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Abstract

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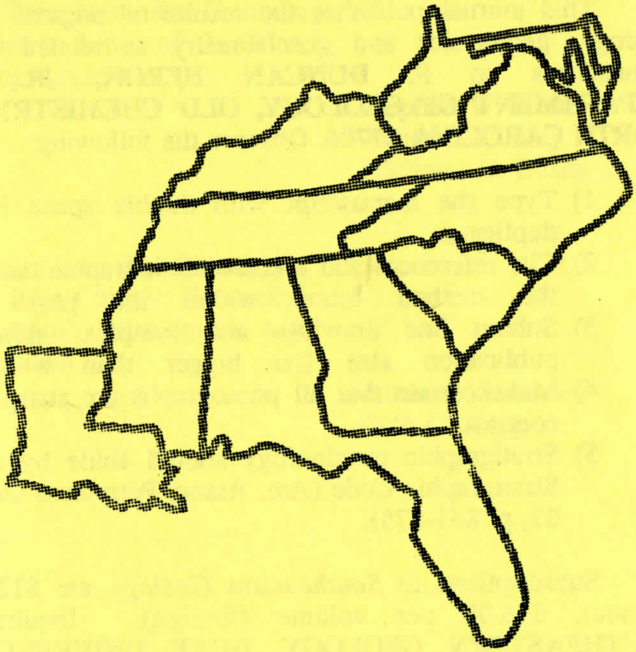

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SOUTHEASTERN GEOLOGY

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TECTONIC SUBSIDENCE HISTORY OF THE CENTRAL APPALACHIAN BASIN AND ITS INFLUENCE ON PENNSYLVANIAN COAL DEPOSITION

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ABSTRACT

The Central Appalachian basin is characterized by a complex geologic history. Tectonic subsidence curves calculated from five wells indicate that this basin initially subsided rapidly as part of the rifting cycle of the North American continent and opening of the Proto-Atlantic Ocean before establishing a relatively slow subsidence rate that was increased periodically by loading caused by collision events corresponding to the Taconic, Acadian, and Alleghenian-Hercynian orogenies. After Proterozoic rifting of Grenville basement between 820 and 610 Ma, the basin underwent thermally controlled subsidence as the Proto-Atlantic Ocean widened. Collisions initiated during the Taconic orogeny (480 to 435 ma) caused development of a foldbelt and foreland basin that was enlarged during both the Acadian (380 to 340 ma) and Alleghenian-Hercynian (330 to 230 ma) orogenies. Tectonic subsidence analysis indicates that post-Taconic basin subsidence was controlled predominantly by lithospheric flexure. The subsidence rate between the Taconic and later orogenic events decreased with time as progressively more rigid crust became involved in basin subsidence. Subsidence curves, sediment accumulation rates, stratigraphic patterns, and sea-level history fit this model. The Central Appalachian basin rock succession is subdivided into lithologically defined sequences the boundaries of which are controlled by tectonic events. Sea-level fluctuations within the Central Appalachian basin collision margin appear to be tectonically controlled, with minor superposed eustatic components.

During Pennsylvanian time, basin subsidence rates ranged from 0.63 to 4.6 m/my, but were particularly slow (0.63 to 0.98 m/my) during Atokan and Desmoinesian time when the thickest and most laterally extensive coals accumulated in the Central Appalachian basin. Occurrence of larger volumes of inertinite content suggests that Desmoinesian and older coals were more oxidized relative to other Pennsylvanian coals in this basin, and sea-level change was minimal. Slow rates of tectonic subsidence combined with a relatively wet tropical climate to control development of these thick, oxidized coals in this particular case and may well control the distribution of thick and laterally extensive coals in other foreland basins.

INTRODUCTION

One of the axioms of coal geology is that foreland basins contain the thickest coal beds and the largest volume of coal resources (Strakhov, 1969; Jones and others, 1987, amongst others). The development of thicker coal is favored in orogenic recesses of foreland basins, as shown by Jones and others (1987), who provided evidence both from the Late Permian Sydney basin of Australia and Pennsylvanian coal fields in the Appalachian basin.

The Appalachian basin of eastern North America contains extensive coal deposits, but it is a successor foreland basin with a longer-term complex history (See Hatcher, 1972; Beaumont and others, 1982; Rodgers, 1983; Tankard, 1986a,b; and Klein, 1987, for summaries). Prior work (Hatcher, 1972; Beaumont and others, 1982; Rodgers, 1983) suggested that both extensional and subsequent compressional forces controlled subsidence of the Appalachian basin. This basin extends from the Adirondack Uplift in New York State southwest into Alabama and contains up to 18 km of sediment (Colton, 1970;Quinlan and Beaumont, 1984; Milici and deWitt, 1988). This long stratigraphic record provides a means to evaluate the control of different tectonic forces on basin subsidence and its role in coal formation and preservation.

The objective of this paper is to analyze the tectonic subsidence of the Appalachian basin using backstripping and decompaction procedures (Steckler and Watts, 1978; Sclater and Christie, 1980; Bond and Kominz, 1984; Sawyer and others, 1987) and to establish the relative importance of both thermal subsidence and crustal stretching (McKenzie, 1978) and foreland flexural loading (Quinlan and Beaumont, 1984) in controlling both tectonic and sedimentation events during the evolution and development of the Central Appalachian basin. An additional objective is to use tectonic subsidence analysis as a test of earlier hypotheses concerning the evolution of the stratigraphic sequences of the Central Appalachian basin and to determine the role of basin subsidence processes and collision on the preserved record of sea-level change. Our final objective was to determine to what extent tectonic subsidence rate and history in this basin influenced the origin and preservation of economic, thick coals.

SUMMARY OF TECTONIC AND SEDIMENTARY HISTORY

Both the sedimentary and tectonic history of the Central Appalachian basin are summarized in Table 1. These sediments were subdivided into separate lithic sequences by Colton (1970; see also Milici and deWitt, 1988) that are referred to herein as "Colton Sequences". Late Precambrian (ca 630 Ma) extension and rifting of North America caused the development of grabens (Bond and others, 1984), which were filled with nonmarine and marine clastic sedimentary rocks before formation of the Proto-Atlantic Ocean caused accumulation of Catoctin Greenstone volcanics over the clastic sediments (Fichter and Diecchio, 1986). These volcanics represent the top of Colton's (1970) Precambrian basement complex. Shallow-marine clastic sedimentary rocks of the Cambrian Chilhowee Group (Simpson and Sundberg, 1987) subsequently were deposited on thermally-subsiding crust until such subsidence slowed, and younger Cambrian sediments were deposited on a stable continental margin (Bond and others, 1984). Rocks deposited on the crust undergoing more rapid thermal subsidence are grouped in the Upper Precambrian stratified sequence (Colton, 1970).

Table 1. Comparison of Orogenic, Stratigraphic, and Depositional Events in the Central Appalachian Basin.

Period	Age ticks	Orogenic events and Appalachian basin phases [1]	Cratonic sequences [2]	Colton sequences [3]	Lithology [4]	Depositional conditions [5]
PENNSYLVANIAN	280, 300, 320	ALLEGHENIAN-HERCYNIAN OROGENY	ABSAROKA SEQUENCE	PENNSYLVANIAN SEQUENCE	Sandstones, shales, limestones, coals. Thickest and coarsest in eastern part of basin.	Alluvial/deltaic systems
MISSISSIPPIAN	340	ACADIAN OROGENY	KASKASKIA SEQUENCE	MISSISSIPPIAN SEQUENCE	Conglomerates, sandstones, shales, limestones. Coarsest in NE, fining to NW.	Alluvial/deltaic system; covered by marine sediments
DEVONIAN	360			DEVONIAN CLASTIC SEQUENCE	Sandstones, siltstones, and shales with local carbonates. (Catskill Delta)	Massive alluvial/ deltaic system
DEVONIAN / SILURIAN	380, 400		TIPPECANOE SEQUENCE	SILURIAN-DEVONIAN CARBONATE SEQUENCE	Central Basin: sandstones, siltstones, and shale. Surrounding areas: shale, limestone, dolomite, anhydrite.	Central Basin tidal flats; later transgressive shelf sedimentation
SILURIAN	420, 440			SILURIAN CLASTIC SEQUENCE; UPPER ORDOVICIAN CLASTIC SEQUENCE	Conglomerates, sandstones. Silurian sediments fine up and to west; carbonates predominate in west.	Alluvial/deltaic systems into NW
ORDOVICIAN	460, 480	TACONIC OROGENY; CLOSURE OF IAPETUS OCEAN				
ORDOVICIAN / CAMBRIAN	500, 520, 540	STABLE DIVERGENT CONTINENTAL MARGIN; DEVELOPMENT OF COASTAL PLAIN WEDGE AND CONTINENTAL SHELF SEQUENCE	SAUK SEQUENCE	CAMBRIAN-ORDOVICIAN CARBONATE SEQUENCE	Limestone unconformably on on dolomites, associated sandstones.	Carbonate shelf on western margin of Iapetus; basal Middle Ordovician unconformity caused by uplift of Waverly Arch and early collision associated with Taconic Orogeny
PRECAMBRIAN	560, 580	SUBMERGENCE OF CONTINENTAL MARGIN		LOWER CAMBRIAN CLASTIC SEQUENCE	Quartzose and arkosic sandstones; local limestone.	Transgressive shelf of sandstones
PRECAMBRIAN	600	THERMAL SUBSIDENCE OF THINNED CONTINENTAL MARGIN	? ? ? ?	UPPER PRECAMBRIAN STRATIFIED SEQUENCE	Quartz arenites and mudstones overlying basalts and arenites.	Alluvial plain, nearshore and shelf sediments
PRECAMBRIAN	620, 640	OCEAN CRUST FORMATION; RIFT VALLEY MARINE FILL; RIFT VALLEY NON-MARINE FILL		PRECAMBRIAN BASEMENT COMPLEX OF METAMORPHOSED SEDIMENTS	Volcanics overlying non-marine and marine clastic rocks.	Subareal rift basalts overlying metamorphosed clastics representing earliest rift fill

1 Glover, et al., 1983; Price & Hatcher, 1983; Fichter and Diecchio, 1986; Secor, et al., 1986; Simpson & Sundberg, 1987.

2 Sloss, 1963.

3 Colton, 1970.

4 Colton, 1970; Fichter & Diecchio, 1986; Simpson & Sundberg, 1987.

5 Wagner, 1966; Oliver, et al., 1967; Colton, 1970; Meckel, 1970; Smosna, et al, 1978; Wrightstone, 1984.

As the Proto-Atlantic continental margin stabilized and submerged, a clastic wedge of sediments (Lower Cambrian clastic sequence) was deposited on the shelf of the northwesterly transgressing Iapetus Ocean. During Middle Cambrian time, carbonate deposition was initiated, resulting in a massive carbonate shelf that thinned to the west (Cambro-Ordovician carbonate sequence; Colton, 1970). The eastern edge of this carbonate shelf marks the original edge of the North American continent (Rodgers, 1968) and is associated with a steep Bouguer gravity gradient in the Piedmont Plateau province (Thomas, 1983; see Figure 1). During the Early to early Middle Ordovician, a reversal of spreading occurred, causing the eastern edge of the North American continent to change from a passive margin to a compressional convergent margin. This initiated the Taconic orogeny (480 to 435 ma; Glover and others, 1983) and exotic terrains and island-arc sediments were accreted onto a passive margin ramp with a crust thinned by erosion and rift-attenuation (Price and Hatcher, 1983; Rodgers, 1983; Fichter and Diecchio, 1986; Tankard, 1986a,b). Collisions resulting from this change in spreading direction were concomitant with a break in carbonate deposition. Subaerial erosion then occurred, resulting in a basin-wide disconformity between Lower and Middle Ordovician sedimentary rocks (Rodgers, 1983). This disconformity is associated with a thin shale unit (Glenwood Shale) that also is associated with uplift of the Waverly arch (Colton, 1970). Along with a foreland basin, this disconformity resulted from lithospheric flexure representing the earliest stage of the Taconic orogeny (Table 1; cf. also Quinlan and Beaumont, 1984; Cloetingh, 1988). Because thin crust was loaded by overthrust sedimentary rocks during the Taconic orogeny, initial basin subsidence was large, and estimates of basin depth range from 700 to 2,000 meters (Tankard, 1986a). Both the Upper Ordovician clastic sequence, consisting primarily of non-marine alluvial sediments (Colton,. 1970; Cook and others, 1983; Leon, 1984), and the Silurian clastic sequence, which was deposited by deltas interrupting marine transgression over these earlier clastic rocks (Colton, 1970; Cotter, 1983), were deposited during subsequent tectonic inactivity. The Silurian-Devonian carbonate sequence above these earlier clastic rocks (Colton, 1970; Cotter, 1983) was deposited in a marine environment that transgressed stratigraphy.

Accretion of additional microcontinents to the northern Appalachians during the Middle-Upper Devonian Acadian orogeny (380 to 340 ma; Glover and others, 1983) resulted in the formation of massive mountain ranges and deposition of a large clastic wedge, the Catskill Delta (Tankard, 1986a). These sediments comprise the Devonian clastic sequence (Colton, 1970) and were deposited by alluvial and deltaic systems. The Acadian Orogeny, however, involved loading of predominantly continental crust (Tankard, 1986b) and less basin subsidence occurred because of increased lithospheric rigidity. Consequently, this successor foreland basin was shallower and wider, with maximum depths estimated to approximate only 200 m.(Tankard, 1986b).

As the Acadian orogeny became inactive, sediment yield decreased and the Central Appalachian basin received mostly fine-grained sediments (Tankard, 1986a,b). During early Mississippian time, the basin was tectonically inactive, and, except for periods of uplift of the Waverly arch, the Appalachian and Illinois basins were yoked (Quinlan and Beaumont, 1984; Tankard, 1986a,b). Both limestones and shales were deposited during Mississippian time, and these rocks contain numerous erosional unconformities caused by periodic uplift of the Waverly arch (Colton, 1970; Quinlan and Beaumont, 1984; Cloetingh, 1988;

Tankard, 1986a,b).

Alleghenian-Hercynian overthrusting (330 to 230 ma; Glover and others, 1983) involved collision of at least two microcontinents with the main North American continent (Geiser and Engelder, 1983; Rodgers, 1983) and most strongly influenced the southeastern part of the northern Appalachians (Rodgers, 1983). Docking of these terrains may represent the final closure of Iapetus and the formation of Pangea (van der Voo, 1983). During the Alleghenian-Hercynian orogeny, fold and thrust belts developed as continental terrace sediments became detached from the basement and were thrust onto thermally mature continental crust (Price and Hatcher, 1983; Tankard, 1986a,b). Because only more rigid continental crust was loaded, subsidence due to lithospheric flexure was of lesser vertical magnitude but of greater lateral extent than during previous orogenies which involved varying thicknesses of attenuated crust (Tankard, 1986a,b). This decreased the magnitude of basin subsidence and caused a much shallower and wider basin to form in which marine embayments generally did not exceed 20 m. in depth (Tankard, 1986a). The analysis of Pennsylvanian cyclothems by Klein and Cloetingh (1989), using a tectonic subsidence model, showed that subsidence rates of the Central Appalachian basin ranged from 0.63 to 4.60 m./my during Pennsylvanian time (Table 2), confirming this orogenic event as involving considerably less basin subsidence than the earlier events. Basin sedimentation patterns fluctuated between over- and under-filled (Tankard, 1986a,b) because of the shallow nature of the basin and variable sediment accumulation rates. The Pennsylvanian sequence consists of all sediments deposited in response to the Alleghenian-Hercynian orogeny and included rocks of both Pennsylvanian and possible Permian age (Colton, 1970). These rocks consist primarily of alternating beds of sandstone and shale with incorporation of limestone and coal, including the well-described Pennsylvanian coal-bearing cyclothems (Colton, 1970; Klein and Willard, 1989) which extended west into the midcontinent.

Table 2. Tectonic Subsidence Rates, Percentage of Coal Reserves, and Range of Maceral Content of Appalachian Basin Coals.

STAGES	APPALACHIAN GROUPS AND FORMATIONS	AVERAGE TECTONIC SUBSIDENCE RATE[1] (m/my)	PERCENT OF APPALACHIAN BASIN COAL RESERVES[2]	PERCENT VITRINITE[3]	PERCENT EXINITE[3]	PERCENT INERTINITE[3]	AVERAGE VITRINITE/ INTERTINITE RATIO
Virgilian	Monongahela	2.33	20.7	86-93	<5	5-11	12.7
Missourian	Conemaugh	1.23	1.8	83-93	<1	7-17	9.3
Desmoinesian	Allegheny	0.63	34.8	63-89	4-12	7-25	7.2
Atokan	Kanawha	0.98	30.6	60-82	2-15	11-33	3.8
Morrowan	New River and Pocahontas	4.60	12.1	72-89	<5	11-27	4.8

[1] From Klein and Cloetingh (1989).

[2] From Phillips, Shepard, and DeMaris (1980).

[3] From Grady (1979).

TECTONIC SUBSIDENCE ANALYSIS

Background

Five wells located in the Allegheny Plateau province (Figure 1) were chosen for tectonic subsidence analysis because (1) all contain relatively complete stratigraphic sections with few unconformities and (2) all penetrated Cambrian rocks. Lithic descriptions and thicknesses were obtained from well logs (Hauser, 1983; Bayles and others, 1956; Fettke, 1961), and absolute ages for formations were obtained from COSUNA charts for the Appalachian region (Patchen and others, 1985a,b); these ages are based on the timescale of Salvador (1985). Lithic composition, absolute age of deposition, and formational thickness were incorporated into tectonic subsidence calculations. These calculations were performed using a computer program based on the approach of Sclater and Christie (1980) and Sawyer and others (1987) in which a backstripping procedure was used that removes sediment loading and progressively decompacts the stratigraphic section (Steckler and Watts, 1978). We assumed that net sediment accumulation rates equal subsidence rates. Also, variation in water depth was not considered in these calculations, because it does not appear to affect the general characteristics of subsidence curves (Heidlauf and others, 1986). Two-dimensional heat flow is incorporated in the program (Sawyer and others, 1987).

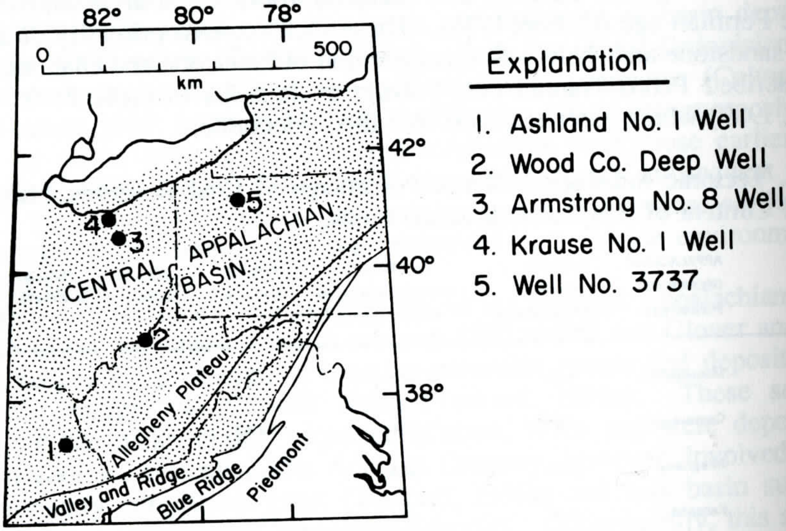

Figure 1. Map of Central Appalachian basin (stippled) showing location of five deep wells used for tectonic subsidence analysis.

To test the tectonic evolution of the Central Appalachian basin that was summarized previously, we compared tectonic subsidence curves constructed for the Central Appalachian basin to theoretical subsidence curves based on a stretching model developed by McKenzie (1978; his figs. 2 and 4). This model characterizes basins formed by rifting and involves stretching of the crust causing crustal thinning and formation of normally faulted grabens and rift valleys. An associated upwelling of asthenosphere into the thinned crustal zone increases heat flow. As lithosphere cools, it thickens and subsides thermally following lithospheric cooling rates (Parsons and Sclater, 1977). Although thermally controlled subsidence shows a linear relationship with the square root of time when vertical heat flow is assumed, this relationship approaches an exponential value about 60 millions years after rifting (McKenzie, 1978). However, lateral heat flow is to be expected also (Watts and others, 1982; Hellinger and Sclater, 1983; Sawyer et al, 1987), causing more rapid central basin cooling and subsidence in the central basin and uplift of the rift shoulders (Hellinger and Sclater, 1983).

An Airy model of isostasy in which sediment is supported only by immediately underlying rocks forms the basis for McKenzie's (1978) model. A second model incorporates lithospheric flexure, in which lateral crustal strength supports the sediments (Watts and others, 1982), and both an elastic lithosphere, known from oceanic crust, and a viscoelastic lithosphere, known generally from continental crust were considered (Watts and others, 1982). Loading of elastic lithosphere results in crustal subsidence and causes an uplifted peripheral bulge to form away from the basin center; a viscoelastic lithosphere responds initially as elastic lithosphere but (shortly thereafter), responds as viscous lithosphere and approaches an Airy model of isostasy (Watts and others, 1982).

The lateral distribution of stratigraphic Colton Sequences of the Central Appalachian basin during a major rift-drift cycle shows progressive overstepping of older strata by younger sediment. This observation can be explained by differential two-layer stretching between the crust and mantle (White and McKenzie, 1988), resulting in a tectonic onlap of these stratigraphic units. With Late Ordovician collision events, this region underwent lithospheric flexure causing development of the Central Appalachian foreland basin, with adjacent dome and arch systems developing inboard of the basin in response to flexure (Quinlan and Beaumont, 1984) . If this is the case, then tectonic subsidence curves constructed from foreland basins will differ from McKenzie's (1978) stretching model and will show little evidence of thermal subsidence.

Results

Tectonic subsidence curves for the five wells (Figures 2 through 6) show little variation in overall trends. Each curve indicates relatively slow subsidence rates with interruption by orogenic events. In the Central Appalachian basin, an oscillatory deviation is evident in each curve around 430 to 410 Ma. This deviation corresponds to fairly thick (up to 200 m.) beds of alternating carbonate and clastic sedimentary rocks, including the Salina Group (Silurian). This deviation is observed regionally in all five wells rather than locally and appears to be caused by sea-level changes (Cotter, 1988).

The Ashland No. 1 well (Figure 2) in eastern Kentucky bottomed in the Cambrian Bonneterre Formation of the Cambro-Ordovician carbonate sequence of Colton (1970). The subsidence curve indicates increased subsidence during the Taconic Orogeny (480 to 435 ma; Glover and others, 1983) followed by a period of relatively less subsidence. Slightly increased subsidence occurred during the Acadian Orogeny (380 to 340 ma; Glover and others, 1983); little subsidence is evident during Mississippian time. When subsidence rate is plotted against time (Figure 2), a disparity between subsidence rate during the Taconic and Acadian orogenies is evident. Subsidence rates in eastern Kentucky were three to four times greater during the Taconic Orogeny than during the Acadian Orogeny

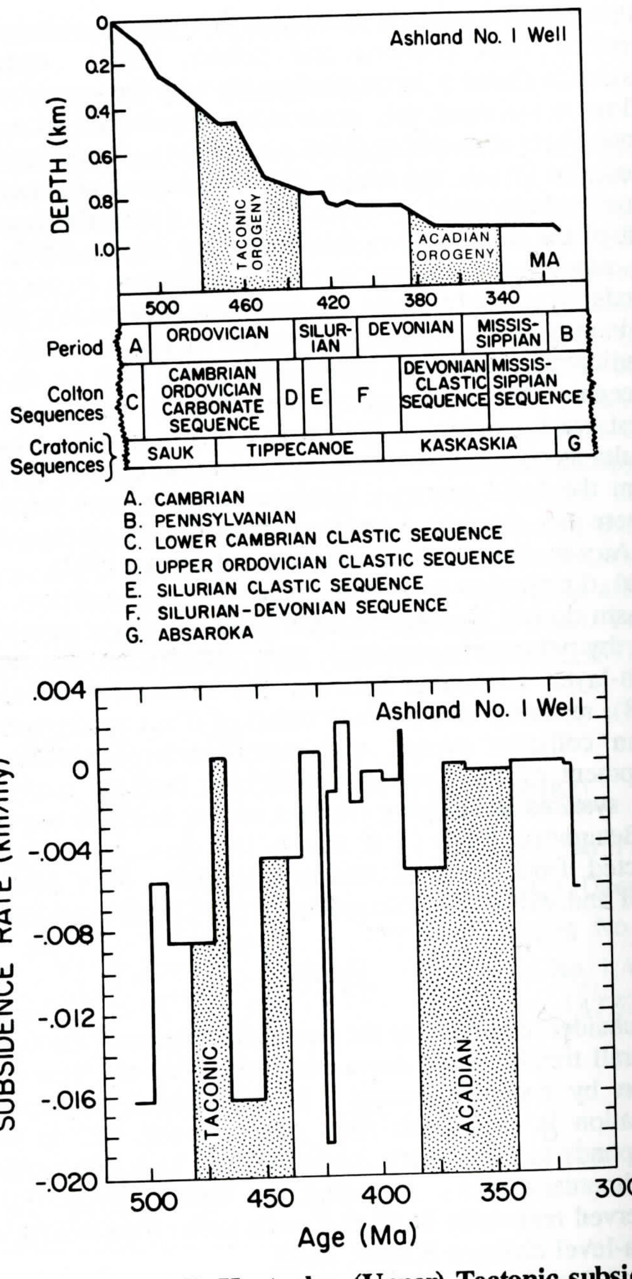

Figure 2. Ashland No. 1. Well, Kentucky. (Upper) Tectonic subsidence curve with depth scale set to middle of Cambro-Ordovician carbonate Bonneterre Formation. (Lower) Comparison of basin subsidence rate with time.

because of increasing lithospheric rigidity (Kuznir and Karner, 1985; Tankard, 1986a). In Figure 2, the indicated time encompassing the Acadian Orogeny is later than the time of increased subsidence indicated by the curves. This discrepancy suggests either that the Acadian Orogeny began earlier than 380 Ma, or this finding is an artifact of using the improved Salvador (1985) time scale used by Patchen and others (1985a,b).

The Wood County Deep Well, also known as the Sandhill Deep Well (Figure 3), in northern West Virginia is the only well in the Appalachian basin that reached crystalline Precambrian basement (Woodward, 1959). Both tectonic subsidence curves and the temporal variation of subsidence rate (Figure 3) show similar subsidence trends to those of the Ashland No. 1 Well, except that the subsidence rates during the Taconic orogeny did not differ as greatly from the Acadian orogeny subsidence rates in the Wood County Deep Well as it did in the Ashland Well. This difference may be caused by a location which is more inboard on the continent.

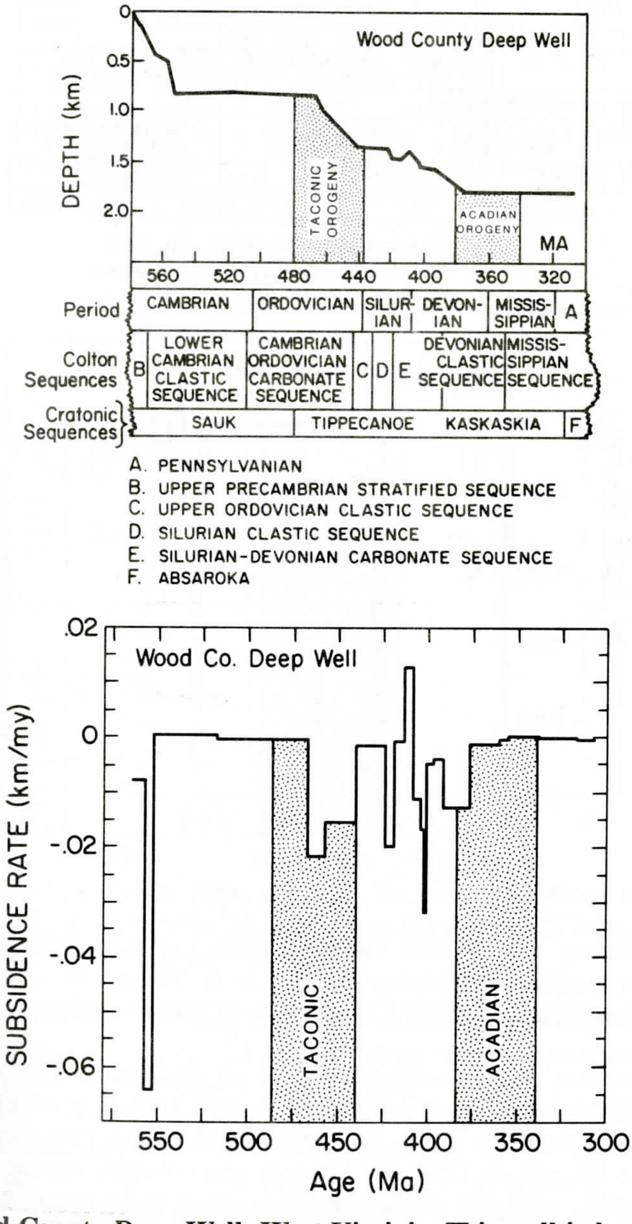

Figure 3. Wood County Deep Well, West Virginia. This well is known also as the Sandhill Deep Well. (Upper) Tectonic subsidence curve with depth scale set to top of crystalline basement. (Lower) Comparison of basin subsidence rate with time.

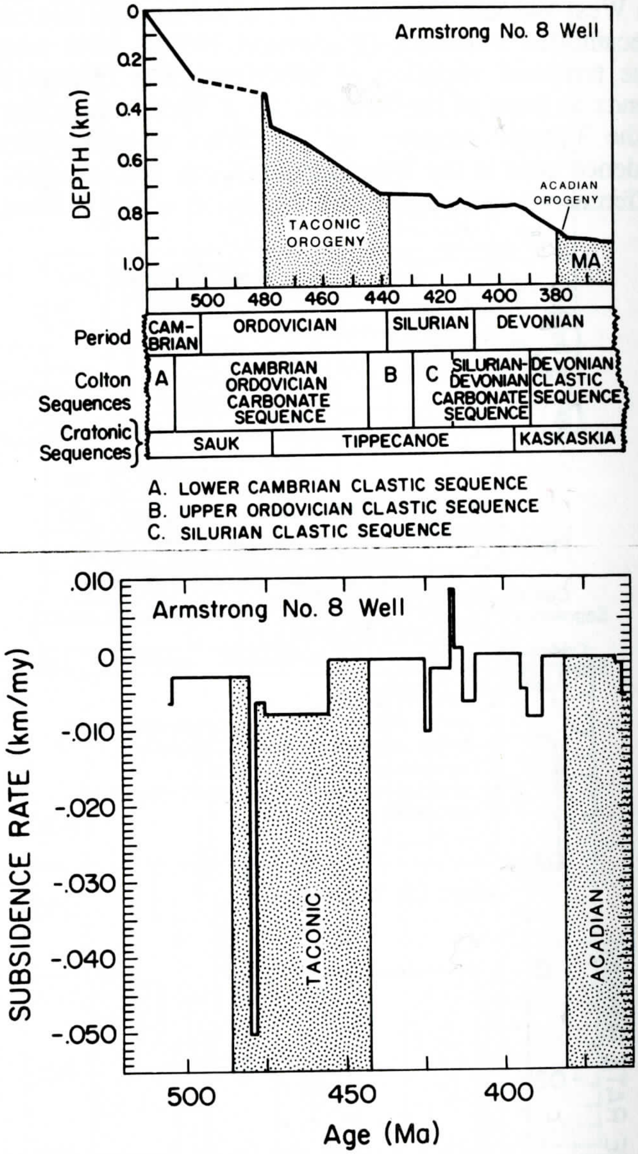

Figure 4. Armstrong No. 8 Well, Ohio. (Upper) Tectonic subsidence curve with depth scale on tectonic subsidence curve set to middle of Cambro-Ordovician Trempealeau Formation. (Lower) Comparison of basin subsidence rate with time.

The Armstrong No. 8 well (Figure 4) in northern Ohio bottomed in the Trempealeau Formation (Cambro-Ordovician carbonate sequence). The tectonic subsidence curve parallels those of the Ashland and Wood County wells, but subsidence rate (Figure 4) does not indicate any unusual increase in subsidence during the Acadian orogenic episode, perhaps because its location is on the basin edge next to the craton.

The Krause No. 1 well of northern Ohio (Figure 5) bottomed into the Cambrian Mount Simon Sandstone. Although its subsidence rate (Figure 5) shows inconsequential changes throughout its entire history, the tectonic subsi-

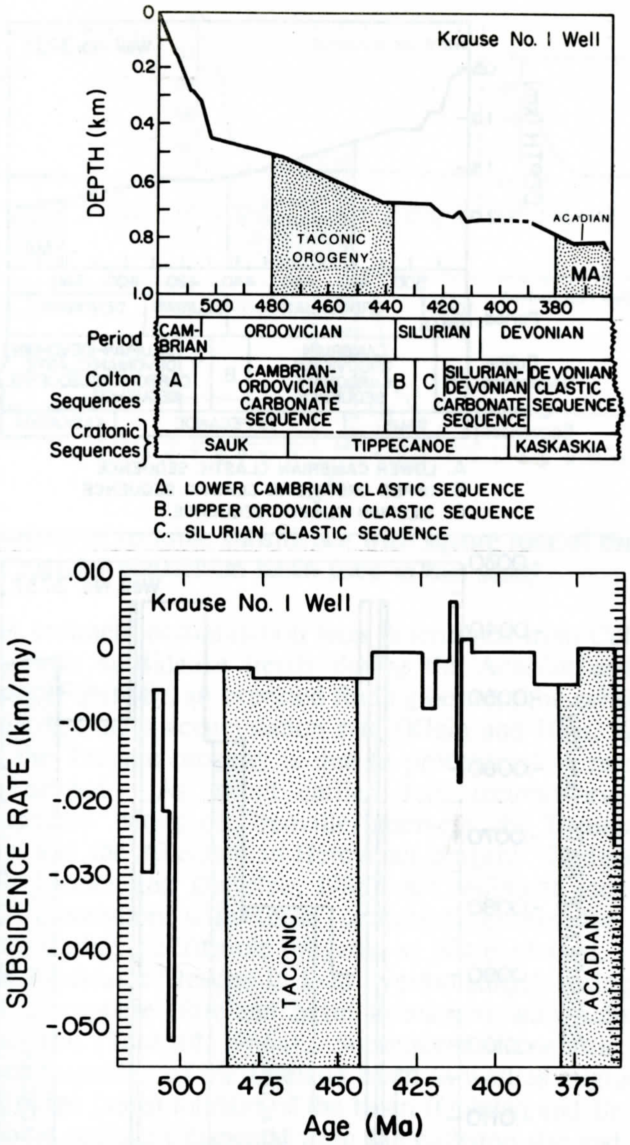

Figure 5. Krause No. 1 Well, Ohio (Upper). Tectonic subsidence curve with depth scale set to middle of Cambrian Mount Simon Formation. (Lower) Comparison of basin subsidence rate with time.

dence curves show slightly increased subsidence during both the Taconic and Acadian orogenies, even though its location is on the basin edge next to the craton and on the flank of the Algonquin-Findlay arch.

Well 3737 of northwestern Pennsylvania (Figure 6) bottomed in the Cambrian Pottsdam Sandstone. Like the Krause No. 1 well, subsidence appears to be constant with few effects from orogenic activity (Figure 6). Subsidence rate (Figure 6) increased during the Taconic Orogeny.

When subsidence is compared to the square root of time (Figure 7), all five curves show the same basic pattern. Rapid subsidence occurred initially during passive margin tectonics (Bond and others, 1984; Quinlan and Beaumont, 1984;

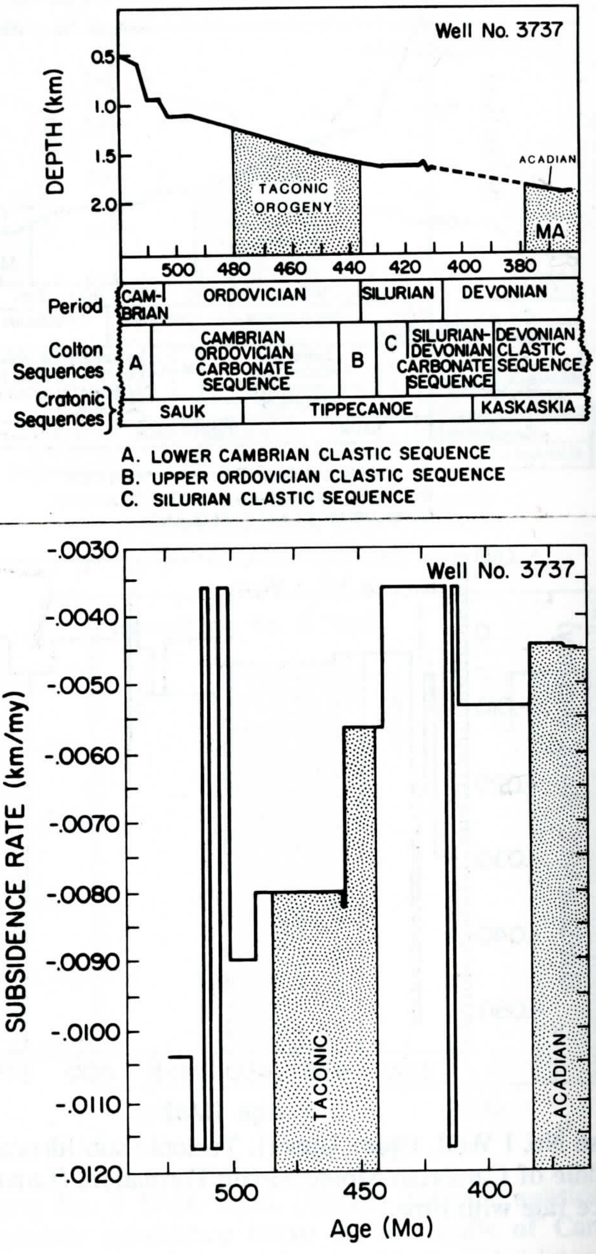

Figure 6. Well No. 3737, Pennsylvania. (Upper) Tectonic subsidence curve with depth scale set to middle of Cambrian Pottsdam Sandstone. (Lower) Comparison of basin subsidence rate with time.

Fichter and Diecchio, 1986; see also Table 1) but slowed to a constant trend as basin development continued. No overall linear relationship exists with the square root of time except during the earliest stage of basin evolution as expected in basins undergoing thermal subsidence (McKenzie, 1978). These curves from wells in the Central Appalachian basin show a horizontal trend (Figure 7) fitting a flexural model of basin subsidence.

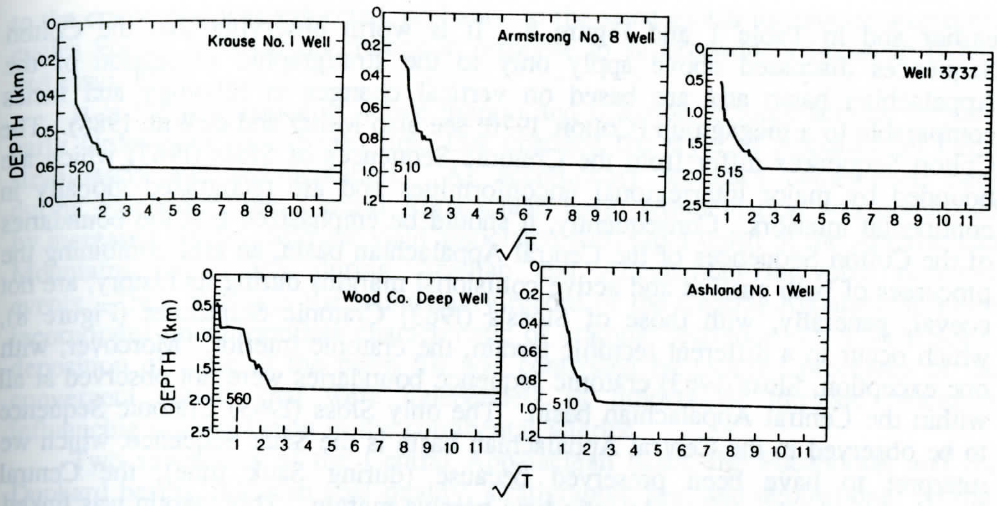

Figure 7. Comparison of tectonic subsidence with square root of time in five deep wells from the Central Appalachian basin used in this study.

Average net sediment accumulation rates determined from Colton Sequences correlate closely with subsidence trends during the Acadian and Alleghenian-Hercynian orogeny (Figure 8), an expected finding because the subsidence curve is the integral of net sediment accumulation rate (Klein and Hsui, 1987). Lack of correlation with the Taconic orogeny is due to probable slow rates of sediment yield from the orogen. As stated earlier, four tectonic events controlled sedimentation patterns: rifting of Grenville basement, the Taconic orogeny, the Acadian orogeny, and the Alleghenian-Hercynian orogeny (Rodgers, 1983; Bond and others, 1984; Fichter and Diecchio, 1986); net sediment accumulation rates show a general correlation with these processes, except for the sediment accumulation rates during the Taconic orogeny, as just explained.

Collision of landmasses results in load overthrusting, causing initial elastic downwarping of lithosphere (foreland basin formation) with sediment dispersal from the collision site (Tankard, 1986a). Subsequent viscoelastic adjustment of lithosphere causes deepening of the foreland basin as well as migration and uplift of the peripheral bulge (arch) inboard of the basin (Quinlan and Beaumont, 1984), causing bidirectional sediment dispersal from the collision site and from the arch (Tankard, 1986b). Colton Sequences correlate directly to these different tectonic events (Figure 8).

Tectonic subsidence rates calculated from subsidence curves generated by Klein and Cloetingh (1989) from Pennsylvanian cyclothems are compared with Pennsylvanian coal reserves, and petrography in Table 2. An inverse relationship appears to exist between subsidence rate and volume of coal reserves. Stratigraphic units representing periods of slowest subsidence (Desmoinesian and Atokan series) are characterized by possessing the largest coal reserves of the Pennsylvanian System of the Central Appalachian basin.

DISCUSSION AND IMPLICATIONS

Results of the above tectonic subsidence analysis are consistent with the tectonic and stratigraphic evolution of the Central Appalachians summarized earlier and in Table 1 and Figure 8. It is worth observing that the Colton Sequences discussed above apply only to the stratigraphic succession of the Appalachian basin and are based on vertical changes in lithology and facies comparable to a megagroup (Colton, 1970; see also Milici and deWitt, 1988). The Colton Sequences differ from the Cratonic Sequences of Sloss (1963) which are bounded by major interregional unconformities and are recognized globally in continental interiors. Consequently, it should be emphasized that the boundaries of the Colton Sequences of the Central Appalachian basin, an area combining the processes of both passive and active collisional margins during its history, are not coeval, generally, with those of Sloss's (1963) Cratonic Sequences (Figure 8), which occur in a different tectonic terrain, the cratonic interior. Moreover, with one exception, Sloss'(1963) cratonic sequence boundaries were not observed at all within the Central Appalachian basin. The only Sloss (1963) Cratonic Sequence to be observed in the Central Appalachian basin is the Sauk Sequence, which we interpret to have been preserved because (during Sauk time), the Central Appalachian basin was an Atlantic-type passive margin. This margin was linked

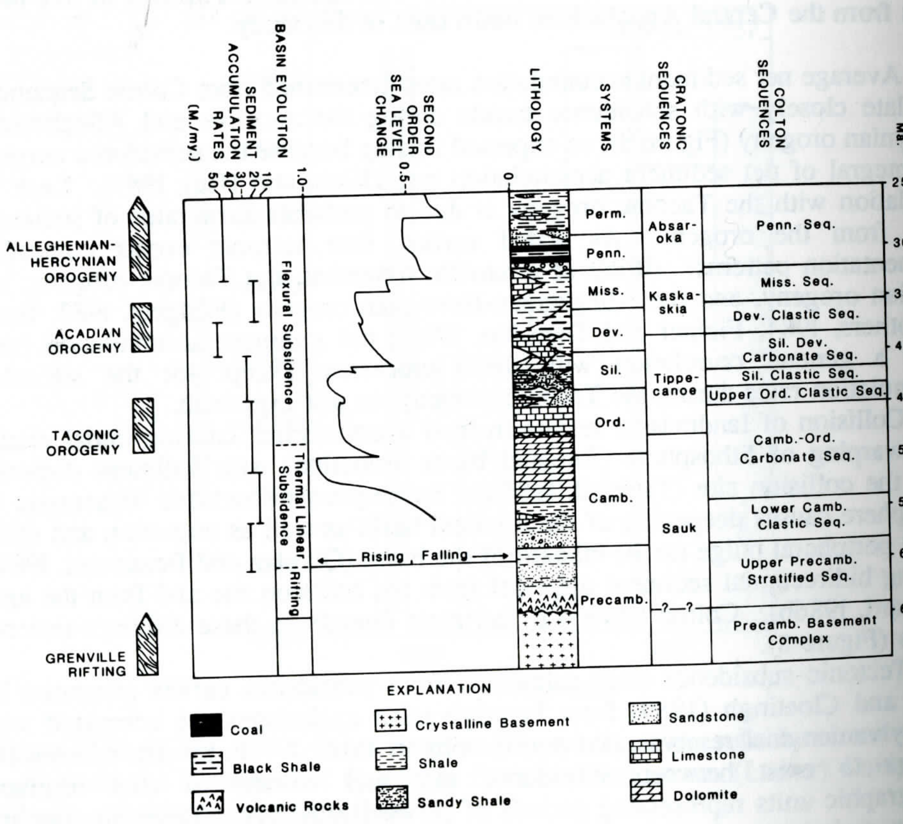

Figure 8. Correlation diagram of the Central Appalachian basin showing geologic time, sedimentary sequences and lithology after Colton (1970), Cratonic Sequences (after Sloss, 1963), sea-level change (after Vail and others, 1977), stages of basin evolution based on tectonic subsidence curves from five wells used in this study, sediment accumulation rates (average from five wells used in this study), and orogenic events (after Rodgers, 1967,1983; Glover and others, 1983; Price and Hatcher, 1983).

to the craton and was subjected, probably, to the same global tectonic-stratigraphic cycles known to occur on cratons in response to supercontinent break-up (Klein and Hsui, 1987).

It is the dominance of active margin and collision tectonic activity in post-early Ordovician (post-Sauk) time that changed the style of Appalachian basin stratigraphy and sedimentation. As accretionary tectonics commenced later during Paleozoic time, Central Appalachian basin sedimentation was dominated by increased clastic-sediment yield associated with rapid uplift of the Appalachian Mountains particularly during Acadian and Alleghenian-Hercynian orogenic events. Thus, although the sedimentary record of the Central Appalachian basin incorporates unconformities characteristic of cratonic sequences on a local scale, deposition of most of the basin fill was controlled by tectonic processes of the convergent margin that were independent of lithospheric intrinsic processes influencing sedimentation in the continental interior.

Two sub-basins of the Central Appalachian basin, the Pocahontas and the Dunkard basins, differ in coal quality, stratigraphic age, and depositional setting. In the Pocahontas basin, a foreland basin, deltaic and prodeltaic sediments are thicker and better preserved and contain more named coals than the peripheral Dunkard basin (Donaldson and others, 1985). These sedimentary features of the Pocahontas basin were interpreted by them as indicative of more rapid subsidence than the Dunkard basin. Computation of tectonic subsidence rate from wells in each of the sub-basins confirms that interpretation. In the Pocahontas basin, subsidence rates range from 0.6 to 8./l m/my throughout Pennsylvanian time, whereas subsidence rates in the Dunkard basin range from 0.5 to 3.0 m/my (Klein and Cloetingh, 1989).

Lowstands of sea level were contemporaneous, generally, with the early stages of collision events, and, in one case, with a Colton Sequence boundary (Figure 8). Early during the Taconic orogeny, reversal in spreading direction caused a lowstand in sea level and a basin-wide unconformity (Hatcher, 1972; Rodgers, 1983). This unconformity occurs at the Sauk-Tippecanoe cratonic sequence boundary, but deposition of carbonate rocks continued until later in the orogeny when increased rate of tectonic uplift caused excessive clastic sediment yield and associated preservation of clastic sequences (cf. Klein, 1985). Thus, the Colton Sequence boundary occurs between carbonate and clastic rocks, and not at the unconformity within the carbonate sequence. Moreover, the Colton Sequence boundaries are independent of major second-order changes in sea level and should not be used to establish either onlap curves or curves of relative changes in sea level on a local or global scale.

The apparent synchroneity of collision events and lowstands of sea level fits a model of sea-level fluctuation in response to continental accretion and fragmentation (Bally, 1982; Watts and others, 1982; Worsley and others, 1984; Cloetingh, 1986,1988). Collision of accreting continents increases both intraplate compressive stress and sediment load by overthrusting; the interaction of these two processes modifies the basin shape and increases the basin volume, causing a relative lowering of sea level (Cloetingh, 1986,1988). Increased intraplate stress also alters the elevation of the peripheral bulge associated with foreland basins. In the Central Appalachian basin, fluctuations in the height of the Waverly arch during Late Devonian and Early Mississippian time recorded such stresses (Tankard, 1986b; Cloetingh, 1988). Periodic uplift of the Waverly arch increased the sediment flux and influenced sea level regionally.

The extensive Pennsylvanian coal reserved which accumulated during both late stages and following the Alleghenian-Hercynian orogeny, show certain correlations between tectonic subsidence rate, coal reserves, petrography (Table 2), and, to some extent, paleoclimate. The largest volume of coal accumulation occurred during Desmoinesian and Atokan time (represented by the upper Kanawha and Allegheny formations (Phillips and others, 1980)). These relatively stable periods, when subsidence was less than 1 m./my , corresponded to some of the wettest intervals of the Pennsylvanian during which time climate was both ever-wet and tropical (Cecil and others, 1983;1985; Phillips and Peppers, 1984). Although these coals accumulated in a relatively wide basin with rigid crust during a time of slow subsidence, conditions appear not to have been optimal for peat preservation. Vitrinite content is small, inertinite is large, and coal ball peats possess greater proportions of roots and degraded plants during this interval of Pennsylvanian time (Phillips and Peppers, 1984). This poor preservation and apparent oxidation may have resulted either from periodic flooding and subsequent exposure of peat (Grady, 1979) or from possible accumulation as ombrogenous (domed) peats (Cecil and others, 1985) during this time interval.

The boundary between the Middle and Upper Pennsylvanian corresponds to a time of increased tectonic subsidence, a decrease in coal reserves, a major extinction among plant groups, and a relatively abrupt decrease in wetness. All these features are related to tectonic activity associated with assembly and repositioning of Pangea and subsequent Gondwana glaciation (Powell and Veevers, 1987). No consistent trends exist among macerals in Upper Pennsylvanian coals, although their vitrinite content generally is greater (Table 2). Drier climatic conditions during the time interval represented by the Conemaugh Formation appear to have contributed to a decrease in coal resources, and possibly caused extinction of most arborescent lycopods and other coal-swamp plants (Phillips and Peppers, 1984). A gradual increase in relative wetness developed between Conemaugh and Mongahela time, allowing greater accumulation of peat, and thus increased coal reserves within the Monongahela Formation. The overall wetness pattern cycle from wet to dry conditions fits a hypothesized model of development of monsoonal circulation beginning with the formation and northward movement of Pangea (Rowley and others, 1985). Resulting seasonality was suggested from floristic evidence observed in Early and early Middle Pennsylvanian rock units but not from younger units (Phillips and Peppers, 1984). Lack of seasonality during the late Middle and Late Pennsylvanian may be due to an increase in elevation of the Appalachians; the mountains could have acted as a high altitude heat source, countering monsoonal circulation and maintaining a relatively wet climate until their gradual erosion and subsequent drying, producing a wet-dry cycle (Parrish and Peterson, 1988; Rowley and others, 1985).

One of the more remarkable consequences of the data in Table 2 concerns its implications for Pennsylvanian sea-level history. If the ratio of vitrinite and inertinite are indeed inversely correlated to degree of oxidation (Harvey and Dillon, 1985) suggesting prolonged exposure of peat, then the Central Appalachian basin must have experienced minimal eustatic sea level change.If these coals were submerged during times of relatively higher stands of sea level, the V/I ratio should be larger, and, as per Harvey and Dillon (1985), oxidation of peat would be diminished considerably. Because these coal beds are associated with Appalachian type cyclothems, one would expect that during times of minimal subsidence rate, sea-level fluctuations would be more of a dominant process in controlling both coal accumulation and deposition of associated cyclothems (Klein and Willard, 1989). We find it most peculiar that the history of early to middle Pennsylvanian coal deposition in the Appalachian basin is a time of extensive oxidation and exposure of coal-forming peat, and suggest that this finding (Table 2) implies that the vertical range of early to middle Pennsylvanian sea-level change was relatively small in the Central Appalachian basin. We interpret these findings to suggest, moreover, that perhaps the vertical magnitude of early to middle Pennsylvanian sea-level change was relatively small, perhaps much smaller than estimated by Ross and Ross (1987). Alternatively, if the vertical magnitude of sea-level change of Early to Middle Pennsylvanian time was of the order of magnitude suggested by Ross and Ross (1987), then the process of sea-level change was characterized by sudden intervals of rapid rate of sea-level rise, followed by a period of slow sea-level rise or fall or a still-stand. This implies, then, that periods of thick coal deposition represent times of minimal sea-level change or a stillstand. Thus the combined effects of a slow change in sea level, or a stillstand, slow basin subsidence rates, and an equitable tropical climate may well favor preservation of thick coals in foreland basins such as the Central Appalachian basin.

CONCLUSIONS

The Central Appalachian Basin is a successor basin that developed through several stages. These stages included both mechanical stretching and thermal subsidence during a rift-basin stage which was followed by a passive margin stage in response to breakup of a major late Precambrian supercontinent from late Precambrian through Early Ordovician time. During the remainder of the Paleozoic Era, the Appalachian region was dominated by active margin and collisional tectonic processes resulting in resurgent stages of foreland basin development during the Taconic, Acadian, and Alleghenian-Hercynian orogenies. Tectonic subsidence analysis from five deep wells confirmed these stages of basin evolution and demonstrated that the rate of subsidence of the three foreland basin episodes decreased through time because the crust underneath the Appalachian basin became progressively more rigid, thus fostering development of a wider and shallower foreland basin evolution through time.

The boundaries between stratigraphic Colton (1970) sequences are defined according to lithology and facies changes controlled by tectonic processes during the evolution of the Central Appalachian basin. In contrast to Cratonic Sequences, where boundaries are interregional unconformities, the Colton Sequences are independent of global sea-level fluctuations; low stands of second-order sea-level changes in the central Appalachians appear to be controlled by and associated with the early stages of collisional events. The eustatic second-order sea-level events represented by the boundaries of cratonic sequences are observed as local unconformities of a scattered and local nature within the Appalachian basin rather than on a regional scale. Consequently, in active margin and collision tectonic terrains the preservation of second order sea level events tends to be masked by clastic sediment yield in response to rapid rates of tectonic uplift during collision tectonics (Klein, 1985).

Subsidence rates calculated from Pennsylvanian age tectonic subsidence curves show that foreland basin subsidence in the Central Appalachian basin were slowest during Late Morrowan through Desmoinesian time. These same time-stratigraphic intervals contain the largest coal reserves in the Central Appalachian basin. We conclude that these slow rates of tectonic subsidence combined with minimal change in eustatic sea level or a still stand, and with an equitable tropical climate, controlled the extensive accumulation of thick coals in this foreland basin.

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HYALOCLASTITE PILLOW BRECCIA IN THE CATOCTIN META-BASALT OF THE THE EASTERN LIMB OF THE BLUE RIDGE ANTICLINORIUM IN VIRGINIA

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ABSTRACT

Two extensive breccia zones have been identified in the Catoctin Formation of the eastern limb of the Blue Ridge anticlinorium. In northern Virginia, breccia occurs at the base of the Catoctin for a minimum strike length of 57 km; in central Virginia a similar breccia extends a minimum of 42 km near the top of the Catoctin. In northern Virginia the breccia zone had previously been mapped and interpreted as agglomerate, but both there and in central Virginia it has textural features and field relationships characteristic of hyaloclastite pillow breccia. Large, poorly-sorted, elliptical to angular metabasalt clasts are dispersed in a matrix composed of ash to fine lapilli-sized chloritized metabasalt grains, with an even more chloritic mesostasis. The matrix grains are elliptical to angular in shape and commonly exhibit concave and convex edges, knife-sharp edges, triangular shapes, slab-like shapes, and in situ brecciation. Pillow metabasalt has been discovered in the breccia zone at three locations in northern Virginia and four in central Virginia. In a large pavement exposure the breccia is in contact with pillow metabasalt, and examples of various degrees of pillow brecciation occur. The structural relationship between pillow lava and pillow breccia is unlike that formed where lava flows into a lake or other shallow body of water, but rather like that formed in subaqueous extrusions. These extensive breccia zones seem to require very large lakes or some sort of marine environment for the paleogeography of the eastern limb of the Blue Ridge anticlinorium near the Precambrian/Cambrian boundary.

INTRODUCTION

An extensive breccia zone in the lower part of the Catoctin Formation in northern Virginia has been mapped by Furcron (1939) and Espenshade (1986) from south of Warrenton to northeast of The Plains, a total strike length of 57 km (Figure 1). A similar breccia zone has been recently mapped by Conley (unpublished map at Virginia Division of Mineral Resources) in central Virginia near the top of the Catoctin. This zone extends a minimum of 42 km from near the Hardware River to northeast of Gordonsville (Figure 1), and has been detected at the same stratigraphic position at Madison Run and in the Unionville quadrangle east of Orange, Virginia (Lou Pavlides, unpublished map). We believe that this breccia will be an important factor in interpreting the paleoenvironment of the latest Precambrian/earliest Cambrian time in the area of central and northern Virginia.

For a number of years a controversy has existed as to the environment of deposition of the Catoctin Formation in the Virginia Blue Ridge. Workers on the western limb of the Blue Ridge anticlinorium have cited primary structures as evidence that the Catoctin originated as a continental plateau basalt (Bloomer and Bloomer, 1947; Reed, 1955, 1969; Gathright, 1976). Some workers have reported mineralogical and chemical evidence that the Catoctin is spilitic and concluded that it was a subaqueous deposit (Bloomer, 1950; Bloomer and Werner, 1955; Brown, 1958). This geochemical argument has been refuted (Reed, 1964; Reed and Morgan, 1971). Other geochemical data has been presented pointing to a continental origin (Blackburn and Brown, 1976; Davis and others, 1978). Brown (1958) considered greenstones interlayered with metagraywacke in the Lynchburg area to be Catoctin equivalents and interpreted a geosynclinal environment for the Catoctin in that area, but correlation of these greenstones with Catoctin Formation has been questioned (Conley, 1978).

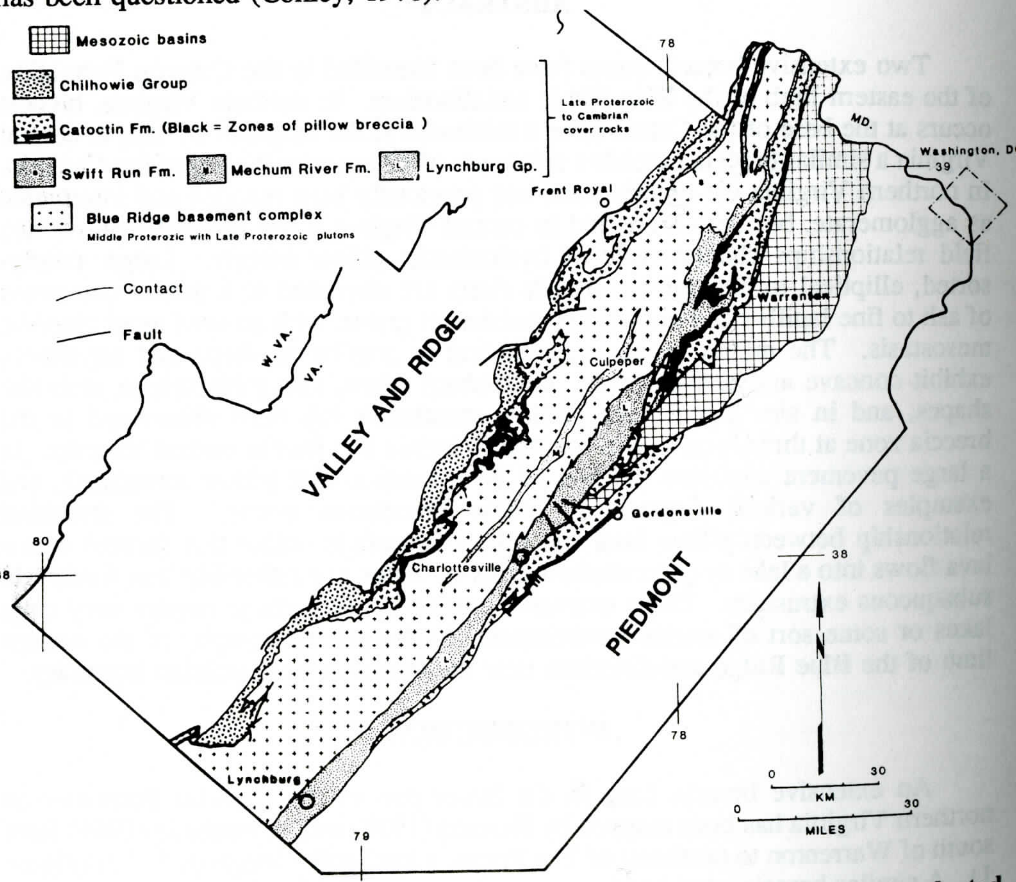

Figure 1. Geology of the Blue Ridge area of northern and central Virginia: adapted from Calver and others (1963), Espenshade and Clarke (1976), Espenshade (1986), and from recent mapping by J.F. Conley. In northern Virginia the pre-Catoctin metasediments, the Fauquier Formation, are included in the Lynchburg Group, following the usage of Wehr (1983). Pillow breccia in the Catoctin has been found only on the eastern limb of the anticlinorium.

Rankin (1975) proposed that rifting of the North American continental margin in the late Proterozoic produced a great graben system with water-filled basins to the east of the present Blue Ridge anticlinorium. Wehr and Glover (1985) presented a similar tectonic model in which the axis of the anticlinorium is considered to approximate the position of the hinge zone of the ancient rift system. The hinge zone was the westernmost limit of substantial rift-induced crustal thinning and related subsidence (LePichon and others, 1983) and controlled the distribution of marine and terrestrial deposits. The Wehr and Glover (1985) model is based partly upon an interpretation, previously voiced by Brown (1970, 1973), that the Catoctin Formation of the eastern part of the Blue Ridge terrane was deposited in a submarine environment and the Catoctin of the western part was subaerial. The interpretation of a subaqueous extrusion of Catoctin lavas was influenced by Nick Evans' discovery of an outcrop of pillow metabasalt in the Catoctin Formation south of Charlottesville. Furthermore, occurrences of the "Warrenton agglomerate" (Furcron, 1939) in northern Virginia have similarities with pillow breccias seen by Glover in the Caribbean (Lynn Glover, 1986, personal communication). These features had not been documented in the literature, however, and no detailed study had been done.

In view of the uncertainty regarding the depositional environment of the Catoctin Formation, a field and petrographic study was undertaken to examine primary structures in the Catoctin of the eastern limb of the Blue Ridge anticlinorium to determine whether these support a continental or a marine origin. This paper presents evidence that the "Warrenton agglomerate" of Furcron (1939) and similar rocks in the Charlottesville area are actually extensive zones of (metamorphosed) hyaloclastite pillow breccia in the Catoctin Formation.

FIELD AND PETROGRAPHIC DESCRIPTION

The Catoctin Formation of the eastern limb of the Blue Ridge anticlinorium consists mostly of massive to amygdaloidal metabasalt, although it also contains minor clastic sedimentary interlayers. Mineral assemblages are typical of the greenschist facies (variable amounts and compositions of tremolite-actinolite, clinozoisite-epidote, albite, chlorite, quartz, sphene, magnetite, calcite, and pyrite). Schistosity defined by metamorphic minerals is variably developed, and in many places primary structures and textures are preserved. In the main body of the Catoctin these structures consist of amygdules, vesicles, porphyritic texture, microlitic texture (seen in thin section), and fragmental textures that mark flow tops.

In the breccia zones in the Catoctin Formation, outcrops are more abundant than where massive metabasalt occurs. The outcrops are, however, rarely continuous over distances greater than about 30 meters in strike length or 10 meters perpendicular to strike. It is difficult to judge how continuous the breccia is between outcrops, and both Espenshade (1986) and Furcron (1939) suspected that some massive metabasalt is interlayered with the breccia but is not exposed due to differential erosion. Recent evidence (discussed below) indicates that pillow lavas occur in some of the covered areas. Espenshade (1986) considers the thickness of the entire breccia zone to range up to a maximum thickness of about 1500 to 1900 meters in the Marshall quadrangle, where he distinguishes two chemically distinct belts, one low in titanium, and one high in titanium. The minimum thickness there is about 300 m. This kind of range appears to be reasonably representative of both the northern and central Virginia breccia zones based on the width of the outcrop belt.

The breccias in both central and northern Virginia consist of poorly sorted assemblages of ellipsoidal to angular clasts of metabasalt scattered about in finer grained matrix (see Figure 2). The clasts may range from less than 1 cm to greater than 1 m in maximum diameter within one outcrop, but sizes more commonly are on the order of 2 to 15 cm. The great majority of the breccias are matrix supported. The clasts may range from massive to vesicular to amygdaloidal, but they may be considered as essentially monolithologic and their mineralogies are similar to other Catoctin metabasalts. Clast mineralogies are normally consistent in a single outcrop but there is some variation between different localities. Relict microlitic textures are commonly preserved in clasts, but in some rocks development of schistosity has obliterated original internal textures.

Figure 2. Outcrop of breccia in Catoctin Formation ("Warrenton agglomerate" member of Furcron, 1939; "Low-Ti metabasalt breccia" of Espenshade, 1986) in pasture 3.7 km south of The Plains, on Virginia Highway 245.

The matrix of the breccias (Figure 3) is composed of 1-10 mm angular to roughly ellipsoidal grains in a mesostasis that may comprise 10-30 percent of the total matrix. Almost invariably a few of the matrix grains are identical in mineralogy and texture to the larger clasts of the same outcrop, but most of the grains are somewhat different in mineralogy. The most typical difference is a distinct increase in chlorite abundance in the common matrix grains relative to the larger clasts, which results in the matrix exhibiting a more foliated appearance than the clasts. Although the common matrix grains differ somewhat from the larger clasts, they are still metabasaltic in character, but rarely have relict microlitic textures. In about half the occurrences examined, many of the matrix grains have small vesicles. Quite commonly sphene is concentrated at the rims of matrix grains giving the rims a bleached appearance in polished slabs. The larger clasts in the breccia and the matrix grains that are mineralogically and texturally like them do not exhibit this rimming phenomenon. In some occurrences, epidote rather than chlorite is enriched in the matrix grains.

The mesostasis is normally even more chloritic than the matrix grains. In places, pure chlorite comprises the mesostasis, but where wider gaps occur between matrix grains, fine clinozoisite-epidote occupies the central part of the mesostasis, in some places along with quartz. Actinolite may partly take the place of chlorite. In some rocks the mesostasis is fine quartz + epidote + calcite. The exact mineralogical composition of the large fragments, the matrix grains, and the mesostasis varies from locality to locality, but at any one place there always is a distinct compositional difference between these three components, most commonly with an increase in chlorite in the matrix grains and mesostasis. Tectonic flattening obscures textures in the matrix at some locations, but even then the three components of the breccia are normally recognizable.

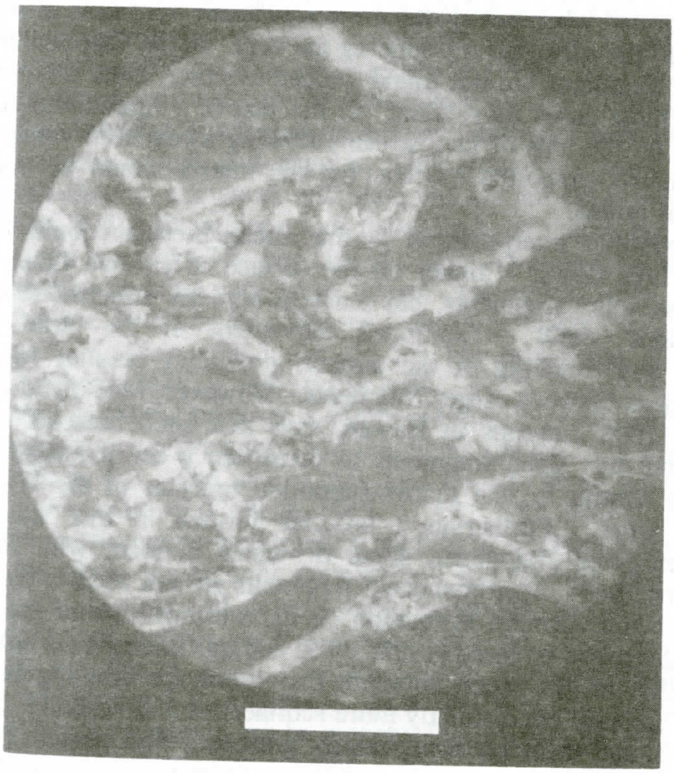

Figure 3. Matrix of sample SK-88-9, Catoctin breccia 1.5 km north of The Plains, Virginia on Highway 626 ("Low-Ti metabasalt breccia" of Espenshade, 1986). Clasts (not seen) range from fist size to <1 cm and are composed of tremolite > epidote > albite(?) > sphene. These have no rimming effect. Matrix grains (those shown) range from almost 1 cm to <1 mm and are composed of epidote > chlorite > albite(?). These have rims enriched in sphene and epidote. The smallest grains are entirely of the rim composition. The mesostasis consists of variable amounts of chlorite + calcite + quartz + tremolite. Scale bar represents 5 mm.

INTERPRETATION OF THE CATOCTIN BRECCIA

Previous Interpretations

The dispersed nature of the larger basaltic clasts amid a matrix of finer fragments is evidence against previous interpretations of the breccia. Furcron (1939) interpreted the Catoctin breccia in northern Virginia to be agglomerate, an accumulation of bombs near eruptive centers. Parsons (1969) gives a summary of field and petrographic criteria for distinguishing between different types of volcanic breccia. One characteristic of agglomerate that he notes is an "almost complete lack of groundmass in coarse breccias" (p. 178). Furthermore, sorting is fairly good in agglomerates and graded bedding commonly occurs. These features are not at all characteristic of the breccias of the "Warrenton agglomerate". Heiken (1974) indicates that basaltic ash from subaerial eruptions is vesicular, but in many of the Catoctin breccias the ash-sized matrix grains are not vesicular.

Espenshade (1986) interpreted the same breccias as "agglutinate". Agglutinate is formed when hot spatter from a volcanic vent falls in a still plastic state and welds to the substrate surrounding the vent. In the Catoctin breccias, the large bomb-like fragments are not found welded together, even where they are in contact, and in most cases they "float" in the matrix. Agglutinate, cinder, and bombs may fall on the tops of lava streams flowing away from vents. This case could form a matrix-supported breccia, but the matrix would be massive lava, rather than small grains.

Another possible interpretation is that these breccias could be flow-top breccias on aa lava flows. Clinker on the surface of aa lava flows commonly consists of loose spiny fragments, but much is actually attached to the massive substrate (Parsons 1969). Continued movement of the flow may cause autobrecciation and rounding of the clinker, producing smaller fragments that lodge in interstices between larger fragments. However, McDonald (1953), who examined over 1000 outcrops in Hawaii, says that such breccias do not have enough sand and dust sized fragments to constitute a continuous matrix. This again stands in contrast to the matrix-supported breccias of the Warrenton and Charlottesville areas. Furthermore, breccias formed in this way commonly contain interconnected fragments, some of which connect with massive lava beneath (Parsons 1969). Interconnected fragments are not seen in these breccias.

Breccias in the Catoctin Formation on the western limb of the Blue Ridge anticlinorium, described by Reed (1955, 1969) and by Gathright (1976) and viewed by the first author of this paper, better fit the criteria for flow-top breccias. The clasts of these breccias are distinctly more scoriaceous and vesicular than those of the eastern limb, and there are fewer matrix granules. The western limb breccias are also altered quite differently. Larger breccia clasts and the smaller granules are commonly replaced by microcrystalline silica and hematite. Many clasts are replaced in part by epidote. The spaces between clasts are chiefly occupied by quartz, epidote, and jasper.

Reinterpretation

The breccias described in this paper better fit descriptions of hyaloclastite pillow breccia. Hyaloclastite is a general term for lava rocks that are brecciated by thermal strain through contact with water or ice. This brecciation results in various pillow breccias and aquagene tuffs. The most helpful references describing pillow breccias for comparison with the Catoctin Formation include descriptions of hyaloclastite in Canadian greenstone belts (Dimroth, 1977; Dimroth and others, 1978) because the greenstone belts are at a metamorphic grade similar to the Catoctin. Carlisle (1963) gives a very thorough description of Triassic pillow breccias in British Columbia that are fresher, but are altered in the breccia matrix. Fisher and Schmincke (1984), Carlisle (1963), and Dimroth and Lichtblau (1979) provide insight into the relationship between textures and mineralogies of recent hyaloclastites and those affected by low-grade metamorphism. The evidence that convinces us that these Catoctin breccias are pillow breccias is based on textural features of the breccias themselves and their association with pillow lavas.

In gross aspect, pillow beccias described by others are similar to the Catoctin breccias in that a matrix of fine granules and globules support dispersed larger clasts (small pillows and pillow fragments) whose ellipsoidal to angular shape may be confused with bombs of an agglomerate and explosion fragments respectively (Carlisle, 1963, pp. 48-49, 55, and Plates 1B and 4A,B [cf. Figure 2 of this paper]; Dimroth and others, 1978, p. 908). Catoctin breccias in northern Virginia, especially in the "low-Ti" breccias of Espenshade (1986), tend to have clasts with size and angularity consistent with "broken-billow breccia" (Carlisle, 1963, p. 63). In broken pillow breccias, clasts that are derived from inner parts of broken pillows tend to be roughly equant and angular, while fragments from pillow rims are commonly slab-shaped, and small unbroken pillows are ellipsoidal (Dimroth, and others, 1978, p. 908). Examples of all these kinds of shapes are common in the clasts of Catoctin breccias. The breccias in central Virginia commonly are similar, but some occurrences have very large "clasts" and appear to be "isolated-pillow breccias" (Carlisle, 1963, p. 55). Most of the large clasts (isolated pillows) are ellipsoidal as in Carlisle (1963, pp. 55-56) but occasional amoeboid pillows (Dimroth and others, 1978, p. 908) occur also.

The matrix of pillow breccias is composed mostly of lava droplets (globules) and fragmented droplets (granules), and fragments spalled from glassy pillow rims (Carlisle, 1963, p. 57). The matrix may also contain small grains derived from interiors of broken pillows (Carlisle, 1963, p. 63). The distinguishing features of pillow breccia matrix grains are largely related to *in situ* brecciation along thermal contraction fractures (commonly conchoidal) and perlitic cracks. Important features include concave and convex faces of grains, triangular shaped grains, knife-like edges, and narrow slabs (Carlisle, 1963, pp. 57-59; Dimroth, 1977, p. 516; Dimroth and others, 1978, p. 910). Many of these features in Catoctin breccia are demonstrated in Figure 3. In a few samples, examples of *in situ* brecciation of small fragments can be recognized, in which the outlines of adjacent grains can be seen to match one another.

It is important to note that photographs of hyaloclastite such as Figure 10-9 of Fisher and Schmincke (1984) may give the impression that all matrix grains of all pillow breccias must be highly angular, but photographs such as Plates 3B and 5A and B of Carlisle (1963) and his descriptions of matrix grains indicate that in many pillow breccias, many grains (especially globules) are not angular. Furthermore, Heiken (1972) describes recent hyaloclastites in which most matrix granules are blocky. In the Catoctin breccias the characteristic features of pillow breccia matrix grains listed above are variable in abundance from sample to sample, but they are present to some degree in nearly every sample we have examined. This is with the exception of some rocks that have matrix globules that are very vesicular, resembling "pumiceous hyaloclastite" of Dimroth and others (1978).

Interpreting the Catoctin breccias as pillow breccias also accounts for the mineralogical difference between larger clasts (plus small ones texturally similar to them), matrix grains, and mesostasis. When pillow breccias are formed, pillow rims, spalled rim fragments, globules, and granules are commonly composed of sideromelane, while the inner parts of isolated pillows and pillow fragments from them tend to be of tachylite or microlitic basalt (Carlisle, 1963, p. 54-56; Dimroth and Lichtblau, 1979, p. 1337). The sideromelane components are preferentially palagonitized early in their history, while other components commonly are not (Carlisle, 1963, p. 68; Dimroth and Lichtblau, 1979). Palagonite also becomes the intergranular cement to the breccia. In many cases the palagonitization process concentrates Ti^{4+} and Fe^{3+} oxides on the rims of globules and granules (Peacock, 1926, p. 394; Dimroth and Lichblau, p. 1324). Through seafloor metamorphism and burial metamorphism the palagonite is commonly chloritized, or it may be prehnitized, silicified, or carbonitized (Dimroth and others, 1978, p. 911; Dimroth and Lichtblau, 1979, p. 1326-1327). The tachylitic or microlitic components, however, are not affected to the same degree (Dimroth and Lichtblau, 1979, p. 1335, 1338).

Differential metasomatic effects like these among the different components of a pillow breccia prior to regional metamorphism could result in a mineralogical variation such as is seen in the larger clasts, the matrix grains, and the mesostasis in the Catoctin breccias. The concentration of sphene commonly seen on matrix-grain rims in the Catoctin breccias is also consistent with the above described scenario. It is significant that the matrix grains that are texturally like the larger clasts in the Catoctin breccias are not at all altered like the common matrix grains even of the same size or larger, indicating that grain size did not control the alteration, but more likely it was controlled by clast composition prior to regional metamorphism. A few samples have large basaltic fragments that have a partial rim with alteration similar to the matrix. These are apparently pillow fragments that contain part of a rim that was once palagonitized sideromelane.

ASSOCIATION OF PILLOW BRECCIA WITH PILLOW LAVA

One of the most important characteristics of pillow breccias is their common association with normal pillow lava (Fisher and Schmincke, 1984; Dimroth and others, 1978; Carlisle, 1963), although this association is not universal (Jackson, 1980; Dimroth and others, 1978). The textures described above are convincing evidence to interpret the Catoctin breccias as pillow breccia, but the association of this breccia with pillow metabasalt strengthens this interpretation. Outcrops of pillow lava have been found within the central Virginia breccia zone at the Hardware River, at Cismont Manor Farm north of Keswick, and at the abandoned quarry at Madison Run a few miles north of Gordonsville, and recognizable pillow structure has been seen in a boulder in a small stream south of Gordonsville. In northern Virginia in the Marshall quadrangle Espenshade (1986) notes two occurrences of pillows in the low-Ti breccia zone. Another outcrop has been located by the senior author of this paper in the high-Ti breccia on the west side of state road 699, 1.5 km south of Old Tavern. Pillow structure has not been found outside of the breccia belts.

The large pavement outcrop of pillow metabasalt on the Hardware River (Figure 4), discovered by Nick Evans, is especially informative. It is polished by river-sediment abrasion, revealing textures that normally can be seen only in cut slab samples. Many features can be seen that exemplify processes involved in pillow breccia formation. The strike of contacts between three distinct zones within the outcrop and the orientation of pillow elongation (apparently a primary feature) is consistent throughout the outcrop and dip is nearly vertical. Stratigraphic up is to the southeast. The northern part of the outcrop is a zone of pillow breccia that occurs along strike with the other breccias in that belt. The breccia zone is conformably overlain by massive metabasalt with chlorite amygdules. This massive horizon has some poorly outlined, but clearly recognizable, pillow structures that are probably "welded" pillows of Dimroth and others (1978). The massive zone has a sharp contact with an overlying zone of pillow metabasalt.

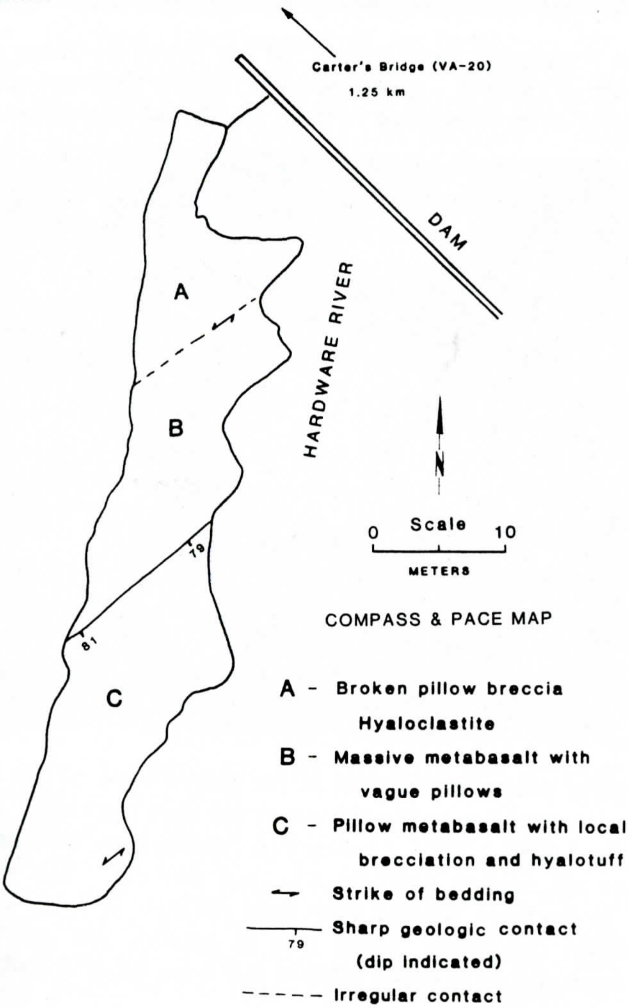

Figure 4. Compass and pace map of outcrop of Catoctin pillow breccia and pillow metabasalt on Hardware River, 1.25 km southeast of Carter's Bridge, Alberene 7.5 minute quadrangle, Virginia.

Figure 5. Pillow metabasalt from Zone C of the Hardware River outcrop (Figure 4). Concentrations of epidote + quartz occur on pillow rims and quartz is the main mineral in the more open interstices (white). Scale bar represents 0.5 m.

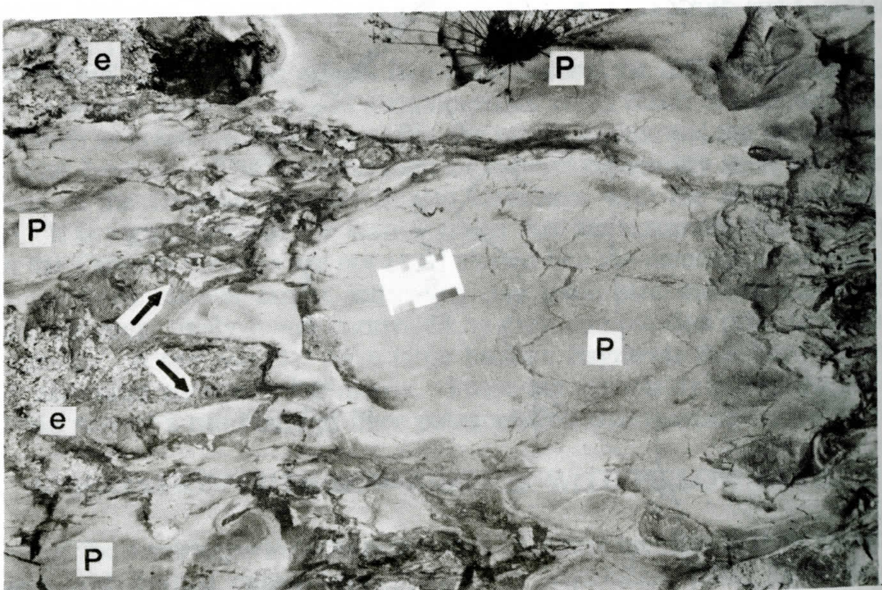

Figure 6. Incipient pillow brecciation in the pillow zone of the Hardware River outcrop. The scale card has both centimeter and inch markings. Interiors of four pillows are marked with a "P". Contraction fractures are abundant. Outer parts of some pillows are broken apart in situ (see arrows). The material between fragments and in fractures (darker gray than pillows) is more chloritic than the pillows and pillow fragments. Areas rich in epidote and quartz (labeled "e") probably were open spaces between pillows that were filled in during hydrothermal alteration or metamorphism.

In the pillow zone (Figure 5) a wide range in pillow size occurs, but most are mattress-type pillows (Dimroth and others, 1978) on the order of 0.3 m X 1.1 m to 1.0 m X 3.6 m. Involuted lobes occur on larger pillows, and interconnections between pillows are common. Pillows are visible because of concentrations of epidote + quartz on pillow rims and in interstices. The pillows are not vesicular. In places epidosite blocks ranging up to a meter across with sharp boundaries are included with the pillows. Pillows at the base of the pillow zone are highly flattened, probably due to original loading while the pillows were still plastic.

Within the pillow zone, and in the pillow breccia zone, there are examples of all stages of pillow brecciation. Dimroth and others (1978) and Carlisle (1963) describe how pillows are brecciated. Pillows cooling in the solid state develop concentric and radial contraction fractures. Many chilled pillow rims spall off, and contraction fractures that penetrate the interior widen, and the pillow is dismembered into angular fragments, most of which do not have a chilled rim. Figure 6 shows incipient pillow brecciation in the pillow zone (Zone C, Figure 4) of the Hardware River outcrop. Contraction fractures appear dark within the lighter pillows. In places, lobes on pillows are broken up, but fragments can be mentally fit back together, demonstrating the in situ brecciation.

Carlisle (1963) shows that where pillow breccias occur as distinct horizons, globules and granules from lava droplets and shattered droplets are mixed with pillow fragments that are formed in the manner described above. Dimroth and others (1978) found an inverse relationship between pillow size and amount of pillow breccia, so that zones of pillow breccia are associated with small pillows. Smaller pillows tend to be ellipsoidal in shape and have thicker glassy rims that disaggregate and become sideromelane shards. These features can be seen in the breccia zone (Zone A, Figure 4) of the Hardware River outcrop. Figure 7A shows small isolated pillows with a peripheral arrangement of vesicles and a thick chilled rim. Nearby, a similar pillow has a partly spalled rim (Figure 7B, see arrow), and several elongate rim fragments occur in the adjacent matrix.

It is quite possible that pillow lava such as that described above is more abundant in the Catoctin Formation than has been previously recognized. As noted in a previous section, rocks occurring between breccia outcrops are rarely exposed because of differential weathering and erosion. The Hardware River exposure is an unusual outcrop where both pillow breccia and adjacent material are well exposed, and the adjacent material is pillow lava. On strike directly across the river is a hillside outcrop of the same rocks, typical of "good" exposures in Virginia. Here, the pillow breccia is readily identifiable, but the pillow basalt is difficult to recognize. Pillow lavas may occur elsewhere in association with the breccias, but are missed because of poor exposure. The other outcrops of pillow lava that have been discovered are relatively small and therefore do not show as many of the features noted above.

One other matter that implies that the Catoctin breccias are pillow breccias is the association with what appears to be "hyalotuff" (Dimroth and others, 1978). In northern Virginia several outcrops were found in which the normal breccia is in contact with what appears to be massive metabasalt. Sawn slabs of the massive-looking material reveal it to be composed of particles on the order of 1-10 mm that look like matrix grains of the pillow breccias. In some, the grains are like pumiceous hyaloclastite, in others they are textureless. Graded bedding in the tuffs (Dimroth and others, 1978; Carlisle, 1963) has not been recognized, however. In some of these tuffaceous rocks, isolated larger clasts occur.

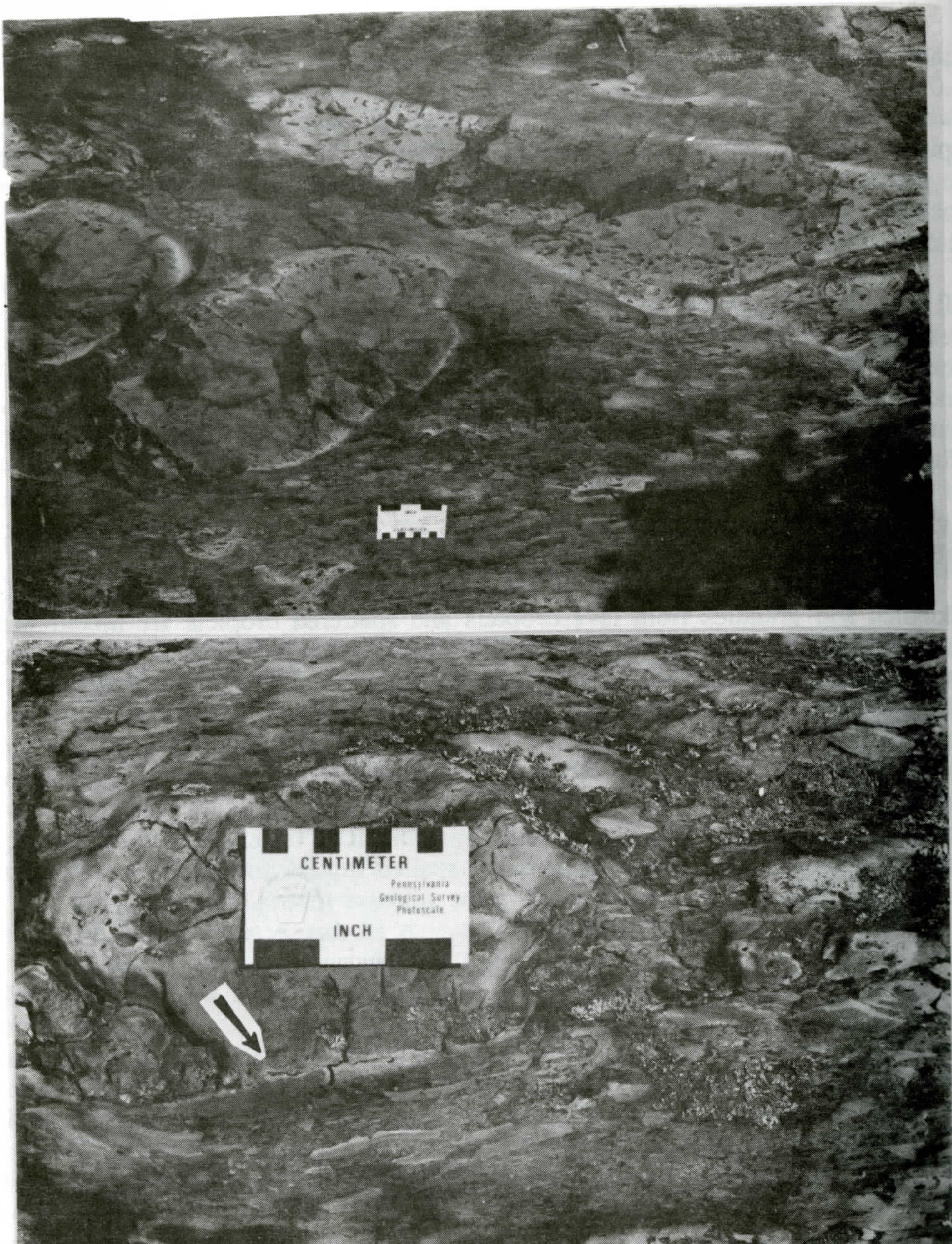

Figure 7. (A, Top) Isolated small pillows in pillow breccia zone of the Hardware river outcrop (Zone A, Figure 4). Scale card has both centimeter and inch markings. A shadow covers the lower right. Note the thick chilled rims expressed as a lighter color and the higher concentration of vesicles near rims. (B, Bottom) An isolated pillow in pillow breccia zone. Arrow points to a chilled rim that was nearly spalled off. Just below it are fragments of spalled rims from this or other pillows. The matrix of both 7A and 7B consists of chloritized globules and granules in a chloritic mesostasis.

PALEOENVIRONMENTAL IMPLICATIONS

Although a paleoenvironmental interpretation of the Catoctin Formation is beyond the scope of this paper, a few comments are appropriate.

The occurrence of extensive zones of pillow breccia in single statigraphic horizons in the Catoctin Formation has important paleoenvironmental and tectonic implications. Models must account for substantial aqueous deposition during Catoctin times in the area now represented by the rocks of the eastern limb of the Blue Ridge anticlinorium. In northern Virginia this aqueous deposition is recognized at least in the lower part of the Catoctin Formation; pillows and pillow breccias have not been found in the rest of the Catoctin in that area. In central Virginia aqueous deposition was at least during late Catoctin time. Breccias in isolated occurrences from the base and elsewhere in the Catoctin have been noted by Conley in reconnaissance work, but these have not been studied in detail, and may or may not be pillow breccias.

Most pillow breccias are formed by either flow of lava from land into bodies of water or by eruption of lava in a subaqueous (or subglacial) environment. Continental flood basalts commonly disrupt drainage patterns forming shallow lakes. Flow of new lavas into these lakes can produce pillows and pillow breccias. When lava flows into a lake, pillow breccia forms at the front of the flow with elongate isolated pillows which define a foreset bedding. The foreset horizon is overridden by massive basalt which is structurally unconformable to the foresets (see Waters, 1960, Plate 2, Figure 1 and accompanying discussion). The thickness of the foreset horizon corresponds with lake depth. The amount of palagonitized hyaloclastite matrix in the breccia decreases upward, and the elongate pillows grade upward into massive lava. On the other hand, where lava erupts under water and pillows and pillow breccia are formed, pillow breccia conformably overlies pillow lava (or massive lava) extruded in the same event (Dimroth and others, 1978). Succeeding lavas normally have a flat to gently undulating contact to the underlying pillow breccia.

In central Virginia, evidence available at present favors extrusion of Catoctin basalts into some sort of marine environment as the mechanism that formed the pillows and pillow breccia, rather than flow of lava into lakes in a terrestrial setting. At the Hardware River outcrop, massive lava (with welded pillows) overlies pillow breccia. The base of the breccia is covered, and it is not known what underlies it. The pillow breccia does not contain elongate pillows, and there is no increase in pillow size or abundance upward, nor a decrease in percent matrix. Furthermore there is no difference in attitude between the pillow breccia and the overlying massive zone. The structural relationship is more like subaqueous extrusion, as observed in submarine basalts in Canadian greenstone belts (Dimroth and others, 1978).

The breccias in the upper part of the Catoctin Formation in central Virginia are also associated with stratigraphically superjacent marine-like metasediments. At the Hardware River, outcrops of massive metabasalt occur above the pillow and breccia outcrop described above. Exposed about 100 m stratigraphically above the pillow outcrop is an apparently conformable contact with medium to very fine grained clastic metasediments with graded bedding. For at least 60 km the Catoctin breccia zone immediately underlies monotonously similar finely laminated phyllites that in places show graded bedding ("Esmont slates" of Conley, 1978, in the area south of Charlottesville; "True Blue Formation" of Pavlides, 1989, in the area east of Orange), with a few lenses of metasandstone. There are no signs that erosion had removed any part of the breccia zone at the top of the Catoctin prior to deposition of the fine clastics.

Wehr (1983) considers the Lynchburg Group, underlying the Catoctin, to be submarine fan deposits and infers a conformable sequence into the Catoctin Formation. If this interpretation is correct, the entire Catoctin Formation in that area could be submarine. One might expect that if that were the case, there would be abundant occurrences of pillow basalt throughout the Catoctin outcrop belt, and in spite of paucity of outcrop there would have been more discoveries of pillows. It is a misconception, however, that extensive pillow development must be present in every submarine lava. In the 2000 m thick sequence of submarine basalts on Quadra Island (Carlisle 1963), 90% of the section is massive amygdaloidal flow and 10% is pillow lava and pillow breccia. Carlisle's map shows small occurrences of pillow breccia widely scattered in the sequence, and one belt near the top of the section where pillow breccias are concentrated. Jackson (1980) describes submarine lavas in the Blake River Group in northeastern Ontario in which massive and hyaloclastite facies are most abundant and pillowed facies constitute only about 20% of the rocks. The typical occurrence is massive flows with hyaloclastite flow-top breccias. Massive lavas seem to be associated with rapid eruption rates and relatively steep paleoslopes, while pillow lavas result from slower flow rates (Dimroth and others, 1978). Further work is needed to determine how much of the Catoctin Formation in central Virginia was actually extruded in a subaqueous environment.

It appears that the pillow breccia in the lower part of the Catoctin Formation in northern Virginia may have formed in a shallow marine environment, however subaqueous extrusion in large lakes cannot be excluded. The outcrops of pillow lava in the Marshall quadrangle have a relationship with pillow breccia that indicates subaqueous extrusion, for in each outcrop pillow basalt is conformably overlain by pillow breccia that is apparently of the same flow. The Fauquier Formation, which stratigraphically underlies the Catoctin in northern Virginia, consists of coarse clastic metasediments at its base and generally becomes finer upwards (Espenshade, 1986). In the Warrenton area black slates underlie the Catoctin (Furcron, 1939), and north of there the uppermost Fauquier units are phyllite, finely-laminated graded-bedded metasiltstone, and discontinuous carbonates (Espenshade, 1986; Parker, 1968). Along Goose Creek in the Lincoln quadrangle, the senior author has located a Catoctin-like metabasalt flow in the lower part of the Fauquier Formation and interbedding of carbonates, metabasalt flows, and phyllites at the Fauquier-Catoctin boundary, showing that mafic volcanism had begun early in Fauquier time and that the Fauquier-Catoctin transition is conformable. The presence of carbonates and fine clastic sediments at the same stratigraphic interval for such a long strike length immediately beneath pillow basalt and pillow breccia seems to suggest that a shallow marine environment existed at the onset of voluminous Catoctin volcanism. However, the Fauquier Formation has been interpreted by others as representing alluvial fan and lacustrine deposits (Espenshade, 1986). The stratigraphic relationships between coarse clastics, fine clastics, carbonates, and hyaloclastites in the Fauquier and Catoctin Formations are not greatly dissimilar to relationships in the Lagoa Feia Formation, which was deposited in the (pre-oceanic) rift valley stage of the Campos Basin of eastern Brazil during separation of Africa and South America in the Cretaceous (Bertani and Carozzi, 1985).

Further work is needed to improve understanding of the paleoenvironments associated with the pillow breccias of the Catoctin Formation of the eastern limb of the Blue Ridge anticlinorium. It is also not clear why pillow breccias are abundant at the base of the Catoctin in northern Virginia (north of where the Culpeper Mesozoic Basin truncates the Catoctin Formation) but occur mostly in the upper Catoctin in central Virginia. This may be an important issue to address because recent regional compilations (Rankin, 1988; Horton and others, 1988) have suggested that the "Catoctin Formation" southwest of the Culpeper Basin and the upper part of the Lynchburg Group (containing ultramafics) together belong to a separate terrane ("Jefferson terrane") from rocks along strike to the northeast. Perhaps the geochemistry of the greenstones in these two areas will reveal other differences or similarities to help resolve this terrane boundary problem.

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DEPOSITIONAL ENVIRONMENT OF THE GLAMORGAN COAL IN SOUTHEASTERN PIKE COUNTY, KENTUCKY, INFERRED FROM GEOCHEMICAL, PETROLOGICAL, AND COMPUTER MAPPING STUDIES

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ABSTRACT

The Lower Pennsylvanian Glamorgan Coal of the Lee and Breathitt Formation (undifferentiated) in southeastern Pike County, Kentucky, is a high volatile A bituminous coal. A low-sulfur ($\bar{x}$ = 0.96%, s.d. = 0.47%), and relatively high-BTU coal ($\bar{x}$ = 13,360, s.d. = 1,757) (dry basis), the Glamorgan is of economic interest, but exhibits thinning and splitting characteristics, making assessment of mineable areas difficult.

Elemental contents of the Glamorgan seam appear to be directly related to depositional subenvironments. To the north and east, where peat accumulation was relatively uninterrupted, Al_2O_3 is enriched in the 1.5 specific gravity float separate (the coal portion), and is interpreted to reflect slow settling of suspended clays in a low-energy, backswamp environment. Na_2O is also enriched in this area in the float separate, but is probably not present in mineral matter. MgO, Al_2O_3, SiO_2, K_2O, CaO, TiO_2 and Fe_2O_3 are enriched in the 1.5 specific gravity sink separate (the sediment-enriched portion) to the south and west, where the accumulation of peat was intermittently interrupted by clastic influx. The low concentration of total sulfur within the seam as a whole ($\bar{x}$ = 0.96%, s.d. = 0.47%) suggests little marine influence during peat accumulation. Lithologies above the seam in the western portion of the study area provide evidence for the presence of marine and/or brackish water which may have been the source of higher sulfur values in the upper split of the Glamorgan.

The Glamorgan is split by a major intra-seam parting in the southern and western portion of the area, reflecting a river-dominated setting. Thickness variations (from approximately 0.5 to 1.5 m) of the Glamorgan are likely due to the buttressing effect of the underlying Gladeville Sandstone during compaction.

The Glamorgan Coal formed on a lower delta plain. Two distinct subenvironments are: 1) a river-dominated setting in the south and west, and 2) a protected backswamp setting in the north and east. Recognition of these subenvironments and their geographical limits should aid in exploration and exploitation of the Glamorgan and other coals formed in similar environments. The subjacent lithology apparently controls the thickness of the seam and the proximity of an overlying marine zone may control sulfur distribution in the seam. Ash and BTU contents are directly related to the subenvironments.

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INTRODUCTION, PURPOSE, AND GEOLOGIC SETTING

The Glamorgan Coal of southeastern Pike County, Kentucky, is a high volatile A bituminous coal. Averages of 64 raw coal analyses within the study area show the seam to contain 0.96% sulfur, 13.10% ash and 13,360 BTU (dry basis). Averages of 70 analyses (Scales, 1985) of the 1.5 specific gravity float separate show 0.81% sulfur, 5.74% ash, and 14,556 BTU (Table 1) (dry basis). The Glamorgan is a low-sulfur, relatively high BTU coal and exhibits thinning and splitting characteristics which make assessment of mineable areas difficult. This study was undertaken to develop a depositional environment model which could assist in the selection of areas for further exploration and development.

Table 1. Proximate analyses of the Glamorgan coal (dry basis).

		All samples			Main and lower split			Upper split		
		n	$\overline{x}$	s.d.	n	$\overline{x}$	s.d.	n	$\overline{x}$	s.d.
%S	raw	64	0.96	0.47	55	0.86	0.30	9	1.62	0.73
	float	70	0.81	0.27	61	0.76	0.24	9	1.11	0.22
%ash	raw	64	13.10	10.94	55	9.86	6.00	9	31.73	12.78
	float	70	5.74	1.03	61	5.74	1.05	9	5.74	0.94
BTU	raw	64	13360	1757	55	13888	970	9	10142	2158
	float	70	14566	180	61	14576	178	9	14497	182

Analyses of individual samples, including %S, % ash, BTU, % volatile matter, % fixed carbon, and fusibility index are presented in Scales (1985).

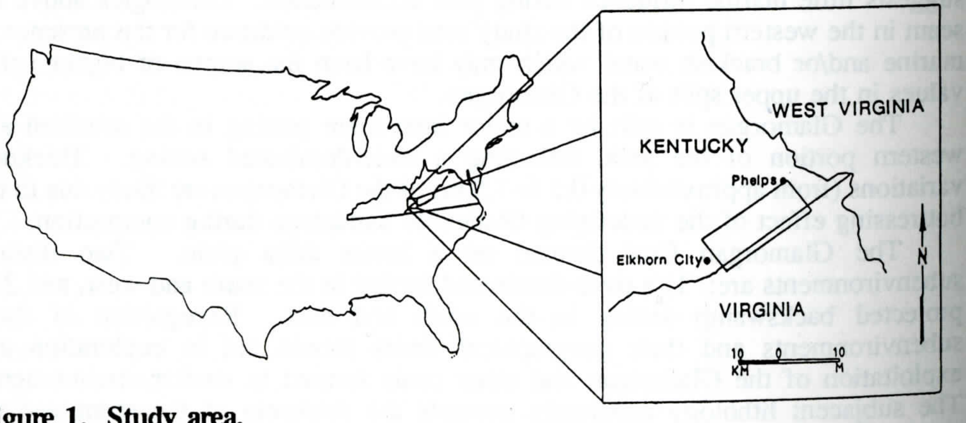

Figure 1. Study area.

The study area, approximately 440 sq. km., is located in southeastern Pike County, Kentucky, and a portion of adjacent Buchanan County, Virginia (Figure 1). The Glamorgan Coal lies within the Lower Pennsylvanian Lee and Breathitt Formation (undifferentiated), which comprises a sequence of interbedded sandstone, siltstone, shale, limestone, coal, and underclay (Figure 2). The

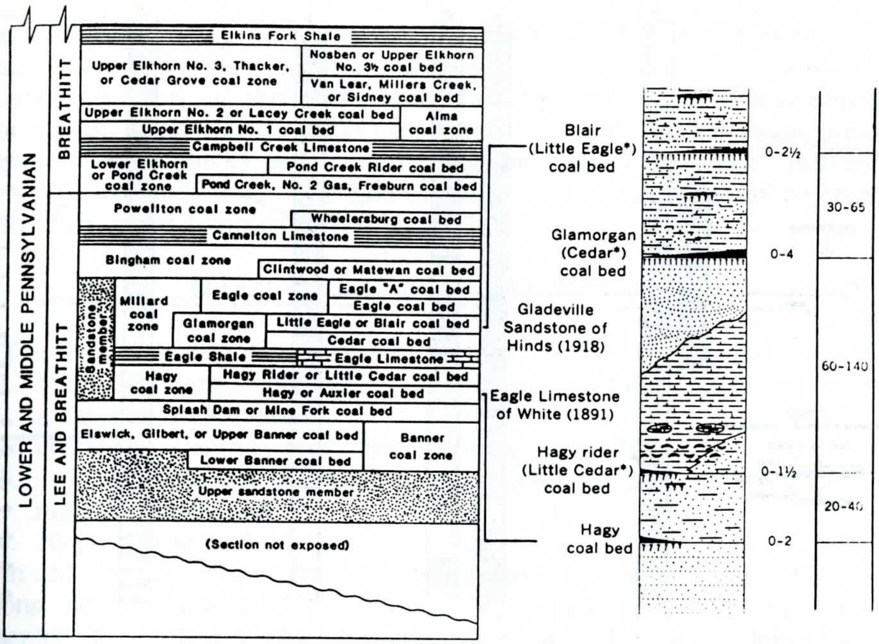

Figure 2. Correlative coal zones (from Rice and Smith, 1980) and generalized stratigraphic column (modified from Outerbridge, 1968). (Thicknesses are in feet.)

Glamorgan overlies the Gladeville Sandstone (Hinds, 1918), which in turn overlies the Eagle Shale (Hennen and Reger, 1914), formerly called the Eagle Limestone (White, 1891). The Hagy zone is the first prominent coal interval below the Eagle Shale. The next prominent coal interval above the Glamorgan Rider is the Blair zone.

To the north and east, the Glamorgan is a single seam with occasional thin, discontinuous partings.[1] In the vicinity of Levisa Fork and to the south and west the seam is split by a major parting (the "Middleman" in miners' terminology) into two thin seams that appear to converge back to a single seam with partings near Elkhorn City. The Glamorgan rider appears to join the upper split in this vicinity (Figure 3).

So far as the authors are aware, no depositional environment studies of the Glamorgan have been made. Chemical analyses had not been performed on the Glamorgan within the study area except for pre-mining evaluation by the Kentucky Berwind Land Company (J. Currens, pers. commun., 1984).

Paleoenvironment depositional models have been proposed for nearby areas.

[1] The terms used in this paper reiterate those used in the initial exploration study of the area and denote potential mineability. "Total seam" refers to the total interval from the first to the last appearance of the coal, including all non-coal lithologies. "Main seam" refers to the single, unsplit seam north and east of the line of splitting. "Upper and lower splits" are the splits above and below a major parting called the "Middleman", south and west of the line of splitting. The "rider" is the thin, probably unmineable seam capping the Glamorgan coal zone. "All-seam" refers to the coal in the area near Elkhorn City where the upper and lower splits and rider converge to a single seam with partings. For further clarification, see Figure 3.

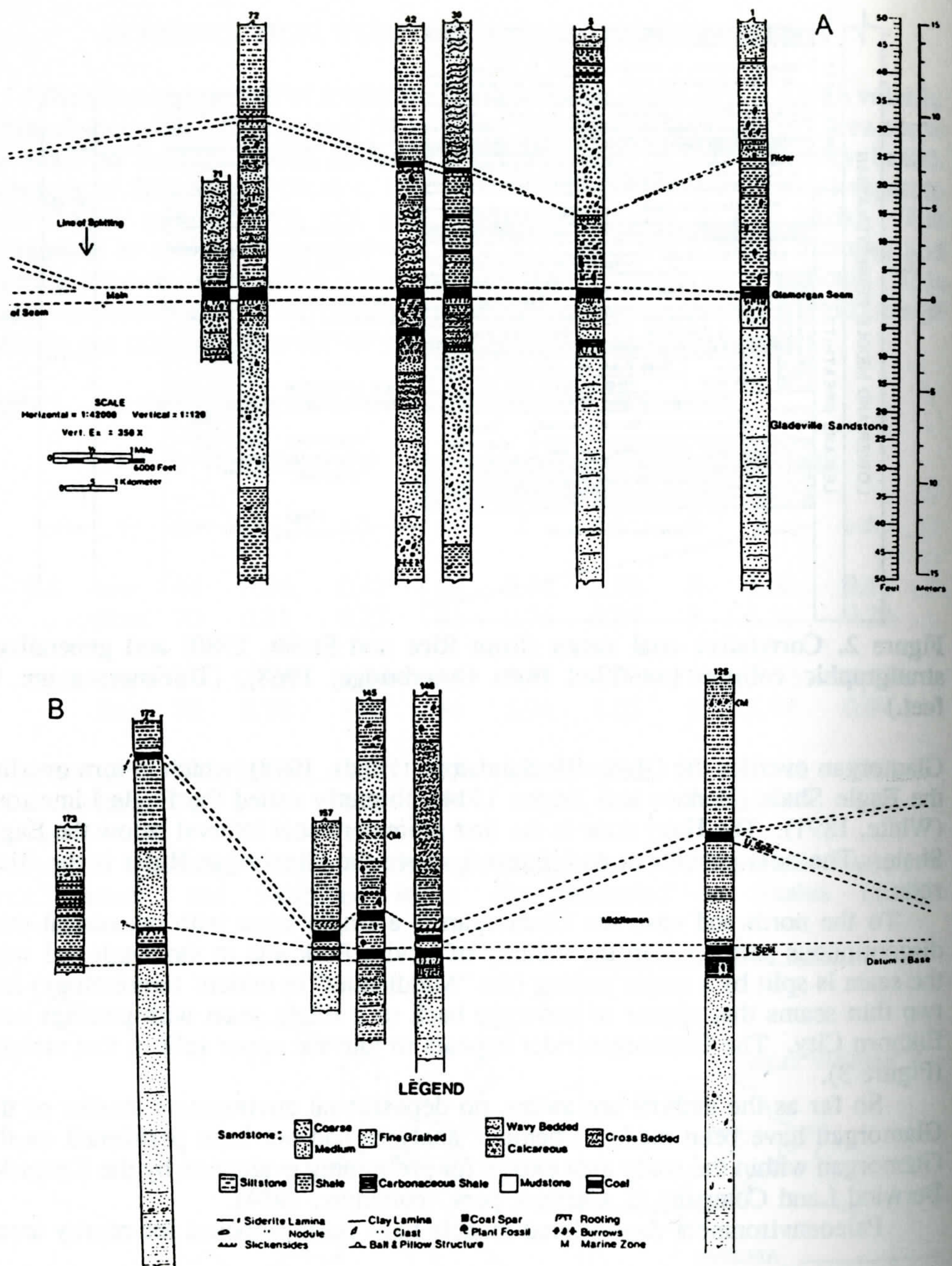

Figure 3. Columnar sections illustrating the splitting characteristics of the Glamorgan coal seam. The line A-B extends generally from northeast to southwest across the study area. The placement of the section line is shown in Figures 4 and 5.

Arkle (1974) reviewed previous studies and characterized the Breathitt Formation as brackish to non-marine sediments deposited in a linear basin, intermittently subsiding to the southeast, and trending northeast-southwest across West Virginia and adjacent areas of Kentucky. Horne and Ferm (1978) described depositional environments for areas near Pikeville, Kentucky, and parts of West Virginia adjacent to the study area. Cobb, *et al*,, (1984) provided field descriptions of areas west and northwest of the study area with detailed depositional environment interpetations. Haney, *et al*., (1984) described structural controls on environments of deposition in Kentucky and noted the Eastern Kentucky Syncline's influence on coals within the study area. The axis of the Eastern Kentucky Syncline coincides with the axis of the basin described by Arkle (1974).

DATA AND ANALYTICAL METHODS

The following data were available (Scales, 1985):

1. maps showing corehole, outcrop and in-mine sampling and measurement locations;
2. corelogs, electric logs and driller's logs for 176 coreholes;
3. coal thickness and elevation measurements for 105 outcrop and in-mine locations, and;
4. 70 proximate coal analyses.

In addition, 71 splits of coal samples obtained by core drilling at 50 corehole locations were available. Twenty-one were raw coal samples, 44 were represented by both 1.5 (specific gravity) float and sink separates, 4 were represented by 1.5 float, and 2 by 1.5 sink separates, only.

A series of contour maps were prepared using the Surface II Graphics System developed by R. J. Sampson (1978) of the Kansas Geological Survey. Surface II can quickly assimilate a large volume of data and requires little user interaction. All data points were assigned X-Y coordinates from an arbitrary origin. Values to be contoured were then tabulated for each data point.

Interested readers may contact A. S. Scales for additional information concerning the Glamorgan Coal.

Coal samples were analyzed using an EG&G ORTEC Model 6110 Tube-Excited Fluorescence Analyzer to supplement the available proximate analyses. Crushed particle pellets were analyzed for maceral, pyrite and mineral matter composition (1000 point counts per sample). Counting errors were assessed using the chart of Van der Plas and Tobi (1965).

LITHOLOGIC DESCRIPTION OF UNDERLYING ROCKS

The stratigraphic interval from the top of the Hagy Coal to the base of the Gladeville Sandstone is the Eagle Shale. The Hagy Coal is typically overlain by a thin, < 1 m, rooted, commonly carbonaceous shale. The remainder of the Eagle Shale interval generally coarsens and becomes more arenaceous upsection although all combinations of interbedded sandstone, siltstone, and shale are noted. Locally, a sandstone overlies the thin shale near the base of the coarsening-upwards sequence, and is a maximum of 9 m thick, medium- to fine-grained, planar- to wavy-bedded or rarely cross-bedded, and commonly with coal spars and/or siderite nodules at its base. A siltstone overlying this sandstone, where present, is up to 6 m thick, planar-bedded and commonly contains siderite and sandstone laminae. Overlying the siltstone (and dominating the interval), the remaining shale is a maximum of 23 m thick and also planar-bedded with siderite laminae and nodules, and sandstone streaks.

The shale and siltstone, especially the more arenaceous portions, are calcareous. A 20 cm argillaceous limestone was reported above a burrowed zone in the basal portion of the shale in corehole 110. Bioturbation and burrowing are common in the basal, typically sandy, portion of the shale or near the siltstone-sandstone contact where the sandstone is present. Fossil shells were reported from these zones in coreholes 4 and 34.

The Gladeville Sandstone lies between the top of the Eagle Shale and the base of the Glamorgan Coal. The Gladeville generally fines upward, from coarse- to medium-, or medium- to fine-grained sandstone. Wavy-bedding and cross-bedding are common. Coal spars, clay clasts and siderite nodules are common, particularly near the base. Clay laminae occur throughout, primarily as ripple troughs. The lower contact is commonly abrupt. Where in contact with the overlying Glamorgan, the Gladeville is commonly rooted, but it can also be separated from the coal by an interval of variable lithology.

The Gladeville varies from 4 m to 31 m in thickness. To the north and east the Gladeville is typically 15 m thick or less, whereas towards the south and west it reaches its maximum thickness, 15 m to 31 m. A structure contour map prepared for the top of the Gladeville Sandstone indicates an average dip of 1 degree to the northwest and a strike of N30° E.

Figure 4 shows the lithologies subjacent to the coal seam. The stippled sandstone pattern indicates areas where the Gladeville Sandstone is in contact with the seam. Where the Gladeville is not in more or less direct contact, i.e., a non-sandstone interval greater than 0.3 m, the non-sandstone lithologies are shown. Non-sandstone subjacent lithologies include shale, carbonaceous shale, siltstone, and mudstone. They contain plant fossils, coal spars and occasional siderite laminae, and are extensively rooted. The maximum thickness of these non-sandstone lithologies is 8.5 m. Non-sandstone lithologies thicken where the Gladeville is thinnest and on the flanks of the thickest part of the sandstone, i.e., where the Gladeville is in contact with the coal.

COAL SEAM CHARACTERIZATION

To determine if there was any structural influence on the coal seam, a structure contour map was generated using base-of-seam elevations from collected core, outcrop, and in-mine measurements, and 55 additional base-of-seam elevations taken from geologic quadrangles in the study area. Average dip in areas of good control is 1 degree to the northwest and the strike is N30° E, identical to those of the underlying Gladeville Sandstone. The low dip of both surfaces suggests that there was little or no local structural influence on the seam.

Several isopach maps were prepared including the total seam, total coal, the main Glamorgan seam (unsplit) and the upper split (Scales, 1985). The total seam to the north and east of the study area is relatively uniform in thickness (app. 76 cm) compared with the highly variable thickness (2-12 m) to the south and west. For the entire study area, the total coal thickness ranges generally from just below to just above 76 cm.

The total coal and main seam thicknesses are generally in agreement in the northern and eastern portions of the study area. To the south and west, the lower split is thin (less than 51 cm) although it appears to thicken at the extreme southwest edge of the study area. The upper split is generally thicker (up to 1 m) than the lower split due to non-coal partings.

The Glamorgan contains a major intra-seam parting called the "Middleman". A

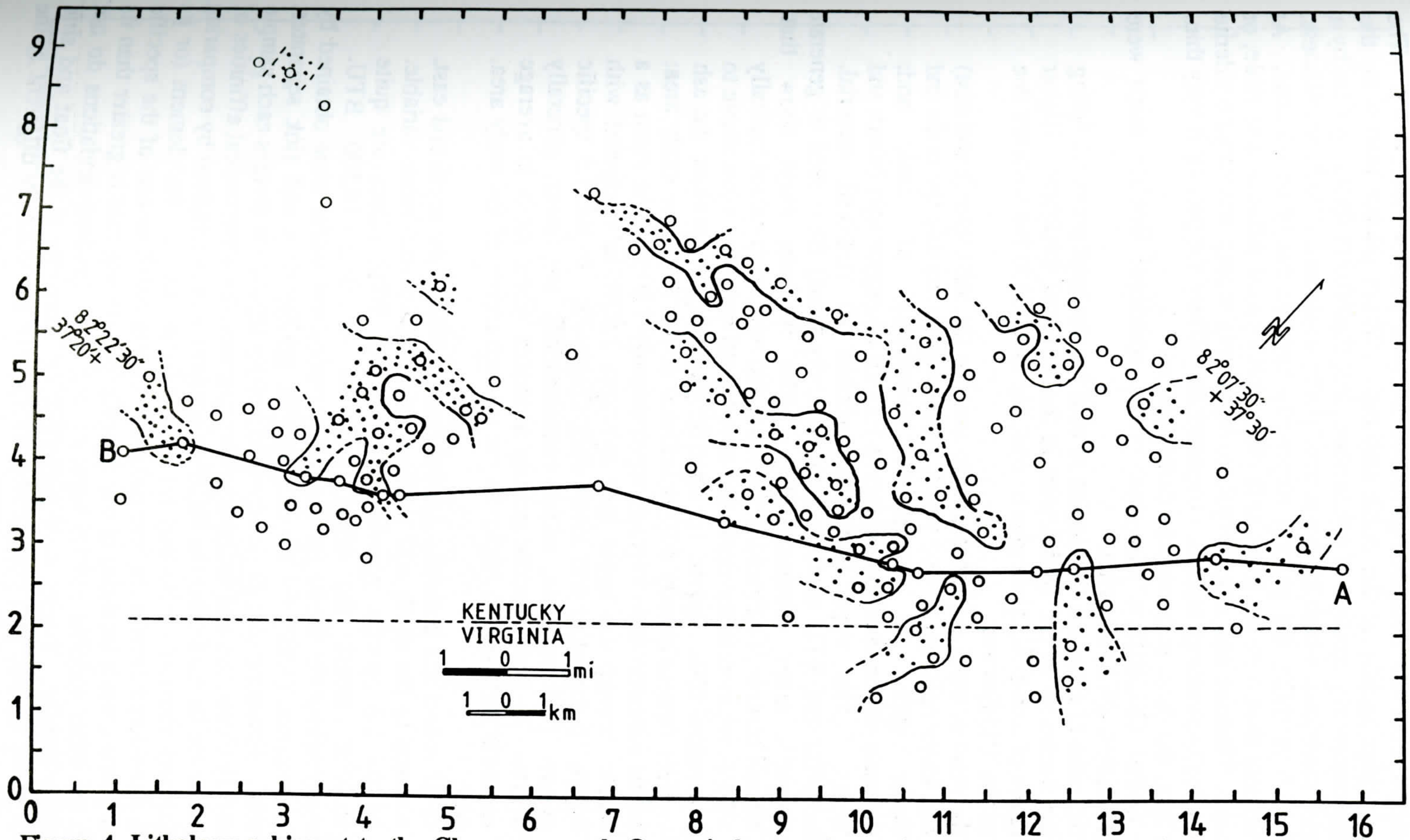

Figure 4. Lithology subjacent to the Glamorgan coal. Open circles are data points used to generate this figure. The stippled pattern indicates where the coal-to-sand interval is less than 1 foot. Areas where this interval is greater than 1 foot have no pattern.

line of splitting divides the study area into two areas of differing character. Figure 5 is modified from the computer-generated isopach map of this parting. The Middleman thins rapidly to the north and east. Where thicker than 6 m, the Middleman commonly consists of a thin shale, less than 0.6 m thick, overlain by a fine-grained sandstone, which in turn is overlain by a siltstone of similar thickness. Where 3 to 6 m thick, the parting commonly consists solely of siltstone. At thicknesses less than 3 m the interval is either interbedded siltstone and shale, or solely shale. Siltstone and shale strata are commonly rooted and contain siderite nodules and/or laminations and plant impressions. Where the parting is very thin, 0.3 m or less, the shale is commonly carbonaceous.

To further characterize intra-seam rocks, additional isopleth maps were generated (Scales, 1985) showing:

1. the ratio of non-coal lithologies to coal for the total seam, including the thickness of the Middleman in the non-coal thickness. Higher ratios of non-coal lithologies to coal (up to 12:1) in the west are due to the Middleman;
2. the ratio of non-coal lithologies to coal for the main (north and east) and lower split (south and west). This map shows that the main and lower split are composed dominantly of coal, and;
3. the ratio of non-coal to coal (up to 1:1) for the upper split (south and west). The upper split is enriched in non-coal material.

Ash, sulfur and BTU contents (raw coal, dry basis) were used to generate several isopleth maps (Scales, 1985). The resulting maps show that:

1. the ash content is elevated in the south and west, with values typically > 25%, whereas to the north and east, values range from just above to just below 10%. In the 1.5 specific gravity float separate, the ash content (approximately 6%) is uniform across the entire study area;
2. sulfur concentrations average approximately 1.00% in the seam as a whole, though the upper split averages 1.62% as compared with 0.86% in the main and lower split (Table 1). In the 1.5 specific gravity float separate, sulfur concentrations are lower, generally 0.75%. In both specific gravity separates sulfur tends to average approximately 0.25% higher in the western corner of the study area, and;
3. BTU values are generally greater than 13,000 to the north and east, whereas to the south and west values are lower and more variable. In the 1.5 specific gravity float separate, BTU values are quite uniform across the study area, at or near 14,500 BTU.

Concentrations (in wt.%) for the major oxides and sulfur were obtained by X-ray fluorescence (XRF) for the 1.5 specific gravity float and sink separates, respectively. Assuming that the 1.5 specific gravity separation divides each sample into relatively clean coal and sediment-rich separates, the geochemical affinities of each element (or its oxide) for the different portions was determined by comparing the means and standard deviations for each sample subset. An element (or its oxide) is considered here to be preferentially concentrated in one of the specific gravity separates when its mean in one specific gravity separate is greater than its mean in the other specific gravity separate and their standard deviations do not substantially overlap. T-tests were performed to determine if the float and sink separates were, in fact, two populations with means significantly different at the 95% confidence interval (Scales, 1985).

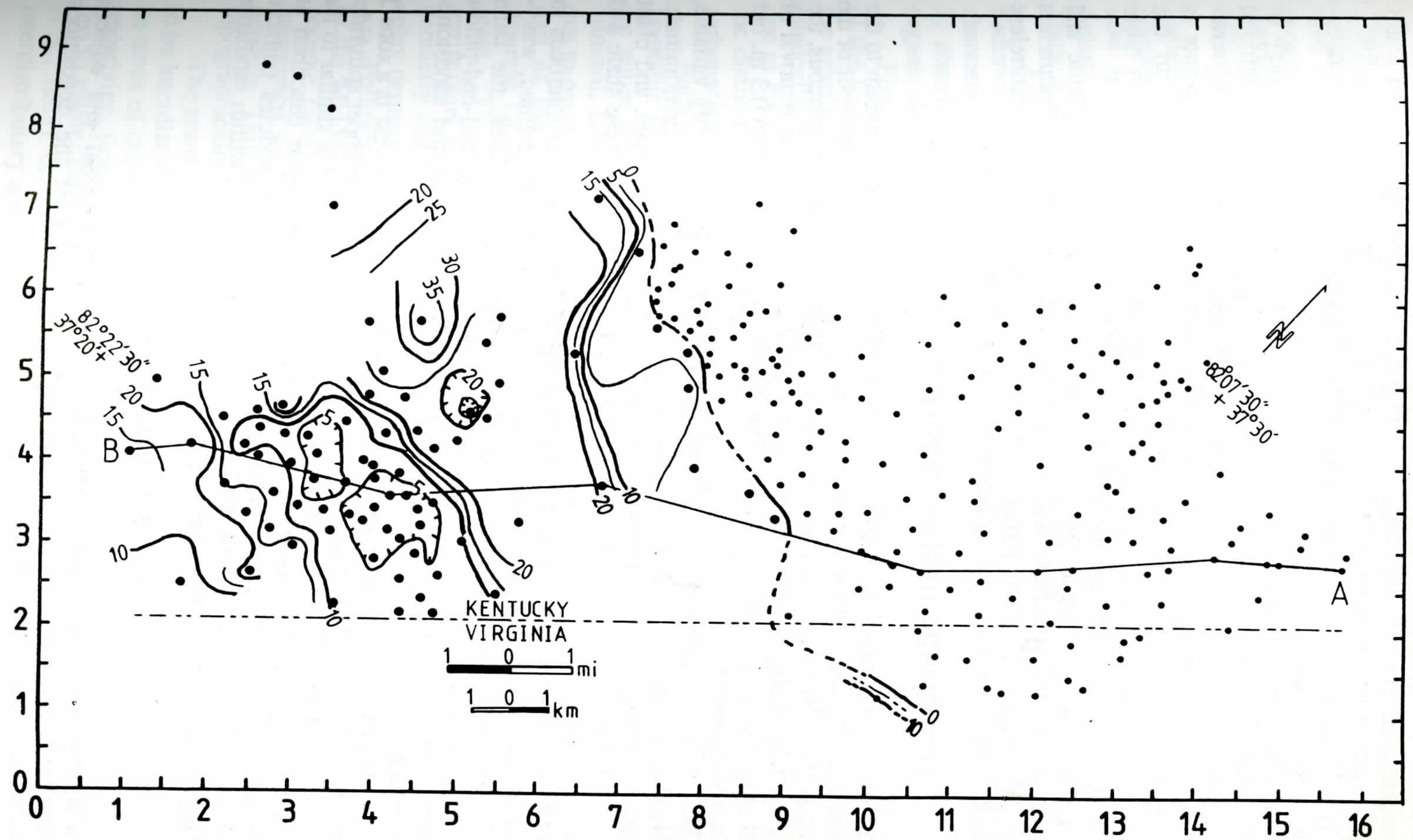

Figure 5. Isopach map of the Middleman intra-seam parting. Darkened circles represent data points used to generate this map. Contour interval = 5 feet (1.52 m).

Within the main seam and the lower split, Na_2O, Al_2O_3, and S are enriched in the float and Fe_2O_3 in the sink. MgO, SiO_2, P_2O_5, CaO, and TiO_2 do not appear to be concentrated in either separate. Within the upper split and all-seam south and west of the line of splitting, S is enriched in the float and MgO, Al_2O_3, SiO_2, K_2O, CaO, TiO_2, and Fe_2O_3 in the sink. P^2O^5 does not appear to be concentrated in either separate.

Twenty-one pellets were analyzed by point counting (1000 points) to determine maceral (vitrinite, exinite, and inertinite), pyrite, and other mineral matter volumetric composition on both raw and 1.5 specific gravity float (mineral-matter-free (mmf)) bases. Although the number and location of data points is limited, exinite appears to increase (from about 12% to about 18%) to the northwest across the study area, roughly paralleling the depositional dip. A more complete discussion of maceral analyses is given in Scales (1985).

Pyrite occurs primarily in two forms, as void fillings (within fusinite and vitrinite retaining cell structure) and as scattered blebs imbedded in featureless vitrinite. Where filling voids, pyrite is dominantly framboidal. In featureless vitrinite, it is either in single, or agglomerates of, subhedral crystals.

LITHOLOGIC DESCRIPTION OF OVERLYING ROCKS

The Glamorgan rider typically lies 5 to 9 m above the Glamorgan seam to the north and east. To the south and west, the rider merges with the upper split of the Glamorgan (J. S. Nelson, pers. commun., 1984). Although mostly continuous, the rider horizon is locally occupied by sandstone bodies, probably due to removal of the peat mass by channel entrenchment. The rider is typically less than 0.3 m, but can attain thicknesses of up to 0.5 m.

The lithologic character of the Glamorgan-to-rider interval is highly variable. Immediately above the Glamorgan Coal, siltstone and shale predominate. Siltstone and shale beds commonly contain siderite laminae, coal spars, and plant fossils. Carbonaceous shale and rooted zones occur locally within these shale and siltstone sequences.

Sandstone strata occur in discrete intervals up to 3 m thick or as interbeds in the silty units. These sandstones are commonly medium- to fine-grained, wavy-bedded and contain clay laminae, clay clasts, and siderite nodules, the latter particularly near the base. Compactional features, such as ball-and-pillow structures and slickensides, are rarely seen at sandstone-siltstone and sandstone-shale contacts.

Lithologies above the rider are similar to those below. Of interest is a zone 3 to 5 m above the rider to the north and east and 3 to 8 m above the upper split/rider to the south and west. This zone is typically 1 to 1.5 m thick, and is similar to the zone in the Eagle Shale underlying the Glamorgan. It has been described as burrowed/bioturbated, calcareous-cemented, containing siderite nodules replaced by calcite, and with sparse fossil shells. This zone typically lies within sandstone or sandier portions of siltstone and shale beds.

DEPOSITIONAL HISTORY

The available information concerning the Glamorgan Coal and adjacent sedimentary rocks was used to interpret the development of the depositional environment of this important seam.

Chesnut (1981) lists known marine zones of the Upper Carboniferous in eastern Kentucky. Marine zones typically coarsen upward, range from a few meters to 37 m thick, and are interpreted as bayfill. The sequence usually overlies a coal seam and grades upward from shale to siltstone to sandstone subjacent to the next coal. Brackish to marine fossils and ichnofossils are common in the basal sequence. Chesnut identified the Eagle Shale as belonging to such a sequence. Sandstone typically caps the bayfill sequence, representing the culmination of the progradational-regressive sequence.

In the study area, the Gladeville Sandstone caps the Eagle Shale bayfill sequence (Figure 6A, B). The Gladeville is greater than 8 m thick, is in abrupt contact with the underlying Eagle Shale, contains siderite pebbles, clay clasts, and coal spars (especially near the base), and contains cross bedding, ripple drift (interpreted as the "wavy bedding" from core log descriptions), and clay laminae. These features suggest a distributary channel/mouth bar origin for this sandstone.

According to Horne and Ferm (1978), interdistributary silt and clay of the backswamp environment are laterally equivalent to the distributary channels. The channels are characterized by three types of fills: 1) active channel sandstone fill; 2) abandoned channel fills of clay and silt; and 3) abandoned channel fills of organic debris.

Figure 6C represents the lower delta plain surface just prior to establishment of the Glamorgan peat swamp. Areas of interdistributary clay and silt appear in the areas occupied by distributary channels (some perhaps abandoned). Where the Gladeville is in contact with the Glamorgan, the sandstone generally fines upward due to the decreasing energy of the channel environment prior to abandonment of the distributary. The coal is underlain by siltstone, shale, and mudstone, and by carbonaceous shale, representing the two types of abandoned channel fills.

Following abandonment of the distributary channel system, conditions favorable for plant growth allowed establishment of a swamp environment across the entire delta plain surface (Figure 6D). The initial influx of sediments was minimal; however, locally the non-coal lithologies:coal ratios are high, probably due to overbank or flood deposits (Figure 6E). Note that the southern and western part of the study area (Figure 6E) represents the lower split, which is thin, whereas the northern and eastern half of the study area contains the thick main seam. The southern and western area was subjected to substantially greater terrigenous sediment influx represented today by the Middleman parting. The geometry and lithologic character of this parting suggest an origin by overbank or splay deposits.

Following abandonment of the channel system that resulted in the deposition of the Middleman, conditions in the southern and western half of the area again became favorable for peat accumulation. This accumulation was periodically interrupted by influxes of terrigenous clastic sediments, probably as overbank deposits (Figure 6F). Some areas within the swamp may have been topographically high and protected from significant clastic deposition.

The seam is thickest where it is underlain by shales and siltstones (interpreted to be interdistributary areas) and thins where underlain by sandstone (interpreted to be areas of channel fill). Two alternatives might explain this relationship: 1) the underlying, linear sand bodies may have acted as buttresses, and the coal thinned due to greater compaction of shale and siltstone away from the channels (Wanless, *et al.*, 1969), or; 2) plant growth and peat accumulation may have started in the interdistributary areas, leading to a greater total thickness than in areas of channels,

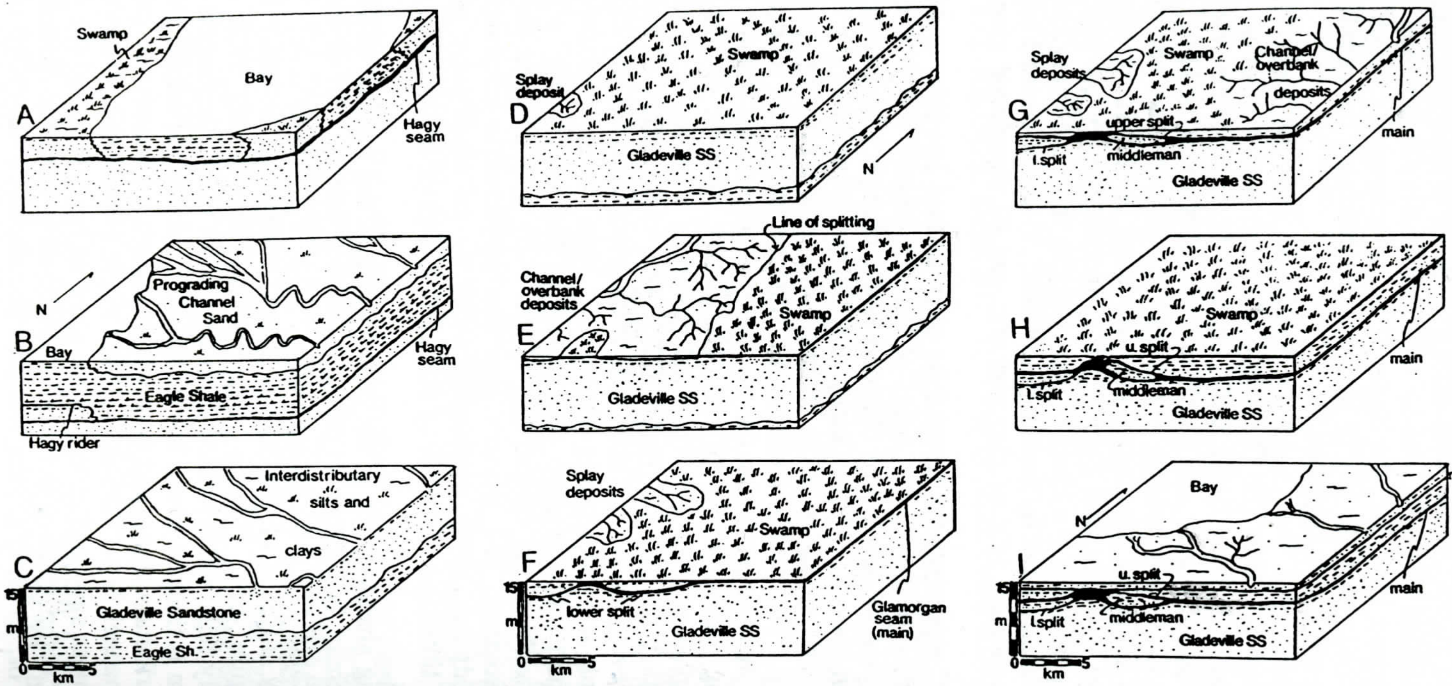

Figure 6. Conditions prior to peat accumulation: A) bayfill sequence; B) distributary channel sand progradation; C) distributary channel abandonment. Conditions during peat accumulation: D) establishment of peat swamp; E) development of Middleman parting and lower split; F) reestablishment of peat swamp. Conditions after peat accumulation: G) development of upper split and end of eastern peat swamp; H) reestablishment of peat swamp preceding rider; I) marine/brackish transgressive sequence. Coal thickness and geometry exaggerated. Schematic only.

which did not become sites for growth and accumulation until after they were abandoned.

Enrichment of sulfur in coal (1.5 float separate) has long been recognized (Thiessen, 1945; Stach *et al.*, 1975; Reyes-Navarro and Davis, 1976). The low total sulfur within the Glamorgan seam suggests little, or no, marine influence on the swamp during peat accumulation (Cecil *et al.*, 1981). The enrichment of MgO, Al_2O_3, SiO_2, K_2O, CaO, TiO_2, and Fe_2O_3 in the sink separate of the upper split reflects the southern and western area's depositional setting which was subject to much greater sediment influx than the northern and eastern half. The enrichment of Na_2O and Al_2O_3 in the float separate of the main (and lower) seam may also be related to depositional setting. The northern and eastern portion of the area was a protected backswamp (low-energy, quiet water setting), allowing the slow settling of suspended clay particles and their incorporation into the peat mass resulting in the enrichment of Al_2O_3 in the float separate. In a study of Western U.S. lignites, Gronhovd *et al.*, (1967), noted that Na_2O was organically bound and did not exist in a mineral matrix, which may help to explain the enrichment of Na_2O in the float separate; however, it is possible that some of the Na_2O may be due to clay particles dispersed through the coal fraction.

Peat accumulation was terminated by a series of newly established channels and overbank deposits (Figure 6 G). Rooting and carbonaceous shales within this interval indicate the cyclical nature of these deposits. With cessation of clastic influx, peat accumulation again occurred, resulting in formation of the rider concurrent with the upper portion of the upper split (Figure 6 H).

Peat accumulation was terminated over the entire area by a marine transgressive sequence (Figure 6 I). Although analyses of the rider were not available, sulfur content is highest on average in the upper split in the western portion of the area perhaps due to the presence of marine/brackish sediments overlying the upper split (Horne *et al.*, 1978).

FUTURE EXPLORATION AND EXPLOITATION

Findings in this study which may be of use in future exploration and exploitation of the Glamorgan Coal (and other coals deposited in similar environments) include:

1. an apparent relationship between coal thickness and subjacent lithology. North and east of the line of splitting, areas in which the coal is *not* underlain by sandstone lithologies, are likely to be areas of greatest coal thickness. The pattern of sandstone bodies shown on the subjacent lithology map (Figure 4) may be repeated west and east of the study area and could serve as a guide for exploration;
2. sulfur contents increase to the south and west in the study area, perhaps due to marine/brackish units overlying the coal in this area. This trend is expected to persist outside of the study area and;
3. within the study area, ash content will be the greatest and BTU content the least south and west of the line of splitting, due to the river-dominated depositional setting. These characteristics are expected to continue outside of the study area south and west of the line of splitting.

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