conditions are known to have appeared in the equatorial regions as far back as the Middle Pennsylvanian (Cecil et al., 2003; Falcon-Lang et al., 2009; DiMichele et al., 2010). The interlayering of callipterid-dominated assemblages with those dominated by typically Pennsylvanian wetland floras indicates greater extremes of climatic oscillation than those characterizing the preceding Monongahela Formation and lower Dunkard – part of a longer term drying trend with repeated oscillations back to wetter conditions (Roscher and Schneider, 2006; Tabor and Poulsen, 2008). The overall pattern is typical of many continental basins throughout the US and Europe wherein the Pennsylvanian-Permian boundary, regardless of how it may be defined paleontologically or by absolute age dates, is lithologically gradational, to the point of being described as "indistinct".

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One day, while looking for fossils in the Dunkard Group.

PLEASANT HILL GAP

William A. DiMichele, Vik Skema, and John A. Harper

During the summer of 1902, David White, then a geologist and paleobotanist with the U.S. Geological Survey, in Washington, DC, made an extended trip through parts of southwestern Pennsylvania, northern and western West Virginia, and eastern Ohio. His primary objectives, based on his locations during this time (determined from his field notebook), were to examine exposures of the Dunkard Group, meet with I. C. White in Morgantown in order to examine existing plant collections and exchange ideas, and acquire enough first-hand information to come to his own conclusions about the age of the Dunkard. David White accepted that a large part of the Dunkard was of Permian age, but disagreed with I.C. White on the particulars about where the boundary should be placed, preferring the base of the Greene Formation to the top of the Waynesburg coal. By the end of his career, however, based on a posthumous publication (White, 1936), he had converged with I.C. White's original conclusion that the entire Dunkard was of Permian age.

Much of the dispute about the age of the Dunkard revolved around the discovery of callipterid plant remains at about the level of the Washington coal. Callipterids are a group of plants that are typical of Permian-age deposits (see DiMichele et al., 2011). In the early 1900s, and until recently, one particular callipterid species, *Callipteris conferta* (now transferred to the genus *Autunia*), was considered to be an index fossil for the base of the Permian Period. As part of his plant collecting, David White found callipterids at several places in the Dunkard, all from the Washington coal zone or higher.

The youngest callipterid plants reported from the Dunkard Group, to this day, are a handful of specimens attributable to *Autunia* (*Callipteris*) *naumannii* from sandy shales that were exposed (in 1902, and perhaps as late as the 1930s, according to Darrah, 1975) on the roadside near Pleasant Hill Gap (Figure 1). The road configuration is slightly different today than in 1902, and we believe the original road can still be detected (Figure 2), though, as can be seen from the photograph, outcrop exposures are somewhat limited! Here D. White, at an elevation of about 1,270 ft (387 m), located these fossils in proximity to a coal bed we presume he identified as the Nineveh coal (Figure 3). White's notes are unclear, however. The word "Callipteris" is prominently displayed, apparently added later, next to the original annotation of "Plants". A pencil line leading from those words to the outcrop diagram appears to place the bed beneath the nearest coal (?) in the section; the dark line above the plant bed may have been a dark shale based on White's annotations ("drk sh" is written near, but not next to the dark line he usually uses to indicate coal beds). "Nineveh Coal" was apparently added as a marginal note at some later time, but exactly which bed this refers to is uncertain (Figure 4A). The callipterids are not preserved in dark shale (Figure 4B1 and B2).

According to W. C. Darrah, in 1932 White provided him (Darrah) with information on the location of this, and many other, Dunkard plant collecting sites (Darrah, 1969, p. 17; Darrah, 1975, p. 86). Darrah returned to the site and located the callipterid bearing bed, but could not locate any limestones or sandstones in the hillside (though note that White does record "LS" in the section, above the plant bed in his note, Figure 4A – we were not able to relocate this or any

DiMichele, W. A., Skema, V., and Harper, J. A., 2011, Pleasant Hill Gap, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 168 -173.

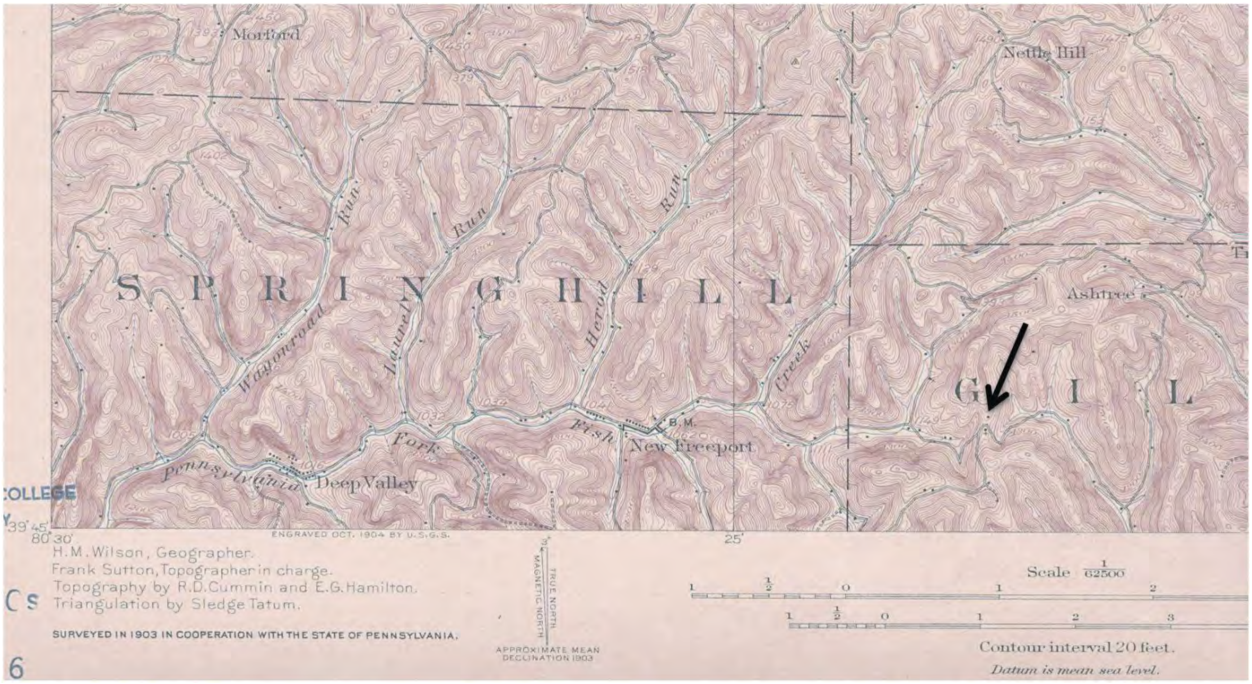

Figure 1. Rogersville 15-minute quadrangle, 1905 (survey of 1903). This map shows the road positions at the time of David White's examintion of this site, in 1902. Site is located at the arrow.

Figure 2. Pleasant Hill Gap, looking downhill to the SW, just below the point at which the modern road crests the hill. This is possibly the road on which David White would have traversed this area in 1902. He notes a "Pine Bank Road" in his field notes – see Figure 3. Note the heavy vegetation cover, making relocation of the plant-bearing bed difficult. A portion of the modern road can be seen down slope to the right.

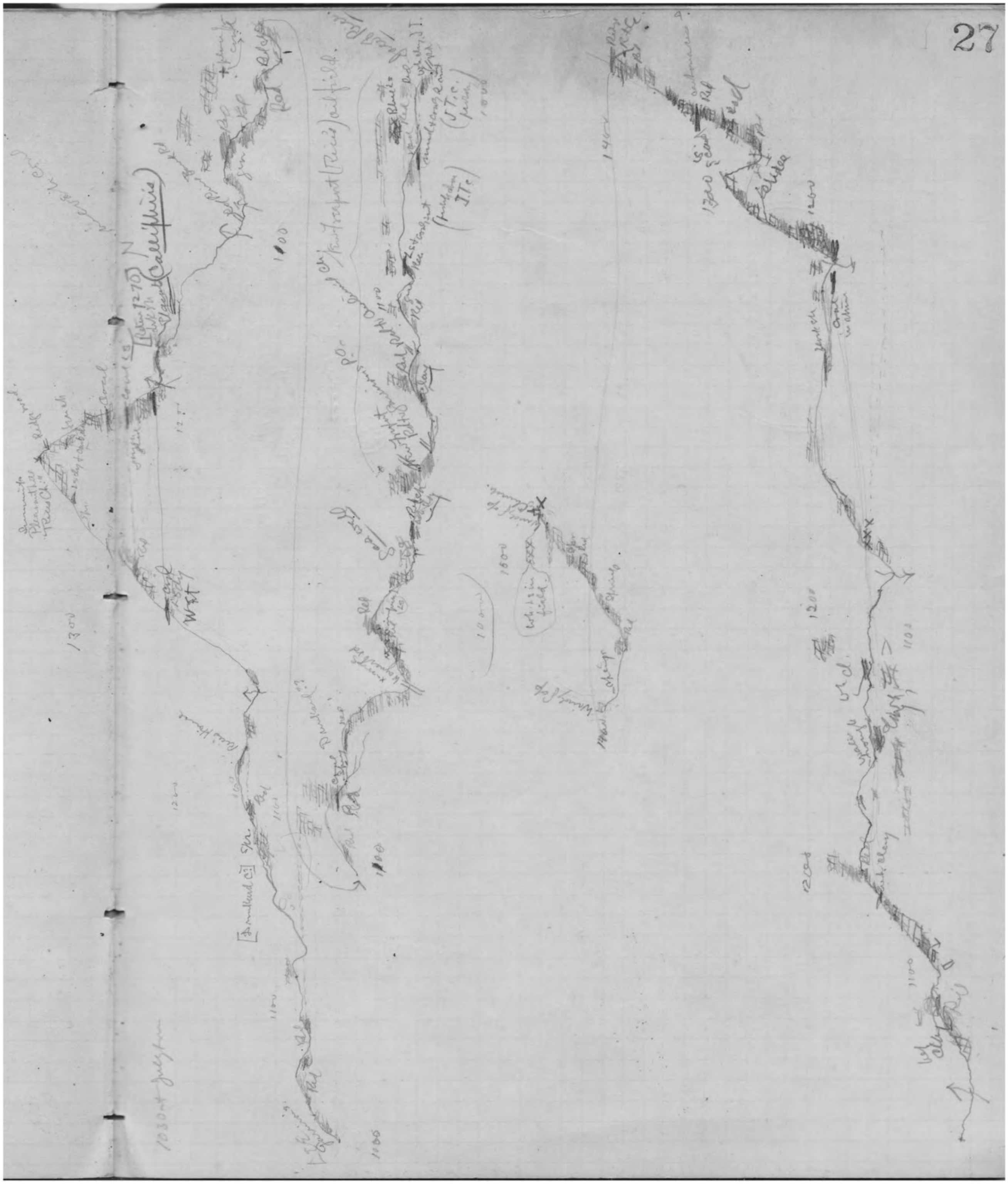

Figure 3. Page 27 from David White's 1902 field notebook, in which he records the presence of "*Callipteris*" in proximity to the Nineveh coal at Pleasant Hill Gap. Plant-bearing beds are noted at arrow.

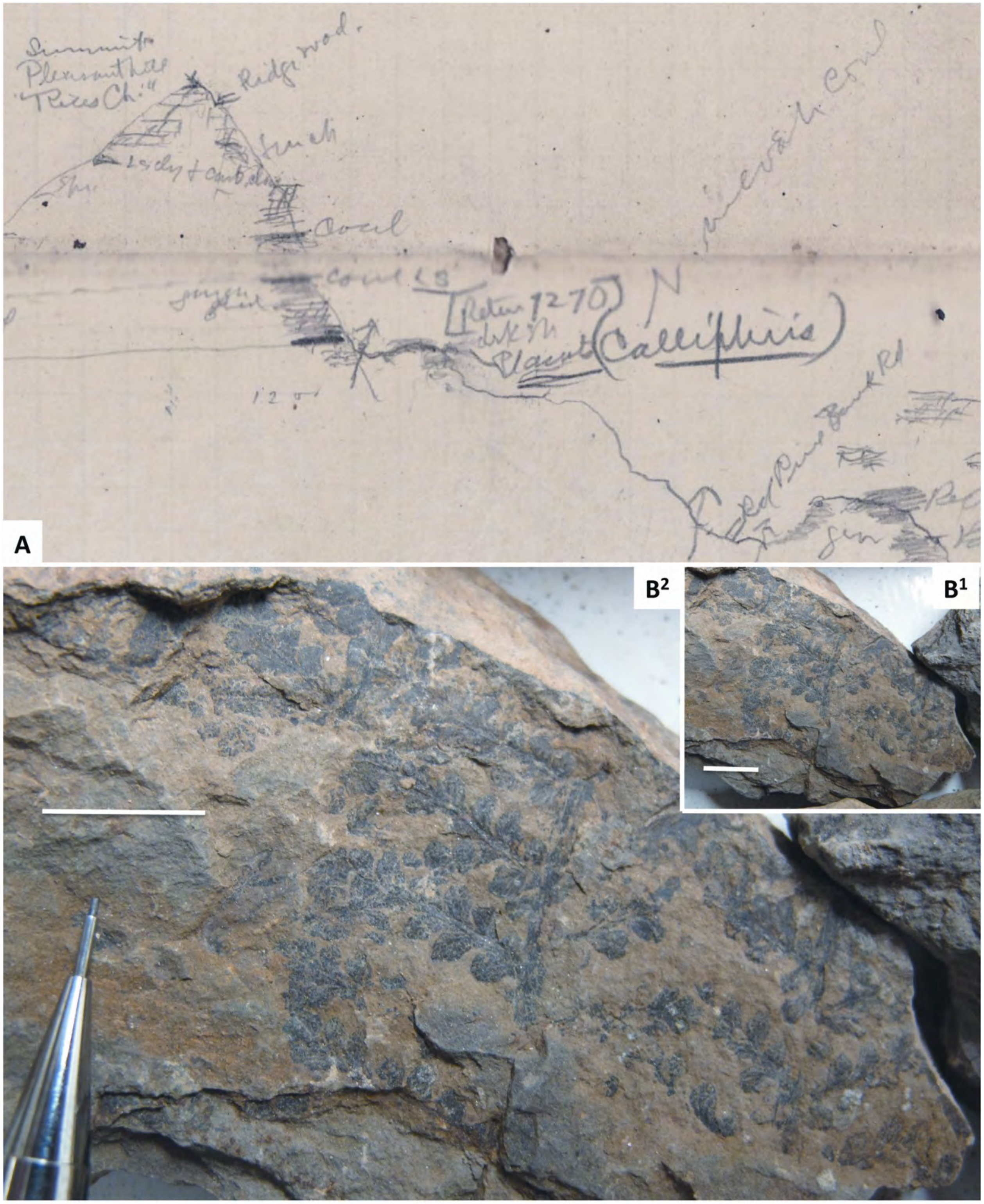

Figure 4. A—Enlargement of portion of page 27 from David White's 1902 field notebook where he describes the occurrence of "*Callipteris*" in proximity to the Nineveh coal. B1—*Autunia naumannii* from the National Museum of Natural History collections, USGS locality number 2915, X1. B2—Specimen illustrated in B1 X3, to show detail of the plant and composition of the rock matrix.

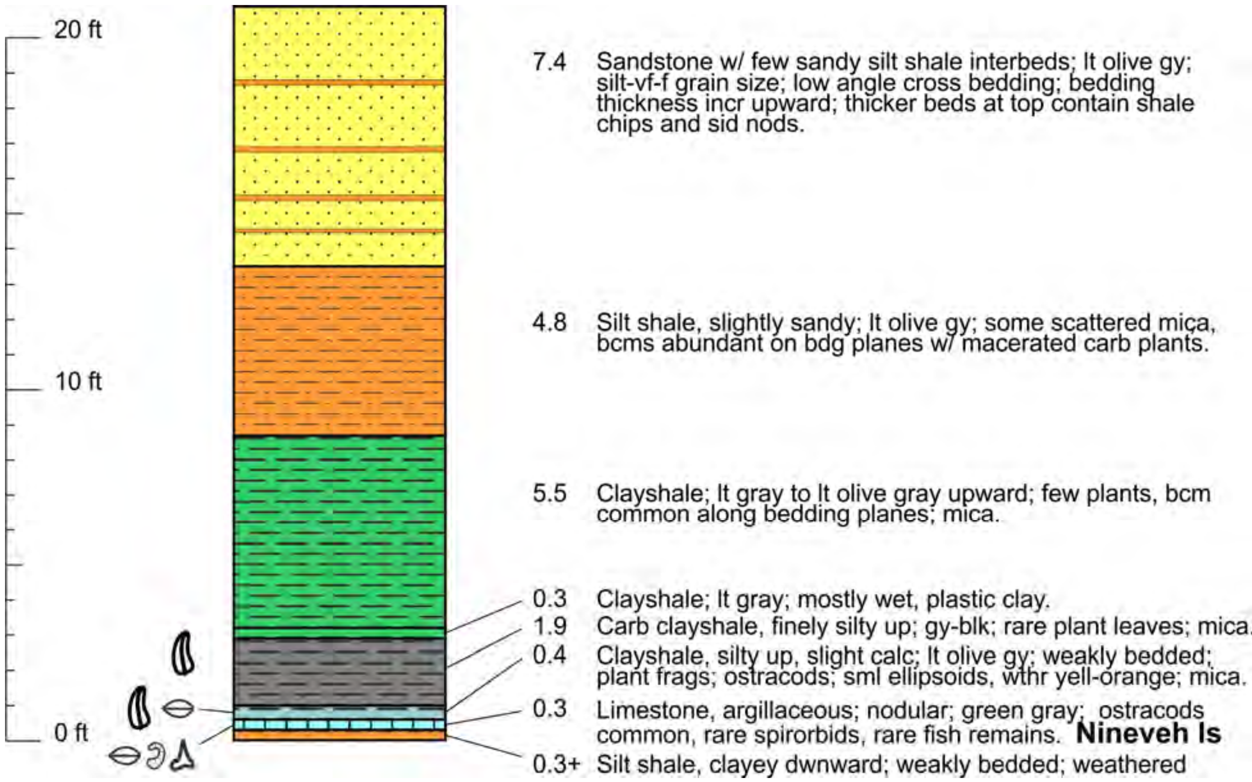

Figure 5. Graphic log of measured section of the lower Greene Formation along Shough Creek Road, about 1 mi (1.6 km) SW of Pleasant Hill Gap. Measured and described by Vik Skema and John Harper.

limestone in the hillside, in place or in float). In his opinion, the beds indeed lie somewhere close to the Nineveh coal horizon (Darrah, 1969, p. 17); he believed they occurred in clays above that coal, but neither the matrix in which the specimens are preserved, nor White's field notes support that assertion. Darrah reports finding two small specimens of fossil conifer (*Lebachia*) in association with the callipterids. This is consistent with the interpretation of callipterid-rich floras as indicators of periods of seasonally dry climate.

Vik Skema and John Harper found and described a better exposure of the same section that includes the Nineveh coal horizon and a thin nonmarine limestone (Figures 5 and 6). The exposure is along Shough Creek Road, a secondary gravel road located ~1 mi (~1.6 km) southwest of White's Pleasant Hill Gap site, and similarly situated on the western slope of the same ridge. The limestone is at about the same elevation as that which White measured at Pleasant Hill Gap. Skema and Harper did not find callipterids.

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Figure 6. Photographs of the Nineveh horizon in the drainage ditch of Shough Creek Road, about 1 mi (1.6 km) SW of Pleasant Hill Gap. A—A carbonaceous clay shale at the horizon of the Nineveh coal. B—The Nineveh limestone.

White, D., 1936, Some features of the Early Permian flora of America: Proceedings of the 16th International Geological Congress (1933), v. 1, p. 679-689.

If I were you, I'd go see a doctor immediately!

TETRAPOD FOSSILS AND THE AGE OF THE UPPER PALEOZOIC DUNKARD GROUP, PENNSYLVANIA-WEST VIRGINIA-OHIO

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The Dunkard Group is a succession of dominantly clastic strata as much as 360 m (1,180 ft) thick that crops out over an elliptical area of ~ 12,800 km^2 (4,940 mi^2) in the central Appalachians--the “tri-state area” of Pennsylvania, Ohio and West Virginia (e.g., Martin, 1998). The Washington and overlying Greene formations make up the Dunkard Group and are mostly sandstone and shale with a few thin beds of coal and freshwater limestone. These Dunkard Group strata are primarily fluvial and lacustrine deposits that contain a diverse fossil record that includes palynomorphs, megafossil plants, insects, ostracodes, conchostracans, nonmarine bivalves and gastropods, fishes and tetrapods (amphibians and reptiles) (see articles in Barlow and Burkhammer, 1975, and see Martin, 1998).

The precise age of the Dunkard Group has been a subject of discussion since the first Dunkard fossil plants were documented in a classic monograph by Fontaine and White (1880). Whether or not the entire Dunkard is Pennsylvanian or Permian, or whether or not the Dunkard encompasses the Pennsylvanian-Permian boundary, have been discussed at length, and a variety of fossils from the Dunkard have been brought to bear on the question. Part of the problem is that the position of the Pennsylvanian-Permian boundary is fixed by marine biostratigraphy (it is now defined by the first appearance of the conodont *Streptognathodus isolatus*: Davydov et al., 1998), and no marine fossils of biostratigraphic significance have been found in the Dunkard Group. This requires what can be called a cross correlation of the biostratigraphy of nonmarine fossils found in the Dunkard with marine biostratigraphy. Various Dunkard fossils (including palynomorphs, megafossil plants, ostracodes, conchostracans and vertebrates) have been interpreted biostratigraphically and my purpose is to summarize what the tetrapod (amphibian and reptile) fossils indicate about the age of the Dunkard Group. The Dunkard tetrapod fossils can be compared to a recently proposed global tetrapod biostratigraphy/ biochronology of the Late Pennsylvanian-Early Permian to conclude that the Washington Formation is very close in age to the Pennsylvanian-Permian boundary (likely Early Permian), and that the Greene Formation is definitely of Early Permian age.

Much of the tetrapod fossil record from the Dunkard Group was derived from collections made by the Carnegie Museum of Natural History during the 1930s (Burke, 1935, 1937), though there were a few earlier published records of Dunkard tetrapods (e.g., Whipple and Case, 1920). Moran (1952) reviewed the stratigraphic context of the then-known tetrapod localities, and Romer (1952) described the tetrapod fossils. Both accepted the then-prevailing conclusion that the entire Dunkard Group is Permian, and Romer (1952) thus compared the Dunkard tetrapods to those from the Lower Permian Wichita Group in Texas. Here, I treat the Washington Formation and Greene Formation tetrapod fossils as two distinct biostratigraphic assemblages, one assemblage from each formation. Note also that the genus is used here as the

Lucas, S. G., 2011, Tetrapod fossils and the age of the Upper Paleozoic Dunkard Group, Pennsylvania-West Virginia-Ohio, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 174-180..

operational taxonomic unit of tetrapod biostratigraphy and biochronology (cf. Lucas, 2006).

Published work on Dunkard tetrapods after Romer (1952), including documentation of newly discovered records, can be found in Beerbower (1963), Stephens (1964), Hlavin (1968), Olson (1970, 1975), Berman (1971, 1978), Berman and Berman (1975), Lund (1975) and Hansen (1996). These authors considered the Dunkard tetrapod fossils to be of Early Permian age; most followed Romer (1952) in stressing the similarity of the Dunkard tetrapods to those of the Lower Permian Wichita Group in Texas. Lund (1975), however, considered the Greene Formation to be correlative with the Leonardian Clear Fork Group in the Texas section.

The tetrapod fossil record of the Dunkard Group (Figure 1) is characteristic of what Olson (1975) perceived of as a Late Pennsylvanian-Early Permian chronofauna of lepospondyl and temnospondyl amphibians, diadectomorphs, primitive amniotes and eupelycosaurs. More specifically, Dunkard amphibians are lepospondyls (the nectridean *Diploceraspis*, the lysorophians *Lysorophus* and *Megamolgophis* and the aïstopod *Phlegethonia*) and temnospondyls (*Broiliellus*, *Edops*, *Eryops*, *Acheloma* [=*Trematops*: Dilkes and Reisz, 1987] and *Trimerorhachis*). The diadectomorph *Diadectes* is present, and the Dunkard taxon "*Limnosceloides*" is based on undiagnostic diadectomorph material (Wideman et al., 2005). The eureptile *Protorothyris* (= *Melanothyris*: Clark and Carroll, 1973) is accompanied by a diversity of eupelycosaurs---"*Baldwinonus*" (a problematic taxon: Lucas et al., 2010), *Ctenospondylus*, *Dimetrodon*, *Edaphosaurus* and *Ophiacodon*.

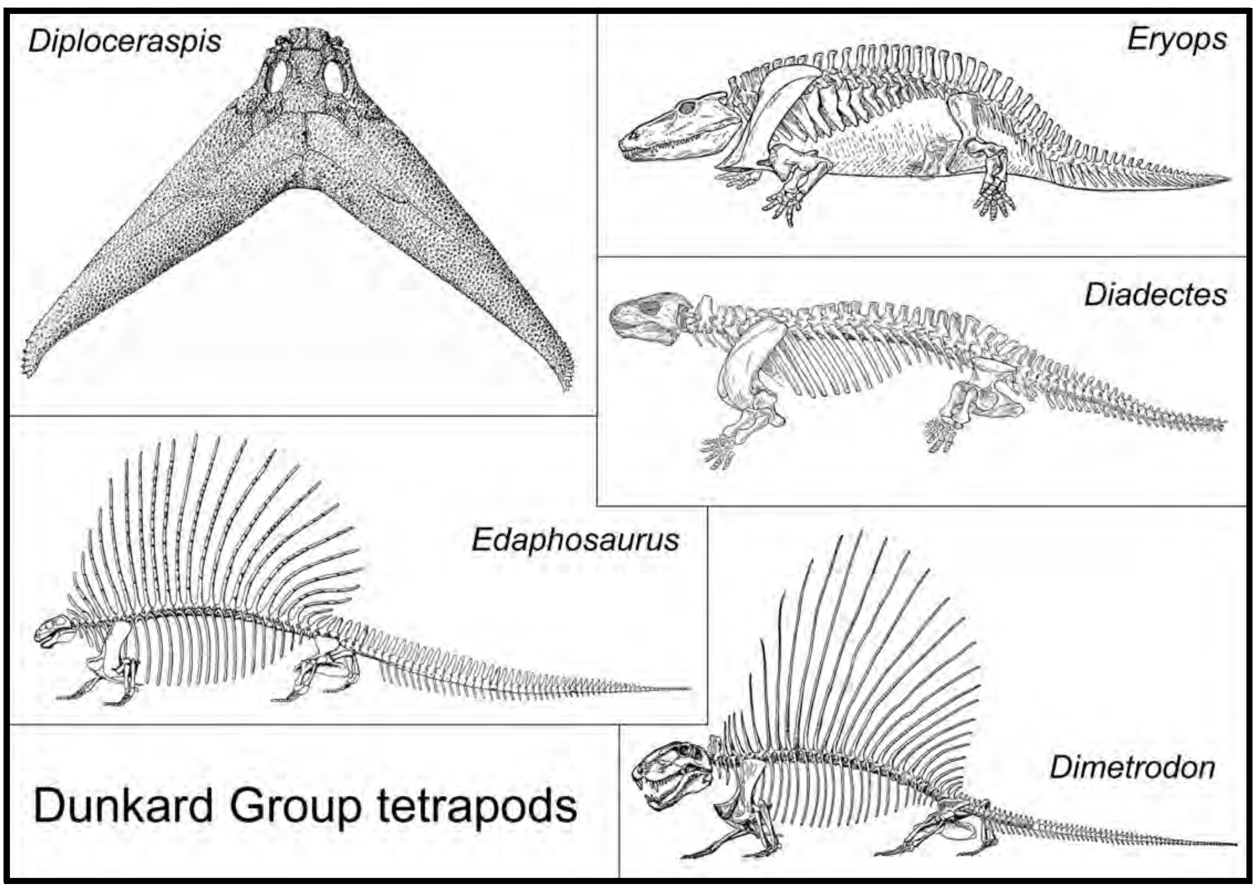

Figure 1. Characteristic Dunkard tetrapods. Skull of *Diploceraspis* (after Beerbower) and skeletons of *Eryops* and *Diadectes* (after Gregory) and of *Edaphosaurus* and *Dimetrodon* (after Romer and Price). Note that Dunkard material of these taxa does not include complete skeletons.

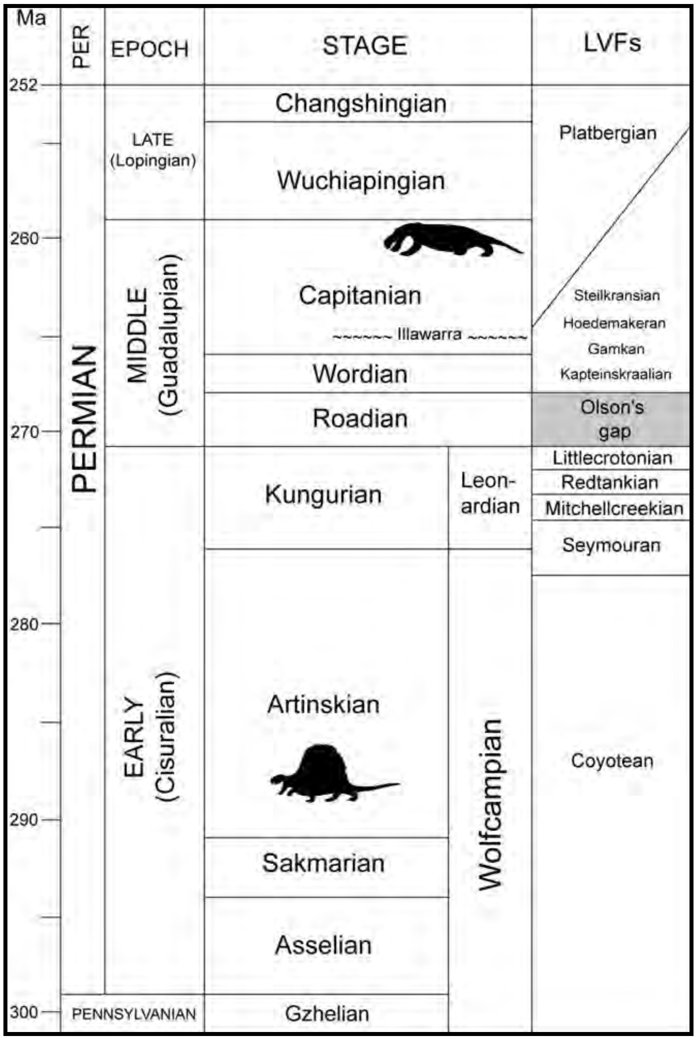

Figure 2. Global Permian tetrapod biochronology and its correlation to the standard global chronostratigraphic scale (after Lucas, 2010). LVF = land-vertebrate faunachron.

Permian tetrapod (amphibian and reptile) fossils have long provided a basis for nonmarine biostratigraphy and biochronology (see reviews by Lucas, 1998, 2002, 2004, 2006). Lucas (2005, 2006) proposed a formal global Permian tetrapod biochronology that recognizes 10 time intervals (land-vertebrate faunachrons: LVFs) (Figure 2). This biochronology is based on the body-fossil record of tetrapods and provides a tetrapod-based timescale that can be used to determine and discuss the temporal relationships of Permian tetrapod assemblages. Parts of it can also be correlated with reasonable precision to the standard global chronostratigraphic scale (SGCS) for the Permian, which is based on marine biostratigraphy (Figure 2).

Substantial fossil records of Permian tetrapods come from the western United States, western Europe, the Russian Urals, northern China and South Africa. The most extensive Lower Permian tetrapod record is from the western United States, especially from Texas, Oklahoma and New Mexico. Tetrapod fossils have been collected from the nonmarine Permian redbeds in north-central Texas since the 1870s, and in northern New Mexico since the 1880s. They were published on extensively by E. D. Cope, E. C. Case, S. W. Williston, A. S. Romer, E. C. Olson and D. S Berman, among others, and provide the basis for most of what is known about the Early Permian evolution of tetrapods. The New Mexican and Texan records were thus used to construct the Early Permian tetrapod biochronology of five LVFs (Figure 2).

The Lower Permian red-bed section in Texas represents fluvial deposition on a broad coastal plain between a Permian seaway to the west and a series of ancestral Rocky Mountain uplifts (Ouachita, Arbuckle and Wichita) to the east and northeast. The nonmarine red beds intertongue with, and are laterally equivalent to, marine strata, allowing cross correlation of nonmarine and marine biostratigraphies. This means it is possible to correlate directly a tetrapod biostratigraphy developed in the Texas red beds with a marine biostratigraphy based largely on fusulinids and ammonoids and for which some conodont data are becoming available (Lucas, 2006). The Texas section thus provides an excellent basis for Early Permian tetrapod biostratigraphy, and this biostratigraphy can be readily correlated to marine biostratigraphy.

Nevertheless, this section has a glaring weakness in lacking an extensive record of tetrapods across the Pennsylvanian-Permian boundary. To remedy this, Lucas (2005, 2006) included the Pennsylvanian-Permian boundary record of tetrapods in northern New Mexico (Rio Arriba County: Berman, 1993; Lucas et al., 2005, 2010) to form a composite standard of New Mexico-Texas for the oldest Permian tetrapod faunachrons. Thus, the New Mexican record superposes tetrapod assemblages that are entirely latest Pennsylvanian, cross the Pennsylvanian-Permian boundary and are of Early Permian age. When combined with the Texas record, the tetrapod succession encompasses the latest Pennsylvanian and the entire Early Permian. This is the standard succession of tetrapod fossils to which the Dunkard Group tetrapods can be correlated (Figs. 2-3).

Most of the Dunkard tetrapod fossils have relatively long temporal ranges from the latest Pennsylvanian (Coyotean LVF) through Early Permian (Seymouran LVF or younger). Tetrapods from the Washington Formation include *Edops* and *Protorothyris*, which are Coyotean index taxa (Lucas, 2006), as well as characteristic Coyotean taxa such as *Eryops*, *Trimerorhachis*, *Diadectes*, *Edaphosaurus* and *Dimetrodon*, and are confidently assigned a Coyotean age. The Greene Formation yields *Trimerorhachis*, *Eryops*, *Edaphosaurus* and *Ctenospondylus*. The eupelycosaur *Ctenospondylus*, reported by Olson (1975) and later documented by Berman (1978) from the Nineveh Limestone in the lower part of the Greene Formation, is an index fossil of the Seymouran LVF (Lucas, 2006). It is also known from the "Belle Plains Formation" (Petrolia Formation) in Texas and the Organ Rock Shale in Utah, both of which are Seymouran age records. This indicates a Seymouran age for the *Ctenospondylus* occurrence in the Greene Formation.

This means that the Greene Formation tetrapods are definitely of Early Permian age, close in age to the Wolfcampian-Leonardian boundary in the North American chronostratigraphic terminology. The Coyotean age that the tetrapods indicate for the Washington Formation is a much

Pennsylvanian	Permian	
	Coyotean	Seymouran
	Washington Formation	Greene Formation
Lysorophus	●	●
Megamolgophis	●	●
Diploceraspis	●	●
Phlegethonia	●	●
Broiliellus	●	
Trimerorhachis	?	●
Acheloma	●	
Edops	●	
Eryops	●	●
Diadectes	●	?
Protorothyris	●	
"Baldwinonus"	●	
Ctenospondylus		●
Edaphosaurus	●	●
Dimetrodon	●	
Ophiacodon	●	

Figure 3. Distribution of tetrapod genera in the Washington and Greene formations of the Dunkard Group and their correlation to the LVFs shown in Figure 2.

less precise age, as the Coyotean encompasses part of the Virgilian (Late Pennsylvanian) through most of the Wolfcampian (Early Permian), so it straddles the Pennsylvanian-Permian boundary (Figure 2). Most of Coyotean time is Early Permian, so this is the most likely age of the Washington Formation tetrapod fossils. However, the possibility of a latest Pennsylvanian age of part or all of the Washington Formation cannot be totally excluded based only on tetrapod biochronology.

Note, though, that the Washington Formation must be younger than early Virgilian, which is the well-established age based on conodont biostratigraphy of the Ames Limestone of the Conemaugh Group (Barrick et al., 2008; Heckel et al., 2008). The Ames is situated ~ 100 m (328 ft) below the base of the Washington Formation. Thus, tetrapod fossils indicate that the base of the Dunkard Group (base of the Washington Formation) is close to the Pennsylvanian-Permian boundary and that the Greene Formation (and probably the entire Dunkard Group) is of Early Permian age (Figure 3).

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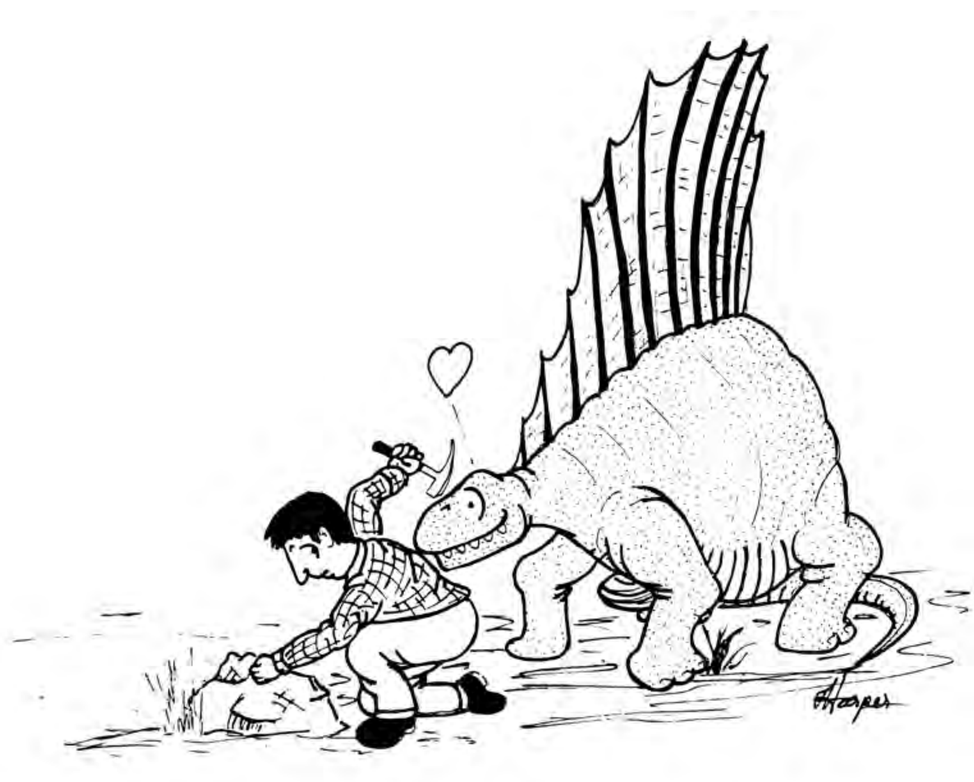

OSTRACODE DISTRIBUTION ACROSS THE PENNSYLVANIAN-PERMIAN BOUNDARY INTERVAL IN THE DUNKARD GROUP OF THE CENTRAL APPALACHIAN BASIN USA

Neil E. Tibert

INTRODUCTION

Ostracoda are bivalved crustaceans that are widely distributed in the nonmarine sedimentary rocks of Pennsylvanian-Permian Age world-wide (Park and Ricketts, 2003). Upper Paleozoic faunas are characterized by populations dominated by the Suborder Darwinulocopina that can be regarded as reliable indicators for Permian age (Molostovskaya, 1979, 1990, 2000; Crasquin-Soleau, 2003). The objective of this report is to summarize the stratigraphic distribution of darwinulocopine ostracodes in Dunkard Group strata in the Appalachian basin (Tibert et al. in press) and argue that the age of the uppermost strata should be regarded as deposits of earliest Permian age. The ostracode distributions presented within are synthesized from the recent work presented in Tibert et al. (in press) and the published reports of ostracodes from Holland (1934), Scott and Summerson (1943), Scott (1944), and Sohn (1975, 1977, 1985).

SAMPLE SITES

The illustrated ostracode material within were collected from the Hundred coal/limestone near Hundred, WV, the Washington coal/limestone (USNM White Collection, USGS 2926) near Browns Mills, WV, and a concretion interval collected from the Windy Gap limestone, Windy Gap, PA (Figure 1). The sampled units from the Dunkard Group include the Waynesburg, Washington, and Greene formations, which contain regionally extensive coal/ limestone intervals that serve as the basis for a general stratigraphic reference herein. DeMichele et al. (2001) reported macroflora that yielded a latest Pennsylvanian-earliest Permian age (Asselian) which is in general agreement with Sohn (1975) who inferred an age near the Carboniferous-Permian boundary. The environment is nonmarine on the basis of abundant red beds with plant fossils and associated freshwater Ostracoda, (Holland 1934; Sohn 1975; DiMichele et al. 2001).

HISTORY OF OSTRACODE RESEARCH IN THE DUNKARD GROUP

Holland (1934) was a pioneer of North American micropaleontology who illustrated Ostracoda from the Nineveh limestone (Greene Formation) in Pennsylvania and West Virginia, including *Whipplella cuneiformis, W. parvula, W. ninevehensis*, and *W. magnitata*. Scott and Summerson (1943) and Scott (1944) followed with studies of nonmarine Pennsylvanian and Permian Ostracoda from Illinois and the Appalachian basins where they revaluated Holland's (1934) type material. The ostracodes reported from these early studies have been widely cited as Permian taxa, likely owing to that age assignment in the Treatise of Invertebrate

Tibert, N. E., 2011, Ostracode distribution across the Pennsylvanian-Permian boundary interval in the Dunkard Group of the central Appalachian basin USA, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 181-189.

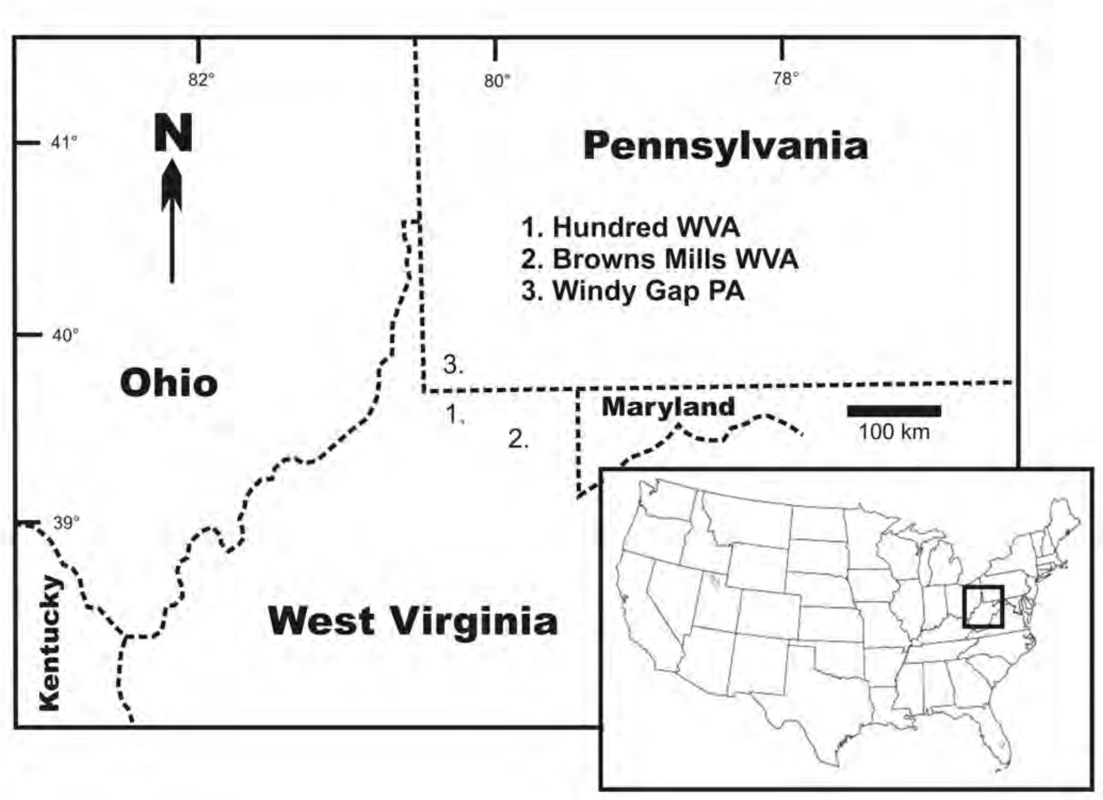

Figure 1. Location of the samples acquired from the Appalachian basin (modified from Eagar and Belt 2003).

Paleontology (Benson et al. 1961). Jones and Clendening (1969) reported *Carbonita* and *Gutschickia* from the Washington and Greene formations and used the orientations of the carapaces to establish northward paleocurrents. Sohn (1975, 1976, 1977, 1985) re-evaluated the nonmarine Ostracoda of the Appalachian basin after Pollard (1966) and Anderson (1970) suggested that all of the central Appalachian nonmarine ostracodes should be placed into synonymy with *Carbonita* from the western European coal measures. Sohn (1975, 1976, 1977, 1985) argued that the distinct adductor muscle scar (AMS) patterns preserved on *Whipplella* and *Darwinula* validated these distinct genera and highlighted their potential biostratigraphic implications. More recently, Sohn and Swain (1999) formally assigned *Whipplella* to the Darwinulocopina on the basis of the AMS.

DISTRIBUTION OF THE METACOPINA AND DARWINULOCOPINA IN THE DUNKARD GROUP

Ostracode Classification

The primary objective of this report is to summarize the distribution of the nonmarine Ostracoda that have been described and published from Dunkard Group localities in the region (Figure 1). Table 1 presents the formal Ostracoda classification system from Tibert et al. (in press). Figure 2 illustrates the patterns of AMS that are diagnostic for suprageneric classification to the Suborder. The Darwinulocopina have AMS patterns of 11-12 spots (Sohn,

Table 1. Ostracode classification (Tibert et al. in press).

Subclass **OSTRACODA** Latreille 1806
- Order **PODOCOPIDA** Müller 1894
  - Suborder **DARWINULOCOPINA** Sohn 1988
    - Superfamily **DARWINULOIDEA** Brady and Norman 1889
      - Family **PALEODARWINULIDAE** Molostovskaya 1990
        - Genus ***Paleodarwinula*** Molostovskaya 1990
    - Superfamily **DARWINULOIDOIDEA** Molostovskaya 1979
      - Family **DARWINULOIDIDAE** Molostovskaya 1979
        - Genus ***Whipplella*** Holland, 1934
- Order **PLATYCOPIDA** Sars 1866
  - Suborder **METACOPINA** Sylvester-Bradley 1961
    - Superfamily **HEALDIOIDEA** Harlton 1933
      - Family **CARBONITIDAE** Sohn 1985
        - Genus ***Gutschickia*** Scott 1944
        - Genus ***Hilboldtina*** Scott and Summerson 1943

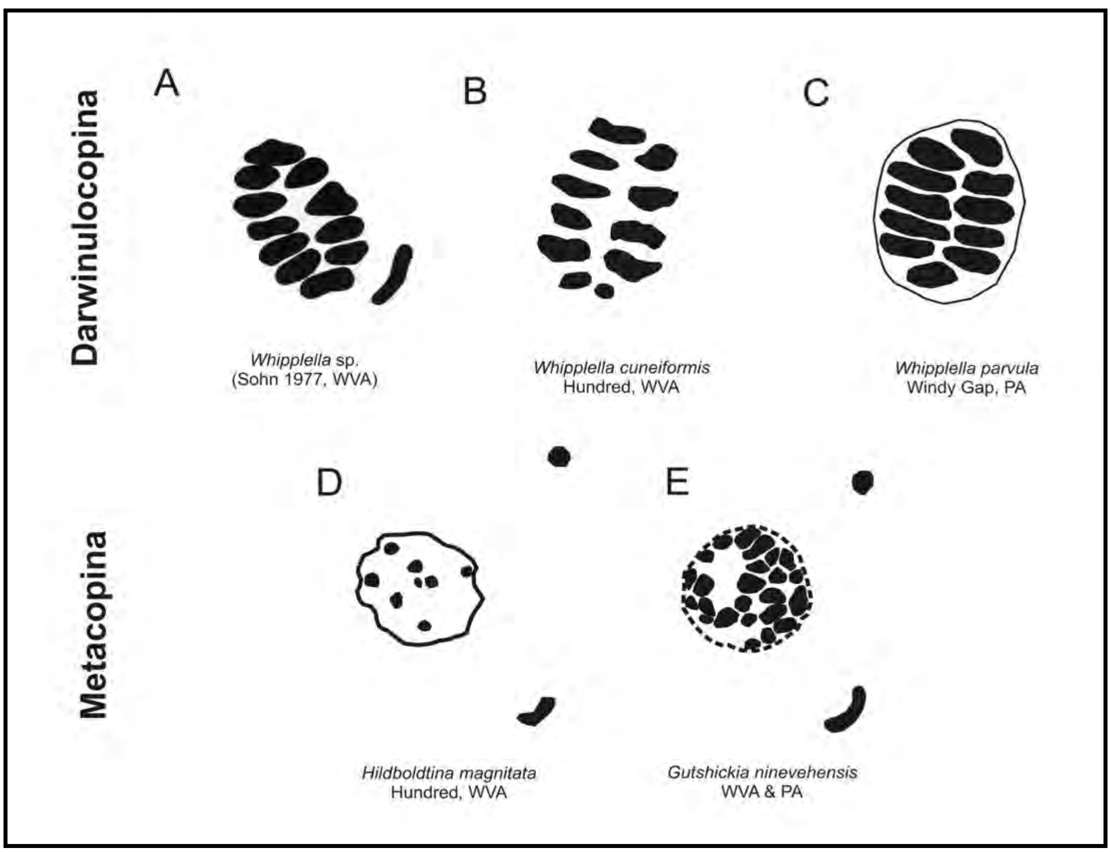

Figure 2. Line tracings of the Adductor Muscle Scars (AMS) for the material described and discussed in this report. A to C – Darwinulocopina - *Whipplella* spp. D and E – Metacopina - *Gutschickia* and *Hilboldtina*.

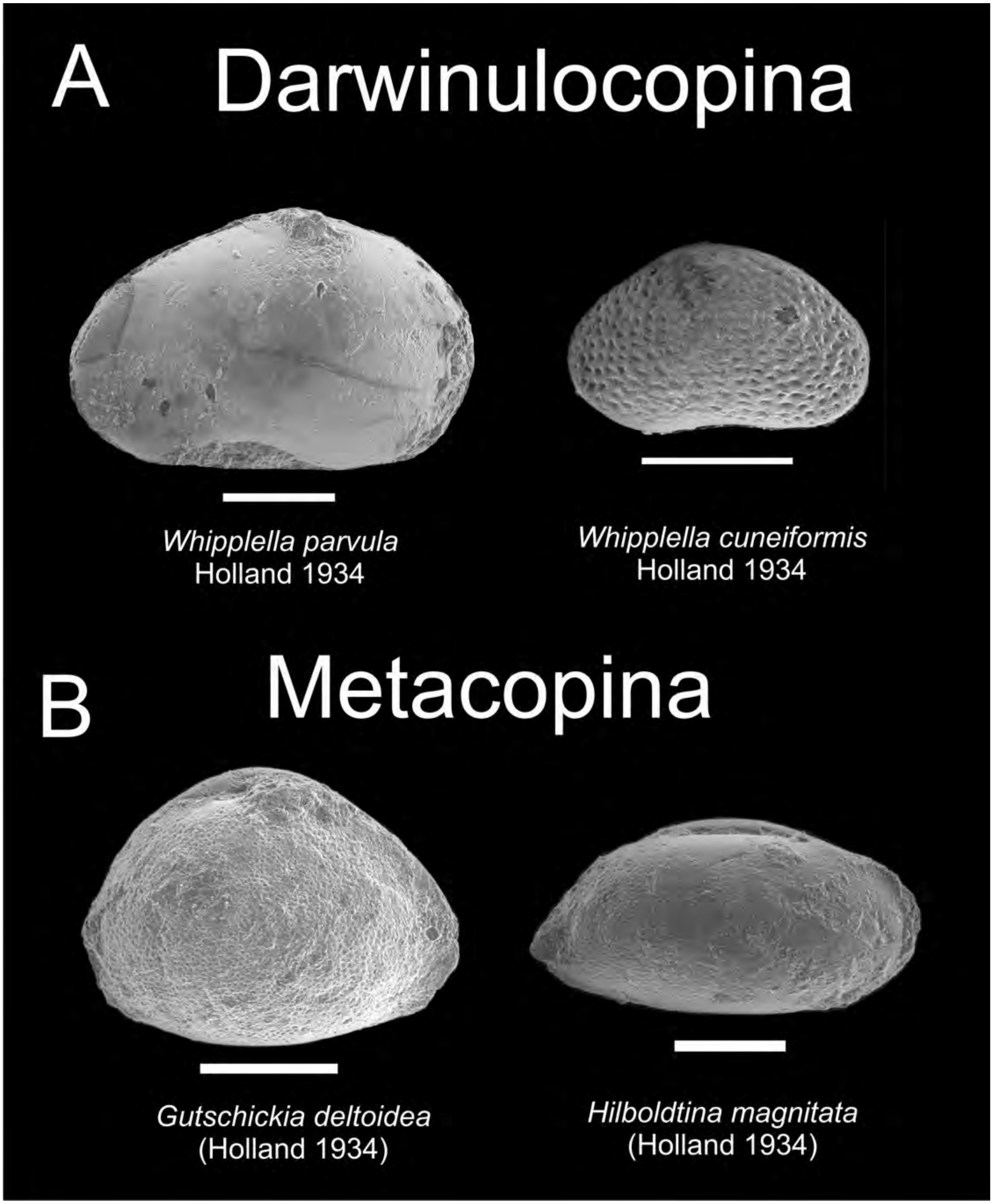

Figure 3. Ostracoda from the Dunkard Group in West Virginia. A – Darwinulocopina - *Whipplella parvula* and *W. cuneiformis*. B – Metacopina - *Gutschickia deltoidea* and *Hilboldtina magnitata*. Barscale = 100um

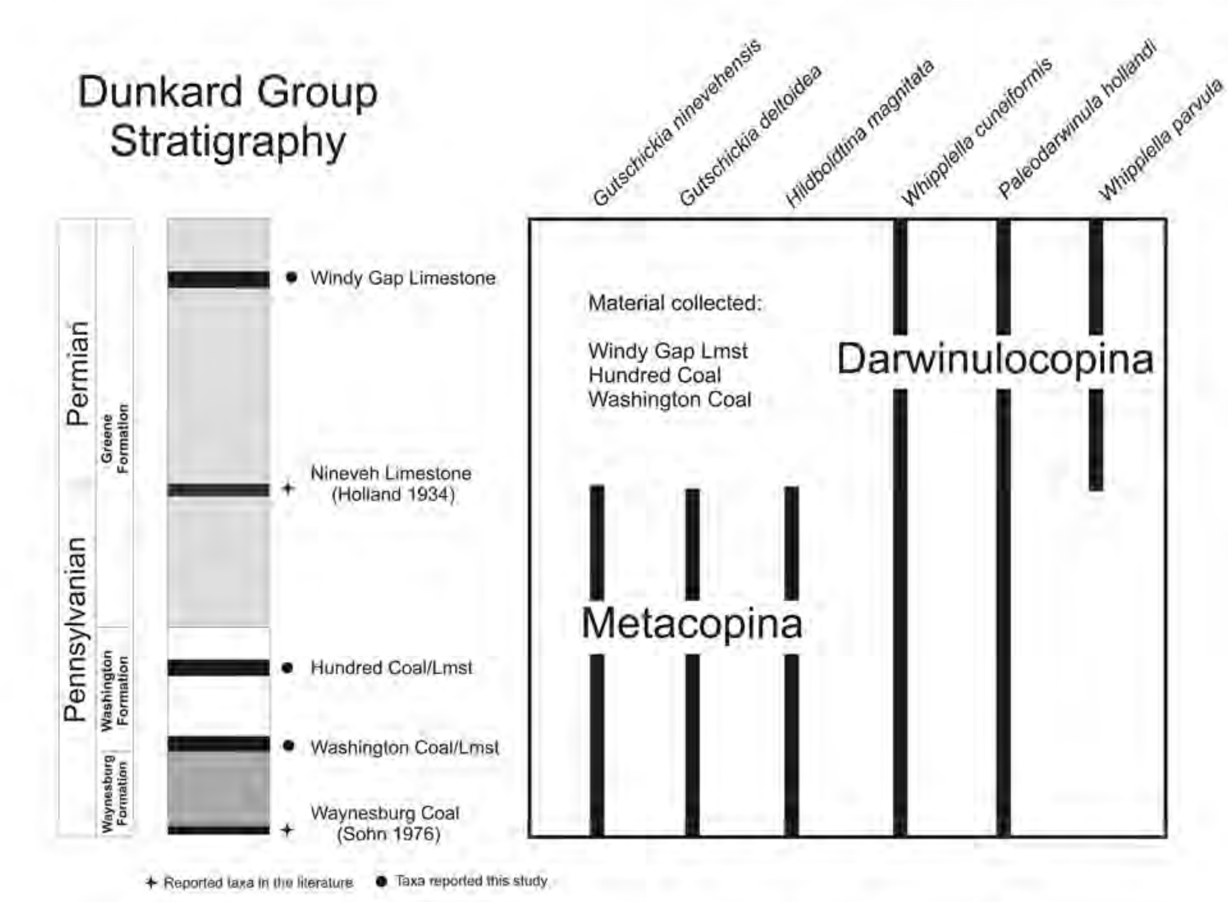

Figure 4. Biostratigraphic distribution of the nonmarine Ostracoda from the Dunkard Group. Samples were collected by Blaine Cecil, Bill DiMichele, and Vik Skema. The stratigraphic reference section is modified from Nick Fedorko (personal communication 2011).

1988) organized into a distinct rosette array of 2 serial rows (Figures 2A-C). In contrast, the Metacopina comprise an aggregate of numerous spots (~20-30) organized in an oblique circular cluster (modified Healdiidae of Sohn 1977); distinct frontal and mandibular scars are commonly visible (Figures 2D-E). The ostracodes from the Dunkard Group (Figure 3) are presented on the basis of their first occurrences relative to the base of the composite stratigraphic section constructed from core logs produced by the WV Geological and Economic Survey (Nick Fedorko, personal communication)(Figure 4).

Stratigraphic Distribution

Waynesburg Formation

Sohn (1976) illustrated well preserved specimens of *Paleodarwinula hollandi* (Holland 1934) from the Waynesburg coal (Figure 4). The illustrated material exhibits a darwinulid biserial rosette AMS, which confirms its assignment to the suborder Darwinulocopina (Tibert and Dewey, 2006; Tibert et al., in press).

Washington Formation

Samples collected from Browns Mills and Hundred contain well preserved specimens of *Gutschickia ninevehensis*, *Gutschickia deltoidea*, and *Hilboldtina magnitata* (Figures 3B, 4). The AMS pattern comprises an oblique central aggregate with frontal scars which confirms an assignment to the Suborder Metacopina (Tibert and Dewey, 2006; Tibert et al., in press). The samples also contain well preserved specimens of *Whipplella cuneiformis* (Figure 3A) characterized by a pitted surface and a biserial array AMS (Figure 2B) diagnostic for the species (Tibert et al., in press).

Greene Formation

Samples from the Windy Gap limestone contain well preserved specimens of *Whipplella cuneiformis and Whipplella parvula* (Figures 3A, 4). Poorly preserved specimens of *Paleodarwinula hollandi* have also been observed. The AMS pattern for the larger, smooth *W. parvula* comprises a biserial array AMS (Figure 2C) diagnostic for the species (Tibert et al., in press). The Nineveh limestone marks the last occurrence of the Metacopina, as documented in Holland (1934), with an exclusive population of the Darwinulocopina persisting to the uppermost unit of the Greene Formation.

DISCUSSION/CONCLUSION

Global Distribution Darwinulocopina

Paleodarwinula and *Whipplella* have been observed in nonmarine deposits in the Wolfcampian Speiser Shale in Kansas (Retrum and Kaesler, 2005). Tibert et al. (in press) recently placed several species of the Kansas ostracodes assigned to *Carbonita* (Kellett, 1935; Retrum & Kaesler, 2005) into the synonymies of *Whipplella cuneiformis*, *W. parvula*, and *Paleodarwinula hollandi*. Sohn's (1977, 1985) efforts to carefully illustrate the AMS for *Whipplella* has confirmed its affinity with the Darwinulocopina and formal placement within the Family Darwinuloididae (Sohn and Swain, 1999). This family is widely distributed in Permian strata on the Russian Platform (Molostovskaya, 2000; Crasquin-Soleau, 2003).

The transition from marine to exclusively fresh water for the Ostracoda presumably occurred during Mississippian-Pennsylvanian times when the Podocopida and Platycopida radiated into the nonmarine waters of the Gondwanan and Laurasian coasts (Tibert and Scott, 1999; Williams et al. 2006). At the end of the Pennsylvanian, the exclusively nonmarine habitats were becoming increasingly competitive and those taxa with pre-adapted feeding and reproductive strategies survived (Horne, 2003). The Darwinulocopina are known for their precocious sexual reproductive strategies, brood rearing of their juveniles, and parthenogenetic reproduction (Lethiers et al., 1996). Specimens of *Whipplella carbonaria* Scott 1944 reported from the North Massif Central in France (Autunian Stage) (Lethiers et al, 1996) have been illustrated with juveniles occupying the posterior brood chambers of adults. Finally, the Permian deposits in North America, Central Europe, and Russia are apparently barren of the Metacopina brackish-to-freshwater taxa (e.g., *Gutschickia* and *Hilboldtina*). Given that the Metacopina are a marine lineage, it is possible that the low salinity tolerant representatives of the Upper Paleozoic apparently were unable to sustain breeding populations in the evolving terrestrial waters of the Permian supercontinent (Tibert and Dewey, 2006).

Location of the Pennsylvanian-Permian Boundary

The nonmarine coal measures of the Dunkard Group can be categorized into distinct faunas on the basis of their associated series/stage assignments (Tibert et al., in press)(Figure 4). Ostracodes recovered from the Waynesburg, Washington, and lowermost Greene formations should be regarded as transitional fauna for the Pennsylvanian-Permian boundary interval, comprising a mixed assemblage of Metacopina and Darwinulocopina (*Hilboldtina-Whipplella cuneiformis* Fauna of Tibert at al. in press) of uncertain age, ranging from latest Stephanian-to earliest Asselian. Ostracodes recovered from the uppermost Greene Formation (above the Nineveh limestone) should be regarded as a nonmarine fauna of likely Asselian age (*Whipplella parvula* Fauna of Tibert et al., in press). I therefore place the Pennsylvanian-Permian boundary at the approximate position of the Nineveh limestone of Holland (1934), which marks the Last Occurrence of *Hilboldtina magnitata*, *Gutschickia ninevehensis*, and *G. deltoidea* (Figure 4). This stratigraphic position is subject to change pending a more detailed sampling of the Dunkard Group.

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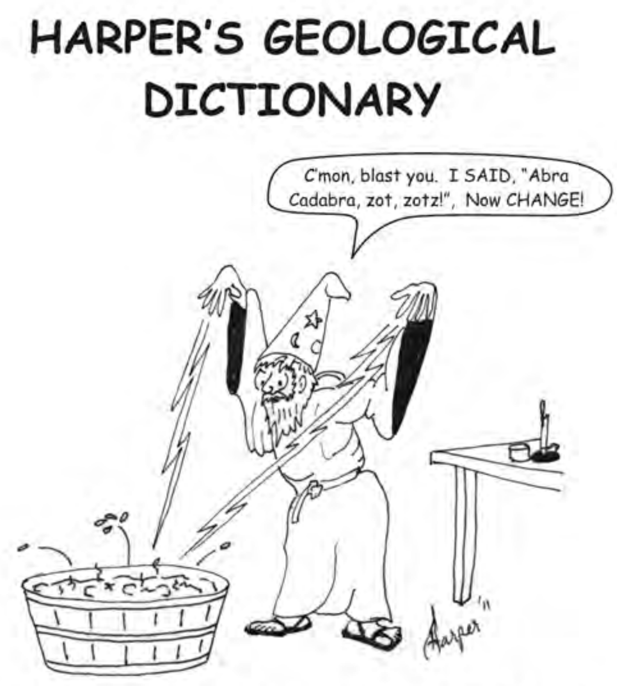

OROGENY - [Sp. Oro - gold + L. Genesis - to be born] The transmutation of base materials to precious metals by means of alchemical procedures.

PERMIAN OUTLIERS IN WESTERN KENTUCKY

W. John Nelson, Scott Elrick, and David A. Williams

A down-faulted outlier of Permian rock has been documented in western Kentucky. Discovered during geologic mapping, the outlier was confirmed by core drilling, which yielded Permian fusulinids. Published findings remain descriptive, lacking interpretation. This is the only confirmed occurrence of Permian sedimentary rocks in the Illinois (Eastern Interior) basin. However, a second graben that probably contains Permian rocks has been identified from oil-test drilling. The Kentucky outliers lie nearly equidistant between the Permian of Kansas and the Dunkard Basin, approximately 645 km (400 mi) from both, and are apparently transitional between the two locations. (Figure 1**)**

BACKGROUND

Kehn (1975) mapped a graben near Cap Mauzy Lake in Union County, western Kentucky where as much as 762 m (2,500 ft) of Paleozoic strata overlie the Springfield (No. 9) Coal. Following up on Kehn's find, the Kentucky Geological Survey drilled a continuously cored test

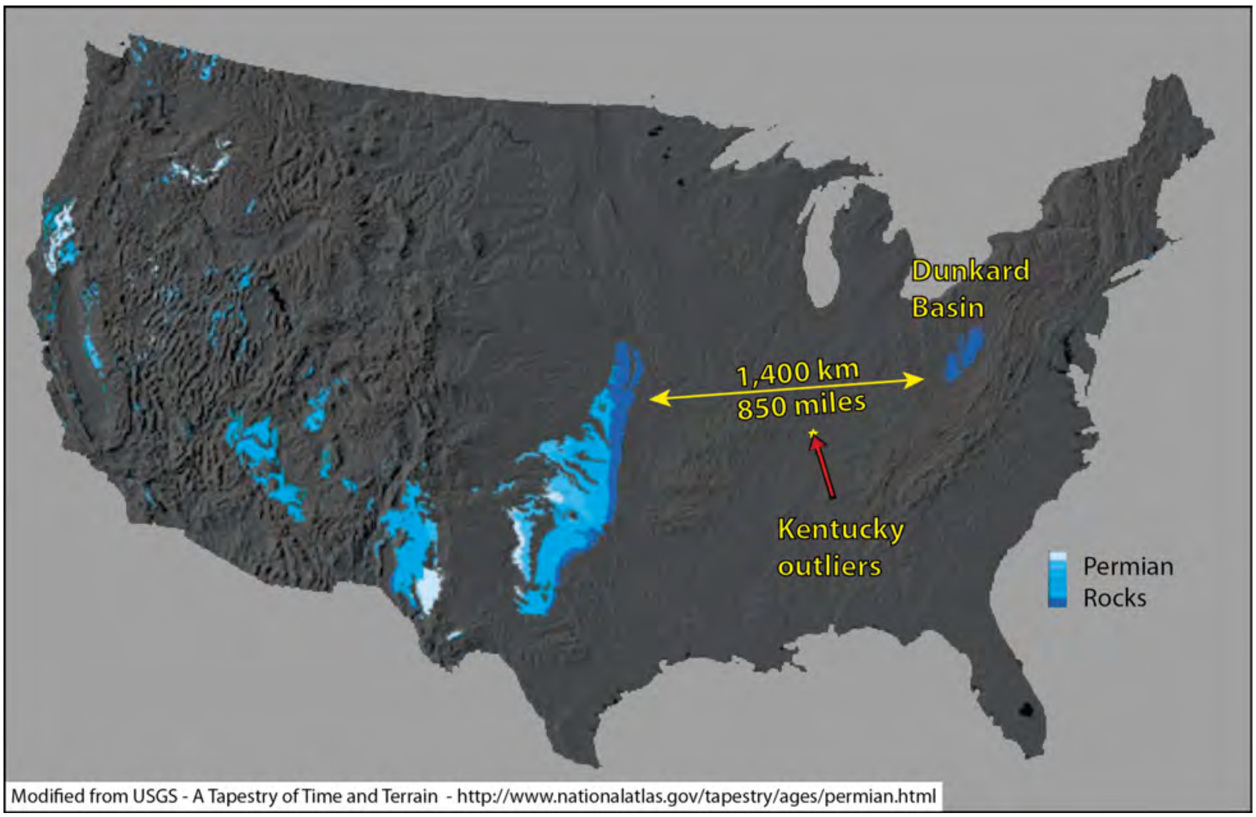

Figure 1 - Map of the United States showing areas of Permian rocks. The Kentucky outliers are midway between the mid-continent Permian and the Dunkard Basin. Source - Modified from USGS - A Tapestry of Time and Terrain - http://www.nationalatlas.gov/tapestry/ages/permian.html

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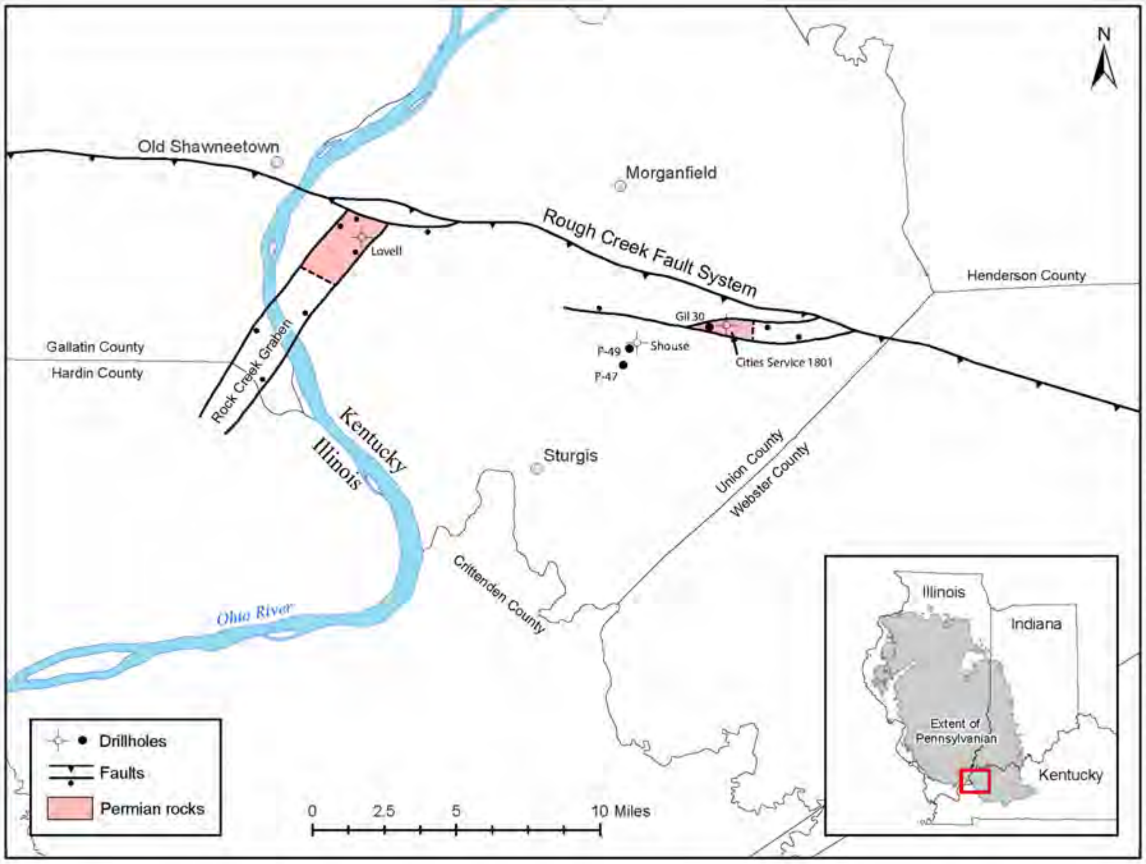

Figure 2 - Map showing locations of Permian outliers in Western Kentucky with key boreholes and selected faults. As shown, both outliers reside within deeply down dropped grabens.

hole, Gil 30, to a depth of 561 m (1,841 ft) (see map, Figure 2). The log of Gil 30 was published by Williams et al. (1982). In February 2011 we redescribed the Gil 30 core; a graphic log is reproduced here (Figure 3).

Three boreholes close to Gil 30 provide significant stratigraphic control. These are Peabody boreholes P47 and P49, which were continuously cored, plus an electric log from the adjacent Shouse oil test hole (Figure 2). Descriptions of the two Peabody cores were published by Smith and Smith (1967). These three boreholes provide depth intervals and stratigraphic context to regionally mapped marker beds in the Carbondale Formation, which the Gil 30 core did not reach.

AGE

Fusulinids of genus *Triticites,* indicating Wolfcampian age, were recovered from limestone at a depth of 59 m (195 ft) in the Gil 30 core (Kehn et al. 1982). Douglass (1987, p. 12) assigned these to *Triticites beardi,* a new species, and stated, "This form represents a developmental stage similar to forms described from rocks of Early Permian (Wolfcampian) age."

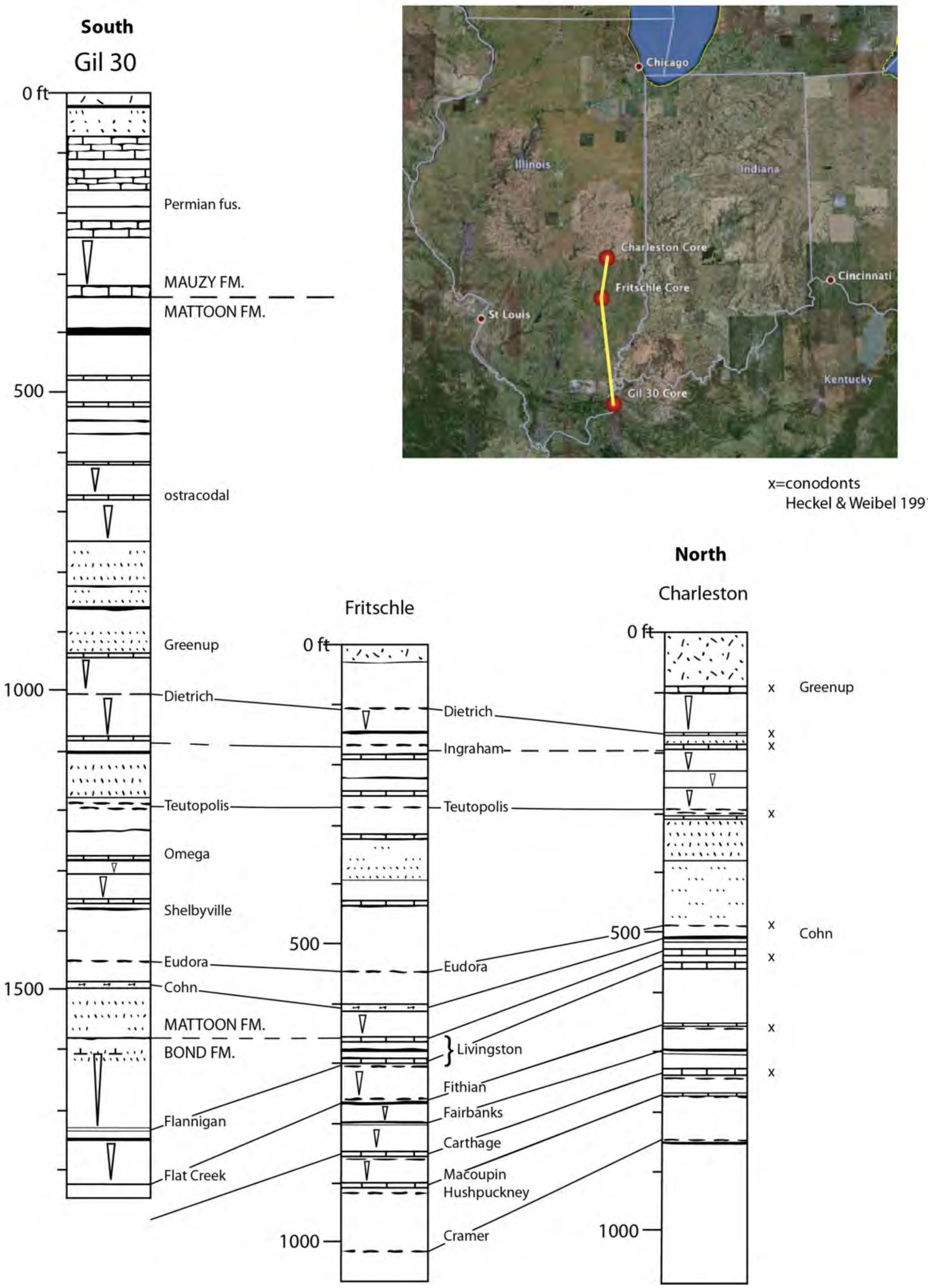

Figure 3 - Generalized graphic log of the Gil 30 core and its correlation to two cores through the youngest Pennsylvanian rocks preserved to the north in Illinois. Conodonts from the Charleston core (Heckel and Weibel 1991) provide regional biostratigraphic correlation.

Fossil spores from coal bed were examined by Peppers (1996, p. 82), who wrote, “Several coal beds at a depth of 390 to 540 feet contain spore assemblages characterized by the abundance of *Thymospora thiessenii*...characteristic of the Pittsburgh coal bed and several other coal beds near the base of the Conemaugh [*sic, should read Monongahela*] Formation in the northern part of the Appalachian region.” These coal beds are Virgilian in age and bracket the Pennsylvanian-Permian boundary between depths of 59 and 119 m (195 and 390 ft) in Gil 30.

MAUZY FORMATION LITHOLOGY

Kehn et al. (1982) assigned rocks above 104-m (340-ft) depth in Gil 30 to a new unit, the Mauzy Formation, based on lithologic differences between these and the older strata. The most significant difference is that the Mauzy contains a much higher proportion of limestone and calcareous mudstone than the underlying Mattoon (formerly called Sturgis) Formation (Kehn 1973). Rocks of the Mauzy may be assigned to three lithofacies.

Nonmarine Limestone and Mudstone

Nearly half the total 91+ m (300+ ft) of the Mauzy Formation consists of interbedded limestone and mudstone believed to be of nonmarine origin. Limestone is light to medium brownish-gray, microgranular, dolomitic, and massive to nodular. The only fossils are rare ostracods and burrows of unknown origin (Figure 4). Some beds display intraformational conglomerate of rounded carbonate clasts; other beds show brecciation that presumably took place while the sediments were soft. Limestone is intercalated and intergrades with mudstone that is greenish to olive gray, calcareous, and massive to weakly fissile. Slickensides, hackly fracture, and irregular carbonate nodules are present. Some mudstone layers contain fine sand and mica flakes.

Marine and Deltaic Strata

Three intervals of the Mauzy in Gil 30 show evidence of marine and deltaic deposition. Together these intervals account for about 41% of the formation thickness. The youngest marine/deltaic interval extends from 48 to 59.6 m (157.5 to 195.6 ft). At the base is limestone, a skeletal wackestone that contains echinoderm fragments, brachiopods, and fusulinids that indicate Permian age. Overlying the limestone is shale

Figure 4 - Photograph of slabbed core from the Gil 30 core at a depth of 85 feet. The limestone is mostly light to medium brownish gray, microgranular, dolomitic, and can range from massive to nodular. Brecciated textures are prevalent.

Figure 5 - Bedding surface of core from Gil 30 showing root traces (below) and bright coal stringer (above). Most of the coal was previously removed for analysis. This was the only carbonaceous interval within the Mauzy, and was found near the base.

that grades upward to interlaminated siltstone and sandstone. Planar lamination below gives way to ripple-cross lamination above. Articulate brachiopods, bivalves, ostracods, and plant fragments are present.

The second marine/deltaic interval directly underlies the first and is about 5.5 m (18 ft) thick. This interval consists of medium to dark gray silty shale and siltstone with laminae of light gray sandstone. Fossils include bivalves, ostracods, a bryozoan, and a rugose coral. Rhythmic lamination in the lower part of the interval suggests tidal activity.

The third and oldest marine/deltaic interval extends from 75.7 to 98.7 m (248.5 to 323.9 ft) depths and is about 23 m (75 ft) thick. At the base is hard calcareous shale to impure limestone that contains chonetid brachiopods. The remainder of this interval grades from clay-shale at the base through silty shale to siltstone in the upper part. Lamination is planar to wavy and lenticular. Bivalves, articulate brachiopods, and a singular goniatite were observed in the lower portion.

Channel Sandstone

The uppermost 11 m (36 ft) of Gil 30 is an upward-fining interval of siltstone to fine-grained micaceous sandstone (litharenite). These rocks exhibit planar to wavy lamination and small-scale soft-sediment deformation. Shale rip-up clasts are common about 3 m (10 ft) above the sharp lower contact.

Coal

The Mauzy in Gil 30 contains a single layer of dull, shaly coal less than 2.5 cm (1 in) thick at the depth of 99 m (324 ft). This layer directly underlies the lowest of three marine/deltaic intervals (Figure 5). Shale below the thin coal lacks strong penetrative rooting and other paleosol features, suggesting the coal may represent either transported plant material or swamp development on top of freshly exposed substrate.

GROVE CENTER OUTLIER

A second graben that probably contains Permian rocks occurs in the Grove Center quadrangle, about 18 km (11 mi) west of the Gil 30 core (Figure 2) in the Rock Creek graben. Logs of two boreholes indicate the Mauzy Formation is thicker here than in the Gil 30 core. The electric log of the John N. Partin # 1 Lovell oil-test hole indicates the base of the thick Mauzy limestones at 180 m (590 ft), and a minimum thickness of 155 m (508 ft) of Mauzy was drilled (Figure 6). The driller's log of a nearby coal-test boring, Donan Engineering # C-2,

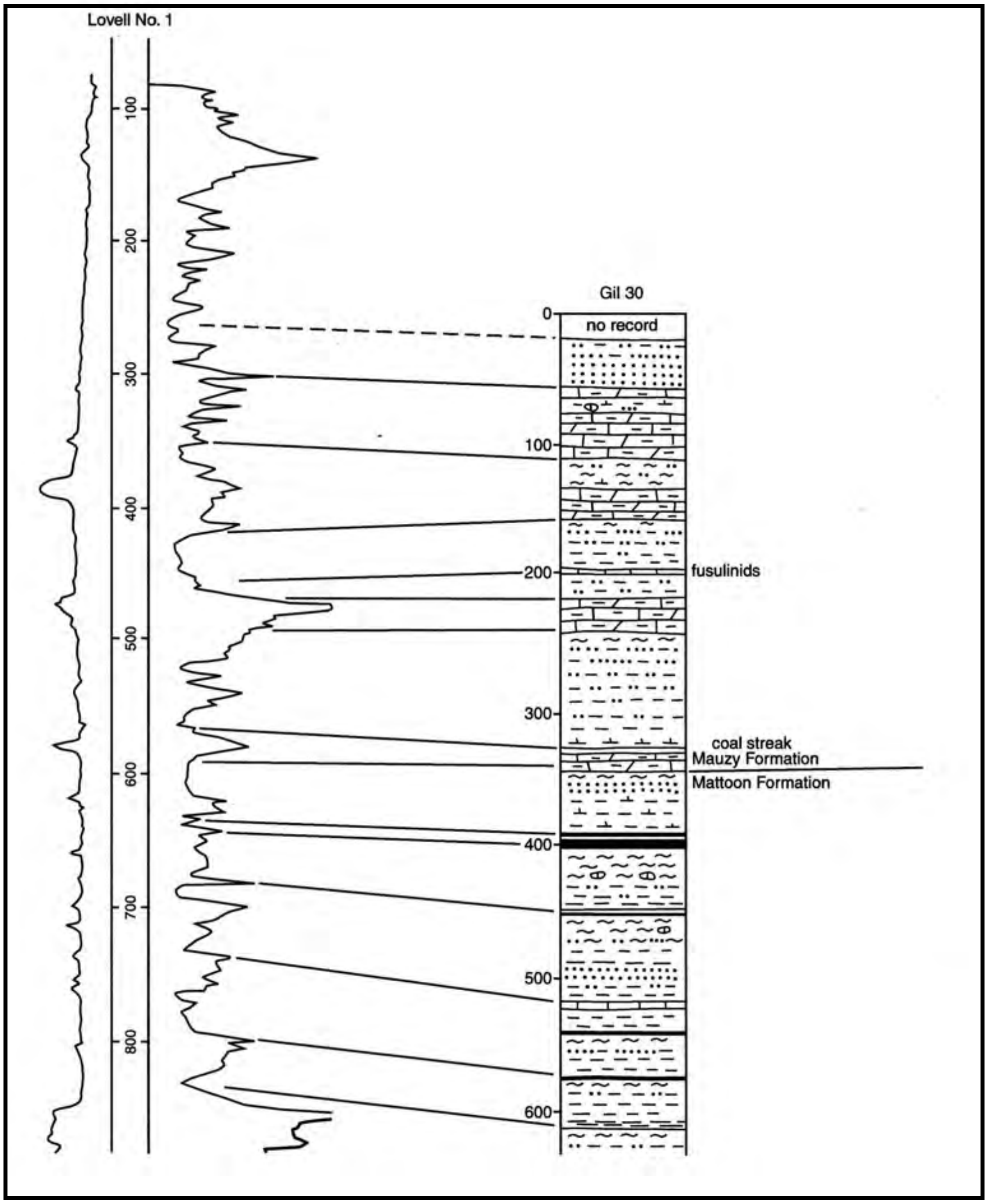

Figure 6 - Comparison of the Gil 30 core to electric log from the Lovell #1 oil-test hole in the Grove Center outlier (see Figure 2 for locations). Approximately 590 feet of Mauzy Formation is present in Lovell #1 as compared to 321 feet of Mauzy in the Gil 30 core.

indicates the Mauzy is 138 m (454 ft) thick. This compares to 113 m (321 ft) of Mauzy in the Gil 30 core.

ARE THESE SYNTECTONIC GRABENS?

Preservation of the Kentucky Permian in two small down-faulted blocks raises the possibility of this Permian deposition being confined to actively sinking tectonic grabens.

Surprisingly little deformation is evident in the Gil 30 core, considering the hole was

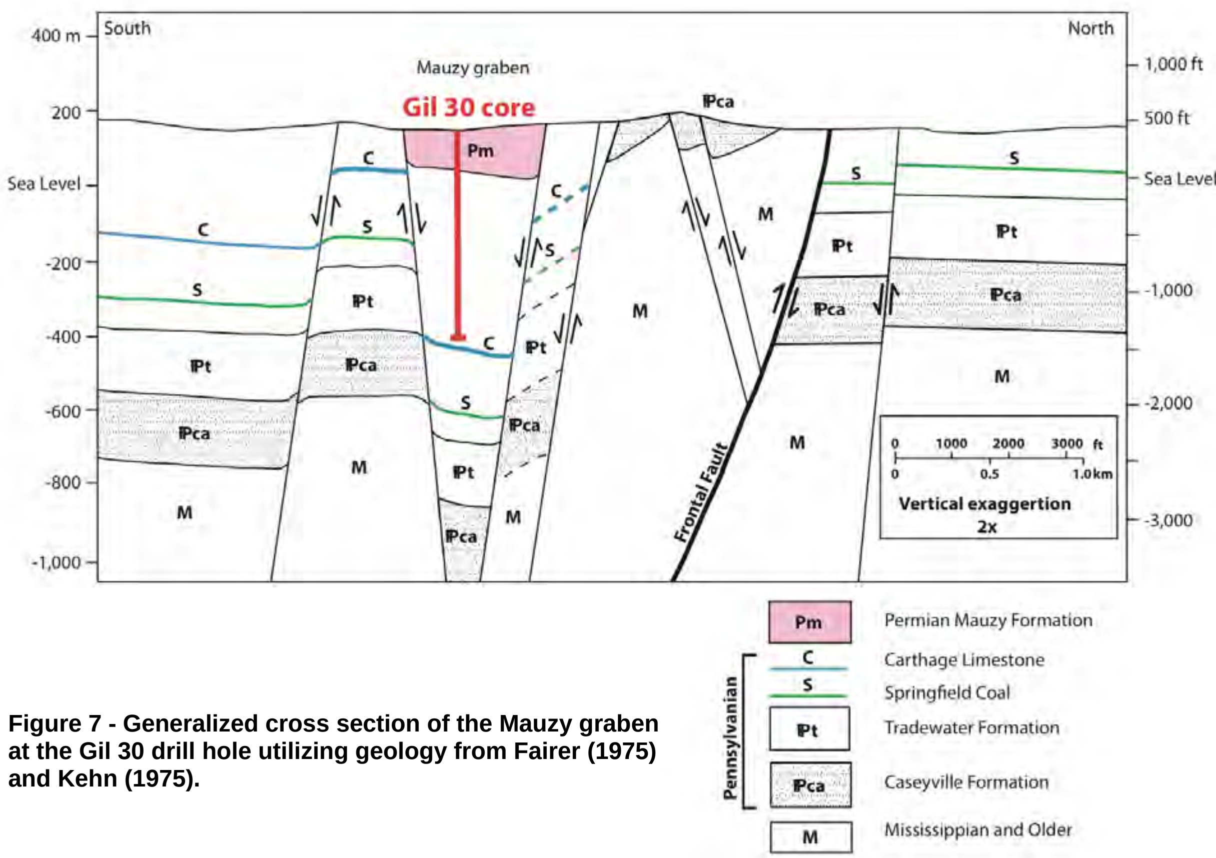

Figure 7 - Generalized cross section of the Mauzy graben at the Gil 30 drill hole utilizing geology from Fairer (1975) and Kehn (1975).

drilled 122 m (400 ft) from a fault that has 457 m (1500 ft) of throw (Figure 7**)**. All cored strata are horizontal or nearly so, and contain only a moderate amount of fractures. Shearing parallel to bedding is apparent in some of the shales, but there is no indications of missing or repeated section. More to the point, there is little of the soft-sediment deformation to be expected were faults active during Mauzy sediment deposition. Moreover, the Mauzy sediments are dominantly fine-grained, lacking conglomerates and breccias that are earmarks of deposition near active faults.

Additional evidence that suggests deposition on a landscape unaffected by concurrent tectonism is that Upper Mattoon and Mauzy strata correlate closely between Gil 30 and the Lovell electric log, in separate grabens 18 km (11 mi) apart.

We conclude that the Permian outliers in Kentucky are fortuitously preserved remnants of deposits that once covered a much larger area. The grabens in which these rocks lie today formed after Mauzy sediments were lithified.

COMPARISON WITH OTHER AREAS

The Mauzy Formation is quite similar to (roughly) coeval rocks in the northern Dunkard Basin, but bears little resemblance to uppermost Pennsylvanian and Lower Permian rocks of Kansas.

In the northern Dunkard Basin, the Monongahela and lower part of the Dunkard Group comprise interbedded limestone, non-fissile mudstone, laminated siltstone, crossbedded sandstone, and coal. Limestone is micritic, argillaceous and silty, and bedded to nodular. It is commonly interbedded with gray to green, non-fissile mudstone in intervals as thick as 15 m (50 ft). These contain dominantly non-marine fossils including ostracods, molluscs, and vertebrate remains. The nearest approach to marine conditions is found in brackish-water *Lingula*, orbiculoid brachiopods, and gastropods in the Elm Grove limestone and Washington coal zone. Whereas coal beds of the Monogahela include the Pittsburgh and others of commercial importance, those of the Dunkard are generally thin, lenticular, and shaly. Clastic sediments came primarily from ancestral Appalachian highlands to the southeast, with a secondary northern source off the Canadian Shield. Depositional setting was a deltaic plain with extensive fresh-water lakes in which limestone precipitated (Martin 1998).

Virgilian and Lower Permian rocks of Kansas display cyclic alternation of marine limestone and stacked calcic and vertic paleosols. Although sandstone and thin coal layers are common in the lower part of the Virgilian, these rock types disappear in the upper Virgilian (Zeller 1968). Paleosol profiles indicate a semi-arid to subhumid, strongly seasonal wet-dry (monsoonal) climate regime (Miller 1994, Archer et al. 1995).

We interpret western Kentucky as having been on the same overall deltaic plain as the northern Dunkard Basin, but in a more seaward position. The area was likely part of a large bay that opened into the epicontinental sea of Kansas (Figure 8). Hence, western Kentucky was submerged during the larger marine transgressions, receiving subtidal limestone and offshore, prodelta sediments. The more northward position of Kentucky during the Pennsylvanian and Permian, may have placed this location in an overall drier climate belt than the Dunkard Basin, therefore making the area less conducive to peat accumulation during the relatively wetter climate phases, and more conducive to limestone precipitation during drier climate phases.

IDEAS FOR FURTHER STUDY

The western Kentucky Permian outliers are the fortuitous remnants of a much larger and extensive deposit, and represents our only stepping-stone strata for potentially connecting the Dunkard Basin and Midcontinent Permian rocks. So far, precious little work has been done on these rocks, and much remains. Suggestions for further work include:

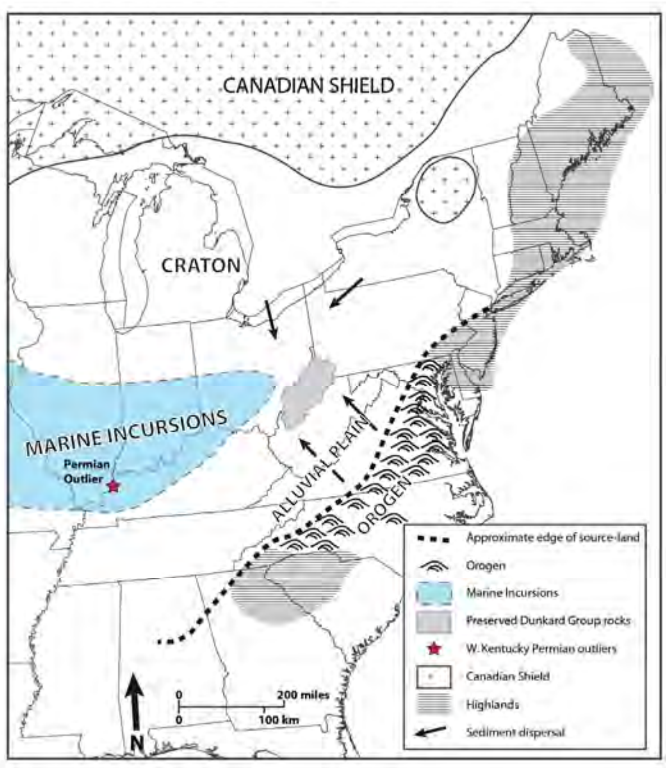

Figure 8 - Interpreted paleogeographic map, modified from Martin (1998), showing the relative landscape position of the western Kentucky Permian outliers relative to the Dunkard Basin. The western Kentucky area was likely part of a large bay that opened into the epicontinental sea of Kansas. During major marine transgressions, subtidal limestone and offshore prodelta sediments were deposited in Kentucky, but not in the Dunkard Basin.

- Drill a core in the Grove Center outlier (Rock Creek graben) perhaps near the Lovell #1 test.
- Examine outcrops of Permian outliers. Geologic maps of Kehn (1975) and Palmer (1976) indicate that outcrops exist. These might yield fossils and bulk samples, sparing precious core.
- Petrographic study of carbonate rocks.
- Further palynology on coal and carbonaceous shale.
- A specialist should examine the paleosols.
- Further microfossil study: ostracods, conodonts.

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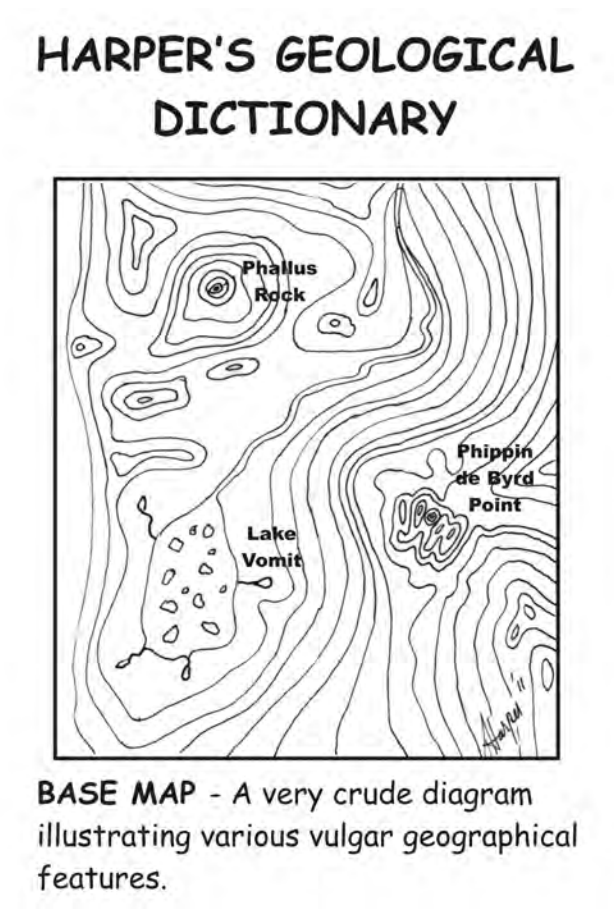

PALEOBOTANY AND STRATIGRAPHY AT THE CARBONIFEROUS-PERMIAN BOUNDARY IN CENTRAL AND WESTERN EUROPE: A SHORT REVIEW

Hermann W. Pfefferkorn

INTRODUCTION

Latest Carboniferous and earliest Permian strata in central and western Europe were deposited nearly exclusively in terrestrial environments. A facies change occurs in this stratigraphic interval from gray-colored strata to red beds and this changeover served as the system boundary well into the 20th century. An old miners' term for the red beds, the "Rotliegend" was used for the red facies and acquired the meaning of "lower Permian." Modern stratigraphic work has established that the lower part of the Rotliegend is Carboniferous in age and the facies change is time-transgressive.

At first, biostratigraphy in these beds was based exclusively on plant macrofossils, and *Callipteris conferta*, a peltasperm foliage type, was used as the index fossil for the base of the Permian. Today it is clear that *Callipteris* and similar forms represent floras living in seasonally drier environments that expanded over time and existed well before the beginning of the Permian. The floras of the latest Carboniferous to earliest Permian interval are known from a number of facies including coal-bearing strata, fluvial red-bed sequences, and forests blown down by volcanic blasts, á la Mount St, Helens, and preserved in the tuff. An interpretation of paleoenvironments and the principals involved in interpreting the interplay between paleoclimate, plant communities, and biostratigraphy can be found in Remy (1975).

A few medium sized and numerous small tectonic basins formed during the late phases or after the Hercynian orogeny that preserved these strata and the floras (Figure 1). Most of these basins were filled with sediment only on an intermittent basis so that they contain a rather incomplete record of time (Roscher and Schneider, 2006).

STRATIGRAPHY

Stratigraphy evolved in Europe between 1760 and 1841, and the geologic periods/systems were named there. Actually, the Permian Period was the last recognized and named when Murchison (1841) recognized rocks representing time between the Carboniferous and the Triassic in the Ural Mountains. Terrestrial and marine rocks from a restricted basin in England and Germany were soon recognized as time equivalent (Lucas et al. 2006). In Germany these rocks were described as two units, the Zechstein and the Rotliegend (Figure 2). Both names are old miners terms from a mining district in central Germany. "Zechstein" is the rock that was being mined, in this case a black shale containing copper and other metals. The Zechstein halite deposits that are mined today had not yet been found. The "Rotliegend" was the red rock lying below the underground mine workings, i.e. the "red-lying."

Pfefferkorn, H.W., 2011, Paleobotany and stratigraphy at the Carboniferous-Permian boundary in central and western Europe: A short review, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 200-206.

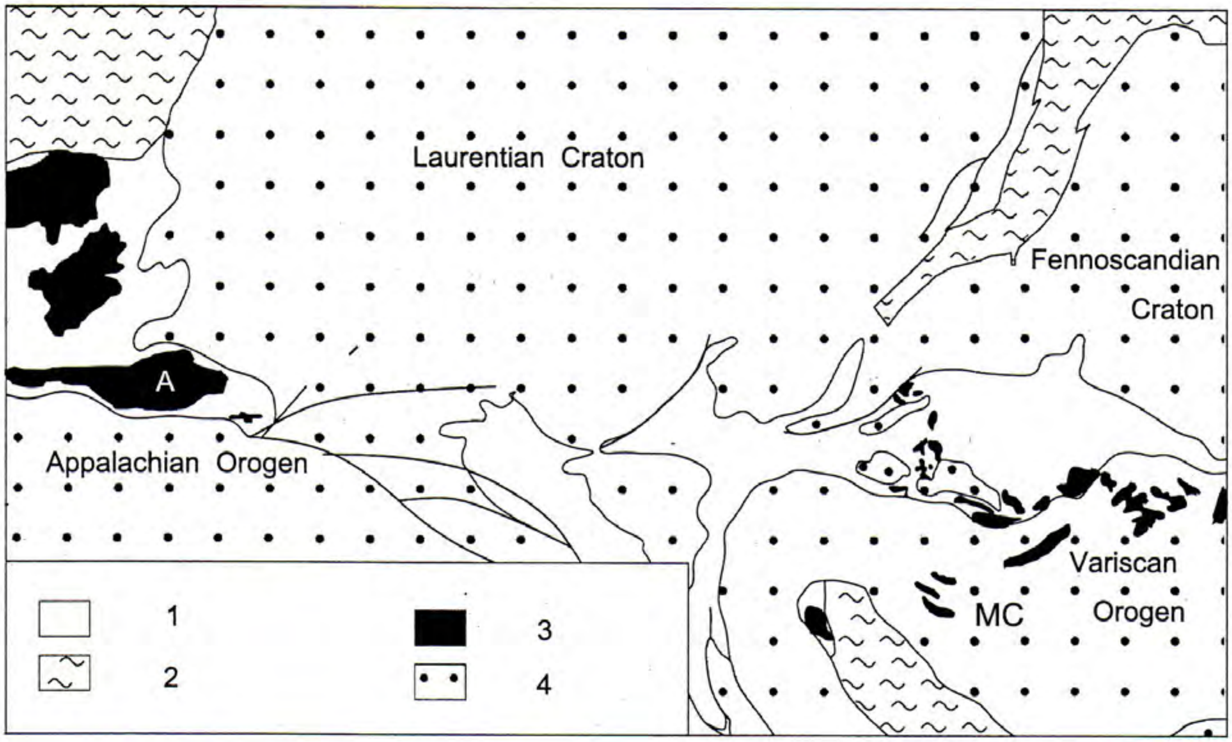

Figure 1 – Map of central part of equatorial Pangaea about 300 million years ago (approximate Carboniferous-Permian boundary) showing the large North American fore-deep and cratonic basins in the West and the much smaller European, mostly intramontaneous, basins in the East. The equator ran approximately in the center of the map from West to East. The distance between the western edge of the Appalachian basin (A) and the Massif Central (MC, outline not shown) was approximately 4,000 km (2,486 mi). 1—Lowlands with or without marine incursions. 2—Shallow marine deposits. 3—Coalfields. 4—Cratons and orogens, i.e. the erosional realm. Not all of the small European coalfields/basins can be shown at this scale. Map modified after DiMichele et al., 2010.

Today subdivisions of the global chronostratigraphic scale are based on marine strata that were rather continuously deposited in time and are rich in fossils with wide distributions (Salvador, 1994). The correlation of terrestrial strata with the marine subdivisions is not easy, but has made great progress that was summarized mostly for Europe with some references to North and South America, in the book by Lucas et al. (2006).

In most European basins with Permian strata, most of the Permian is terrestrial and only the latest Permian (Lopingian, Zechstein) is marine, but these beds were deposited in a basin that was not in continuous connection with the open ocean. The terrestrial strata deposited represent different parts of the stratigraphic column and are rather discontinuous (Figure 3).

OCCURRENCES AND BASIN SIZE

The basins in Europe that contain Permian strata formed during the last phases of the Hercynian orogeny (= Variscan orogeny, especially in central European literature). Many of these basins are small and fault bounded. Figure 1 cannot show all of these basins at that scale. The largest basin, in northern Germany, northern Poland, and under the North Sea, is not shown because it is mostly covered by a thick cover of younger sediments and known nearly exclusively from drill cores and salt domes.

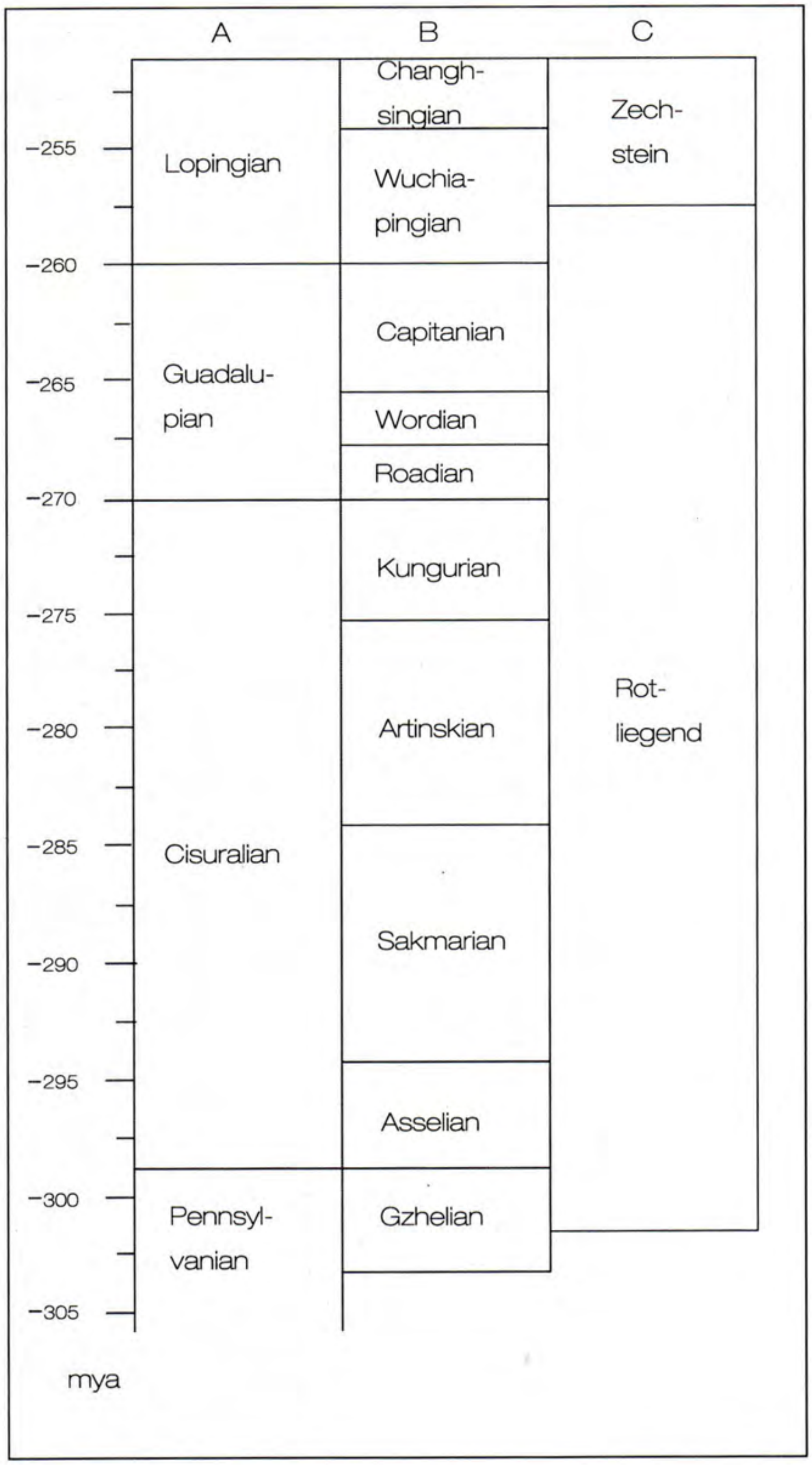

Figure 2 – Stratigraphy of the latest Carboniferous and Permian showing international epochs (column A) and stages (column B) on the left and the German division of the Permian into Rotliegend and Zechstein on the right (Column C). Please note that the lower part of the Rotliegend is Carboniferous in age (Anonymous, 2011).

Figure 3 – Latest Carboniferous and Permian stratigraphy of selected basins in France (columns A and B) and Germany (columns C-F) demonstrating the incompleteness of the stratigraphic record (modified after Roscher and Schneider, 2006). Vertical stripes indicate gaps in the record; white areas indicate that strata are present for this time interval. Lithostratigraphic names have been omitted for clarity. A – Basin of Lodève. B—basin of Autun. C—Saar-Nahe Basin. D—Thuringian Forest Basin. E—Ilfeld Basin. F—Döhlen Basin. C—coal deposits. V—volcanic deposits.

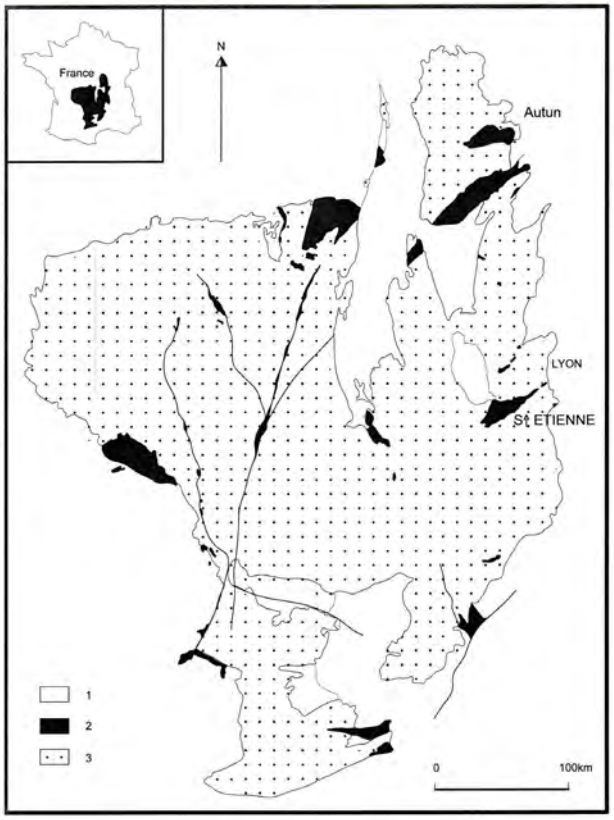

Figure 4 – Map of the Massif Central in France showing the numerous small fault bounded basins containing latest Carboniferous and earliest Permian strata (modified after Doubinger et al. 1995). The original type area for the Stephanian (late Pennsylvanian) is the basin of St. Étienne and that for the Autunian (early Permian) at Autun. 1—Post-Permian strata. 2—Coal basins with latest Carboniferous and early Permian strata. 3—Basement, i.e. rocks deformed during the Hercynian and earlier orogenies.

Figure 4 shows a map of the Massif Central in France, which consists of older rocks deformed by the Hercynian orogeny, with small fault-bounded sedimentary basins containing late Carboniferous and/or Permian sediments that often include mineable coal seams. It becomes clear from this map how difficult the stratigraphic correlation was. The work of the French colleagues who elucidated the stratigraphy was admirable.

PLANT FOSSIL PRESERVATION

Plant fossils occur in many of the latest Carboniferous and earliest Permian strata. However, it also has to be mentioned that there are large intervals of the stratigraphic column that are barren red beds. In some cases, collecting over decades has produced sizeable collections. In other cases, lagerstätten with exceptional preservation have been found. There are six main types of preservation listed below. The name of the locality or basin has been added where the preservation is not widespread:

Carbonaceous compressions in coal-bearing, gray facies
Permineralisations in coal-bearing, gray/black facies
Volcanic ash fall tuffs in coal-bearing facies (Döhlen Basin)
Adpressions and mold/cast in red bed facies, sometimes occurring in gray or greenish lenses
Permineralization in volcanic ash-flow deposits (Chemnitz)
Adpressions in volcanic ash-flow deposits (Chemnitz)

The basins in France and Germany are classical areas of paleobotany and numerous and large works have been published for nearly two centuries on the fossil plants found in these basins. Only a few can be mentioned here: Remy (1975); Doubinger et al. (1995); Barthel (1976, 2009); Kerp et al. (2007); Roessler (2006).

DISCUSSION AND CONCLUSIONS

Defining or finding the Carboniferous-Permian boundary in the terrestrial sequences of Europe has been as difficult as in the Dunkard Basin (Remy, 1975). Until recently there did not even exist an internationally accepted boundary stratotype, so that the definition of the boundary differed between schools of thought, authors, decades, or regions. Currently, one can summarize the research on this time interval in Europe as follows:

- Several working groups are investigating this interval intensively in different basins and a good summary can be found in the book by Lucas et al. (2006)
- The time of the latest Carboniferous and earliest Permian is characterized by fluctuating climate change within an overall warming and drying trend
- The Carboniferous-Permian time interval is characterized by discontinuous sections in numerous basins that are mostly relatively small
- Extirpation of plants seems to occur at different times in different basins
- No specific macrofloristic change has been found at the Carboniferous-Permian boundary
- However, a characteristic flora occurs during the time interval into which the Carboniferous-Permian boundary falls

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Oh, that's Dewey K. Byrd, the great geologist, and those are some students who have just made his acquaintance.

PLANT LIFE DURING THE CARBONIFEROUS-PERMIAN TRANSITION ON THE NORTH CHINA BLOCK

Wang Jun and Hermann W. Pfefferkorn,

INTRODUCTION

The late Paleozoic cratonic basin on the North China Block covers an area of 1.2 ×106 km^2, in which the Carboniferous-Permian strata are extensively preserved and represent the major coal-accumulation deposits of North China. During late Paleozoic time the North China Block was situated to the east of Pangaea as a micro-continent in tropical latitudes in the Panthalassic ocean (Figure 1). Plant fossils are plentiful in the terrestrial deposits and have been studied to clarify the stratigraphy and aid the coal industry. The fossil flora consists of numerous taxa also known from the Euramerican Floral Realm while at the same time including a significant number of endemic elements that justify the separation as the Cathaysian Floral Realm. Here we discuss the floral assemblage around the Carboniferous-Permian transition from the Taiyuan Formation as a contrast with that from the Dunkard Group of the Appalachian Basin situated in western Pangaea.

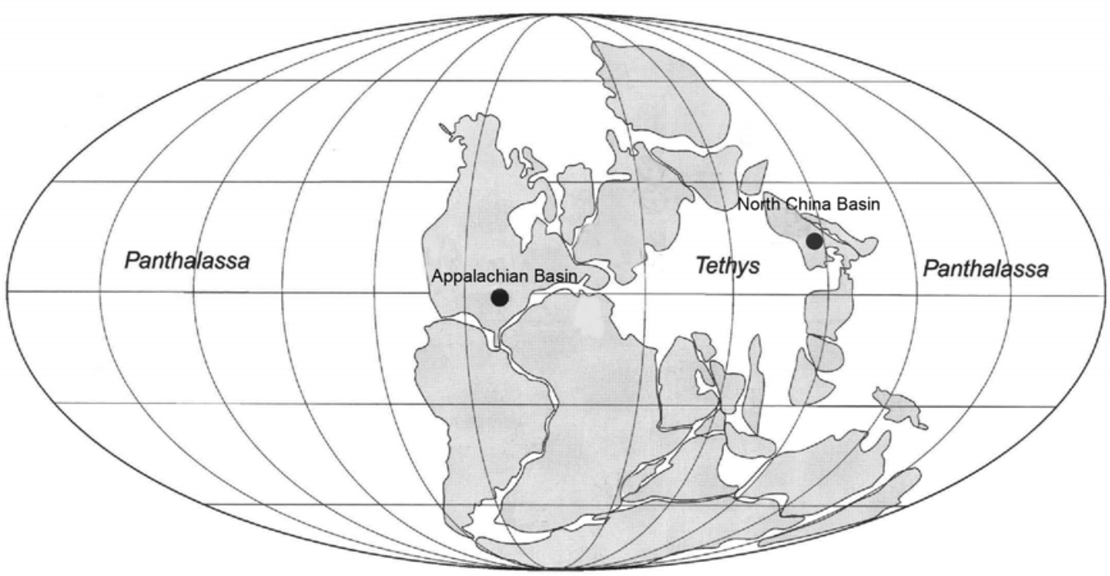

Figure 1. Map of the Earth at the Carboniferous-Permian transition showing the positions of the cratonic basin on the North China Block and the Appalachian Basin (modified after Lucas, 2006).

LITHOLOGY AND CHRONOLOGY

On the North China Block late Paleozoic deposits are widely developed and preserved. The Carboniferous Permian boundary occurs within the Taiyuan Formation which occurs above the Benxi Formation and is itself overlain by the Shanxi Formation. The Taiyuan Formation

Wang Jun and Pfefferkorn, H. W., 2011, Plant life during the Carboniferous-Permian transition on the North China Block *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 207-218.

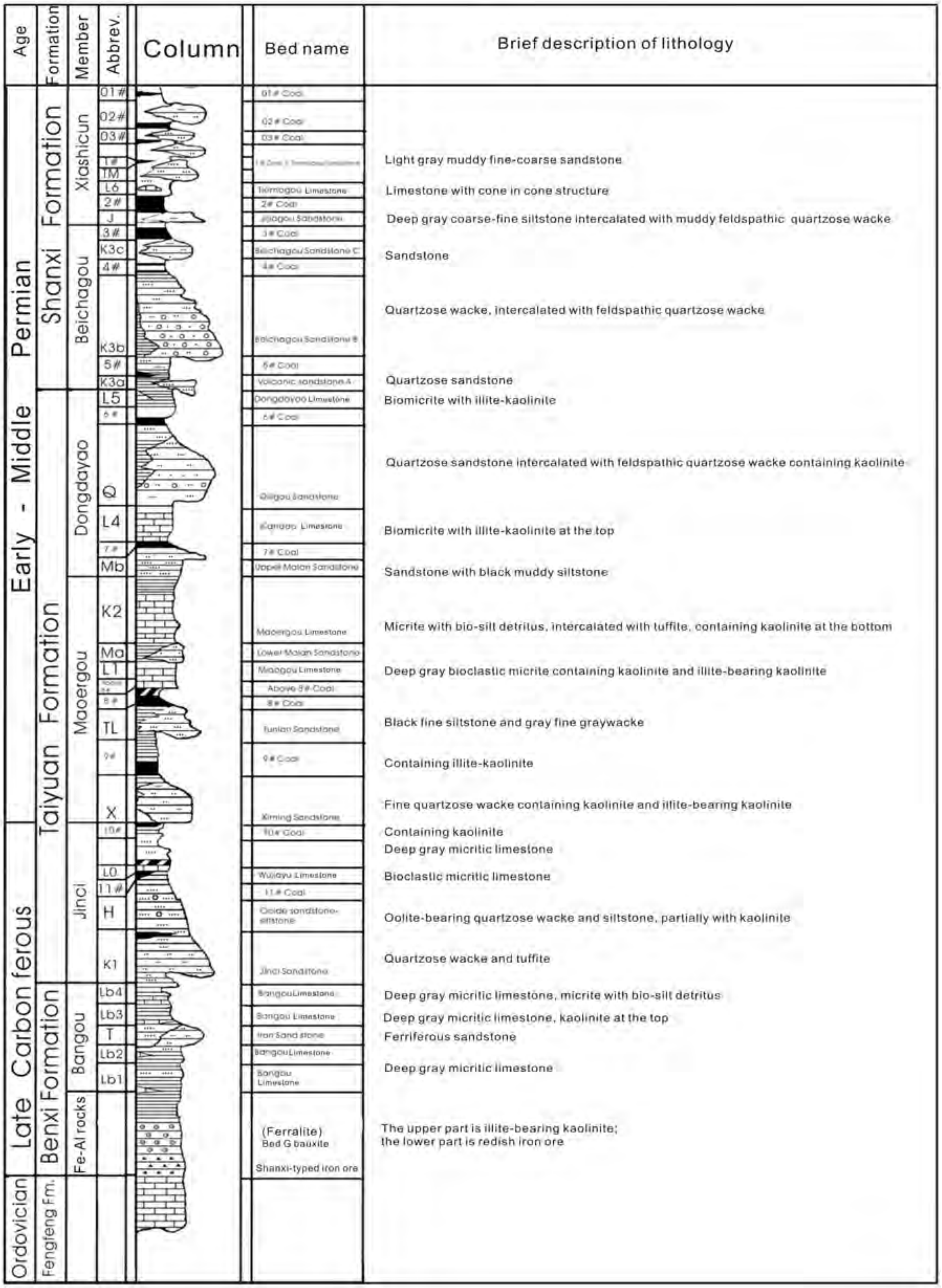

Figure 2. General stratigraphic section of the Carboniferous-Permian deposits in the West Hill Coalfield, Taiyuan (modified after Yang et al., 1987).

consists of cyclic repetitions of coals, mudstones, limestones and sandstones as an alternating marine and terrestrial facies. As a typical area of late Paleozoic deposits in North China, the West Hill Coalfield near Taiyuan has been investigated for over a hundred years during coal industry development and is therefore geologically and paleontologically one of the best known areas in North China. A generalized geological column of this area shows the stratigraphic profile of the North China Block (Figure 2). The Taiyuan Formation is about 100 m (328 ft) thick and is subdivided into three members, based on paleontological data and facies differences described here from the base to the top:

Jinci Member: The lowermost unit within this member is the Jinci Sandstone, medium to coarse, quartz-rich graywacke and tuffite. The middle part consists of dark grey, fine-grained clastic rocks containing pyrite and siderite nodules and oolite bearing fine sandstone or siltstone, intercalated with 1-2 coal seams. The uppermost part is formed by the Wujiayu Limestone which varies markedly in thickness and lithology.

The Maoergou Member is composed of clastic rocks, carbonates and coal beds. Two workable coal beds (i.e. No. 8 and 9 coal) occur in the lower part. There are three major sandstone layers. The lower coarse quartzose graywacke is named the Ximing Sandstone. The middle lithic graywacke between No. 8 and 9 coal beds is called Tunlan Sandstone. The upper one, the Lower Malan Sandstone, occurs between the Miaogou and Maoergou limestones. The Miaogou Limestone, a bioclastic micrite, forms the roof of No. 8 coal.

The Dongdayao Member consists of clastic rocks intercalated with carbonates and coal beds No. 6 and 7. Two major sandstones are the Upper Malan Sandstone below the No. 7 coal and the Qiligou Sandstone below the No. 6 coal. Two major carbonate layers are the biomicritic Xiandao Limestone and Dongdayao Limestone.

In all three members numerous thin layers of mudstone and siltstone are intercalated in which fossil plants are preserved. In addition, plentiful fossil faunas occur in all the limestones.

Before the 1980s, the whole Taiyuan Formation was thought to be entirely of late Carboniferous age. Later in the 1990s, an effort was made to recognize the Carboniferous-Permian boundary based on international definition. Thus, the limestones yielding the fusulinids *Pseudofusulina*, *Pseudoschwagerina* and *Sphaeroschwagerina* in the middle and upper part of the Taiyuan Formation were attributed to the Lower Permian (Asselian), while the lower part yielding the fusulinids *Montiparus-Triticites* remained in the late Carboniferous (Stephanian, Late Kasimovian to Gzhelian, Wu, 1995). The fusulinid and conodont assemblage in the Taiyuan Formation of the West Hill coalfield are summarized in Figures 3 and 4 respectively.

FLORAL ASSEMBLAGE

When the whole Taiyuan Formation was thought to be entirely of late Carboniferous age before the 1980s, one floral assemblage was proposed (e. g. Li, 1963a, 1963b), namely the *Neuropteris pseudovata – Lepidodendron posthumii* assemblage. After the Permo - Carboniferous boundary was set to conform to the international definition in the 1990s, the fossil plants derived from the lower part of the Taiyuan Formation and those from the middle and upper Taiyuan Formation were attributed to two different assemblages (e.g. Li et al., 1995;

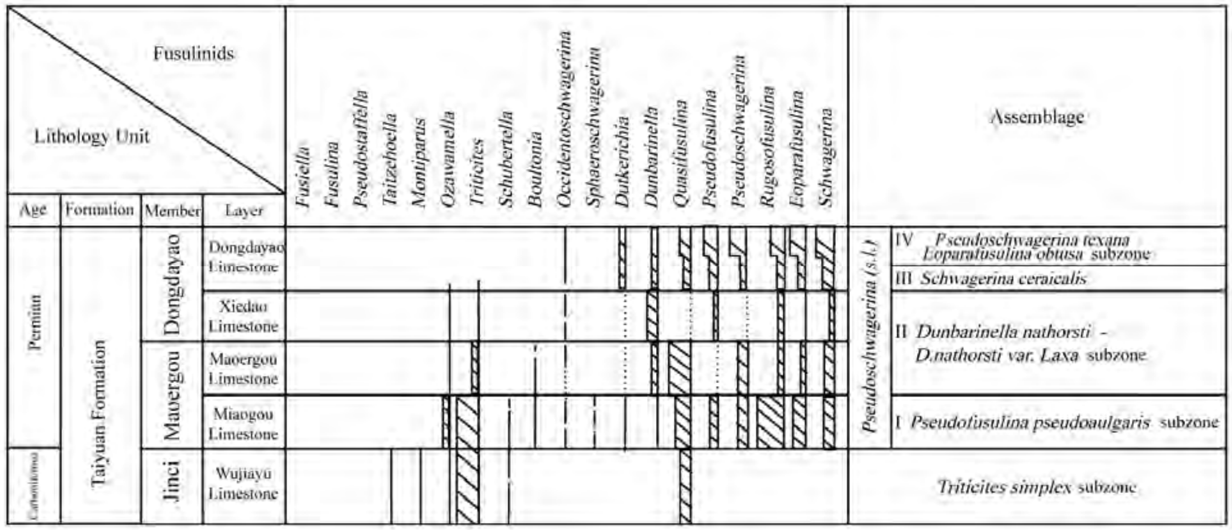

Figure 3. Diagram showing ranges of fusulinids in the Taiyuan Formation of the West Hill Coalfield (modified after Chen and Niu, 1993).

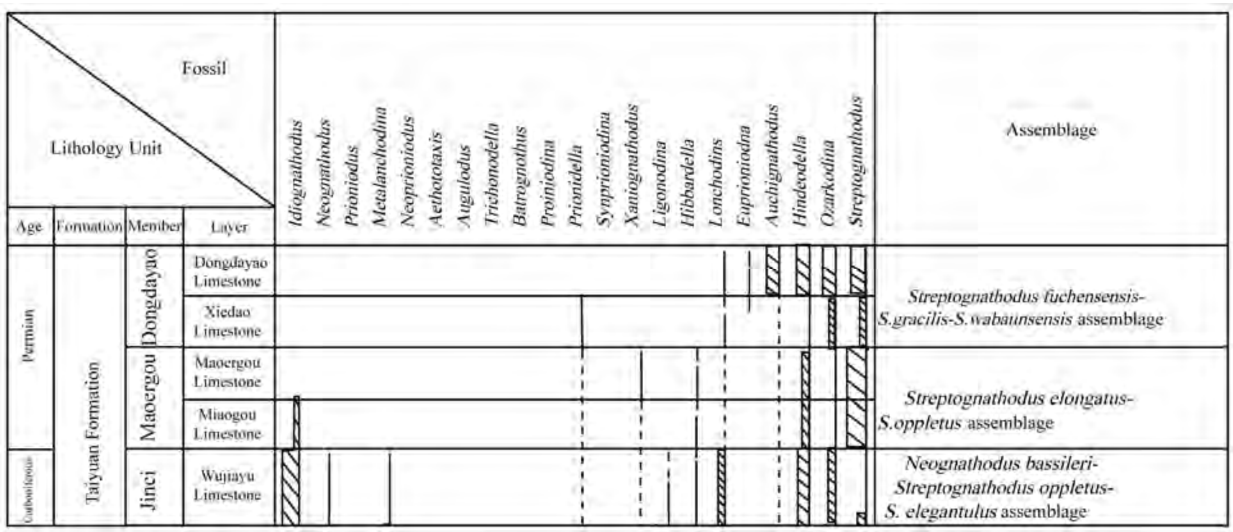

Figure 4. Diagram showing ranges of conodonts in the Tayuan Formation of the West Hill Coalfield (modified after Chen and Niu, 1993).

Li, 1997; Liu et al., 1996). However, as a matter of fact, the flora found in the lower part of the Taiyuan Formation cannot be distinguished from that of the upper part of the Taiyuan Formation In the well known Western Hill coalfield, nearly all of the species occurring in the lower part of the Taiyuan Formation are also present in the upper part of the formation (Zhang, 1987). According to Liu et al. (1996), in middle and northern Shanxi, there are 27 species derived from the lower part of the Taiyuan Formation, of which only 4 do not extend into the upper part of the Formation. In the Jushuihe Section of the Weibei coal field (Wang, 2010), there is no obvious floral change throughout the whole Taiyuan Formation Among the total 16 species occurring in the Taiyuan Formation, only 3 in the lower part, do not range into the upper part of the Taiyuan Formation The species not present in one or the other part of the formation are rare taxa that might be missing or present due to sampling biases. They are not common forms that would be found in any sustained collecting. Thus, it is evident that the flora of the

Taiyuan Formation does not change across the Carboniferous-Permian boundary.

The representative member of the assemblage *Neuropteris pseudovata* (Plate 1, d) has mostly been recorded from the Taiyuan Formation in northern China, and often selected as the naming element of plant assemblages in this interval. In Chinese paleobotanical literature, this taxon also has been called *Neuropteris ovata* based on gross morphology. *N. ovata* is widely distributed in the Westphalian D of Euramerica. However, according to Mosbrugger (1989), the cuticle structure of *N. pseudovata* from Baode, Shanxi is certainly different from that of the European *N. ovata*. Therefore, it is suggested that the name *N. pseudovata* is preferable for materials of *N. ovata*-like pinnules in Cathaysian floras. Furthermore, the representative members *Lepidodendron posthumii* (Plate 1, g) and *L. szeianum* (Plate 1, f) are also often selected as the naming taxa of this assemblage. Other common elements of the assemblage include *Lepidodendron oculus-felis*, *L. ninghsiaense*, *Cathaysiodendron nanpiaoense* (Plate 1, e), *C. incertum*, *Ulodendron tienii*, *Sphenophyllum oblongifolium* (Plate 1, b), *Sph. emarginatum, Sph. kawasakii* (Plate 1, a), *Sph. laterale*, *Calamites cistii*, *C. suckowii*, *Annularia pseudostellata*, *Tingia carbonica*, *T. hamaguchii*, *Pecopteris arborescens*, *P. candolleana*, *P. polymorpha*, *P. cyathea*, *P. hemitelioides*, *Nemejcopteris feminaeformis*, *Neuropteris plicata*, *Alethopteris huiana* and *Cordaites volkmannii*.

Lycopsids, sphenopsids and pecopterids dominate the flora. *Lepidodendron oculus-felis*, *L. posthumii* and *Cathaysiodendron nanpiaoense* are typical oriental lepidophytes (Li et al., 1995), whereas *L. szeianum* is similar to the Carboniferous *L. gaudryi* (Li, 1963a) of western Europe. Among the sphenopsids, all species of *Sphenophyllum* do not have a middle vein in their leaves. Except for *Sphenophyllum laterale* which is endemic to Cathaysia, all of the other sphenopsids are found in both the Euramerican and Cathaysian floral provinces. *Pecopteris candolleana* and *Nemejcopteris feminaeformis* occur in both Cathaysia and Euramerica, whereas the occurrence of *Tingia carbonica* emphasizes the Cathaysian characteristics.

The permineralized floral assemblages, specifically the floras preserved in the coal balls and tuff beds of the Taiyuan Formation in the North China Block have been extensively investigated (Li et al., 1995; Wang et al., 1995, Tian et al., 1996; Hilton et al., 2001). The coal ball floras are dominated by cordaitopsids and lycopsids, with the marattialean fern *Psaronius* as a common element. This is generally consistent with the known macrofloral assemblages.

THE WUDA TUFF FLORA IN INNER MONGOLIA AND ITS IMPLICATION FOR THE DIACHRONOUS CHARACTER OF THE TAIYUAN FORMATION

An exceptionally well-preserved peat-forming flora was reported from a volcanic tuff layer of the uppermost part of the Taiyuan Formation in Wuda, Inner Mongolia (Pfefferkorn and Wang, 2007). Six groups of plants make up this peat-forming plant community (Plates 2-4). Lycopsids are represented by *Sigillaria* cf. *ichthyolepis* (Plate 2, e-g, Wang et al., 2009; Pfefferkorn and Wang, 2009). *Sphenophyllum* (Plate 2, c-d), a dwarf shrub, and a calamite are the sphenopsids that were encountered. Some species of Marattialean tree ferns (Plate 3, a, c, e-f) have been found which represent the most abundant plants. Herbaceous ferns present are *Nemejcopteris feminaeformis* (Plate 3, b) and *Sphenopteris* (Plate 3, d). Noeggerathiales are represented by several species of *Tingia* (Plate 4, d) and *Paratingia* (Plate 4, c) (Wang, 2006; Wang et al., 2009). *Taeniopteris* (Plate 4, e) and *Pterophyllum* (Plate 4, b) can be interpreted as early relatives of cycads. *Cordaites* (Plate 4, a) trees were early coniferophytes.

In terms of the biostratigraphy, fossil fauna including fusulinid and conodont are not available in this area because marine beds are not developed in the section for this time interval. Lithostratigraphically, the volcanic tuff is located in the uppermost part of the Taiyuan Formation, so that the flora would represent an Early Permian Sakmarian flora according to established biostratigraphy (Li, 1963a, b, Li et al., 1995). However, the plant fossils that are identical to those found in Euramerica (*Sigillaria* cf. *ichthyolepis*, *Nemejcopteris feminaeformis*, *Sphenophyllum oblongifolium*, and several species of *Pecopteris*) indicate a latest Carboniferous or earliest Permian Asselian age. This is further supported by a dating of the tuff, 299-298 My (to be published). Therefore, the Taiyuan Formation is Carboniferous in age in this area. For the first time, the diachronous nature of the Taiyuan Formation of the North China Block is revealed by floral composition in combination with isotope dating.

CONCLUDING REMARKS

The North China Block was a micro-continent in tropical latitudes surrounded by oceans. On this block interlayered marine and terrestrial strata are preserved and named the Taiyuan Formation, representing the Carboniferous-Permian transition. Traditionally, the Carboniferous -Permian boundary was placed in the middle of the Taiyuan Formation while our own investigations in the Wuda coalfield in Inner Mongolia have shown that the newly defined Carboniferous-Permian boundary at 299 Ma coincides at least in that coalfield with the boundary of the Taiyuan Formation with the overlying Shanxi Formation. This indicates that a number of floras that were formerly described as lower Permian are in reality late Carboniferous in age. The diachronous nature of the Taiyuan Formation and the persistence of the flora need to be systematically investigated.

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Plate 1. Selected elements of the Taiyuan Formation of the North China Block (Refigured from Li, 1963).

a. ***Sphenophyllum kawasakii*** **Stock. et Math. Naopaokou, Tachingshan, Inner Mongolia. Scale bar = 1cm, PB3039.**
b. ***Shenophyllum oblongifolium*** **(G. & K.). Naopaokou, Tachingshan, Inner Mongolia. Scale bar = 1cm, PB2034.**
c. ***Alethopteris huiana*** **Li. Sishan (West Hill), Taiyuan. Scale bar = 1cm, PB 3158.**
d. ***Neuropteris pseudovata*** **Goth. et Sze. Naopaokou, Tachingshan, Inner Mongolia. Scale bar = 1cm, PB3193.**
e. ***Cathaysiodendron nanpiaoense*** **Li. Nanpiao, Liaoning. Scale bar = 1cm, PB 3082.**
f. ***Lepidodendron szeianum*** **Li. Nanpiao, Liaoning. Scale bar = 1cm, PB 3065.**
g. ***Lepidodendron posthumi*** **Jong. et Goth. Pinglu, Shansi (Shanxi). Scale bar = 1cm, PB 3060.**

Plate 2. Representative elements of the earliest Permian Wuda Tuff Flora in Inner Mongolia

a. ***Asterophyllites longifolius*** **(Sternberg) Brongniart. Scale bar = 2cm. PB 21425.**
b. **Strobilus of *Palaeostachya*-type associated with *Asterophyllites longifolius.* Scale bar = 1cm. PB 21426.**
c. **c, d. *Sphenophyllum oblongifolium* (G. & K.). c, Scale bar = 1cm; d, Scale bar = 2cm. PB = 21427-21428.**
d. **e. Leaves of *Sigillaria*-type associated with *Sigillaria* cf. *ichthyolepis* (Presl) Corda. Scale bar = 1cm. PB = 21429.**
e. **f. *Sigillaria* cf. *ichthyolepis* (Presl) Corda. Scale bar = 2cm. PB 21106.**
f. **Strobilus of *Sigillaria*-type associated with *Sigillaria* cf. *ichthyolepis* (Presl) Corda. Scale bar = 1cm. PB = 21430.**

Plate 3. Representative elements of the earliest Permian Wuda Tuff Flora in Inner Mongolia

a. ***Pecopteris candolleana*** **Brongn. Scale bar = 1cm. PB = 21431.**
b. ***Nemejcopteris feminaeformis*** **(Schloth.) Sterz. Scale bar = 5mm. PB = 21432.**
c. ***Pecopteris orientalis*** **(Schenck) Potonie. Scale bar = 1cm. PB = 21433.**
d. ***Spenopteris*** **sp. Scale bar = 1cm. PB = 21434.**
e. ***Pecopteris lativenosa*** **Halle Scale bar = 1cm. PB = 21059.**
f. ***Pecopteris arborescens*** **(Schloth) Sternb. Scale bar = 3cm. PB = 21435.**

Plate 4. Representative elements of the earliest Permian Wuda Tuff Flora in Inner Mongolia

a. ***Cordaites*** **sp. Scale bar = 2cm. PB = 21436.**
b. ***Pterophyllum*** **sp. Scale bar = 2cm. PB = 21437.**
c. ***Paratingia wudensis*** **Wang et al. Scale bar = 2cm. PB = 20781.**
d. ***Tingia unita*** **Wang Scale bar = 2cm. PB = 21438.**
e. ***Taeniopteris*** **sp. Scale bar = 2cm. PB = 21439.**

Plate 1

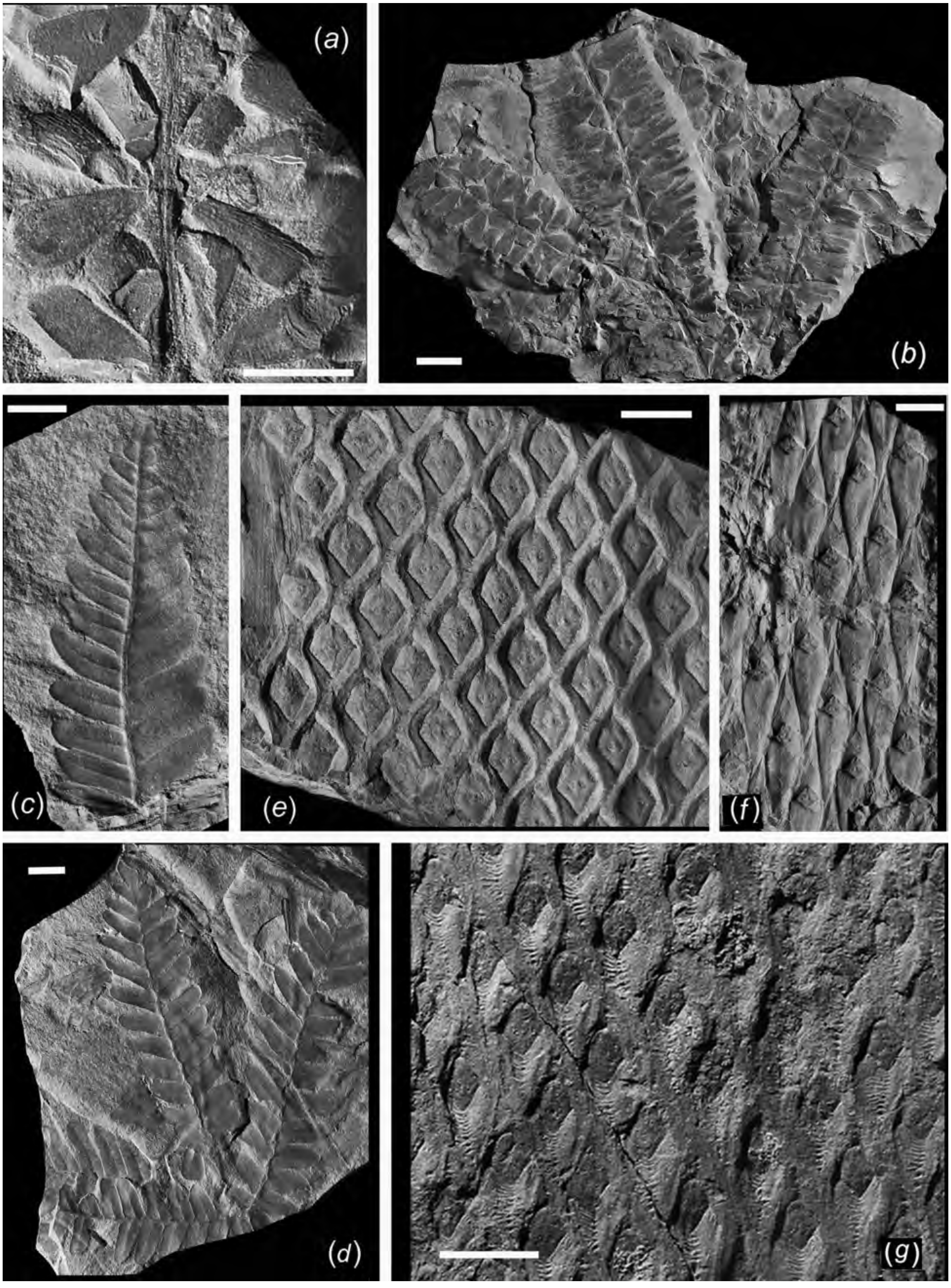


Plate 2

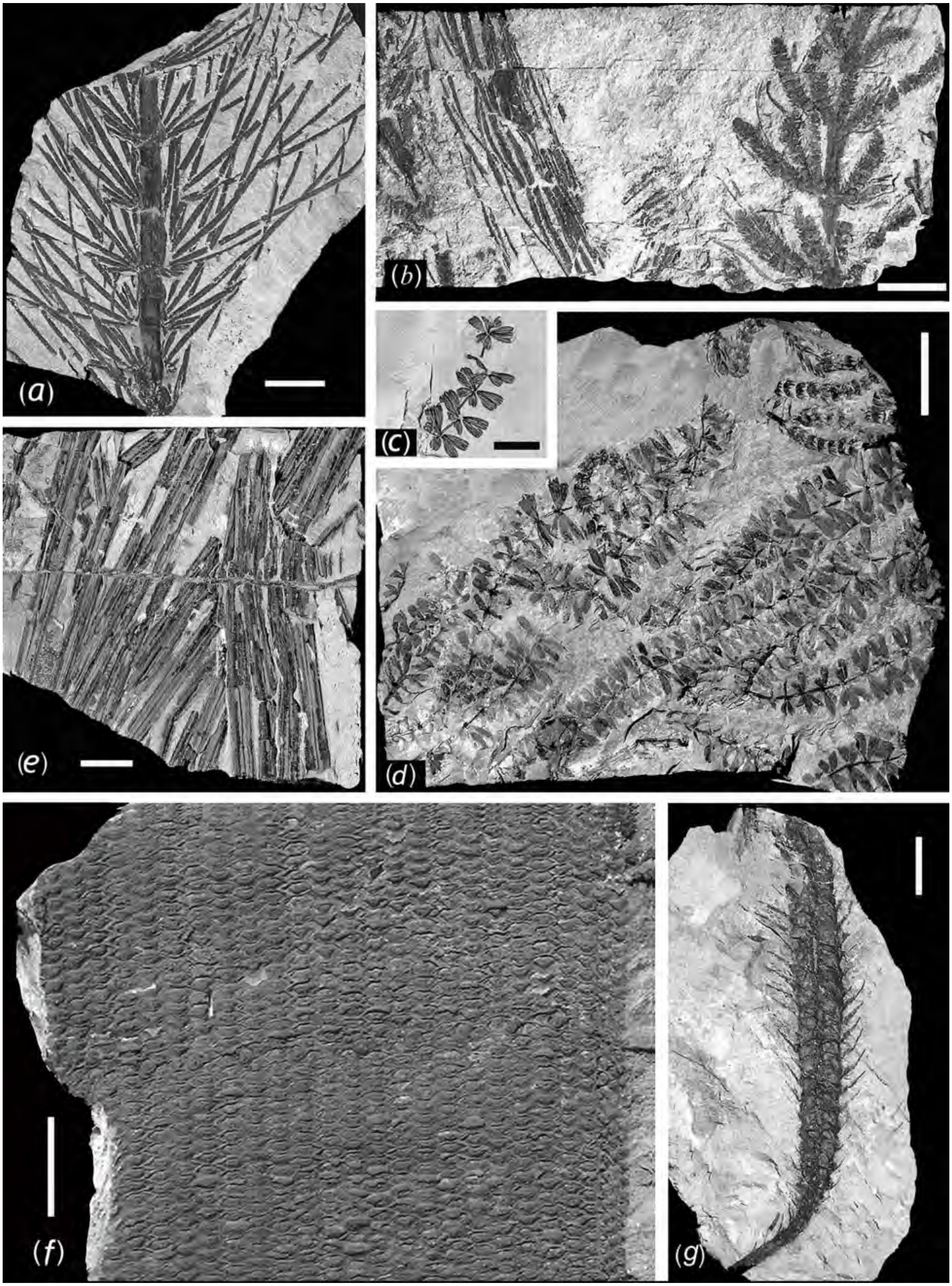


Plate 3

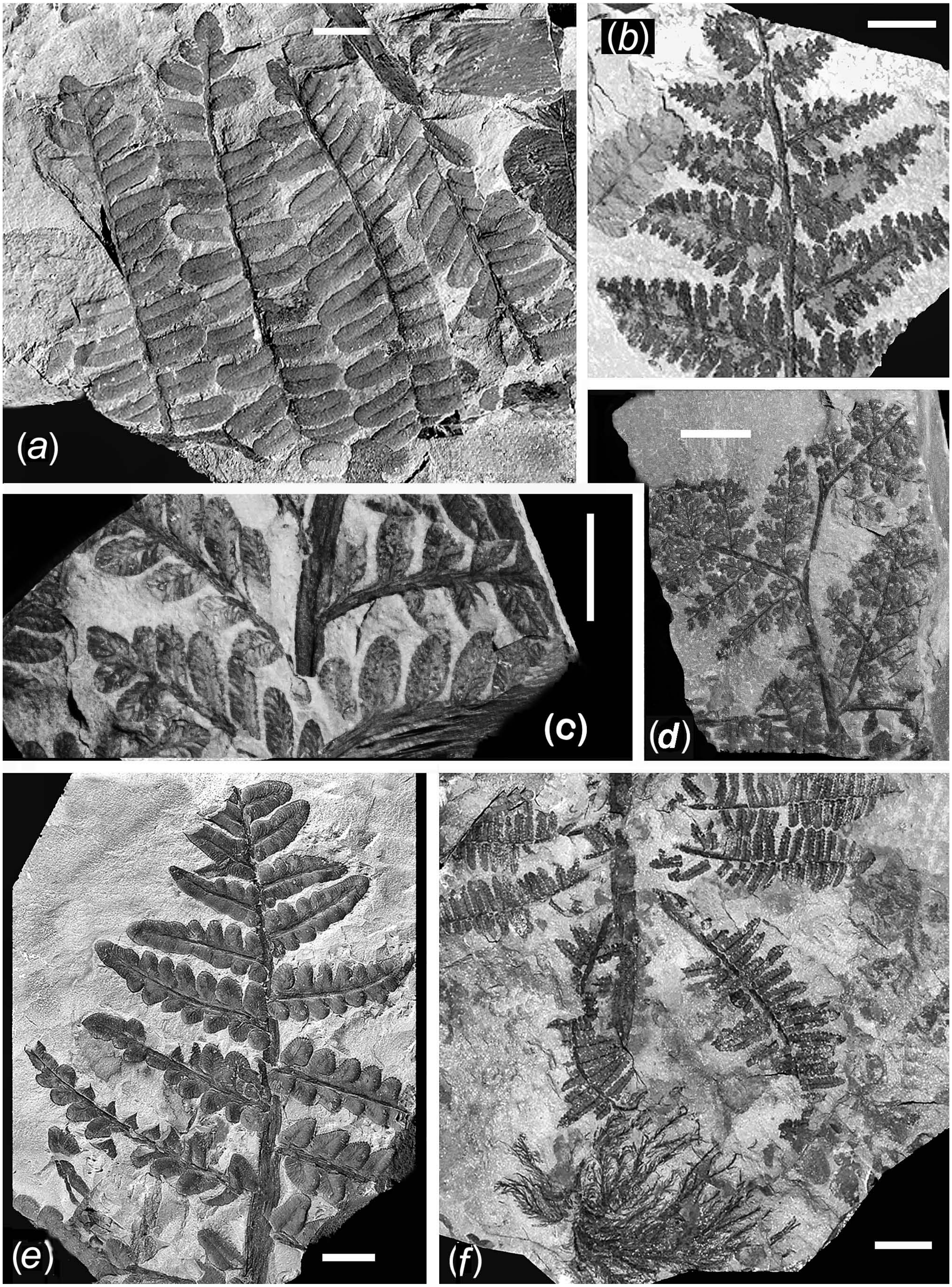
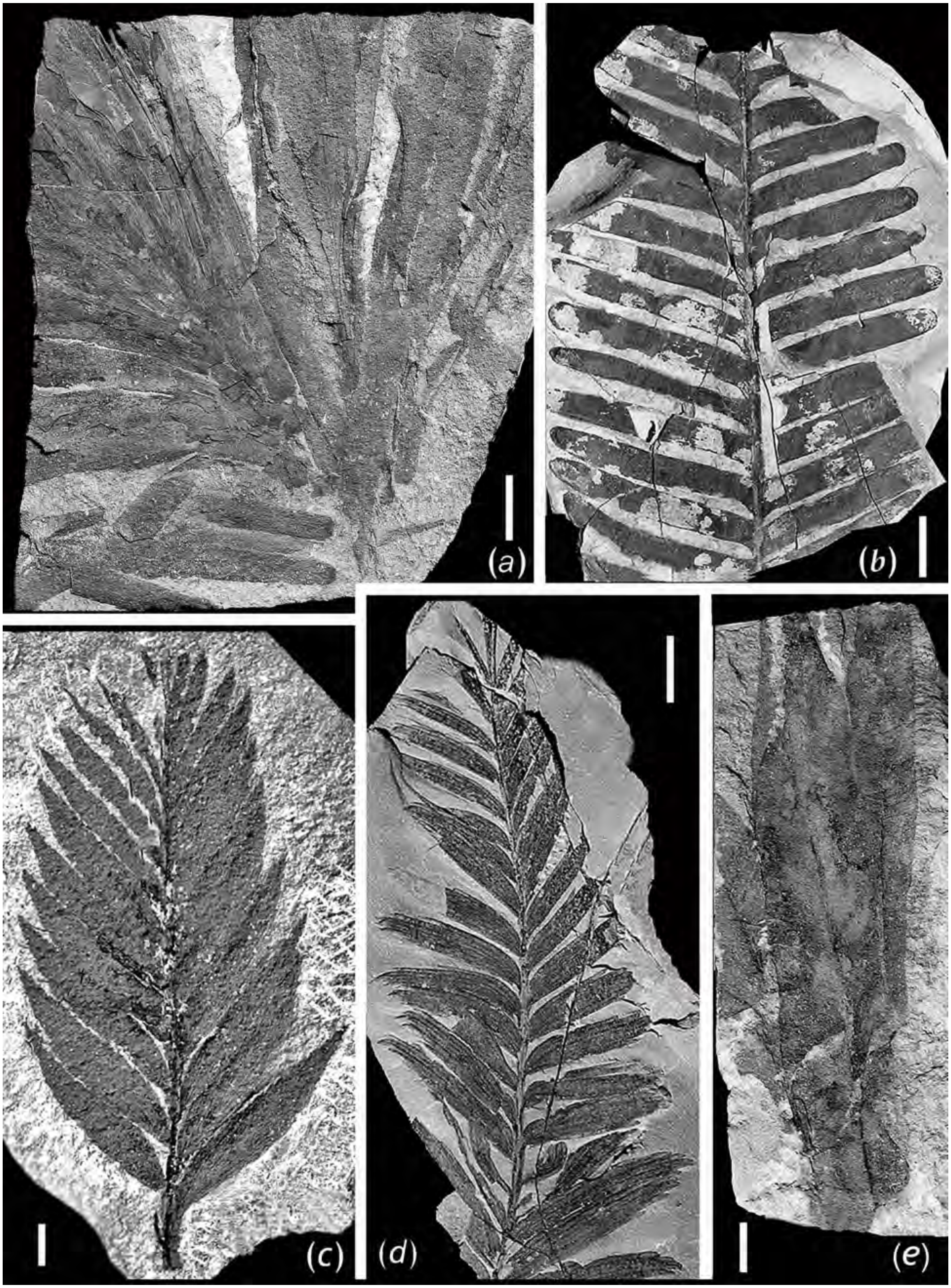


Plate 4



LANDSCAPE OF THE GREENE AND WASHINGTON COUNTIES AREA

W. D. Sevon

INTRODUCTION

This brief discussion of the general landscape of Greene and Washington counties in southwesternmost Pennsylvania is based on limited work that involved (1) examination of Greene and Washington counties 1:50,000-scale topographic maps, (2) examination of the Geologic Shaded Relief Map of Pennsylvania (Miles, 2003), (3) three days of field examination, and (4) minimal literature review. The landscape is unique for Pennsylvania and deserves more research.

THE LANDSCAPE

The landscape of Greene and Washington counties lies entirely within the Waynesburg Hills Section of the Appalachian Plateaus Physiographic Province (Sevon, 2000). The landscape comprises low hills separated by narrow valleys with moderate to steep slopes. Relief ranges, in general, from 200 to 600 ft (61 to 183 m) and probably averages 300 to 500 ft (91 to 152 m). Elevations of hill tops are generally less than 1,600 ft (488 m) while valley bottoms are generally more than 1,000 ft (305 m).

The landscape is generally moderately to totally covered by woodland vegetation, as is shown in Figures 1 and 2.

The landscape is the result of erosion by 10 drainage systems that have their headwaters somewhere within the section. An irregular, N-S drainage divide occurs west of the section

Figure 1. Typical Greene County landscape with abundant woodland vegetation.

Sevon, W. D., 2011, Landscape of the Greene and Washington counties area, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 219-232.

Figure 2. Typical Greene County landscape showing the considerable extent of woodland vegetation that obscures the upper slopes and hill tops.

center. There are no streams that cross the drainage divide naturally. East-flowing Tenmile Creek appears to connect with west-flowing Enlow Fork Wheeling Creek along a course that follows the Greene-Washington County line. This connection is an artifact of Pleistocene stream damming that caused Tenmile Creek to overflow a col into Enlow Fork Wheeling Creek and deepen the col by erosion. South Fork Tenmile Creek in Greene County has the largest drainage basin and the most dendritic pattern. The landscape of the two-county area is unique in Pennsylvania because of the intricacy and extent of the drainage system that has caused all the erosion. Nowhere else in Pennsylvania is the drainage system as intense and intricate as this system.

The total drainage system is shown in Figure 3. In addition to showing and naming all the streams in the two-county area, the map also shows the primary drainage divides that occur. Within the area are smaller drainage divides that separate local drainage basins, but do not change the major stream, either Monongahela or Ohio, into which they flow. The complexity of this drainage system and the degree of its development are the result of an erosional history that is not dealt with here because I lack the knowledge necessary to do so.

The stream patterns shown give a good indication of the complexity of the drainage system. Some closer views will show not only details of the patterns, but also the overall nature of the landscape. Figure 4 centers on Aleppo Township that occurs in the southwestern part of Figure 3. This area drains westward into the Ohio River and has probably the best development of dendritic drainage pattern in the two-county area. Note the interconnection of the many streams and the general lack of preferential orientation in all but the southeasternmost part of the map. The rocks in this area are presumably flat-lying and not controlling drainage development through structure.

Note the well-displayed topography shown by the 20-ft (6-m) contours. Narrow hilltops are generally irregular in shape and, at least when this map was created (1982), free of

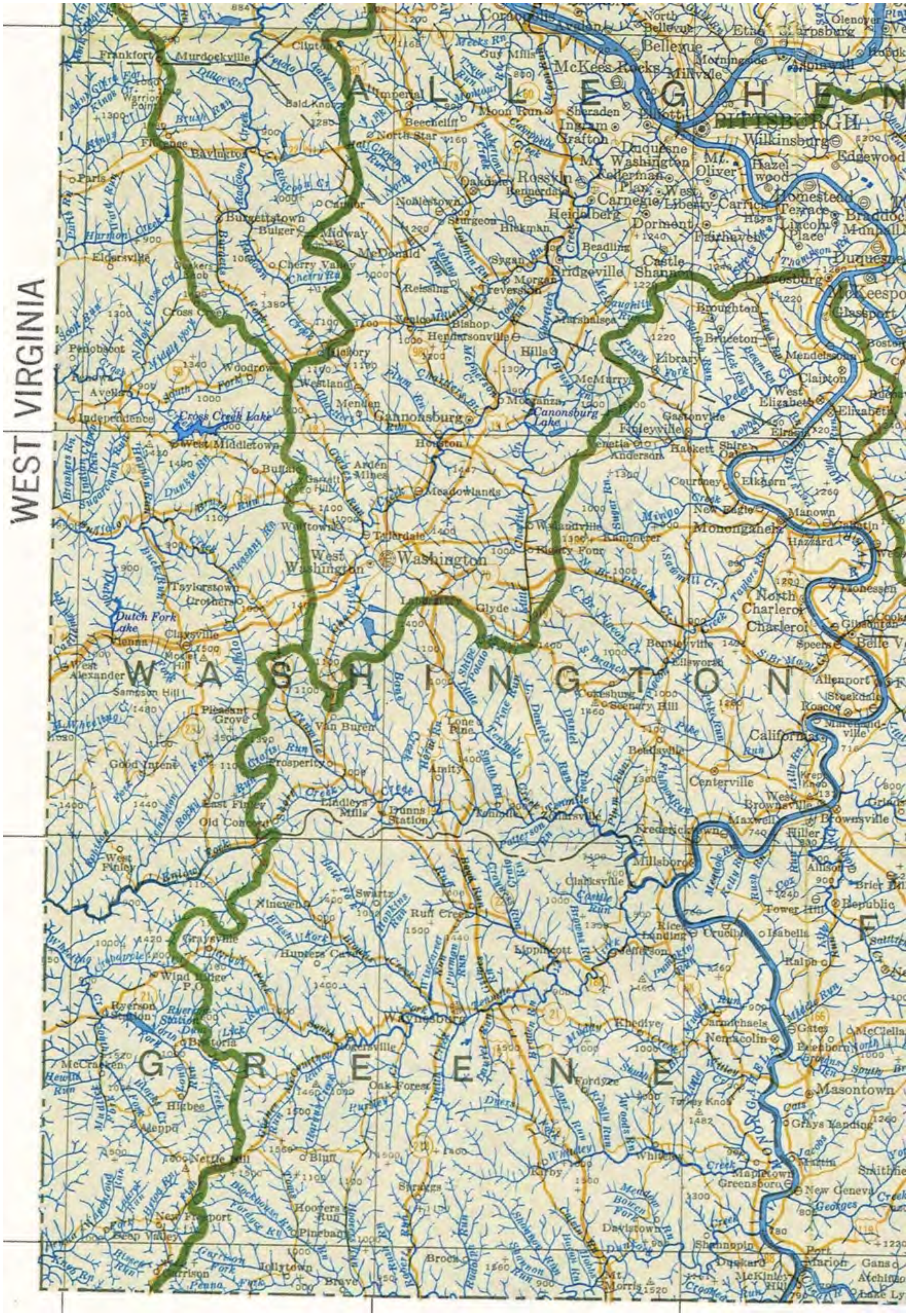


Figure 3. The stream map of the Greene and Washington counties area. This is part of Professor Higbee's Streams of Pennsylvania at 380,160 scale.

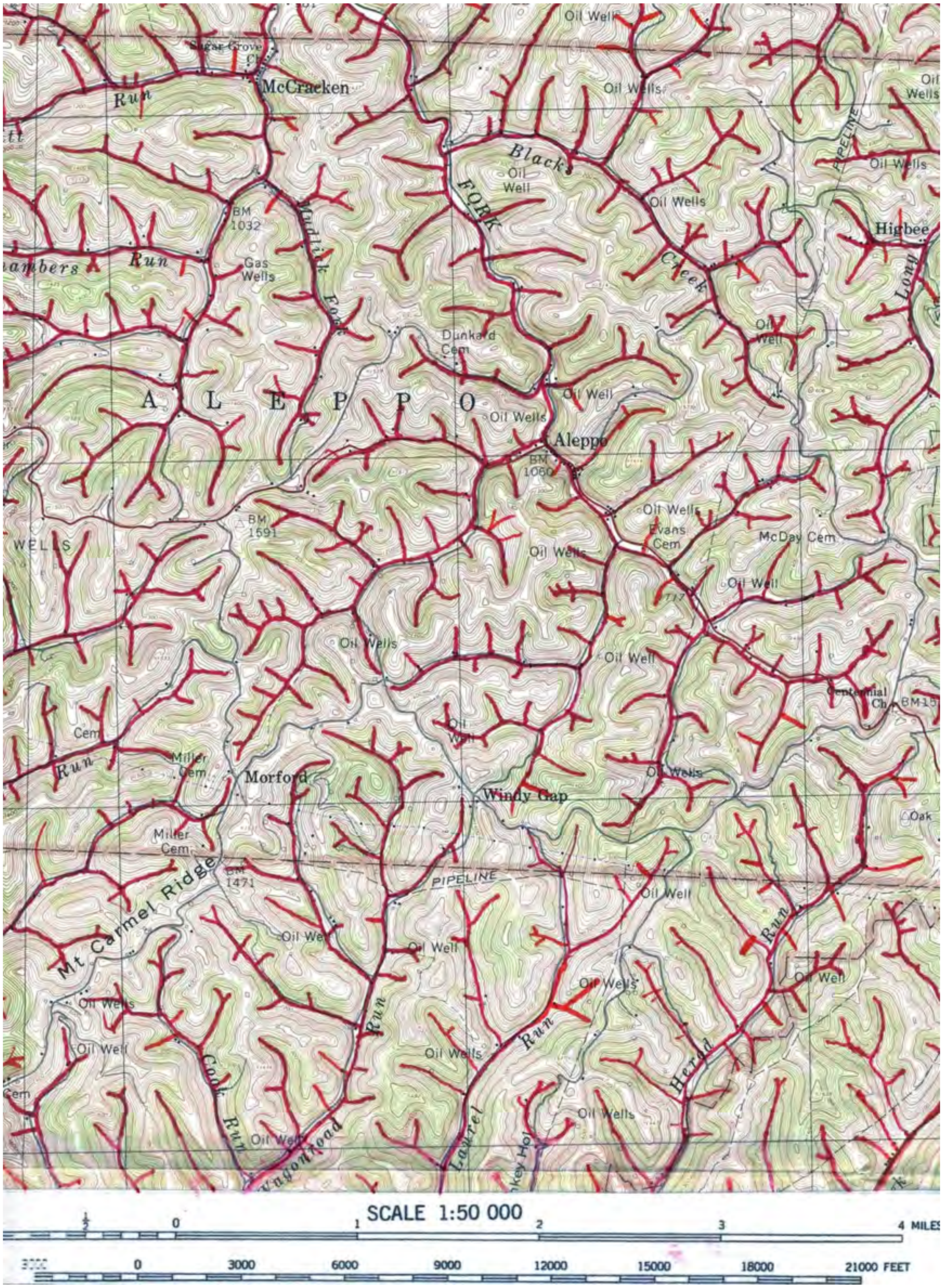


Figure 4. Aleppo Township area of Greene County. Map is from the Greene County 1:50,000-scale topographic map. Red lines delineate valleys.

woodland vegetation. Tributaries to larger streams that terminate in a specific valley have well defined valleys. Slope angles are uniform and slopes are smooth. Upland elevations are generally 1,400 to 1,500+ ft (427 to 457+ m) and relief is generally 300 to 400 ft (91 to 122 m).

In the southeasternmost part of the map, there is a strong tendency for north side tributary asymmetry (NSTA). Look at Laurel Run, Herod Run, and the next stream southeast. These longer northeast-oriented streams have large tributaries entering from the northwest and few, very short streams entering from the southeast. The overall aspect of the topography does not change, just the stream pattern. This change presumably indicates a change in bedrock control. The contrast between the two patterns, dendritic and NSTA, is striking and is as well displayed here as anywhere else in the two county area.

A contrasting area is that shown in Figure 5 that shows East Finley Township. This township occurs in Figure 3 in southwest Washington County just west of the major drainage divide. In this area there is a dominance of NSTA as shown along Templeton Fork, Rocky Run, and (Enlow) Fork. The long tributaries entering these northeast-oriented streams are all entering on the north side of the major streams and are oriented either north or slightly northwest. Note in the northwesternmost part of the figure that the drainage pattern has a slight tendency toward dendritic, but still has some NSTA character. This drainage pattern has considerable contrast with that in Figure 4.

Because of the NSTA character of the landscape, the hilltops and adjacent valleys have an elongation not present in the Aleppo Township. However, the overall topography, except for the drainage pattern, is very similar to that of Aleppo Township. Upland elevations are generally 1,300 to 1,400 ft (396 to 427 m) and relief is generally 300-400 ft (91 to 122 m). The topography is smooth, hilltops rounded, slopes steep, and much of the topography is covered with woodland vegetation.

The third area shown is the Jefferson Township area in northeast Greene County with Muddy (Run) centered on the right (Figure 6). Here is an area with extensive woodland vegetation and mixed stream patterns with both dendritic and NSTA. Note that Muddy (Run) has long tributaries entering from the north and short tributaries from the south. Note in particular the many, very short drainages entering tributaries of all sizes from all directions. Close examination of the map will show that most of these are not pronounced drainageways, but merely slight indentations on the slope. These are probably not real water channels, but scars from landslide activity that will be discussed later.

Relief here is generally 300 ft (91 m) or slightly less, but locally gets up to 400 ft (122 m). Elevations are locally up to nearly 1,500 ft (457 m) and along the South (Fork Tenmile Creek) drop below 1,000 ft (305 m). More will be said about South (Fork Tenmile Creek) later. The overall appearance of the topography is the same as for the other map areas (Figures 4 and 5).

The overall character of the topography in the two-county area is similar throughout. Slopes are basically smooth in character and, on longer interfluve noses, project well into adjacent valleys (Figure 7). Shorter noses between adjacent smaller tributaries are steeper and project less far into the adjacent larger valley (Figure 8).

GEOLOGY

The geologic shaded relief map in Figure 9 shows the rock units underlying the landscape of the two-county area. The uppermost and primary rock unit is the Greene Formation. The

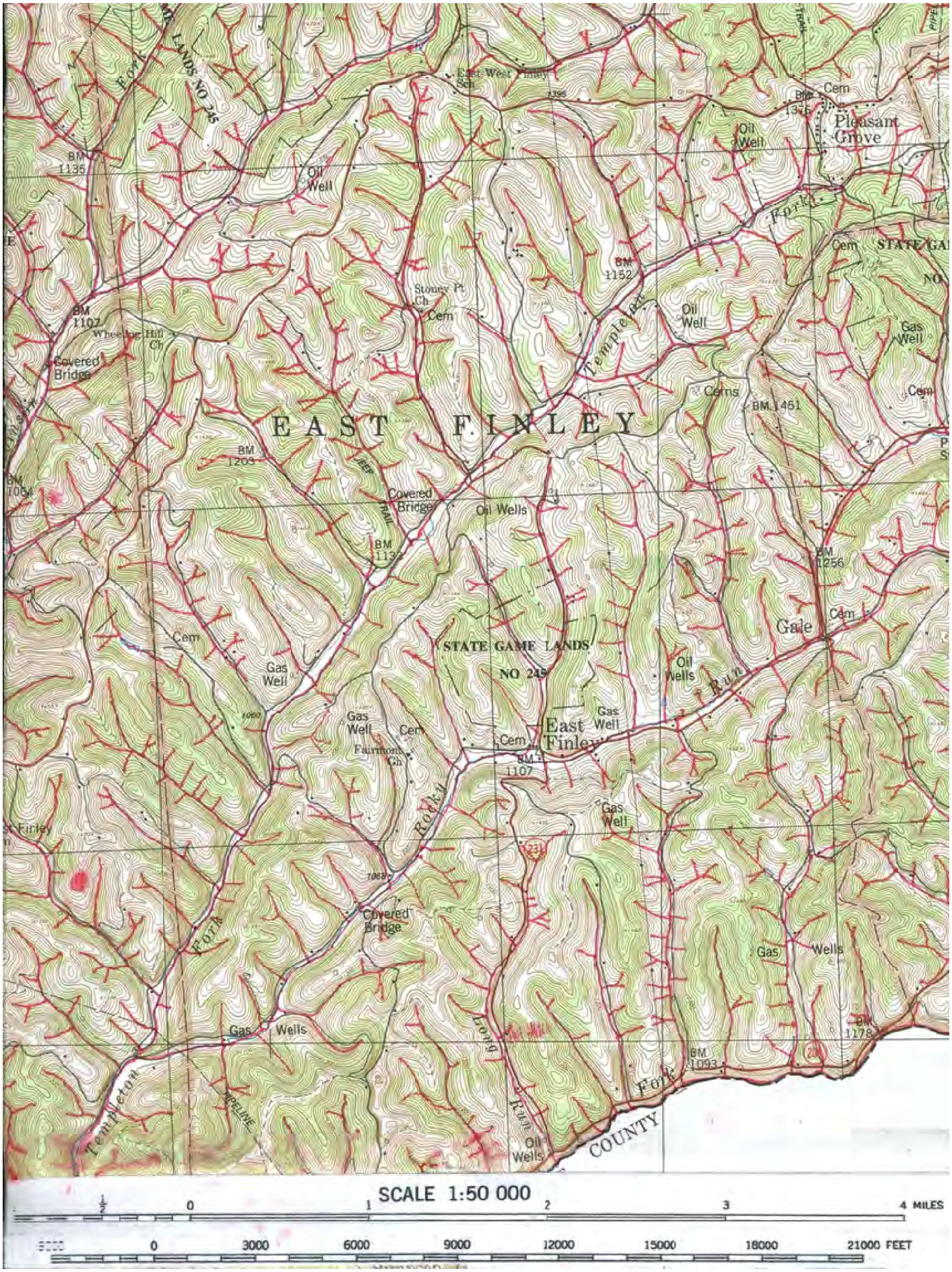


Figure 5. East Finley Township area of Washington County. Map is from the Washington County 1:50,000-scale topographic map. Red lines delineate valleys.

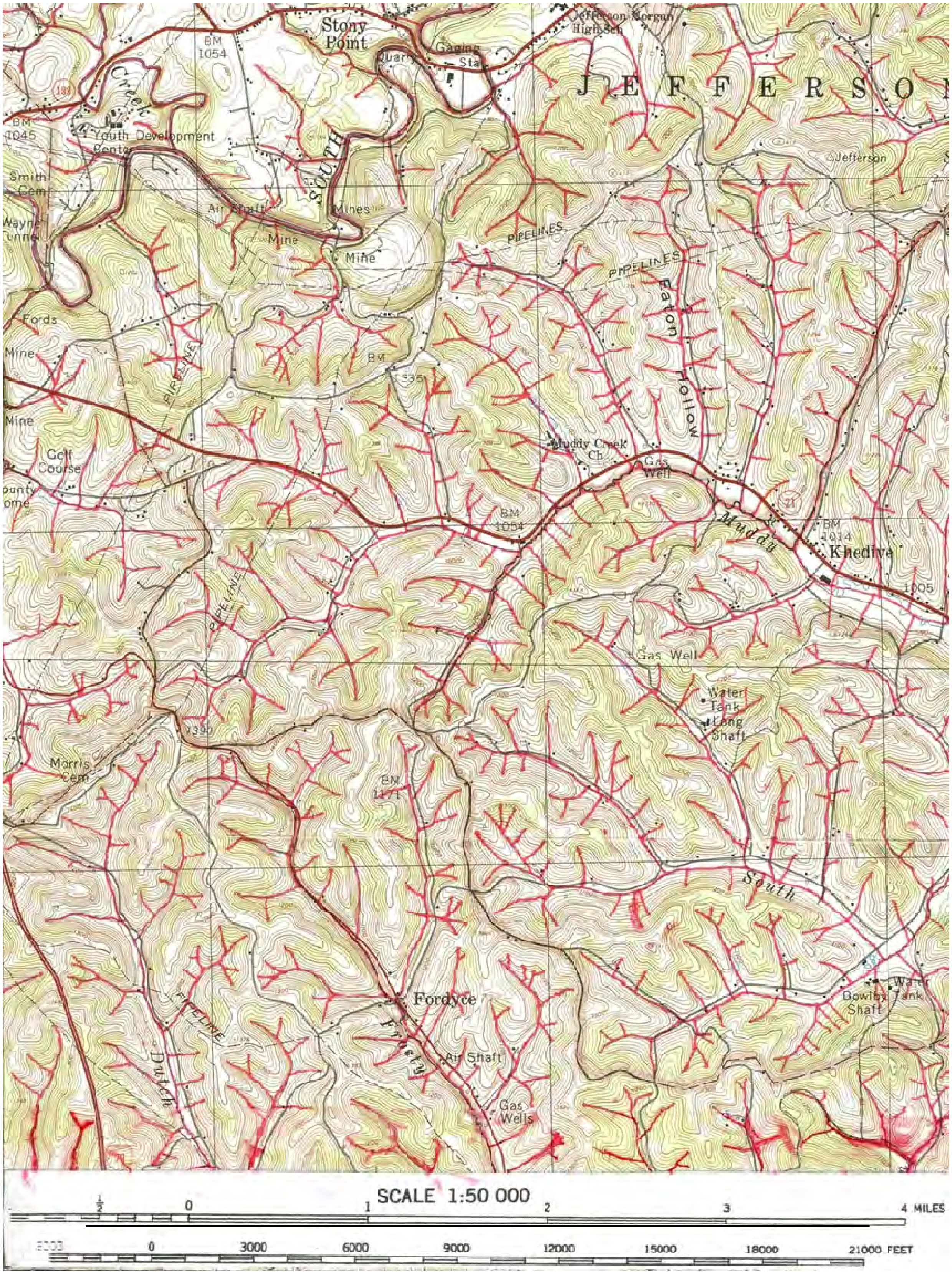


Figure 6. Jefferson Township, 1;50,000-scale topographic map, Greene County.

Figure 7. Long nose of typical terrain slope projecting into area between a larger valley, foreground, and a large tributary on right rear.

Figure 8. Here are the noses of two terrain slopes separated by small tributary with all coming into a larger valley. Note the floodplain top in the foreground, the very vague suggestion of an alluvial fan surface between the two noses, and the low colluvial slope between the floodplain and the lower nose surface.

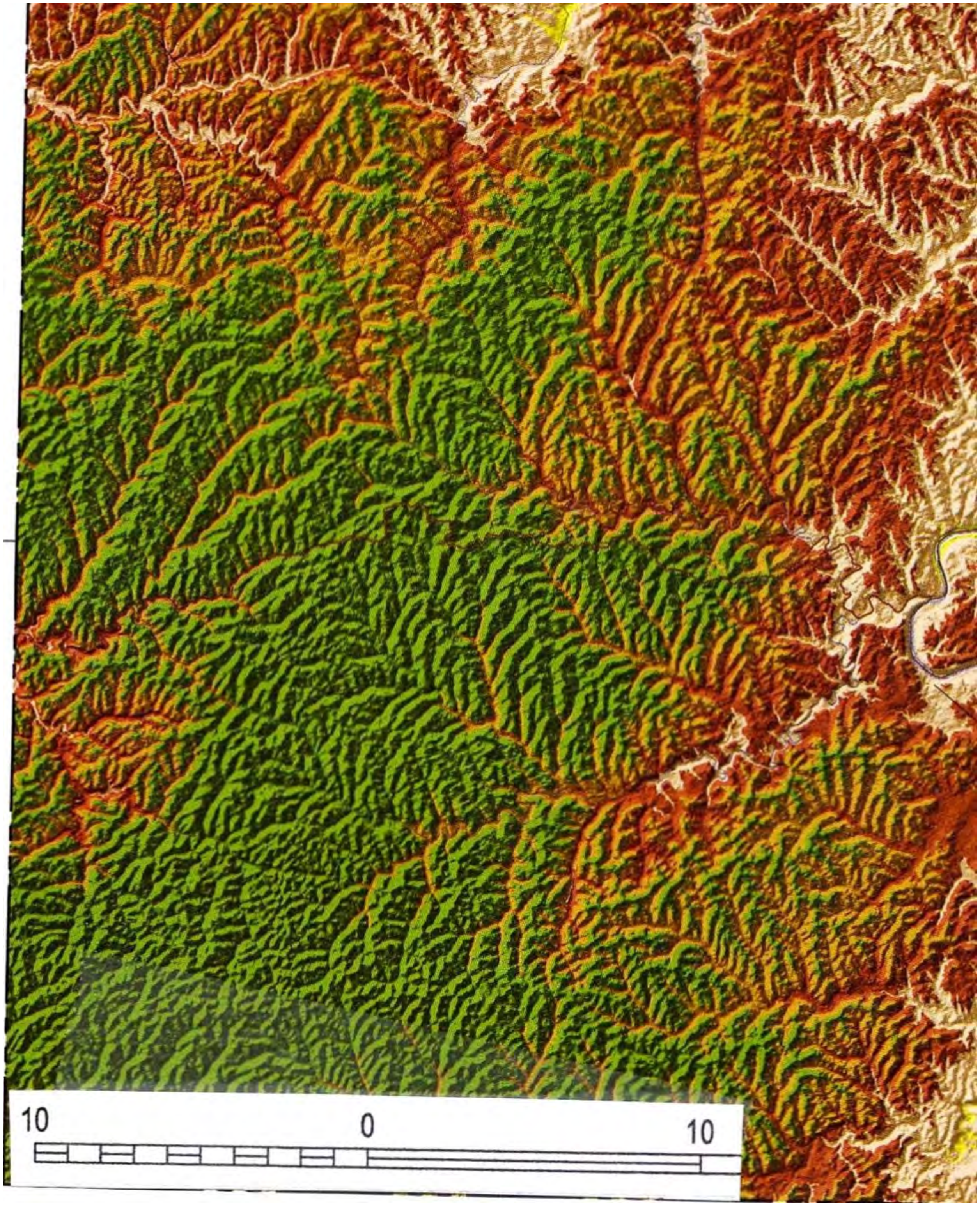


Figure 9. Geologic shaded relief map of the Greene and Washington counties area. Colors: green = Greene Fm., barely visible light brown = Washington Fm., red brown = Waynesburg Fm., and very light brown = Monongahela Formation. Map from Miles (2003). Compare the topography shape with that in Figures 4, 5, and 6. Scale is in miles.

Greene Formation consists of shale, some siltstone and sandstone, and minor amounts of shaly limestone, claystone, and thin coal. The various units are interbedded and the formation is up to 500 ft (152 m) thick at its thickest, but is generally thinner because of the lack of high elevations. Vertical fracturing is common. Underlying the Greene Formation is the Washington Formation that consists of cyclic sequences of sandstone, limestone, shale, claystone, some thin coal beds with thin underclay. The formation ranges from 160 to 234 ft (49 to 71 m) thick and has well developed closely spaced joints in the finer grained rocks and more blocky joints in the sandstones. The Waynesburg Formation underlies the Washington Formation and comprises cyclic sequences of sandstone, shale, limestone, siltstone, claystone, and coal. Joints are irregular in pattern, poorly formed, and highly abundant. The unit is 100 to 245 ft (30 to 75 m) thick. These first three units are very late Pennsylvanian to Permian(?) in age. The lowermost unit involved in the landscape development is the Pennsylvanian Monongahela Group that occurs only on the outermost margins of the area. The unit consists of cyclic sequences of shale, limestone, sandstone, siltstone, claystone, coal, and underclay. Bedding is well developed and joints are poorly to moderately well developed. Thickness is 270 to 350 ft (82 to 107 m). The above data is from Geyer and Wilshusen (1982.

The above described units are those upon which the landscape of Greene and Washington counties has developed. The Greene Formation is the primary unit. One feature very notable within the area is the lack of outcrop. Natural outcrops are essentially non-existent. Outcrops occur primarily in artificial road cuts and are mainly at lower elevations, particularly along larger stream valleys. A few outcrops occur along the margins of larger streams.

Because of the overall character of the rock in the area, it weathers rapidly and deeply. As a result loose, fine-grained soil material is developed. This material has moved down slope by colluviation and colluvial deposits thicken down slope. Figures 7 and 8 show slopes that are covered with colluvium of unknown thickness. In Figure 8 the lowermost end of the slopes merge gradually with the floodplain surface of the stream valley. An alluvial fan surface is barely detectable at the mouth of the tributary coming out of the valley between the two noses. In all valleys of size larger than miniscule, the colluvial slopes blend smoothly with the valley bottom that may or may not have a well defined floodplain. The mouths of these valley bottoms generally have a small alluvial fan that blends well into the floodplain surface of the larger valley into which the tributary merges.

The larger floodplains in larger valleys are generally being eroded by small permanent streams that often have cut-banks showing 2 to 3 ft (0.6 to 0.9 m) of alluvium. Most of this alluvium is fine grained, but sandstone cobbles often occur scattered in the alluvium. All told, the landscape of the two-county area is smooth surfaced with narrow hilltops, moderately steep to steep hillsides, narrow valley bottoms with no to some floodplain development, and an overall uniformity throughout the area. This uniformity can be attributed mainly to the character of the rock from which the landscape was eroded and the apparently consistent erosional process.

Going back to Figure 6, take a close look in the upper northwest part of the map at the South (Fork Tenmile Creek). Note the well developed meander pattern and the several meander cores. Note in particular the large abandoned meander immediately southeast of the word SOUTH. This large abandoned meander is real, the slopes are considerably colluviated, and the near perfect roundness is not common. The larger bottom area of the meander at 940 ft (287 m) elevation is 40 to 60 ft (12 to 18 m) above the adjacent stream. The road across the northwest edge of the meander has abundant road cuts in sandstone and there are also stream margin outcrops in the South Fork Tenmile Creek. The other meander cores are closer in elevation to the present creek level. This pattern occurs here and also on three other major drainages in the area: Dunkard Creek in southeasternmost Greene County, Tenmile Creek in southernmost Washington County, and Buffalo Creek in east central Washington County (see Figure 3 for locations). Each of these drainages with meanders and meander cores is both similar and different from the other. They represent an aspect of erosional history that is not obviously reflected in the remainder of the streams in the area. All told there is a complex erosional history here that I can't discuss because I do not have sufficient information or understanding. Such would make a fine research project.

LANDSLIDES

Greene and Washington counties are an area of extensive landslide activity despite the fact that landslides are not particularly visually evident. To get an idea about the interpreted extent of landslides, go to http://www.dcnr.pa.us/topogeo/hazards/landslides/slidepubs.aspk. Scroll down on this site to the landslide inventory maps section and pick an area. You will find a listing of 1:24,000-scale topographic maps on which are drawn the landslide areas interpreted from aerial photographs and rapid field checking by the USGS in the late 1970s and early

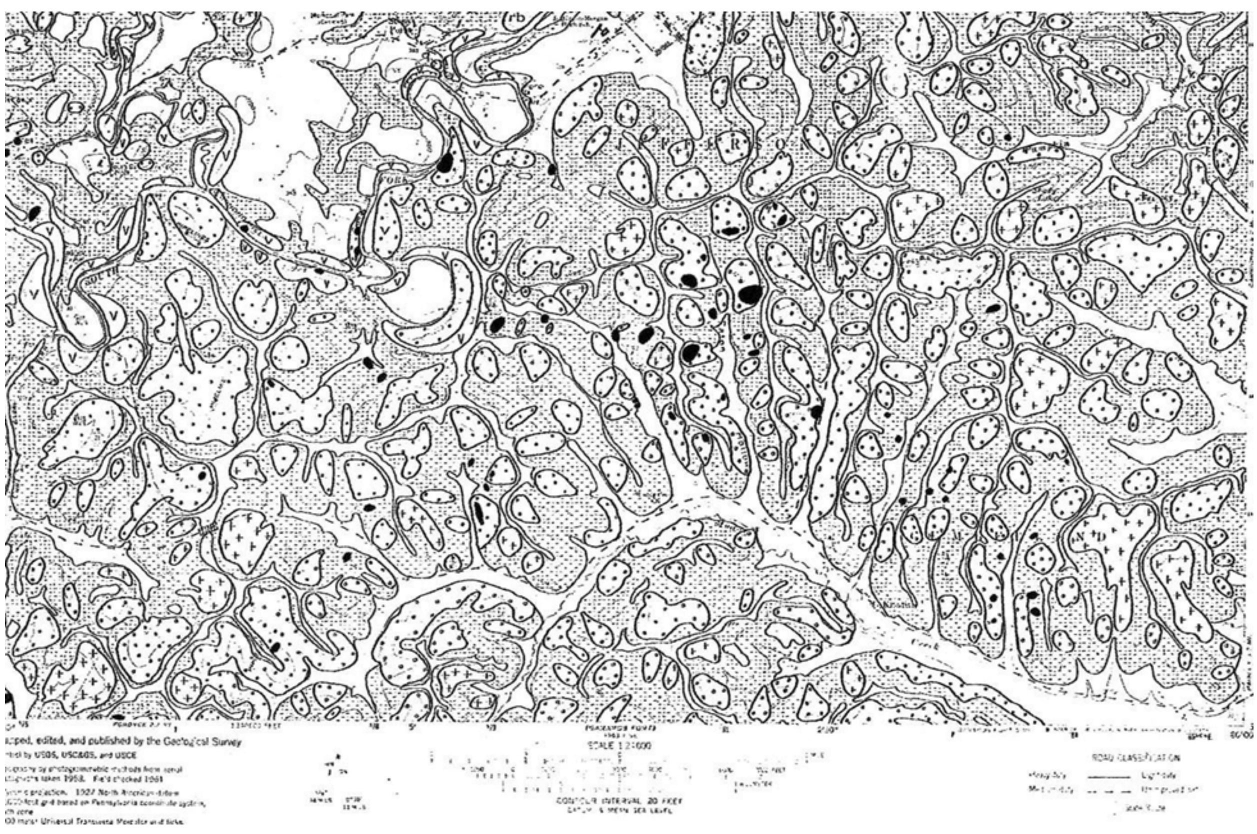


Figure 10. Part of the Mather 1:24,000-scale topographic map showing interpreted landslide areas. Solid black areas are active or recently active landslides. Enclosed areas with black dots are old landslide. Enclosed areas with plus signs are old and recent landslides. Enclosed areas with V's are colluvial slopes. (from Hackman and Thomas, 1978).

Figure 11. Excellent landslide with good hummocky terrain in foreground and more landslide higher on the slope in the extreme left.

1980s. Part of one of these maps, the Mather quadrangle, is shown in Figure 10. This map covers the same area as shown in the upper half of Figure 6. This map gives an idea about how extensive landslide activity is interpreted to be. A basic problem to the non-landslide expert is recognition of the landslide area in the field. This is particularly exacerbated by the abundant woodland vegetation that covers much or all of many slopes. However, some landslides are clearly visible and others can be interpreted readily if one knows what to look for.

Figure 11 is an excellent example of a landslide area on a typical slope. The hummocky area in the middle distance is all moved material. Note the lack of of woodland vegetation in the upper part of the slope above the landslide area. The area to the left of the main slide area shows considerable hummocky topography and is almost certainly more landslide area. In my short time of field work, this is the only really good landslide I saw.

Figure 12 shows a different landslide area. Note the hummocky area on the lower slope coming down into the tributary valley on the left side. Probably an old landslide, but possibly still moving. In the right foreground there is good evidence of very recent downslope movement that is well defined by the parallel lines of topography. A small area at the top of this movement suggests some subsidence. Upslope are some very low-relief slope irregularities that may be related to either old landslide movement or colluvial movement. Finally, note the trees on the left that come down to the valley bottom. The trees at the upper end of this mass have a different height than those immediately above. This I interpret as a landslide mass with a probable upper end scarp.

Figure 12. Landslide areas on the left lower slope, the right lower slope.

Figure 13. Foreground trees that come all the way to the valley bottom have a marked tree height separation at their upper end from the trees higher on the slope. This is interpreted to delineate the upper area of landslide activity for this landslide—tree marked height.

Figure 13 show an area of heavy woodland vegetation that comes down to the valley bottom but has open slopes on either side. This is interpreted as a landslide area. Note that the woodland area that goes upslope from the valley bottom has a distinct separation in tree height from the trees at the top of the slope. This separation is interpreted to demark the area of landslide separation from the original position. This may be a recent landslide. Similar tree height separations occur throughout the two-county area.

More information on development of landslides in the area can be found in Gray and Gardner, 1977.

SUMMARY

The landscape of the Greene and Washington counties area consists of a complex of narrow valleys with moderate to steep slopes separated by narrow uplands. The landscape has been eroded primarily into the variety of relatively soft rocks comprising the Greene Formation. The drainage system responsible for the erosion is a complex of minor tributaries joining larger tributaries that in turn join larger tributaries all of which eventually drain into the Monongahela or Ohio Rivers. The drainage pattern varies from dendritic to asymmetric, NSTA. In NSTA, main tributaries enter larger streams from the north because of a preference to erode up dip on harder rocks (Lattman, 1954; Sevon et al., 2006). The overall appearace of the landscape is much obscured because of extensive woodland vegetation. However, the general character of the hill and valley pattern is very consistent over the whole area. Of particular note is the large amount of landslide material throughout the area. Although often not visible because of the vegetation, the amount of landslide is extensive and ongoing. This area has a topography that is unique within the state.

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A BRIEF OVERVIEW OF LANDSLIDES IN GREENE AND SOUTHERN WASHINGTON COUNTIES, PENNSYLVANIA AND ADJACENT AREAS

H. L. Delano

Southwestern Pennsylvania and adjacent areas in West Virginia and Ohio are well known for a high incidence of landslides (Lessing et. al. 1976; Radbruch–Hall et. al., 1982). Although Pittsburgh and nearby areas underlain by redbeds in the Conemaugh Group are probably the best known landslide areas, southern Washington and Greene Counties have considerably higher densities of mapped slides. Greater urbanization in the Pittsburgh area leads to more damages to buildings and infrastructure, increasing costs and perceived level of incidence.

Landslide is a general term that encompasses a wide variety of types of slope movement, ranging from shallow, rapid earth and debris flows, usually triggered by concurrent rainfall, to deep-seated slow-moving slides and slumps in thick colluvial deposits or bedrock. Sudden rockfalls and slower moving complex slides that combine sliding and flow round out the list of most common slope movements in the area of our field trips. U.S. Geological Survey landslide inventory maps for Pennsylvania also identify areas of old landslides, colluvial slopes and other areas particularly susceptible to landslides. An introduction to landsides in Pennsylvania including types of landslides is in Delano and Wilshusen (2001).

The Pittsburgh Regional Environmental Geology Study in the 1970s mapped landsides and other features in six counties surrounding Pittsburgh. Washington County clearly stood out as having the largest density of mapped landslides (Pomeroy, 1982a). Quadrangle-scale landslide mapping of most of the rest of western Pennsylvania and portions of adjacent states was conducted by the USGS in the late 1970s and early 1980s (Hackman and Thomas, 1978).

Isopleth (contours of percentage of area covered by landslide deposits) mapping of landslide density in Washington County showed a pattern of high occurrence in southern Washington County that roughly follows the distribution of the Washington and Greene Formations (Pomeroy, 1978). In the high-density areas, between 50 and 80 percent of the surface was covered by landslide deposits. Later work in the Oak Forest quadrangle in Greene County reinforced the identification of the rocks above the Washington Coal with high incidence of landslides (Pomeroy, 1986). Detailed mapping in the quadrangle identified more than 1200 recent and 900 older landslides. The landslides are mostly small (less than 30 m [98 ft] in diameter), shallow (less than 3 m [10 ft] thick) earthflows and slump-earthflows in colluvial or residual non-red clay to clay-silt soils or weathered rock derived from claystone, mudstones and shale.

Kent (1972) noted an association between elevations of heads of landslides and limestone layers with interbedded layers of claystone and black shale in the Greene Formation and the upper limestone member of the Washington Formation. The limestones are generally highly permeable, the shales nearly impermeable, causing water to move laterally to the surface of the hillside where it can saturate colluvium. Pomeroy (1978) concurs that impermeable bedrock layers commonly control locations of slide initiation, but notes that limestones are not always present.

Delano, H. L., 2011, A brief overview of landslides in Greene and southern Washington Counties, Pennsylvania and adjacent areas, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 233-235.

Other factors noted in areas of high landslide density are steeper slopes (15 to 60 percent), high maximum relief (90 to 150 m [295 to 492 ft]), and presence of clay minerals illite, vermiculite, kaolinite and interlayered minerals (Pomeroy, 1978). Nakamura et. al., (2010) related residual strength in a variety of landslide soils to mineralogy of fine-grained (sub-425µm) soil fraction. They found that soils with total content of layer silicate minerals prone to preferred orientation (smectite, vermiculite, chlorite and mica) above 50 percent showed residual strengths less than 10 degrees. Although some earlier work related slide prone soils to content of minerals prone to shrinking and swelling with changes in moisture (Ciolkosz, et. al. 1979), the new data suggests that platy grain-shape that is easily aligned by shearing may be enough to explain low strength in landslide soils.

In southern Washington County, more than 75 percent of recent slides were on north-facing slopes, which tend to be steeper than their south-facing counterparts (Pomeroy, 1982b). Unlike slides in the Pittsburgh area, Pomeroy found that slides in southern Washington and Greene Counties were not commonly related to human activity such as cutting or loading slopes or changing drainage conditions. It will be interesting to see if this pattern continues as more development and disturbance of slopes occurs.

The Ohio River valley in Pennsylvania, the northern panhandle of West Virginia and adjacent Ohio is well represented in the landslide and engineering geology literature because of the need to remediate large landslides or design excavations and foundations in slide-prone materials derived from the Conemaugh and younger rocks. The industrial history of the valley is a major factor, with locks and dams, coal mining and handling facilities, mills and cities disturbing already unstable slopes. Significant papers include D'Appolonia et al. (1967), Gray and Gardner (1977), and Gray, et.al. (1979), which provide detailed descriptions of individual sites, geotechnical properties and remediation histories.

Radiocarbon dating of organic materials from old slide surfaces indicates that landsliding in the Ohio valley is at least as old as Pleistocene and continues to the present. These include materials older than 40,000 years at Weirton, WV, and about 9,750 years at the Pike Island Locks near Wheeling, WV (Gray, et.al, 1979).

Almost all of the preceding information is gleaned from work more than 30 years old. Recent availability of high-resolution Lidar-derived elevation data for all of Pennsylvania and Ohio offers opportunities to reexamine landslide inventories, and using GIS, to more easily analyze patterns and relationships. Development pressure from increased natural gas drilling and related activity may trigger inadvertent opportunities to study new slide events along roads and pipelines. There is plenty of opportunity for further investigations of landslides and slope stability issues in southwestern Pennsylvania and surrounding areas.

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"Let's move to Florida," she said. "I think Winter Park would be nice," she said. "It would be SO nice to retire there," she said. Yeah, so nice - so VERY nice!!!

A BRIEF DESCRIPTION OF THE MASONTOWN, PA KIMBERLITE INTRUSION

Henry S. Prellwitz and Michael Bikerman

INTRODUCTION

The Masontown kimberlite intrusion cuts through late Pennsylvanian sedimentary strata in the eastern portion of Fayette County, Pennsylvania near the village of Adah (Figure 1). The dike, which averages 1 m (3.3 ft) wide, has a vertical dip and a strike of N51°W. Sporadic surface outcrops can be found for about 3 km (1.9 mi) along strike. Exposures of the dike in the now abandoned underground coal mines were followed for over 5 km (3 mi) (Hickock and Moyer, 1940). A pre-existing fault zone, normal to the axes of the regional folding, provided a conduit for the intrusion (Roen, 1968).

The mineralogy of this kimberlite is consistent with other kimberlites found in the Eastern United States. The Masontown kimberlite contains sedimentary, metamorphic, and igneous rock xenoliths; some of the peridotite xenolith samples may represent rocks that occurred near the base of the continental lithosphere. Geochronology studies indicate a complex Mesozoic age for the kimberlite emplacement. See below.

This kimberlite was first described by Kemp and Ross (1907), followed by a more thorough report by Smith (1912), aided by many new exposures resulting from underground coal mining in the Pittsburgh seam. Sosman (1938) calculated an intrusion temperature of 550° to 600° C (1,022 to 1,112° F), based on laboratory coal coking experiments. Hickock and Moyer (1940) described the mineralogy of the kimberlite, as part of a Pennsylvania Geological Survey County Report. A more detailed description of the minerals in this kimberlite was provided by Hunter and Taylor (1983 and 1984) along with some trace element geochemical work on the phlogopite mica and garnets. A further petrographic description of the kimberlite and the contained xenoliths was given by Prellwitz (1994), and Prellwitz and Bikerman (1994).

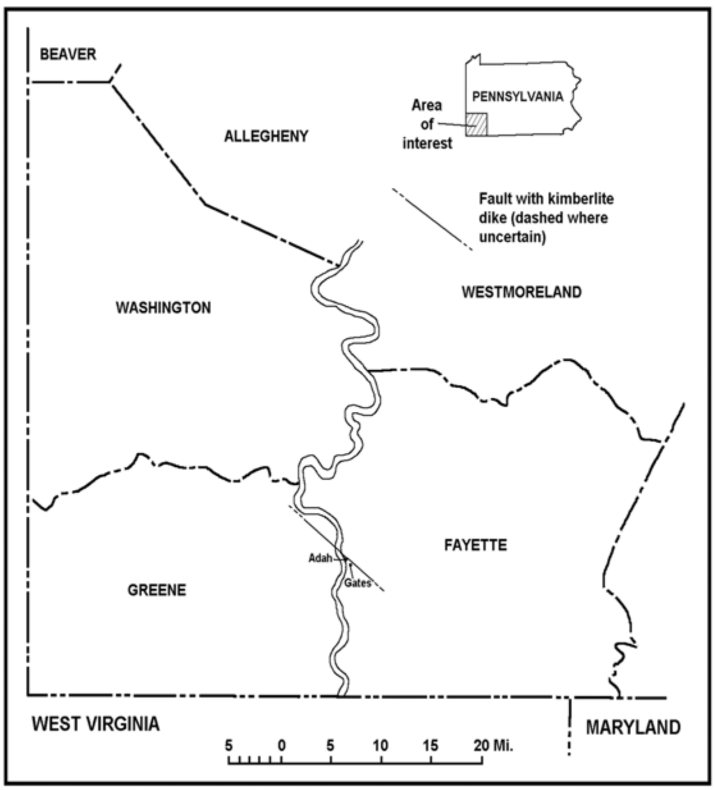

Figure 1. Map of southwestern Pennsylvania showing the location of the

Prellwitz, H. S. and Bikerman, Michael, 2011, A brief description of the Masontown, PA kimberlite intrusion, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 236-242.

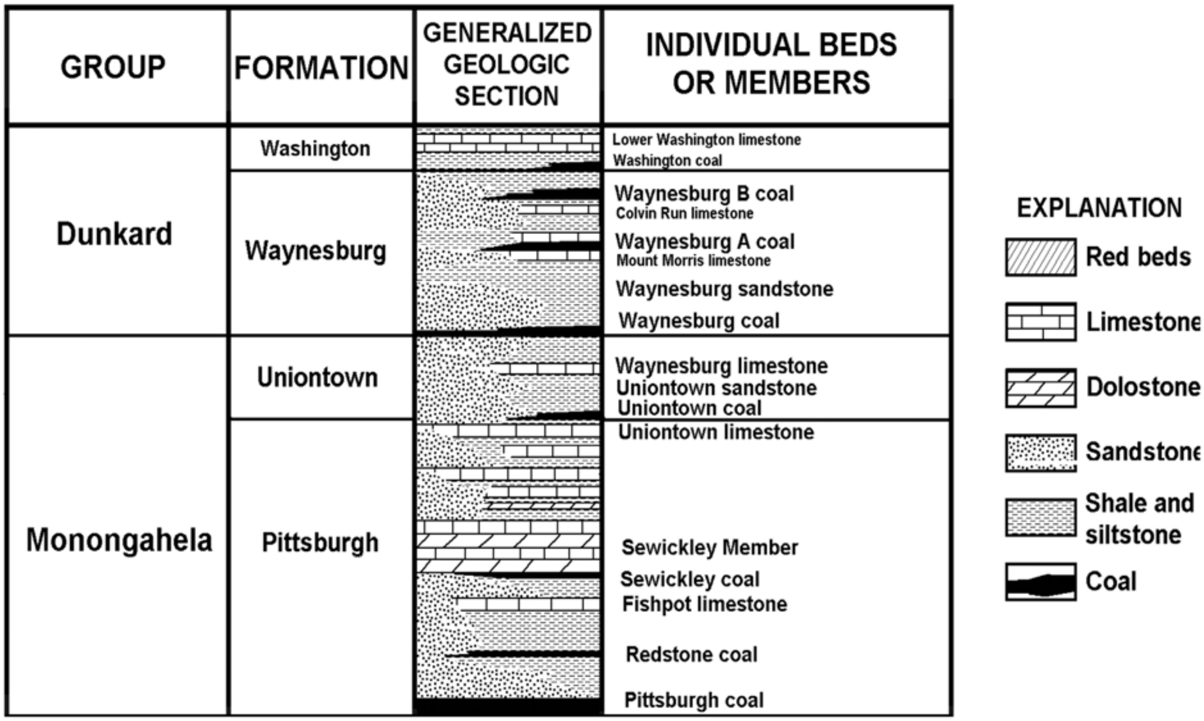

Figure 2. Generalized geologic column of the country rock exposed in the area of the Masontown kimberlite dike.

GEOLOGICAL SETTING

The Masontown kimberlite intrudes through sedimentary rocks of the upper Pennsylvanian Monongahela Group and Waynesburg Formation (Figure 2). These beds consist mostly of shale, siltstone, sandstone, fresh-water limestone, and coal.

The attitude of the kimberlite dike is vertical, and strikes N51°W in a pre-existing fault zone (Roen, 1968). Other parallel fracture zones can be seen in the immediate area, and one contains a small 6 cm (2.5 in) wide kimberlite dike that is about 30 m (100 ft) NE of the main dike. The main kimberlite dike averages 1 m (3.3 ft) wide (Figure 3). Outcrops are scarce, as the kimberlite decomposes at a faster rate than the surrounding country rock. There is no field evidence of contact metamorphism, except in an outcrop of the Waynesburg Coal, which was coked slightly from the heat of intrusion. Shale, siltstone, sandstone, and limestone contact areas show no mineral changes in thin section.

Some portions of the kimberlite have been extensively hydrated, while other parts of the dike appear fresh and unaltered. The more altered sections of the kimberlite weather and decompose more quickly than the unaltered areas; most “outcrops” of the kimberlite appear as trenches filled with red-orange mud, representing the decomposed dike. The few rare outcrops consist of the more competent unaltered portions of the dike (Figure 3). Occasional surface outcrops have been traced for more than 3 km (1.9 mi) (Prellwitz 1994), and over 5 km (3 mi) in the abandoned underground coal mines. Northwest of the kimberlite surface outcrop area, the fault itself (devoid of igneous rock) can be seen in the south bank of Muddy Creek (Roen, 1968).

Figure 3. Photograph of an unaltered portion of the kimberlite dike in outcrop. Rock hammer for scale.

GEOCHRONOLOGY

Age determination on igneous rocks from a deep-seated origin is a tricky proposition. The parent isotopes incorporated in the deep source generated significant daughter isotopes during the lengthy pre-eruption residence time. Depending on the nature of the host material - mineral, magma, or gas - the daughter isotopes may be retained entirely in their host which, for a robust mineral, would later give an apparent age of first formation of the mineral rather than the eruption or emplacement age. Alternatively source materials open to migration of parent and/or daughter *in the magma chamber* but not after eruption may be datable for the eruption time. To have a valid date any carry over daughter would either have to have expelled in the eruption, or be accounted for in the analysis and calculation of the date. Finally any true age would not have had any change in isotope composition post emplacement. Of course many intermediate possibilities exist often without a ready way of identifying them. The results of the many dating attempts summarized below suggest that this was true to some degree.

K-Ar on phlogopite	Dates in Ma		References
Two coarse	368+/-18	408 +/- 20	Zartman et al., 1968
fine	184+/- 10		Pimental et al., 1975
Coarse / very fine	353 +/- 2.2	147 +/- 1.5	Prellwitz, 1994
$^{40}Ar/^{39}Ar$ on fine phlogopites – laser step heating		161 to 176	Bikerman & Phillips, 2000
$^{40}Ar/^{39}Ar$ on fine phlogopites – laser spots		149 to 167	

Rb-Sr dating - phlogopite	149 +/- 5		Alibert & Albarede, 1988
Coarse/fine	188 +/-0.7	170 +/- 1.3	Bikerman et al., 1997

Sm-Nd scatterchron [garnet, calcite, whole rock]	145+/- 11	Bikerman et al., 1997

Since the emplacement age must postdate the Early Permian or about 280 Ma the 147 +/- 5 average of K-Ar, $^{40}Ar/^{39}Ar$, Rb-Sr and Sm-Nd is possibly the best estimate of the actual final emplacement of the dike. Possibly the concentration of dates in the 160 to 180 Ma range reflects incomplete purging or closure of some isotopic systems prior to and during eruption, or as Bikerman and Phillips have postulated (2000) there might have been two eruptions – one around 180 Ma and the other at 147 Ma with some partial resetting of systems in the interim. For most purposes the latter date may be used as the emplacement age.

MINERALOGY AND PETROGRAPHY

The major minerals included in the Masontown kimberlite are olivine, phlogopite mica, titanium-rich ilmenite, magnetite, pyrope garnet, and perovskite in an aphanitic carbonate groundmass. The olivine and phlogopite both occur as phenocryst, xenocryst, and groundmass phase minerals. Alteration minerals include serpentine (from alteration of olivine) and secondary vein-filling calcite.

The olivines are magnesium-rich, and compositions range from Fo81 to Fo93. Most of the fresh, unaltered olivines are rimmed with magnetite. Many olivine crystals are partially or completely altered to serpentine, especially when in close proximity to secondary calcite vein material. The fractures, now filled with the calcite, probably provided a conduit for hydrating solutions to move through the kimberlite while still hot. Phlogopite mica occurs as phenocrysts and groundmass lathes, and is essentially unaltered, except for some minor local chloritization. Titanium ilmenite is seen as phenocrysts, and was mistaken in earlier reports as coal inclusions. The pyrope garnets, usually a deep blood red, are surrounded by a kelyphitic alteration rim, that is mostly chlorite, and very small perovskite grains (< 1 mm.) are scattered in the calcite groundmass.

Earlier reports (Hickock and Moyer, 1940) interpreted the calcite groundmass as an alteration product, produced by the contact of the ascending kimberlite with sedimentary carbonate beds. However, carbon and oxygen isotope evidence from a related kimberlite in Indiana County, PA indicate a primary, igneous carbonate source versus a sedimentary calcite (Deines, 1968). The initial $^{87}Sr/^{86}Sr$ ratios of the Masontown Dike whole rock [0.70513] and of an included vein calcite [0.70365] are similar to primary igneous and well below the normal limestone values (Bikerman and others, 1997). These findings are consistent with similar data from African kimberlites.

The texture of the kimberlite is porphyritic, with olivine up to 5 cm (2 in), phlogopite up to 4 cm (1.5 in), ilmenite up to 2 cm (0.8 in), and pyrope garnet up to 1.5 cm (0.6 in) as the phenocryst and xenocryst phases. The groundmass is aphanitic calcite, with small perovskite grains, and some rare apatite crystals. Small phlogopite mica lathes are found in the

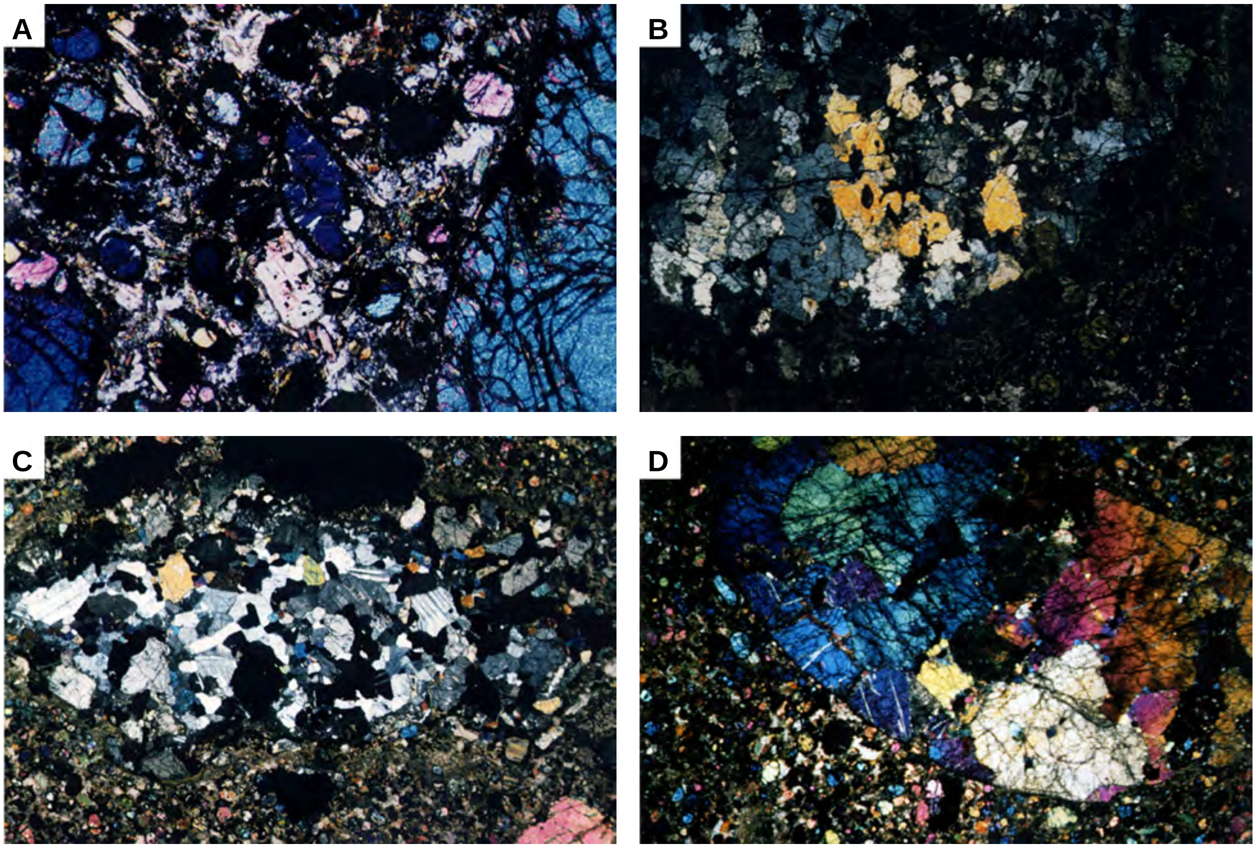

Figure 4. Photomicrographs of the Masontown kimberlite, all under cross polarizers. A – Unaltered kimberlite. Field of view = 4 mm. B – Granitic gneiss xenoliths in the kimberlite. Xenolith is 2 cm (0.8 in) long, and the quartz grains have an anomalous yellow color due to an overly thick slide. Field of view = 4 mm. C – Eclogite xenoliths in kimberlite. Field of view = 4 mm. D – Spinel dunite xenoliths in kimberlite. Field of view = 3 cm.

groundmass (up to 1 mm) and are usually aligned with the flow direction of the kimberlite. The small mica lathes are also seen in the contact area between a xenolith or large phenocryst, tangentially encircling the larger crystal. The large purple and blue crystals in Figure 4A are olivine with magnetite rims, the pale laths and blocks are phlogopite mica, and the groundmass is calcite. There is no evidence of hydration in this sample.

XENOLITHS

The xenoliths found in the Masontown kimberlite represent all three groups of rocks – sedimentary, metamorphic, and igneous. These xenoliths also provide a "window" to the rock types that would be encountered if one could drill a very deep core in southwestern Pennsylvania.

Numerous sedimentary xenoliths were found when kimberlite material was broken or slabbed. Many xenoliths of the Uniontown shale (Monongahela Group – see Figure 2) are located near the dike walls; sedimentary rock samples from greater depths include several samples of an oolitic limestone, and a very coarse immature quartz sandstone. Inferring what the host formation depth is from xenolith samples is very difficult, and interpretations are tenuous at best! Since there are no oolitic limestone occurrences at the surface in southwestern

Pennsylvania, a depth lower than the Pennsylvanian System is inferred. A large cobble-sized xenolith of very coarse quartz sandstone was encountered in the kimberlite; the quartz grains are angular, and the sample shows evidence of some applied stress. The cobble is divided by slickensided surfaces, and contains a limestone xenolith with a Silurian coral fossil. This is an odd example of a xenolith within a xenolith in the kimberlite. This sandstone could represent the Lower Devonian Oriskany Sandstone.

The minerals in the granitic gneiss xenolith, shown in Figure 4B**,** are quartz, albite, biotite, and minor microcline. The sample from which the thin section was made clearly shows typical gneissic banding, and most probably represents the Grenville basement that is buried at least 3 mi (5 km) deep in southwestern Pennsylvania.

Another xenolith type commonly encountered in the kimberlite is eclogite. The minerals in Figure 4C are pyroxene and garnet; the many 120^{o} grain boundaries suggest a metamorphic rock. The composition of this mafic sample contrasts greatly with the granitic gneiss xenolith. This specimen may represent material in the continental crust that is below the MOHO boundary, and can be interpreted as the highest grade metamorphic rock type known.

The most common xenoliths found in the Masontown kimberlite are peridotites. One study examined over 30 samples (Prellwitz, 1994) and classified them into several groups including garnet lherzolites, spinel dunites, harzburgites, and dunites. The peridotite samples could represent material from the bottom of the lithosphere, at about 75 km (47 mi) depth. Temperature/pressure/mineral assemblage studies (Hunter and Taylor, 1984) indicate that the kimberlite itself could have originated at a 140 km (87.5 mi) depth, and may represent an asthenosphere melt. Figure 4D is a dunite containing olivine and a few very small blebs of spinel, which appear black in this figure. The olivine grains show a slight amount of undulatory extinction, indicating some strain was applied to this sample in its past history. Note the sharp contact between this xenolith and the surrounding kimberlite; there is little alteration, in contrast to alteration seen along the boundaries of more sialic xenoliths.

RECOMMENDED FURTHER STUDIES

As always, when a new study is initiated, and a particular topic is to be resolved, many new questions arise, which include:

- What is the carbon/oxygen isotopic composition of the calcite groundmass?
- Can the reason for conflicting old isotopic dates be resolved unambiguously?
- What are the radiometric dates on the xenolith minerals?
- How would trace element analysis on xenolith minerals compare with kimberlite minerals?
- Is there a variation in trace elements or isotopic ratios at the contacts?
- What would a true emplacement temperature be?
- Was this rock intruded as a liquid or a CO_2 rich gas?

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PRE-CONFERENCE FIELD TRIP: MASONTOWN KIMBERLITE DIKE, FAYETTE COUNTY, PA

Leaders: Henry S. Prellwitz and Michael Bikerman

Participants will meet at the front entrance of the Ramada Inn in Washington, PA, Thursday September 29, 2011, at 12:00 noon, and will return by 5:00 PM. Space will accommodate a minimum of 5 and a maximum of 20 persons. If you want to examine the streambed exposures in comfort, please bring waterproof boots or other footgear suitable for wading in up to one foot of water. The tributary valley outcrop(s) will require strenuous climbing in some places; these valleys have steep walls. Be prepared for about one mile of walking in difficult terrain. If you want to collect samples, you will need a heavy hammer, unless you want only small chips. Please do not hammer on the *in situ* kimberlite outcrops, as we want to preserve these exposures for the future. We will be carpooling, so we will try to arrange transportation by e-mail/phone prior to the trip. Large-scale maps of the area showing outcrop locations will be provided.

ROAD LOG

Miles Int.	Cum.	
0.0	0.0	**Leave** the Ramada Inn, Washington, PA, 12:00 Noon
0.3	0.3	Turn **Left** on Federal Route 40
0.3	0.6	Exit on to I-70, pass through Washington, PA
5.8	6.4	Exit from I-70 East on to I-79 South, towards Morgantown
19.9	26.3	Exit from I-79 South at PA Route 21
0.2	26.5	Turn **Left** on PA Route 21 East
9.0	35.5	Turn **Right** on PA Route 21 East, in Carmichaels, PA
2.6	38.1	Intersection with PA Route 88
1.0	39.1	Power plant on left, once served by the Nemacolin Mine
0.6	39.7	Cross Monongahela River
2.7	42.4	Turn **Left** on State Route 166 North, Ball Hill Road
0.5	42.9	Pass through village Edenborn, PA
1.3	44.2	Turn **Left** on Middle Run Road
0.9	45.1	Outcrops of Uniontown Limestone on right
1.8	46.9	Intersection of Croushare Road on right, park cars here.

STOP 1

From the Croushare Road parking area, walk approximately 130 feet downstream in Middle Run. Two small (6” and 1” wide) outlier dikes are exposed in the streambed, on the

Prellwitz, H. S. and Bikerman, M., 2011, Pre-conference field trip: Masontown kimberlite dike, Fayette County, PA, *in* Harper, J. A., ed., Geology of the Pennsylvanian-Permian in the Dunkard basin. Guidebook, 76th Annual Field Conference of Pennsylvania Geologists, Washington, PA, p. 243-244.

North bank, intruding shale. ***Please! No hammers at this site, as it is small and fragile.***

STOP 2

From the Stop 1 exposure, continue to walk downstream in the bed of Middle Run, about 120 feet. This exposure (if we are lucky) will show the complete width of the main kimberlite dike in the North bank of Middle Run. ***Please! No hammers at this site.***

Int.	Cum.	
0.5	47.4	Proceed up the hill on Croushare Road, and park cars in the public hunting parking area on the left.

After inspecting this exposure, we will walk back to the parked cars at the Croushare Road intersection.

STOP 3

We will walk approximately 1,000 feet west into a ravine, where there is a good exposure of the full width of the intrusion. ***Please do not hammer on the in-place exposure.*** Below the exposure are several large blocks of kimberlite, if one wants to obtain specimens. There is plenty of specimen material here.

Return trip to the Ramada Inn is the reverse of the outgoing trip.