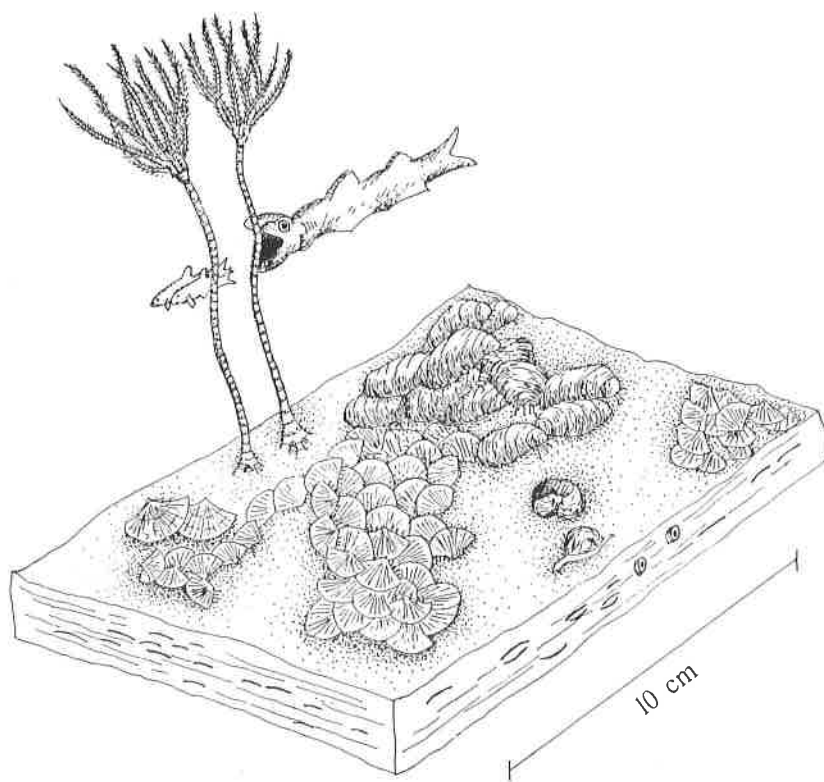


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- 2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
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LATE CARBONIFEROUS PALEOECOLOGY ALONG A TECTONICALLY ACTIVE BASIN MARGIN: AMES MEMBER NEAR HUNTINGTON, WEST VIRGINIA

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ABSTRACT

The Huntington area is situated along what was the tectonically active southern margin of the Ames seaway in Late Pennsylvanian time. As such, the deposits within the Ames Member in this area are the most nearshore, most brackish, and most provenance-influenced Ames rocks that have been described anywhere within the basin. Three major lithofacies with their associated biofacies are clearly defined in the region: green calcareous shales of lagoonal origin (green chonetid shale) that were probably originally brachiopod-mollusk dominated; red fissile shales of intertidal origin (red platy shale) with a sparse biota of inarticulate brachiopods and scallops; and a massive calcareous sandstone of bar-barrier origin (calcareous crinoidal sandstone) dominated by rheophilic crinoids and bryozoans. Conodont faunas, although all a part of the brackish *Cavusgnathus*-biofacies, reflect salinity changes associated with geographic, stratigraphic, and lithogenetic position.

INTRODUCTION

The Conemaugh Group, a lithostratigraphic unit roughly 150 m thick and of Missourian-Virgilian age, crops out on the flanks of the Dunkard Basin in the Northern Appalachian Plateau (Figure 1). Only locally does it contain mineable coals, resulting in its earlier designation as the "Lower Barren Measures" (Wilmarth, 1938). A distinguishing feature of the Conemaugh is a series of marine beds in the lower half of the unit, the most persistent of which is the Ames Member that occurs near the middle of the group. The top of the Ames is

used to separate the Conemaugh into a lower Glenshaw Formation that contains marine beds and an upper Casselman Formation that generally does not (Figure 2). Over most of the outcrop belt in Ohio, the Ames Member is 0.5-2.0 m thick and is composed mostly of limestone, commonly with thin under- and overlying shales. Eastward and southward the proportion of shale increases and limestone decreases, so that in southernmost Ohio and adjacent West Virginia and Kentucky, as well as in the more easterly outcrops in central Pennsylvania, northeastern West Virginia and Maryland, the Ames is mostly shale and locally exceeds 10 m in thickness. In southernmost Ohio, Condit (1912) reported only three Ames localities, two of them calcareous shales and the third a calcareous sandstone. Between Huntington, West Virginia and Louisa, Kentucky the marine character of the Ames is lost (see also Merrill, 1988, Figure 2). Eastward from Louisa to a point approximately 100 km east of Charleston no marine fossils have been reported at the Ames position and none have been found in recent work. Marine, fossiliferous Ames rocks are present on the flanks of the Burning Springs Anticline in West Virginia southward to where the plunge of the fold takes the Ames position below the cover of younger rocks. The Ames has not been recognized in cores about 25 km north of Charleston. This suggests a range of possible shorelines through no more than 30 km in this part of the subsurface.

In most earlier reports both the description of Ames rock types and genetic interpretations were general, but more recently Brezinski (1983) and Saltsman (1986) have described sequences in central and western Pennsylvania that record distinct episodes of transgression,

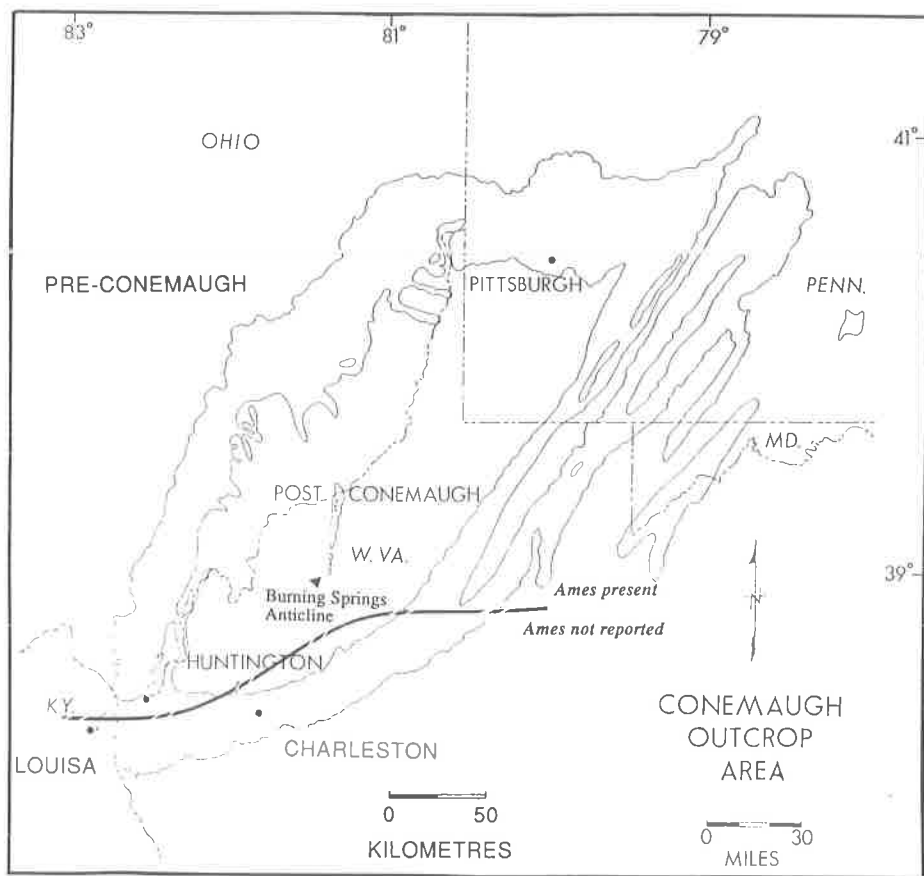


Figure 1. Map of the Dunkard Basin showing the outcrop area of Conemaugh rocks and approximate position of the Ames shoreline.

stillstand, and regression in Ames rocks. Over much of the outcrop area, especially in northern Ohio and Western Pennsylvania, the Ames rocks seem highly uniform, changing only gradually over considerable distances, although control is commonly widely spaced. Southward and eastward it is evident that lateral changes occur with closer spacing, but control generally remains sparse. However, in the Kentucky-Ohio-West Virginia tristate area near Huntington, extensive highway construction has provided an excellent record of these transitional facies in an extreme nearshore setting.

It is the objective of this paper to present the record of these facies, both lithofacies and biofacies, and to interpret their origin by application of their geometry, petrology, megafauna,

ichnofauna, and conodont microfauna.

THE AMES IN THE HUNTINGTON AREA

In the Huntington area exposures of the Ames are found mostly in roadcuts along both sides of the valleys of the Ohio and Big Sandy Rivers (Figure 3) where Glenshaw rocks are exposed on the converging end of the Pittsburgh-Parkersburg-Huntington Synclinorium. The structure in this area, previously described by Merrill (1973a, 1988) consists of a broad, nearly flat-bottomed syncline with dips as steep as $3\frac{1}{2}^{\circ}$ on the northern limb and even steeper dips of up to 9° (Martino, personal communication, 1990) on the southern. The roles of the various structures on the distribu-

LATE CARBONIFEROUS PALEOECOLOGY - AMES MEMBER

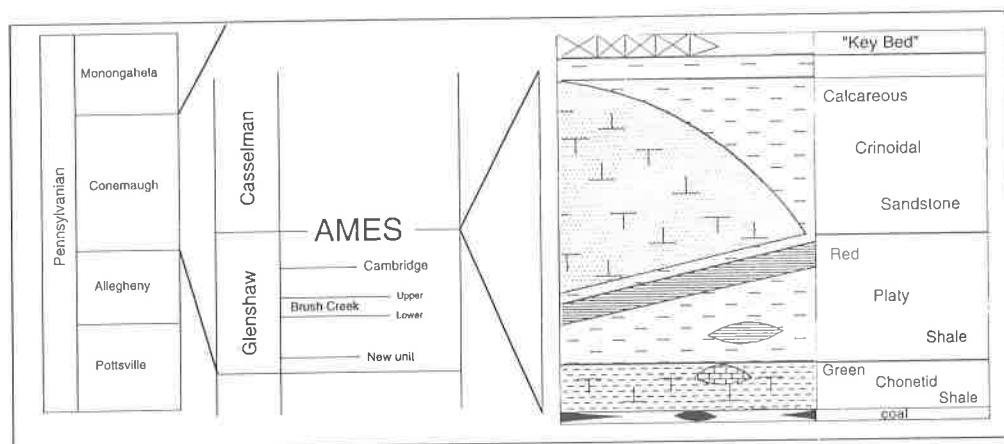


Figure 2. Generalized stratigraphic columns. The left column portrays the four major lithostratigraphic units (groups) in the Pennsylvanian System in its type region. Approximate height of the column is 340 m. The middle column portrays the extent of Pennsylvanian rocks studied in the Huntington region comprising uppermost Allegheny Group (Breathitt Formation in Kentucky) and Glenshaw and Casselman Formation rocks, the Glenshaw containing several marine units as shown. Approximate height of this column is 168 m. The right column portrays the common marine lithosomes in the Ames Member in the Huntington region. Approximate height of this column is 10 m.

tion of Ames rocks has previously been described (Merrill, 1988) and it is suggested that these structures were active during Ames deposition.

Ames lithofacies in the Huntington area are dominated by terrigenous materials, some relatively coarse. Several distinctive lithosomes can be distinguished and their general vertical and lateral relationships are shown in Figure 2. Most of these terms are self-explanatory, but the "key bed" needs further explanation. It is a dark, bluish-green, siderite-cemented sandstone that weathers rounded and blood-red. Its persistence (it can be traced for several kilometres along the Big Sandy River) and its presence just above Ames marine rocks of whatever lithology, cause its designation. The base of the Ames Member is usually sharply defined by a thin coal (up to 40 cm) or, where the coal is absent, by a bluish-gray underclay. Like most Conemaugh marine units, the top of the Ames is commonly difficult to locate precisely; shales become siltier upward and the fossil content declines and disappears. Thus, the exact position at which the record of marine conditions ceased is difficult to define, and indeed, the freshening of the waters during

progradation from marine to brackish to fresh was probably so gradual that no sharp demarcation is to be expected.

Although individual Ames lithosomes are less than completely persistent throughout the area, the Ames marine interval as a whole is much more so. At only one locality (F5, Figure 3) where the Ames is expected, have marine fossils not been found. At this locality a fissile red shale without any apparent fossils may represent its position. Previous studies (Merrill, 1988) have shown that each of these lithosomes varies in thickness across the area as does the entire Ames interval. Some of these changes are associated with the minor structures, but the southernmost Ames outcrops are terminated as steep dips carry the Ames position above the hilltops. Near Louisa, Kentucky (Figure 1), rocks at the Ames position are brought back into the hilltops by a gentle syncline, but no marine fossils have been found (Connor and Flores, 1978). It seems certain that the southern Ames shoreline has been crossed.

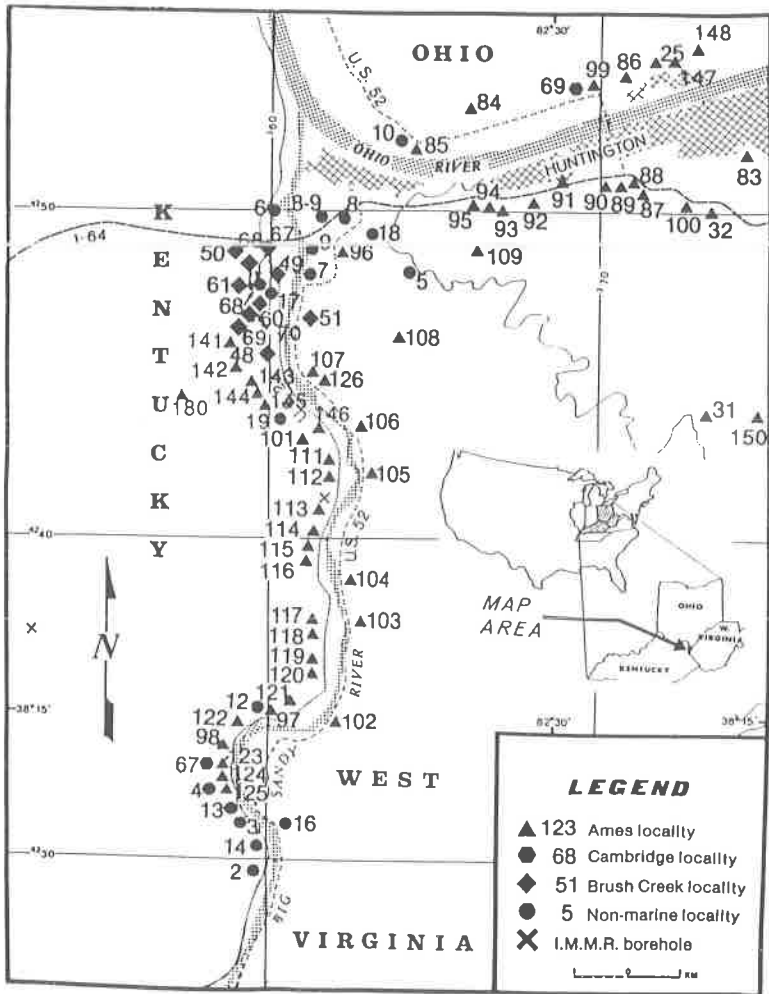


Figure 3. Ames localities in the study area. Modified from Merrill (1986).

LITHOFACIES AND MEGAFaUNA

Green Chonetid Shale

Rocks of this facies are made up of soft, greenish gray to bluish gray, mostly calcareous shale. The most distinctive feature of this litho- and biofacies is the very large number of chonetid brachiopods (*Neochonetes granulifer*). Actual counts have disclosed more than 1000 individuals per litre of original sample.

Within this lithosome are numerous thin layers and nodules of impure limestone (Figure 4G). The latter are biomicrites and biomicrosparticles with faunas that are commonly more diverse than those of the surrounding

shales. All of these limestones are sideritic and the amount of siderite generally increases upward in the section and southward along the outcrop. In some of the more southerly localities the limestones are up to 30 cm thick and are 4-5 m apart with the intervening shale having a reduced fauna. These are similar to the "lower" and "upper" Ames described by Hennen (1912) in West Virginia.

Although dominated by the chonetids, the green chonetid shale contains the most diverse biota of any of the Ames lithosomes in this study. Second in abundance to the chonetids is the shell-cementing brachiopod *Derbya*, followed by stilt-spine supported productid bra-

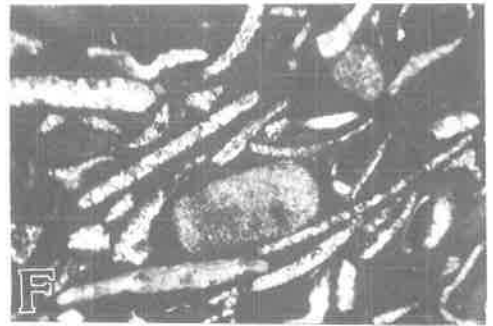
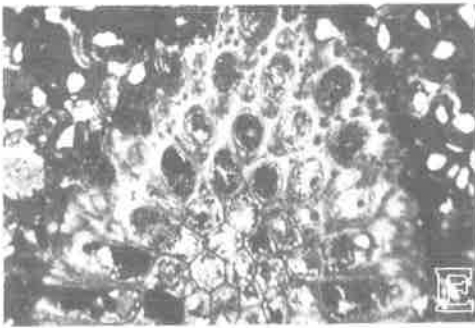
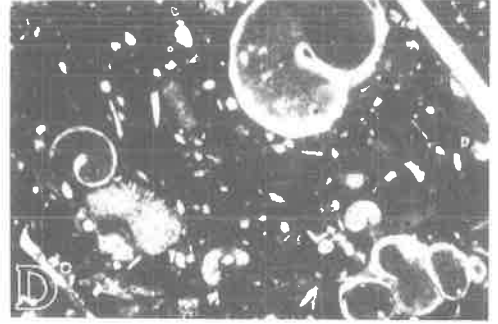
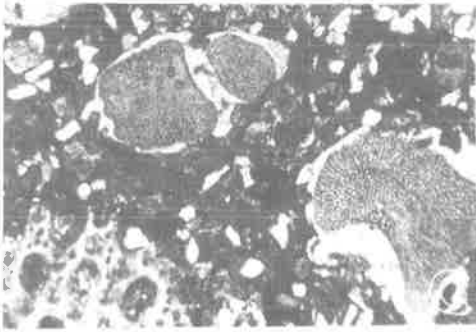
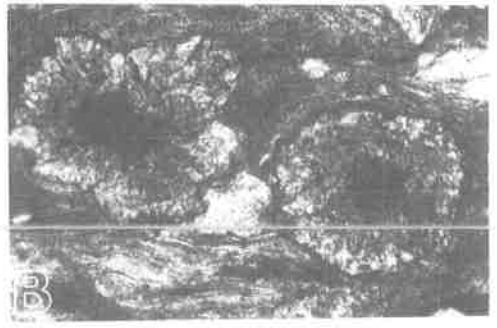
chiopods such as "*Dictyoclostus*" and spine-attaching (to crinoids?) brachiopods such as *Linoproductus*. Other benthic elements include crinoids, calcareous (pseudophthalmid) foraminifers, ostracods, and one thin section showed a trilobite. Locally abundant is the bysally-attached mussel *Orthomyalina* and at one locality (98) these formed a mussel shoal more than a metre high that was traceable for 10-15 m along the outcrop. Snails, especially *Glabrocingulum* and several kinds of bellerophonaceans such as *Pharkidonotus* and *Euphemites*, are commonly present. Unfortunately *Straparollus*, used by Saltsman (1986) for the interpretation of epigenetic alteration of community structure, is rare to absent. At locality 150 the Ames biota consists of clam and snail steinkerns and little else. The most spectacular fossils from the green chonetid shale are the spines of the giant shark *Ctenacanthus*. Reconstructed spine lengths well in excess of 25 cm indicate total body lengths of these sharks in the 5-6 m range or greater (Hansen, personal communication, 1990).

Although the faunas of the limestones are generally similar to those of the surrounding shales there are some exceptions. The dominant fossil in the majority of the limestones is *Neochonetes granulifer*. In some limestones, however, other faunal elements dominate or at least share dominance with the chonetids. Several of the limestones are rich in fossil mollusks (Figures 4D,F,H) although the surrounding shales contain few if any. These observations suggest support for Saltsman's (1986) model for an originally chonetid-mollusk biofacies, the aragonitic mollusks having been leached from the muds before diagenesis converted them into shales, while early cementation of some of the limestones preserved the mollusks and permitted their aragonite to invert to calcite. Under this hypothesis, it is likely that leached molluscan shells provided much of the carbonate that led to the early cementation in these limestones and the differences between shale and limestone portions in this lithosome could reflect timing and locations of carbonate removal and/or introduction rather than major

sedimentologic differences. Saltsman (1986, p.224,230) used the dual mineralogy of the shell of the gastropod *Straparollus* to determine whether or not aragonite had been leached from the surrounding sediment and whether a shale that today is dominated by chonetids represents an original *Neochonetes*-biofacies or a *Neochonetes*-mollusk-biofacies with the mollusks destroyed. Mollusk-rich limestones and the unusual preservation of abundant mollusk steinkerns at locality 150 make it highly likely that many of the green chonetid shales of the Huntington area were originally faunal mixtures with abundant mollusks.

Olive Silty Shales

Although the lithosomes with this description (Figure 2) probably constitute the greatest volume of Ames marine rocks at most localities, they lack distinctive lithic characteristics. In general they are an upward extension of the green chonetid shale with colors varying toward bluish-gray that weathers olive, buff and tan as the reduced iron minerals are oxidized. They also characteristically show a reduced biota, although still mostly dominated by chonetids, and an increase in particle size from clay-shale to mud-shale to silt-shale (Ingram, 1954). *Planolites*-like burrows are occasionally present, usually filled with siderite. Shales of this lithology and biota occur not only above the green chonetid shale, but are also found at other positions within the Ames succession. They have been observed above the red platy shale, between the red platy shale and the calcareous crinoidal sandstone, and above the calcareous crinoidal sandstone (Figure 2). Indeed, after the initial deposition of the green chonetid shale, subsequent deposition that did not result in highly distinctive lithosomes like those described below would probably have resulted in such nondescript silty shales. The characteristic development and greatest thickness of these rocks normally is directly above the green chonetid shale. In many respects this litho- and biofacies is exactly what is generally encountered as the



terminating coarsening-upward transition from fully marine Appalachian Pennsylvanian units into the overlying rocks lacking conspicuous marine fossils. These coarsening-upward deposits apparently represent infilling from progradational regression. In the Huntington area, however, this coarsening-upward succession does not represent the termination of the Ames marine event, because these silty shales are succeeded by rocks of more clearly marine aspect.

Red Platy Shale

Another distinctive rock type within the Ames Member in the Huntington region is a fissile, maroon clay-shale to mud-shale, herein designated the "red platy shale." The stratigraphic positions of these rocks are sufficiently different in different parts of the region that it can be reasonably assumed they represent a recurrent rock type rather than a single lithosome. In the northern part of the study area along the Ohio River, however, a single body of this rock type can be traced for several kilometres. In this area it averages somewhat over

half a metre thick and occurs about five metres above the green chonetid shale and is separated from it by the previously described olive, silty shales. Less extensive bodies of similar lithology are found nearly in contact with the green chonetid shale farther south along the Big Sandy River. Although the red platy shale is a conspicuous rock type on fresh exposures, the color quickly fades upon weathering and the shales become the same olive color as the associated mudrocks. This makes positive identification difficult and it is possible that some of the presently olive-colored shales were originally maroon. Like the associated shales, these red platy shales are silty and sparsely fossiliferous and yellow-weathering siderite nodules are especially common. Although the distribution of rocks with this lithology appears patchy, the transient nature of its coloration has probably led to it remaining unidentified in some places and it is probably more widespread than is currently realized.

Fossils are not particularly abundant in the red platy shale, but are probably more common than in the underlying silty shales. Chonetid brachiopods are occasionally present, but the

Figure 4. Oriented thin sections of Ames rocks from the Huntington area. A - Packed arenaceous crinoidal biomicrite/calcareous sandstone, sample 92ALSA, 26X, crossed polars. A single large crinoid ossicle occupies the upper half of the field and a smaller one below it has an attached foram. The matrix contains fine angular quartz sand and a little clay. Cements include spar, micrite and siderite. B - Argillaceous arenaceous siderite, 96ALSC, 130X, plane light. The matrix is clayey with cement that is mostly fine siderite with some calcite. Much angular quartz, some of it coarse, is present. The "sunflowers" have deep red, iron-rich cores of siderite pellets/crystals, surrounded by euhedral petals of highly birefringent carbonate (calcite?) with decreasing amounts of stain distally. The crystals grew in soft sediment and pushed it aside during growth. C - Packed arenaceous crinoid-bryozoan biomicrite/calcareous sandstone, sample 101ALS, 26X, crossed polars. Fine angular quartz sand and clay in the matrix. Micrite and siderite cements. Dominant allochems are crinoids (some with syntaxial overgrowths) and bryozoans (lower left hand corner and

E below). D - Sparse mollusk biomicrite, sample 106ALSC, 26X, plane light. Although snails dominate this particular field, bryozoans, crinoids, and brachiopods are also present. E - Same thin section as C above, 101ALS, 26X, crossed polars. Detail of bryozoan, note finely preserved delicate structure. F - Packed mollusk fragment biomicrite, sample 105ALSC, 26X, crossed polars. This is a dense micritic matrix with apparent bivalve fragments, but some may be codiacean algae. G - Sparse to packed brachiopod biomicrosparite, sample 111ALS, 26X, crossed polars. The matrix is patchy siderite-rich and siderite-poor microspar with pyrite (opaques in field shown). Fairly quiet water conditions are indicated by the tented geopetal beneath the concave (brachial) valve of this chonetid brachiopod. H - Same thin section as F, 26X, crossed polars. In addition to mollusks or algae (?) as listed above, some crinoids are present. Figures D,F,G and H are from limestones within the green chonetid shale, A,C, and E are from the calcareous crinoidal sandstone and B is from the "key bed" siderite-cemented sandstone.

dominant fossils both in terms of number of individuals and numbers of places where they occur are linguloid brachiopods, pectinoid bivalves, and ostracods, many belonging to the genus *Cavellina*. Martino (personal communication, 1990) found a small nautiloid (*Pseudorthoceras*) in this facies. Trace fossils consist of pale green, essentially horizontal sinuous burrows that are conspicuous in contrast to the maroon background on which they occur.

Calcareous Crinoidal Sandstone

One of the more distinctive marine lithosomes is a lenticular body of sandstone occurring in the upper part of the interval. At most places it is a calcareous sandstone, although locally the carbonate, largely in the form of crinoid columnals (Figures 4A, C), may exceed 50%. It is also rich in siderite that causes its color, light to medium bluish gray when fresh, to weather to brown, red, or deep orange. Weathering yields a pattern that is rounded, honeycombed, and spongy because of the leaching of the calcite-rich, generally crinoid-rich, portions. The crinoid columnals commonly form low-amplitude (<10 cm), northward-dipping cross-beds and placers. In addition to the crinoids, several kinds of bryozoans, most of them ramose, are next in abundance (Figure 4E). Other bryozoans such as fenestrates are also present as are fairly common brachiopods (*Neospirifer*, *Neochonetes*, various productids) as well as a few mollusks. In addition to these body fossils, this is a heavily trace fossil-dominated community. The sandstone is intensively and pervasively burrowed from bottom to top, in many places to the degree that bedding and other abiotic sedimentary structures have been obliterated. The base of the unit commonly carries large (1-2 cm diameter) burrows of *Planolites*-type in hyporelief into the underlying shale. The high degree of cross-cutting of these burrows makes it impossible to determine whether other ichnotaxonomic assignments could be made if they were not repeatedly cut into such short segments. The bulk of the unit is bioturbated pri-

marily and most recently by relatively small (mostly less than 5 mm) burrows that are dominantly vertical (*Skolithus*, *Arenicolites*, *Rossetia*), although these near-vertical burrows also cross-cut one another repeatedly, obliterating parts of older vertical, and perhaps other, burrows. In the area where the sandstone body reaches its maximum thickness (locs. 111, 112) the very top of the sandstone preserves what were clearly maintained burrows that appear to have formed networks and had vertical shaft entrances. These burrows average about a centimetre in diameter and branch, but exposures were not good enough to determine whether or not they form polygonal networks. No swellings could be seen at their junctions. These are tentatively assigned to *Thalassinoides*, but whatever ichnogeneric assignment is given them, they appear to have been formed by crustaceans, probably in the swash zone. The halos outlining some of them, and the common infilling with greenish-stained material from above makes these stand out from the surrounding rock. Several identifiable ichnotaxa are present as hypo- and endoreliefs (Table 1).

This unit is not present in the northeastern part of the study area, appears as a knife-edge at about 82°30'W (loc. 91). It then thickens westward and southward to a thickness maximum of more than five metres along the axis of the Pittsburgh-Parkersburg-Huntington Synclinorium at locality 101. Southward it thins, splits, and both benches disappear within two kilometres of the area of maximum thickness. Thus it is a northeast-southwest-trending, wedge-shaped rock body more than six kilometres wide and at least 12 km long that thins gradually northward and abruptly southward from its thickness maximum. Rocks of this type are not common in the Ames north of the Huntington area, but the sandstone reported by Condit (1912) in southernmost Ohio may be of this type.

Sideritic Sandstone (Key Bed)

A short distance (less than a metre) above the calcareous crinoidal sandstone and sepa-

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Table 1. - Taxonomic Abundance in Major Ames Lithosomes

	Green Chonetid Shale	Red Platy Shale	Calcareous Crinoidal Sandstone
Monera			
unidentified cyanophytes	p	-	?
Protista			
calcitornellids	a	-	c
incrusting agglutinates	c	-	c
Animalia			
Ectoprocta			
<i>Rhombopora</i> (?) sp.	r	-	aa
<i>Septopora</i> (?) sp.	-	-	c
" <i>Fenestella</i> " sp.	r	-	a
Brachiopoda			
<i>Lingula</i> sp. cf. <i>L. carbonaria</i> Shumard, 1858	p	a	-
<i>Derbya</i> sp. cf. <i>D. crassa</i> (Meek and Hayden, 1858)	aa	-	-
<i>Neochonetes granulifer</i> (Owen, 1852)	aaa	c	c
" <i>Dictyoclostus</i> " sp. (= <i>Antiquatonia</i> sp.)	c	-	?
<i>Linoproductus</i> sp.	r-c	-	r
<i>Neospirifer</i> cf. <i>cameratus</i> (Morton, 1836)	p	-	c
<i>Crurithyris</i> cf. <i>planoconvexa</i> (Shumard, 1855)	p	-	r
Mollusca			
<i>Orthomyalina</i> sp. cf. <i>subquadrata</i> (Shumard, 1858)	a	-	-
<i>Aviculopecten</i> sp.	-	a	-
miscellaneous bivalves	c	-	-
<i>Pharkidonotus</i> sp.	c	-	-
<i>Euphemites</i> sp.	c	-	-
<i>Cymatospira</i> sp.	c	-	-
<i>Amphiscapha</i> sp.	r	-	-
<i>Glabrocingulum</i> sp. cf. <i>grayvillense</i> (Norwood and Pratten, 1855)	c	-	-
miscellaneous snails	c	-	r

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Table 1. - Taxonomic Abundance in Major Ames Lithosomes

	Green Chonetid Shale	Red Platy Shale	Calcareous Crinoidal Sandstone
<i>Pseudorthoceras</i> sp. cf. <i>knoxense</i> (McChesney, 1860)	c	r	-
Arthropoda			
<i>Ditymopyge</i> sp. cf. <i>decurtata</i> (Gheyselinck, 1937)	r	-	-
<i>Trypetesa</i> sp. cf. <i>caveata</i> Tomlinson, 1963	c	-	-
<i>Cavellina</i> sp.	a	c	?
other ostracods	c	a	p
Echinodermata			
<i>Delocrinus</i> sp.	?	-	p
crinoid fragments	a	-	aaa
Conodonta			
<i>Cavusgnathus lautus</i> Gunnell, 1933	a	c	a
<i>C. merrilli</i> von Bitter, 1972	r	-	-
<i>Aethotaxis</i> sp. cf. <i>advena</i> Baesemann, 1973	r	-	-
<i>Hindeodus ellisoni</i> (Merrill, 1973)	c	-	-
<i>Streptognathodus elegantulus</i> Stauffer and Plummer, 1932	c	r	c
<i>Ellisonia conflexa</i> (Gunnell, 1933)	c	r	c
Chordata			
<i>Ctenacanthus</i> sp.	r	-	-
miscellaneous vertebrate remains	c	p	r
Ichnotaxa			
<i>Arenicolites</i> ichnosp.	-	-	aa
<i>Planolites</i> inchoosp.	c	-	aa
<i>Rosselia</i> inchoosp.	-	-	aa
<i>Skolithis</i> inchoosp.	-	-	aa
<i>Thalassinoides</i> (?) inchoosp.	-	-	c
miscellaneous small traces	c	c	c

aaa= extremely abundant, aa= very abundant, a= abundant, c= common, r= rare, p= present, ?= uncertain

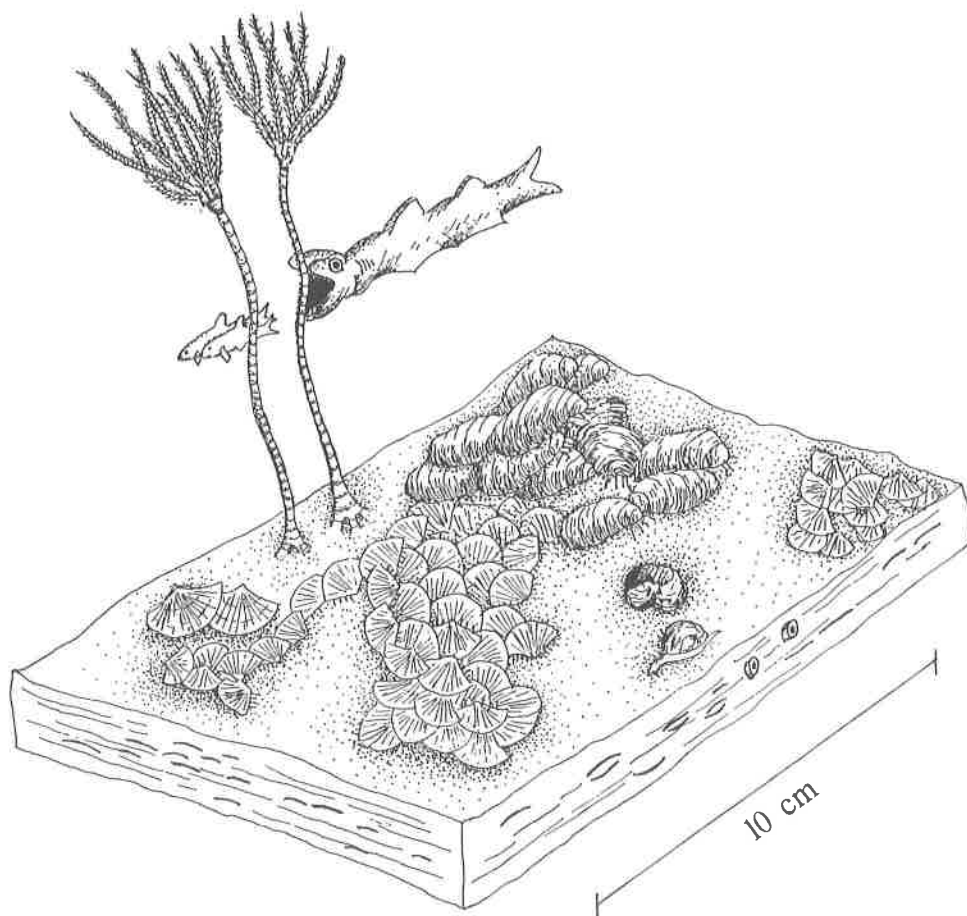


Figure 5. The inferred lagoonal paleocommunity represented in the green chonetid shale. Dense clumps of *Neochonetes* dominate the lagoon floor with larger shells of *Derbya* at the lower left. In the upper left is a mussel shoal of *Orthomyalina*. An empty *Glabrocingulum* shell is passed as a live *Euphemites* crawls by it. A pair of short-stemmed crinoids is along the left margin of the block and behind them, in the far distance, a large ctenacanthid shark chases a pair of palaeoniscids.

rated from it by olive or gray silty shale, is a distinctive siderite-cemented sandstone (Figure 4B; Merrill, 1986, Figure 6D). It is dark bluish-green when fresh and weathers blood-red with rounded shapes. It has not been found in the Ohio River Valley, but it is traceable for more than five kilometres along the Big Sandy River. At most localities it is less than 15 cm thick and contains no marine fossils. At locality 144, however, it is in contact with the top of the calcareous crinoidal sandstone and there, like the lower unit, contains crinoid columnals. In all probability it represents the last phase of the

Ames marine basin.

A MODEL FOR NEARSHORE AMES DEPOSITION

In contrast with Ames rocks described elsewhere that include nearshore as well as more offshore deposits (Brezinski, 1983; Saltzman, 1986), those in the Huntington area are entirely of nearshore origin. In the transgressive-regressive couplets described by Saltzman, nearshore environments are clearly represented near the base and top of the Ames

interval, but more offshore environments are represented in the stillstand portion of the couplet and shoreline was remote during their deposition.

In the Huntington area, however, there is no evidence to support the interpretation of any "offshore" deposits. Absence of marine fossils at the Ames stratigraphic position at Louisa indicates that shoreline was between the southernmost locality in the study area (loc. 125) and Louisa. Within this 12 km interval its exact position is unknown, but the rate of thinning of the member and the coincidence of this thinning with the steep northward dip at locality 125, which appears to be partially contemporaneous with sedimentation, suggests a position only slightly south of locality 125.

Additional evidence for an extremely nearshore setting for all the Ames rocks in the study include the greatest thickness of the coarsest terrigenous rocks within the Ames Member known anywhere within the basin, a high siderite content throughout that suggests brackish conditions (Curtis, 1967), a fauna, both mega- and conodont and ostracod microfauna, which suggests dramatically reduced salinities.

Green Chonetid Shale and "Olive" Silty Shale Environment (Figure 5)

In keeping with Saltsman's (1986) environmental interpretation of rocks of similar lithology with similar biota and in a comparable stratigraphic setting, these shales almost certainly represent quiet water, basal transgressive deposits as seawater invaded the preexisting coal swamp now represented as a thin coal and underclay. The absence of any significant current velocity, indicated by the very fine grain size, could result simply from the low paleoslope and great distance travelled by the transgression (the distance from the most offshore Ames deposits seen in northern Ohio to locality 125 is in excess of 250 km). On the other hand, the sand bodies described by Condit (1912) north of Huntington could have served as barriers, albeit of low amplitude. On

these lines of evidence, the green chonetid shales are tentatively assigned a lagoonal origin.

A quite high biologic productivity is evident in this lithosome. The high density of megafossils suggests this, but could be an artifact of lowered ("condensed") rates of sedimentation. The conodont frequencies (see below) attest to rapid deposition, however, and this community seems to represent high-density, low-diversity populations of opportunistic pioneers invading former coal swamps with high nutrient levels (Rollins, and others, 1979). Chonetid brachiopods seem to have been especially well adapted to migrating with the initial transgression (Dennis and Lawrence, 1979).

The silty, less fossiliferous, olive shales overlying the green chonetid shales probably were also deposited in a lagoonal setting, but the coarser grain size indicates increasing terrigenous influx, perhaps with some freshening of the water. Both factors would further dilute a declining benthic population.

Red Platy Shale Environment (Figure 6)

Still situated within an overall lagoonal setting, these shales are clearly oxidized in comparison with the blue, green, and gray colors of the associated rocks. Although red coloration in mudrocks may result from the introduction of red clayey detritus or diagenetic alteration following burial, the patchy distribution of these shales, both vertically and laterally, suggest local conditions of oxidation within the general Ames milieu. In the absence of clear criteria of diagenetic change, oxidation in this case is interpreted to be a product of local, probably periodic exposure in an area that would support *Lingula*, ostracods, and scallops (the last probably pseudoplanktonic, Clark, 1978), but would exclude rooted plants and other aspects of soil formation found in massive, red Conemaugh claystones ("red beds") on one hand and also exclude characteristically marine benthos on the other. The presence of an occasional chonetid brachiopod

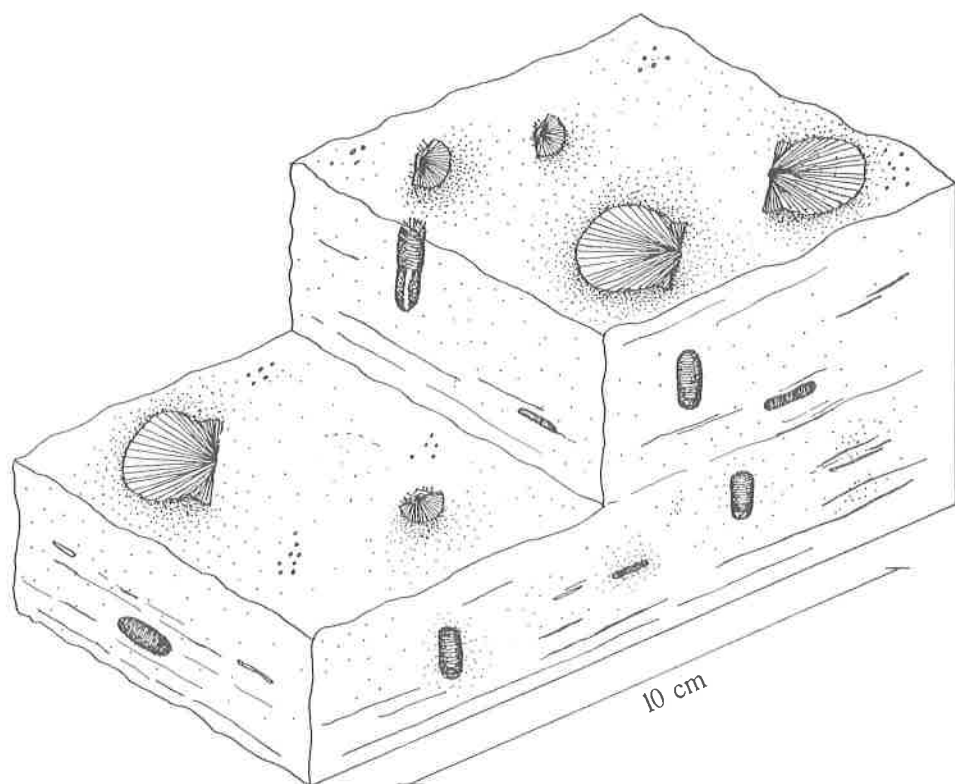


Figure 6. Tidal flat paleocommunity represented in the red platy shale. The most widespread shells are pectenoid bivalves, although they tend to be concentrated on particular bedding planes. *Lingula* occurs throughout the section, some retaining their life position. Occasional chonetid brachiopods are scattered as are individual and small clusters of ostracod carapaces.

suggests proximity to the more normally marine, shallow water Ames deposits. The stratigraphic settings of these lithosomes, especially those almost directly below a sand body of proposed barrier origin may indicate an intertidal area on the lee side of the barrier.

Calcareous Crinoidal Sandstone Environment (Figure 7)

The relatively coarse grain size in this rock body suggests the highest energy of any of the Ames depositional environments in this area. Lateral facies relationships further suggest its contemporaneity with lagoonal deposits probably including both the green chonetid and red platy shales. The gross anatomy (6 × 12+ km and up to 5 m thick, the longest dimension more or less parallel to shoreline), cross-bed-

ding, shell placers, and biota (a crinoid meadow with bryozoans) all attest to more or less consistent, relatively high energy currents plus moderate waves. The relatively large size of the detrital particles (medium sand) in this rock is particularly significant. Embedded as this unit is in a mass of marine sediment that rarely exceeds silt size suggests that substantial hydraulic energy would be required to so completely separate the particle sizes. At any given time the relief on this sand body above the surrounding sea floor was probably not great, undoubtedly far less than its maximum thickness of 5(+) m. Nevertheless, a barrier of this magnitude in very shallow water would impede circulation and produce lagoonal and tidal flat areas in its lee.

This bar/barrier appears to have remained active while the lagoon landward of it was

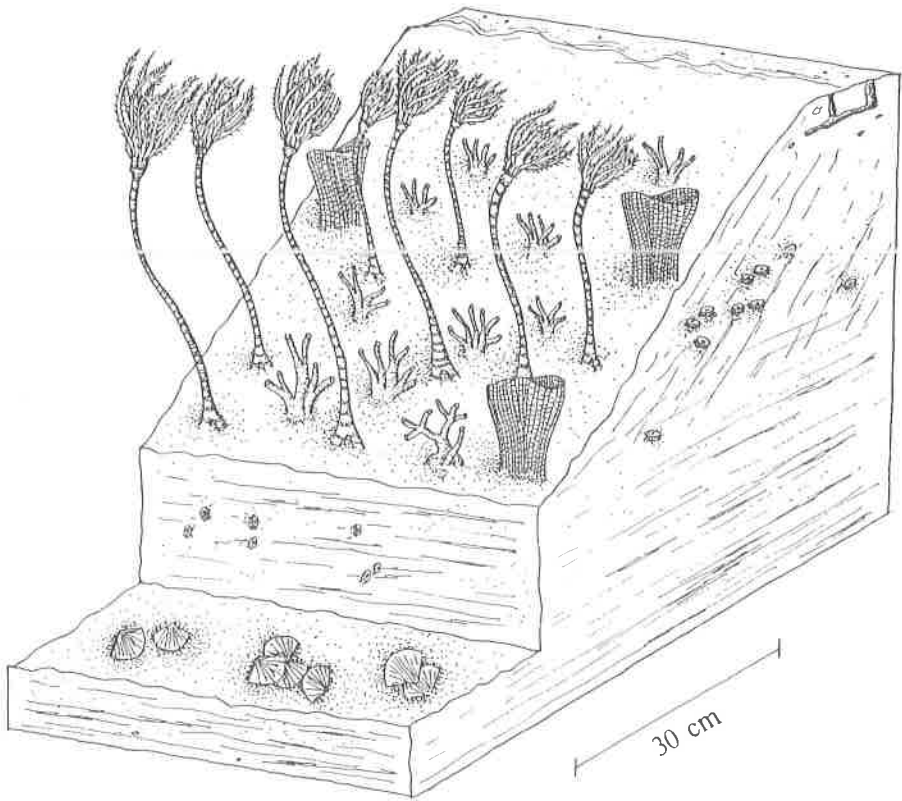


Figure 7. Barrier paleocommunity represented in the calcareous crinoidal sandstone. An exaggerated beach face slope is dominated by long-stemmed, rheophilic crinoids, interspersed among which are many ramose bryozoans and a few fenestrate ones. Crustacean burrows are abundant in the swash zone. Placers of crinoid debris occur throughout the sand body, and local layers of brachiopod shells, probably as reworked lagoonal deposits, occur near the base of the sand.

filled with silty olive mud, after which progradation built non-marine sediments across both lagoon and barrier. The benthic community represented in this sandy lithosome was heavily dominated by two elements: crinoids and ramose bryozoans. Saltsman (1986, p. 226-227) has argued that crinoids in Ames environments were dominantly offshore organisms and their concentrated remains in nearshore deposits resulted from onshore transport. Although significant onshore transport cannot be ruled out for some crinoids, the very large particle size of many ossicles places restrictions on the currents or waves necessary to put them in motion. All were probably moved around *within* the environment before burial, but it is not likely that the larger ones (5+ mm) were moved *into* the environment while other,

smaller, offshore skeletal debris was not. Moreover, the concept of indigenous, high-energy "crinoid meadows" is well established for rocks of several ages (for the Pennsylvanian see Heckel and Cocke, 1969; Merrill, 1975). The argument based on modern analogs that ancient crinoids were "stenohaline," deep water, or at least "offshore" organisms is probably spurious because of the paucity of truly comparable modern crinoids both in abundance, diversity, and habitat. Other modern echinoderms, for example starfish, can be found in quite low salinities and the diversity of nearshore deposits in which Pennsylvanian crinoids are found suggests that, as a class, the Crinoidea should not be so stereotyped in terms of their environmental tolerances. Furthermore, Meyer and Meyer (1986) in dealing with mod-

North

South

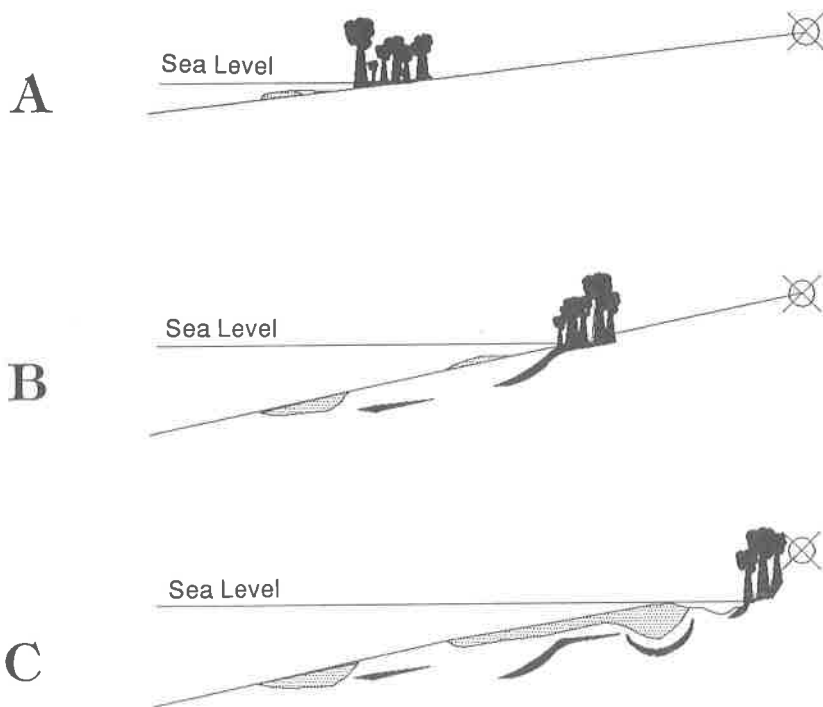


Figure 8. Developmental model for the Ames transgression in southern Ohio, northern Kentucky and north-western West Virginia. North of this area and prior to A, the land surface of very low relief was rapidly overrun by marine water with insufficient time for the establishment of significant coal swamps or nearshore brackish marine conditions and the Ames deposits appear to be fully marine. The marine deposits contain little coarse terrigenous matter and are mostly carbonates formed in low energy, normal marine waters. In A, the rate of transgression slows as the hinge line of the downwarping located near Louisa, Kentucky, is approached. This slowdown leads to the deposition of wave-reworked sands (barrier) and the formation of more brackish conditions and peat accumulations in southern Ohio. In B continued slowing of the downwarping and transgression increased brackishness with more continuous sand bodies and coals. In C deformation of the basin margin localizes the barrier sands along the syncline axis and limits the extent of marine transgression. A stillstand coal swamp developed adjacent to the hinge.

ern rheophilic crinoids (current velocities of 50 cm/s and greater), found no significant post-mortem transport of these ossicles and cast serious doubt on the role of "buoyant transport" as an explanation for concentrations of crinoid debris in ancient limestones.

All of the ichnotaxa are compatible with a shallow, nearshore origin, the heavy domination through the bulk of the unit by vertical burrows suggests adaptation to a rapidly shifting, mobile substrate in a relatively high-energy setting. These trace fossils are entirely

compatible with a *Skolithus*-ichnofacies assignment and interpretation of the environment as a nearshore bar/barrier complex. This is consistent with interpretations by Chamberlain (1978, p. 182,183) and Ekdale and others (1984, p. 192).

Sideritic Sandstone ("Key Bed") Environment

Occurring as this lithosome does as the uppermost part, if not above the top, of the

truly marine Ames interval, it is clear that the origin of this rock body is closely related to the termination of the Ames marine episode. The concentration of siderite as particles and cement suggests a salinity range from brackish to essentially fresh water. Nevertheless, there was enough energy in the environment to introduce and selectively sort fine to medium sand. Although a tidal flat origin for this lithosome is suggested by its stratigraphic position with no definitely marine rocks occurring higher, the general lack of fauna and its distinctive mineralogy are factors difficult to reconcile with this origin. The lack of any sedimentary structures, biotic or abiotic as well as widespread distribution argue against formation in some small tidal pool and the grain size denotes at least moderate transport energy within the environment. In one manner or another, however, the rock unit must be related to the prograding Ames shoreline.

The mechanisms for the origin and distribution of these major environments and their deposits in time and space are shown in Figure 8.

EVIDENCE FROM AMES CONODONT BIOFACIES

Nearly all the rocks that bear marine megafossils in the Huntington region also bear conodonts (Figure 9). In all 179 samples (151 productive) from 40 localities yielded 22 113 conodonts. Many of these samples were difficult to disaggregate and the residues were even more difficult to concentrate. In addition, the conodont yields were commonly extremely low, often fewer than 10/kg. The frequencies generally parallel the relative abundances of megafossils. Nevertheless, in spite of extraction difficulties, low yields, and poor preservation, the conodonts provide valuable paleoecologic information.

All sampled rocks belong to the *Cavusgnathus*-biofacies as defined by Merrill (1970) and von Bitter (1972) in that more than 10% of the platform elements belong to that genus as opposed to the generally more widespread

Streptognathodus. In fact the highest percentage that *Streptognathodus* attains in the Huntington area is less than 50% of the platform total and *Cavusgnathus* is the dominant genus throughout the region. *Ellisonia*, a *Cavusgnathus*-biofacies associate, is third in abundance and more specimens are known from the Ames in the Huntington area than from any other unit at any other place in the Carboniferous. The number of specimens per sample, however, is low. *Hindeodus* is rare, *Aethotaxis* rarer, and *Diplognathodus* and *Idioprioniodus*, known from Ames localities farther north, have not been found near Huntington. Like the megafauna, the conodont microfauna is one of low diversity.

Environmental controls are manifest by the distribution of conodonts in two obvious ways. The first is taxonomic at the generic level, mostly in differing ratios between *Cavusgnathus* and *Streptognathodus*, but also in the exclusion of genera such as *Diplognathodus* and *Idioprioniodus*. A lesser taxonomic effect is within *Cavusgnathus*. Forms called *C. lautus* dominate throughout the region, but in those samples with a higher percentage of specimens of *Streptognathodus* there is a concomitant increase in the proportion of *cavusgnathids* referred to *C. merrilli*. This same effect was noted by von Bitter (1972) who named *C. merrilli* from such occurrences in the Virgilian of Kansas. It is not certain whether *C. merrilli* is genetically distinct from *C. lautus* or merely an ecophenotypic variant of the more common species. Whether taxonomically or ecophenotypically different, the responses in the populations of *cavusgnathids* to salinity are reasonably clear. *Cavusgnathus merrilli* is absent from most samples. The total distribution of *Streptognathodus* and *Cavusgnathus* is summarized in Figure 9.

Other evidence of environmental stress produced by lowered salinities is found in the reduction of proportions of ramiform elements from the apparatuses of the common genera. As pointed out by Merrill and von Bitter (1984) it was common for *Streptognathodus* to have lost ramiform elements while the genus

LATE CARBONIFEROUS PALEOECOLOGY - AMES MEMBER

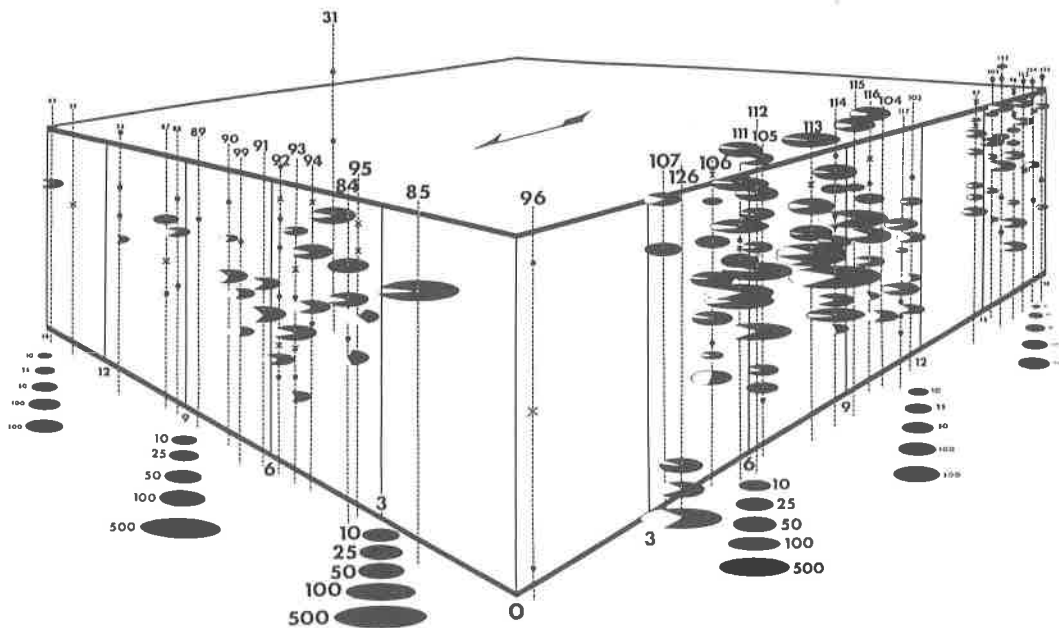


Figure 9. Conodont distribution diagram. Perspective drawing looking southeast with the near corner of the block representing the confluence of the Ohio and Big Sandy Rivers. Bold numbers beneath solid vertical lines are distances in kilometres away from this confluence along each river. Numbers at the tops of the dashed vertical lines are the locality numbers and the ellipses represent pie diagrams seen in perspective with their centers at the approximate stratigraphic position of the sample within in section. Vertical distance between the heavy bounding lines at the top and bottom of the block is approximately 15 ft (5 m). Diameters of circles (now major axes of ellipses) represent conodont frequencies per kilogram of original sample as shown beneath each 3 km mark. Samples containing fewer than 10 platform elements are shown by solid circles and entirely barren samplers by the x symbol. Within each pie diagram, the proportion of *Streptognathodus* is shown in white and of *Cavusgnathus* in black.

remained common and for *Cavusgnathus* to begin losing its ramiforms as *Streptognathodus* dwindled in numbers. Thus in the most stressed brackish environments, in this area in the red platy shale, the final result are rocks containing only *Cavusgnathus lautus* platforms and perhaps a few specimens of *Ellisonia*.

Salinity reductions in the Ames Member near Huntington probably arose from not fewer than three separate, but not totally unrelated, causes. Two of these, transgressive attenuation and progradational dilution are commonplace in other Appalachian Pennsylvanian marine units, but the third, although probably not unique to the Ames, is considerably less common in other units and areas. The third is barriered impedance (Merrill, 1982). In contrast to tropical arid regions where nearshore hypersa-

linity favors evaporite deposition, in humid tropical regions, waters of a marine transgression become more dilute as they advance because of increasing distance from the open sea and substantial freshwater runoff from the land that is being transgressed. This effect is particularly well shown by the Ames Member basinwide as the highest percentages in *Streptognathodus* decline from locality to locality southward toward the paleoshoreline (Merrill, 1970). The second way in which salinities become diluted is called "progradational dilution" in which the substantial load of sediments that extends the shoreline seaward and fills the basin carries with it an influx of fresh water that dilutes the marine waters. This is particularly characteristic of widespread marine units in the Appalachian basin where coarsening-

upward regressive successions are accompanied by reduction in marine fauna, both in kinds and numbers. This is also true among the conodonts where the reduction in numbers is commonly accompanied by an increase in *Cavusgnathus*. Indeed, upward increase in cavusgnathids is the most common nature of the replacement of the *Streptognathodus*-biofacies by the *Cavusgnathus*-biofacies in the Appalachians (Merrill, 1973b, Figure 5). Most characteristically, the shales in the upper part of a Conemaugh marine unit have increases in grain size, reduction in conodont frequency, and increase in the proportion of cavusgnathids, and the Ames near Huntington is no exception. The third method by which salinities may be reduced is termed "barriered impedance" because a physical barrier plays an important role in restricting circulation and altering salinity in its lee. In the Huntington area the calcareous crinoidal sandstone apparently formed an effective barrier to circulation and deposits south of it show a higher proportion of *Cavusgnathus* than stratigraphically equivalent rocks north of it.

SUMMARY AND CONCLUSIONS

The transitional deposits from normal Ames marine strata in northern Ohio to a non-marine depositional setting southward are well shown in the valleys of the Ohio and Big Sandy Rivers near Huntington, West Virginia. Fossil-bearing rocks and their lateral non-marine equivalents range in thickness from less than three metres to more than nine metres and consist generally of a lower green chonetid-dominated shale that contains an abundant and relatively diverse fauna. Although its conodont fauna is dominated by *Cavusgnathus*, it contains the highest proportion of streptognathodontids among the rocks in this study, although the proportion dwindles southward. These inferred lagoonal deposits grade southward and upward into olive gray shales averaging about three metres in thickness with a diminished fauna. Limestones like those farther north that represent offshore, stillstand

conditions, are absent in the Ames in this area, but a thick crinoidal barrier sandstone, ranging in thickness from zero to more than five metres in thickness, may be equivalent. Red shales, some over a metre thick, are common in the upper part of the marine interval a short distance below the crinoidal calcareous sandstone, and less common in other positions. They contain a much reduced biota dominated by *Lingula* and scallops and a very heavily *Cavusgnathus*-dominated conodont micro-fauna. These are interpreted to represent intermittently exposed intertidal deposits. A siderite-cemented sandstone caps the interval over a considerable area. Although generally unfossiliferous, it probably marks the final transition from marine to nonmarine conditions marking the beginning of the end of the last major marine event in the Paleozoic of eastern North America, the Ames transgressive-regressive episode. Structural controls on Ames deposition near Huntington and possibly elsewhere have been described by Merrill (1988). Thickening, thinning, and total pinchout of lithosomes such as the calcareous crinoidal sandstone that correspond with gentle, and some not so gentle, local dips suggest that the extent of transgression, geometry, and distribution of marine lithosomes were structurally controlled.

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LITHOFACIES RELATIONS AND CONTROLS ON DEPOSITION OF FLUVIAL-DELTAIC ROCKS OF THE UPPER POCAHONTAS FORMATION IN SOUTHERN WEST VIRGINIA

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ABSTRACT

Previous work has suggested a deltaic environment of deposition for the Lower Pennsylvanian Pocahontas Formation based on regional stratigraphic syntheses. This paper addresses the sedimentology of individual lithofacies of the upper part of the Pocahontas Formation in part of the central Appalachian basin. Rocks of the upper part of the Pocahontas Formation have been divided into two general facies assemblages: a channel-belt sandstone and a coal-bearing overbank-floodplain complex. Facies assemblages comprise six objectively defined lithofacies: A) fluvial channel sandstones, B) crevasse-splay sandstones and siltstones, C) backswamp-floodplain mudstones and siltstones, D) levee-overbank and distal crevasse-splay deposits consisting of interlaminated sandstone, siltstone, and mudstone; E) rooted mudstone paleosols, and F) coal and carbonaceous black mudstone deposited in peat swamps. The results of Markov chain analysis emphasize four preferred lithic transitions: 1) fining-upward transitions from channel-belt sandstones to levee-overbank deposits to backswamp-floodplain deposits; 2) a coarsening-upward transition, reflecting progradation of crevasse-splay deposits over backswamp-floodplain deposits; 3) a two-way transition reflecting interbedding of crevasse-splay deposits with backswamp-floodplain deposits; and 4) a uniform (constant grain size) transition from root-penetrated mudstone paleosols to coal. Deposition of lithofacies between the Pocahontas No. 6 and 8 coal beds is attributed to fluvial-deltaic depositional processes on a

transitional mid- to upper-delta plain that was structurally modified by subtle down-to-the-south contemporaneous growth faulting.

INTRODUCTION

The Lower Pennsylvanian Series in southern West Virginia consists of the Pocahontas Formation and overlying New River Formation, a coal-bearing sequence containing interbedded sandstone, siltstone, mudstone, and mudstone paleosol (underclay). Exposed Lower Pennsylvanian strata in the vicinity of Bluestone and Garwood, West Virginia, and Pocahontas, Virginia, constitute the foundation of a proposed Pennsylvanian System stratotype (Englund and others, 1979, 1986a), a global reference section for strata of Pennsylvanian age. Regional stratigraphic syntheses have suggested that the Pocahontas Formation was deposited as a deltaic clastic wedge that prograded northwestward into the central Appalachian foreland basin (Ferm and Cavaroc, 1969; Englund, 1974; Ferm, 1974; Miller, 1974; Englund and others, 1986b, 1986c; Ferm and Weisenfluh, 1989; Englund and Thomas, 1990). Evidence for a deltaic interpretation is primarily based on a Late Mississippian to Early Pennsylvanian transition from marine facies, to mixed coal-bearing continental and marine facies, to dominantly continental facies. In general, Late Mississippian depositional trends reflect regression from persistent marine limestone to mixed limestone and siliciclastic facies that contain marine fossils, continental plant fossils, and thin coal beds (Englund and others, 1986b, 1986c). Latest Mississippian and earliest Pennsylvanian strata reflect depo-

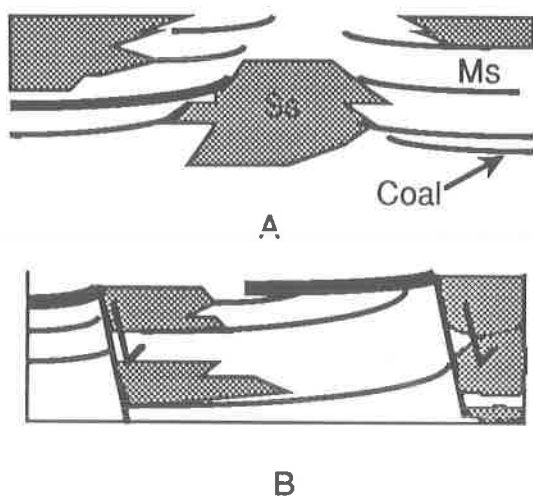


Figure 1.- Two generalized depositional models for coal-bearing strata (diagrams are only approximately scaled). A) Nontectonic-controlled model for *en echelon* offset of sandstone bodies generalized from Ferm and Cavaroc (1968). Central sandstone body is approximately 10 km x 18 m for scale. B) Fault-controlled depositional model generalized from Weisenfluh and Ferm (1984). Faults are spaced approximately 20 km apart and sandstone bodies are on the order of 10 m thick for scale.

sition of a sequence primarily composed of carbonaceous and calcareous mudstones containing fresh- and brackish-water faunas (Englund and others, 1986c). The Lower Pennsylvanian Pocahontas Formation overlies this succession and has been interpreted as a series of overlapping coal-bearing delta-lobe complexes that are generally comprised of distributary sandstone and interdistributary mudstone (Englund, 1974; Englund and others, 1986c).

Despite the development of a regional stratigraphic framework and designation of a stratotype section, little work has been done concerning the detailed sedimentology of rocks within the stratotype area. This is especially true in the uppermost strata of the Pocahontas Formation which historically have been studied in less detail because of lack of thick, economically important coals. Although regional studies (e.g., Englund, 1974; Ferm, 1974; Miller, 1974) inherently predict the types of depositional subenvironments that may be present,

rocks representing these subenvironments have never been systematically identified and described. The goal of this study is to describe and interpret individual lithofacies of the upper part of the Pocahontas Formation, develop a lithofacies sequence model describing vertical rock relations, and develop a depositional model describing rock relations through time.

General Depositional Models for Coal-Bearing Strata

Depositional models describing the distribution of coal-bearing facies on local scales are varied, but a distinction can be made between tectonic- and nontectonic-controlled models. Nontectonic models generally rely on some form of depositional autocyclicity to explain variations in the distribution of sedimentary facies. For example, Figure 1A is a model (Cavaroc, 1963; Ferm and Cavaroc, 1968) developed for upper-deltaic strata of the central Appalachian coal fields. In this model, major channel-form sandstones are stacked in an *en echelon* array. The distribution of sandstone bodies reflects fluvial channels that migrated laterally as higher and more favorable local depositional gradients became available through time in adjacent, topographically depressed backswamp areas. Similar facies relations have also been observed in recent deltaic environments (Gould, 1970).

Tectonically controlled depositional models may rely on synsedimentary structures, such as growth faults, to explain the distribution of facies. An example developed for coal-bearing strata in the Black Warrior basin is illustrated in Figure 1B (Weisenfluh, 1982; Weisenfluh and Ferm, 1984). In this model, major channel-form sandstones are terminated against a synsedimentary normal fault that modified the local depositional gradient and allowed channel sandstones to preferentially stack in a roughly *en echelon* pattern on the downthrown fault block. Although factors such as regional transgression and regression may influence nonmarine stratal geometry on a basin-wide scale (e.g., Ryer, 1984; Aubrey, 1989), the two

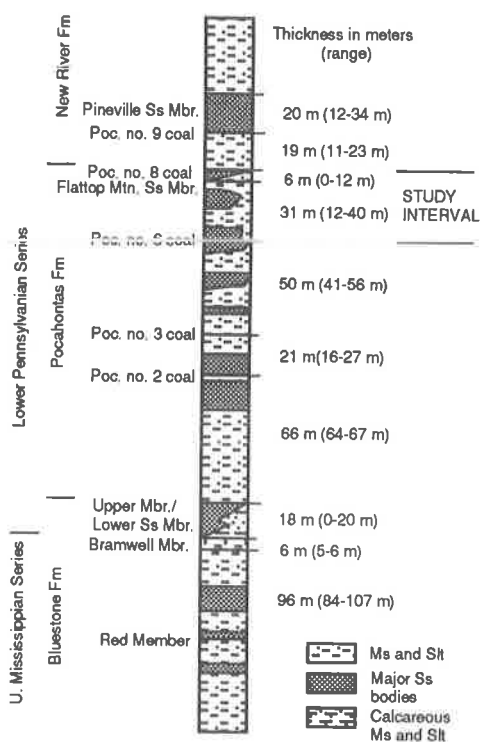


Figure 3. Idealized stratigraphic column for part of the Upper Mississippian and Lower Pennsylvanian Series in the vicinity of Crumpler quadrangle, WV. Study interval lies between the base of the Pocahontas No. 6 coal bed and the Pocahontas No. 8 coal bed at the base of the New River Formation (modified and redrawn from Stricker, 1981).

cies across the study area. The vertical relations of lithofacies in the rock succession was quantitatively determined using single-dependency embedded Markov chain analysis to statistically test observed trends in the lithofacies sequence (Miall, 1973, 1982). The technique is well proved and has been successfully employed as a test for cyclicity in several recent studies of coal-bearing strata (*e.g.*, Casshyap and others, 1987; Ghosh, 1987; Kreuser, 1991). Details of the statistical technique are outlined in a following section.

RESULTS

Facies Analysis and Correlation

The basic rock types observed in the field can be further subdivided into six objectively defined lithofacies, designated A through F (descriptions in Table 1). To simplify discussion, each lithofacies has been assigned a sub-lithofacies code, or series of codes using the scheme developed by Miall (1978; Table 1). These lithofacies are: A) a fine- to medium-grained, trough and subordinate planar cross-bedded sandstone, commonly with a deeply scoured basal contact (Se, St, Sh, Sp); B) a fine-grained, trough to low-angle cross-bedded sandstone, that exhibits a relatively broad and shallow scoured basal contact (Ss, Sl); C) laminated to massive deposits of siltstone and mudstone with vertical rootlets and plant fossils (Fsc); D) finely laminated interbedded sandstone, siltstone, and mudstone (Fl); E) vertically rooted mudstone paleosol (Fr); and F) coal and thin organic-rich black mudstones (c).

A north-south cross section of the study area illustrates general field and subsurface relations between lithofacies and the lateral continuity of the thick sandstones (Figure 4). Persistent sandstones of lithofacies A, such as the Flattop Mountain Sandstone Member, have been correlated across most of the study area. The Pocahontas No. 6 and No. 8 coals have been accurately correlated across the entire study area. The accuracy of coal correlations was confirmed by comparison with published geologic maps (Englund, 1968; Stricker, 1981) and by physical tracing of the horizons in the field. Other lithofacies are indicated on the cross section but are not correlated because of their lateral discontinuity.

Lithofacies A (Sublithofacies St, Sh, Se, Sp) - Channel-Belt Sandstone

Lithofacies A consists of a thick (>3 m), fine- to medium-grained, micaceous, litharenitic sandstone in a characteristic vertical succession of sublithofacies (Table 1; Figure 5). This lithofacies is interpreted as a fluvial channel-belt sandstone. A fluvial origin is supported by the presence of a sharp erosional base, lag deposits and rip-up clasts, continental

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Table 1.- Description of lithofacies A-F. Sublithofacies codes (after Miall, 1978) described under sedimentary structures heading where applicable. Cross-bedding types after Allen (1963).

- A LITHOLOGY:** Sandstone (Sublitharenite to litharenite); grayish orange; fine to medium grained; moderately sorted; micaceous
THICKNESS: 3-12 m
BEDDING: 4-50 cm (generally decreasing up)
SEDIMENTARY STRUCTURES: Sublithofacies St - medium-scale grouped trough cross-beds (pi); solitary trough cross-beds (theta); rare large-scale lateral accretion cross-beds (epsilon). Sublithofacies Sh - horizontal lamination intercalated with trough cross-beds. Sublithofacies Se - crude trough cross-bedding; erosional scours; basal intraclasts (mudstone, plant fossils). Sublithofacies Sp - tangential tabular-planar cross-beds, solitary (alpha) or grouped (omikron).
CONTACTS: Basal contact abrupt and scoured (concave down); upper contact gradational to abrupt; some internal erosional scours.
FOSSILS: Stem and axial plant fragments, lycopods, *Calamites*. Commonly as intraclasts of Se.
- B LITHOLOGY:** Sandstone (lithic wacke to litharenite); grayish orange; very fine to medium grained (commonly fining up); poorly to moderately sorted; micaceous; argillaceous.
THICKNESS: <3 m
BEDDING: variable, <40 cm
SEDIMENTARY STRUCTURES: Sublithofacies Ss - broad shallow scours (eta cross-stratification). Sublithofacies Sl - low-angle trough cross-beds (poorly defined).
CONTACTS: Basal contact abrupt and scoured (generally concave up, locally incised by troughs); upper contact planar and abrupt to gradational.
FOSSILS: Lycopods (rare); upper surface commonly root-penetrated.
- C LITHOLOGY:** Mudstone and siltstone; dark gray to pale brown; micaceous; carbonaceous; sideritic nodules (<1.5 cm; solitary and in discontinuous bands).
THICKNESS: <8 m
BEDDING: <1 cm (fissile to flaggy)
SEDIMENTARY STRUCTURES: Sublithofacies Fsc - Laminated (planar to wavy) to massive; laminations defined as alternating brown and black sideritic and carbonaceous layers, and light gray layers of coarser silt; local (rare) convoluted laminations (cosets <0.2 m).
CONTACTS: Basal contact generally gradational; upper contact generally abrupt.
FOSSILS: Abundant organic debris including whole leaf and stem molds; root penetration sporadically distributed throughout (most abundant in mudstone-dominated facies). Marine fossils reported from one locality in roof shale of the Pocahontas No. 8 coal (Henry and Gordon 1979); brackish and freshwater fossils reported in the roof shale of the Pocahontas No. 6 coal bed (Hennen and Gawthrop 1915, p. 220; Krebs 1916, p. 366; Price 1916; Reger 1926, p. 276).
- D LITHOLOGY:** Interlaminated sandstone, siltstone, and shale; grayish orange to medium gray; sandstone very fine to fine grained; micaceous.
THICKNESS: <1.5 m
BEDDING: <3-4 cm
SEDIMENTARY STRUCTURES: Sublithofacies Fl - fine laminations and ripples (poorly defined); locally massive.
CONTACTS: Basal contact nondistinct (gradational to abrupt); upr. contact generally gradational
- E LITHOLOGY:** Silty claystone; mottled light gray to pale orange; some very-fine quartz sand.
THICKNESS: <1 m
BEDDING: Nondistinct or "listric" internal texture
CONTACTS: Abrupt upper contact, gradational to abrupt basal contact.
FOSSILS: Vertical rootlets (*Stigmara*); dispersed leaf impressions including pteridosperm foliage (*Neuropteris* sp., *Mariopteris* sp.).
- F LITHOLOGY:** Coal and interlaminated very carbonaceous shale; black; bright/reflective.
THICKNESS: <1 - 62 cm
CONTACTS: Abrupt upper and lower contacts.
FOSSILS: Plant foliage locally present in shaley intervals within coal beds.

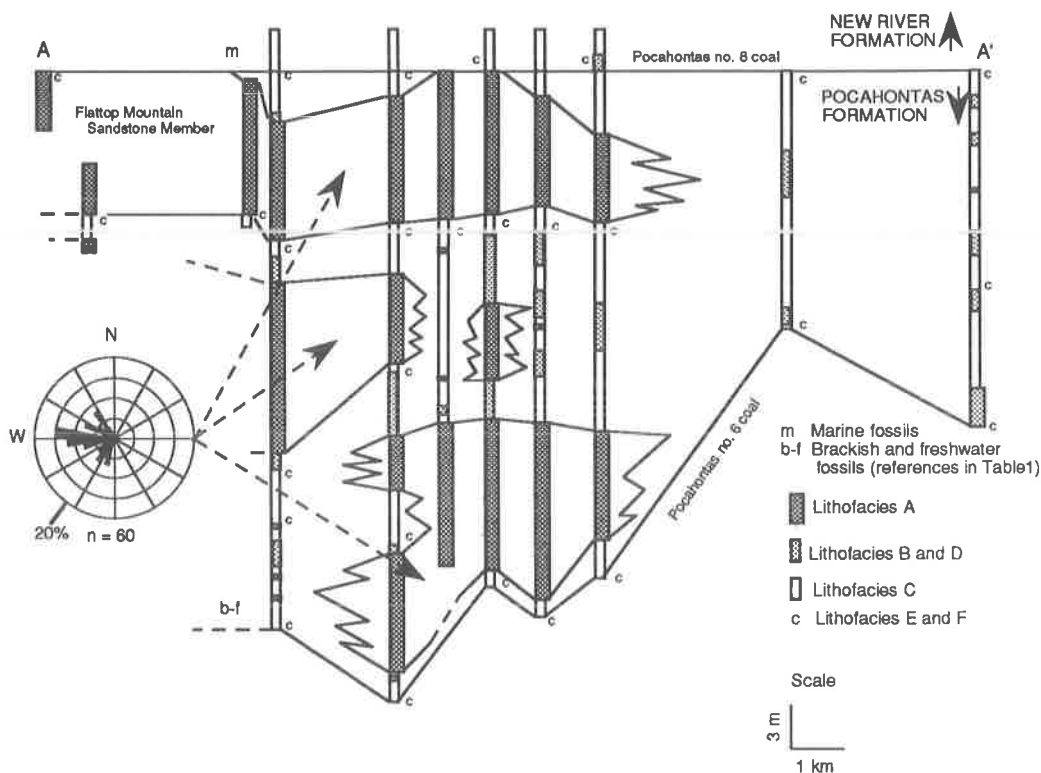


Figure 4. North-south cross-section of study area (refer to Figure 2 for location A-A'). Datum is base of Pocahontas No. 8 coal bed (top of the Pocahontas Formation). Lithofacies A = channel-belt sandstone, B = crevasse-splay sandstone and siltstone, C = backswamp/floodplain mudstone, D = levee/overbank (and distal crevasse splay) siltstone and very thin sandstone, E = mudstone paleosol, and F = coal. Rose diagram summarizes paleocurrent data from lithofacies A sandstones.

plant fossils, and cross-bedding characteristics described in this section and Table 1. Paleocurrent trends derived from trough cross-bed axes and preferred intraclast orientation data indicate a local westerly trend (Figure 4). Examination of published geologic maps (Englund, 1968; Stricker, 1981; Trent and Spencer, 1990) and limited outcrop and subsurface data (Figures 2, 4) suggests that sandstone bodies are arranged in east-west-striking elongate belts that are on the order of several kilometers or more wide. Sandstones are erosively based (Se) and commonly have a basal matrix-supported lag of mudstone intraclasts, claystone intraclasts, fossil stem and leaf fragments, and coaly casts of larger tree fragments. A variety of cross-bed structures are present and commonly superimposed upon one another. These

include abundant trough and irregular trough cross-beds (St, Se), minor tangential-based tabular-planar cross-beds (Sp), and rare or poorly defined epsilon cross beds. Flat to irregular planar-lamination sets (Sh) are irregularly distributed throughout the sandstones. The lithofacies is locally capped by 1 to 11 cm thick rooted zones.

Sublithofacies Se forms both a basal channel lag and internal erosional scours. The internal scours form erosive coset-bounding surfaces (Figure 6) that are commonly overlain by casts of leaf and tree fragments. These surfaces do not show evidence of extended sub-aerial exposure (such as rooted zones) and are therefore interpreted as bar accumulations, rather than a new basal channel lag produced in response to large-scale channel migration and

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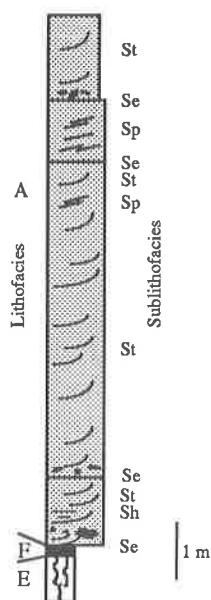


Figure 5. Sublithofacies relationships for strata of lithofacies A. Note three internal erosional scours (Se) and basal lag (Se) of organic detritus, bark and tree molds, and mudstone clasts commonly overlain by trough cross-beds of sublithofacies St. Sedimentary structures include grouped (pi) trough cross-beds in St and some solitary (theta) trough cross-beds typically associated with Se. Sublithofacies Sh (planar lamination) is randomly intercalated with both St and Se. Sketch is representative of outcrop of the Flattop Mountain Sandstone Member near Coaldale, WV.

superposition. In some exposures (e.g., Figure 6), coset bounding surfaces appear to be down-paleocurrent-descending (Haszeldine, 1983).

Sublithofacies Sh was produced by planar bed flow which can occur in both upper and lower flow regimes. The duality of paleohydraulic conditions under which planar flow can occur result in a randomly dispersed, intercalated network of preserved planar-lamination sets. Therefore, parallel laminations are commonly found vertically distributed throughout fluvial-bar deposits (Walker and Cant, 1984).

The overall characteristics of sublithofacies Se, St, Sp, and Sh, such as the upward decrease in scale of sedimentary structures, the stratigraphic position of St and Sh above a basal

scour and lag, and the random intercalation of plane bedding (Sh) and tabular-planar cross-beds (Sp), support a fluvial point bar or medial bar origin (Jackson, 1976; Miall, 1978, 1982). Basal scour and channel-lag deposits (Se) are overlain by channel-floor and lower-bar sands deposited as migrating subaqueous dunes. Middle-bar deposits are represented by medium-scale trough cross-bedding or local lateral accretion beds. These beds, in turn, are overlain by upper-bar deposits characterized by planar lamination, tabular-planar cross-bedding, and small-scale trough cross-bedding. The sequence is locally truncated by coset-bounding surfaces (Se).

This lithofacies was deposited as a series of variably sinuous fluvial channels in laterally restricted channel-belts. Vertical bedform sequences, sedimentary structures, and the overall geometry of lithofacies A sandstones (e.g., high ratio of width to depth) are very similar to rocks that have been interpreted as variable- and low-sinuosity channel-belt sandstones in the Upper Carboniferous coalfields of northern England (Haszeldine, 1983; Fielding, 1986). High-sinuosity channels are implied by the presence of sigmoidal-shaped lateral accretion bedding (epsilon cross-beds), and low-sinuosity channels are implied by the predominance of down-paleocurrent-descending (Haszeldine, 1983) bar deposits. Features of both high and low sinuosity could have been preserved in lithofacies A sandstones if they were deposited in high-sinuosity channels during periods of low discharge, and in migrating low-sinuosity channels across the width of channel belts during normal discharge and flood-stage conditions.

Lithofacies B (Sublithofacies Ss, Sl) - Crevasse-Splay Complex

Lithofacies B consists of thin (<1.5 m), very fine- to medium-grained lithic wackes and micaceous litharenites, commonly with a high content of argillaceous matrix (Figure 7). Based on features described in Table 1, this lithofacies is interpreted as a crevasse-splay

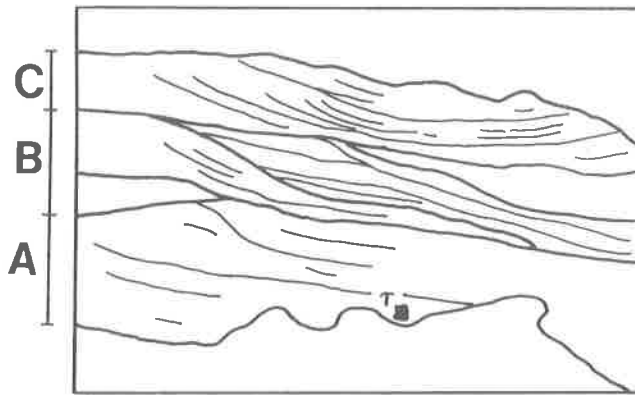


Figure 6. Outcrop photograph and interpretive sketch of the Flattop Mountain Sandstone Member (typical of lithofacies A). Rock hammer and day pack at base of exposure for scale; cross-bed coset a = 5.8 m thick, coset b = 3 m, coset c = 3.7 m. This outcrop displays several characteristic bedforms: 1) two internal erosional scours bounding cosets a-b and b-c (type 1 coset-bounding surfaces of Haszeldine, 1983); 2) cosets a, b, and c thicken down-paleocurrent to the right (west); 3) overall upward-thinning trend in bedset thickness (mean set thickness 40-50 cm in coset a, 4-10 cm in coset b, and 4-6 cm in coset c; 4) prominently west-dipping planar cross-bed sets; 5) superimposed trough cross-beds that strike subparallel to the inclined coset bounding surfaces (subperpendicular to their strike). These features are characteristic of down-paleocurrent-descending cross-beds (Haszeldine, 1983).

complex including both crevasse-channel and proximal-crevasse-splay deposits. This interpretation is supported by a low ratio of width to depth observed in individual channels, an irregular basal scour, overall lenticular geometry in outcrop sections, and variable sedimentary structures (Galloway, 1981; Guion, 1984; Ghosh, 1987). Lithofacies B was deposited

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through a combination of accretion and progradation.

In general, proximal-crevasse-splay deposits have a scoured and broadly undulatory base commonly associated with irregular cross-bedding, and crevasse channels are trough cross-bedded and have a deeply scoured basal contact. Overall scour geometry is broad and shallow, but close inspection reveals local troughs (on the order of 2-3 m wide by 0.5-1.5 m deep) incised into underlying strata below the mean basal contact plane. These local troughs probably indicate ephemeral high-flow events that produced small crevasse channels.

Lithofacies C (Sublithofacies Fsc) - Backswamp Complex

Lithofacies C consists of interbedded carbonaceous and micaceous mudstone and siltstone (Table 1; Figure 8). Sediments of this lithofacies were primarily deposited in a fresh- and brackish-water backswamp-floodplain environment through a process of slow suspension deposition. Evidence includes the presence of thick-laminated mudstone, well preserved plant debris, and the presence of freshwater and brackish fossils. Silt-dominated or rooted mudstone-dominated intervals are presumed to represent subenvironments on or proximal to a floodplain, whereas unrooted mudstone-dominated intervals probably reflect subaqueous deposition in shallow interchannel lakes, bays, or watery marshes. A lack of calcareous concretions and mudcracks along with an abundance of well-preserved organic debris including plant fragments and foliage indicates predominantly subaqueous deposition. Similar deposits have been interpreted as characteristic of backswamp environments by other workers (e.g., Coleman and Gagliano, 1965; Ferm and Cavaroc, 1968; Horne and others, 1978; Cecil and Englund, 1985, 1989).

Lithofacies D (Sublithofacies Fl) - Levee-Overbank and Distal Splay

Lithofacies D consists of heterolithic depos-

its comprising interlaminated sandstone, siltstone, and mudstone, typically less than 1.5 meters thick (Table 1; Figure 8). Deposits of lithofacies D are interpreted to be of either levee-overbank or distal splay origin. This interpretation is based primarily on close association with lithofacies A and C. For example, D commonly overlies the channel-belt sandstone succession (lithofacies A) and can be seen in cross section (Figure 4) to merge laterally into backswamp environments of lithofacies C. Intervals in close association with lithofacies A are interpreted as levee-overbank sequences, whereas intervals in association with lithofacies C are probably distal splay deposits. These deposits are difficult to distinguish from the crevasse-splay complex (lithofacies B), but are generally thinner, finer grained, internally monotonous, and have a less distinct or nonexistent basal scour. In general, beds of lithofacies B are differentiated from lithofacies D by their coarser grain size, prominently scoured base, and the presence of some cross-bedding. These deposits are compatible with those described by other workers as levee and minor distal-splay deposits (Coleman and Gagliano, 1965; Morgan, 1970; Ethridge and others, 1981; Guion, 1984; Ghosh, 1987).

Lithofacies E and F (Sublithofacies Fr and c) - Peat Swamp

Lithofacies E and F consist of a basal mudstone paleosol overlain by coal (Table 1). Lithofacies E represents development of a paleosol horizon similar to the gley deposits described by Besly and Fielding (1989). Evidence of subaerial exposure, such as mudcrack development, was not observed in lithofacies E; therefore, paleosols probably represent water-saturated accumulation under near-surface water-table conditions. Lithofacies F consists of coal and organic-rich black mudstone deposited in peat swamps that developed on paleosol horizons. In general, the presence of these units in a stratigraphic succession implies that conditions were free of continuous detrital influx. Similar deposits have also been



Figure 7. Outcrop photograph of lithofacies B near Garwood, WV. Hammer at left center for scale. Exposure has an abrupt and prominently scoured basal contact, gradational upper contact, upward thinning in bedset thickness, and is characteristic of a crevasse-channel deposit. Well-defined small channels extending below the mean plane of the basal contact locally incise underlying fissile mudstones of lithofacies C (a 1.5 m x 2.5 m trough cross-bedded channel is present to the right just outside of the photograph).



Figure 8. Outcrop photograph of lithofacies D silty sandstone underlain by fissile mudstone and siltstone of lithofacies C near Garwood, WV. This exposure of lithofacies D is characteristic of a distal crevasse-splay deposit based on the presence of a broad undulatory basal contact, distinct lenticular form (unit pinches out to the right), and close association with lithofacies C.

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Table 2.- Transition count matrix. Stratigraphically lower lithofacies in rows, immediately overlying lithofacies in columns. Derived from measured sections and borehole data described in appendix A and figures 11, 12, and 13 of Allen (1990).

Lithofacies	A	B	C	D	E	F	row sum
A	0	0	12	14	2	1	29
B	0	0	6	0	5	0	11
C	16	9	0	20	8	11	64
D	4	1	19	0	8	3	35
E	0	1	3	0	0	22	26
F	8	0	29	3	3	0	43
Column sum	28	11	69	37	26	37	208

described by other workers (*e.g.*, Ferm and Cavaroc, 1968; Horne and others, 1978; Cecil and Englund, 1985, 1989; Besly and Fielding, 1989).

Vertical Lithofacies Relations

Statistical Technique

The initial step involved the compilation of a transition count matrix (Table 2), utilizing complete and near-complete upper Pocahontas sections in order to summarize upward transitions among lithofacies. This results in a tabulated two-dimensional array in which the stratigraphically lower lithofacies of each transition pair is listed by row number and the overlying lithofacies by column number. Lithofacies transitions involving a unit overlain by itself (such as unit A overlain by unit A) are not recognized.

A second sequence of calculations was performed to create an independent trials probability matrix and a transition probability matrix. A difference matrix is generated by subtracting the former from the latter (Table 3). Details of this process are summarized by Miall (1973, 1982). Positive numbers in the difference matrix indicate a frequency of transition which is greater than that of chance. It is then assumed positive values may indicate significant, genetically related lithofacies associations. In order to further test for significance and eliminate the possibility of chance or non-Markovian processes as a control on transi-

tions, a chi-square test was performed on both the entire transition count matrix (using equation 4 of Miall, 1973; as recommended by Hiscott, 1981) and on individual positive cells of the difference matrix (using the equation suggested by Hobday and others, 1975). Resulting chi-square statistics greater than a minimum critical value dependent upon degrees of freedom in the matrix indicate correspondingly higher levels of confidence that the transitions exhibit Markovian properties, such as preferred transitions or cyclicity, and are not due to chance.

Results and Interpretation

Results of a chi-square test on the transition count matrix yield a statistic of 131.7 with 24 degrees of freedom. This value is well above the minimum critical value required for a 99.5% confidence interval and suggests that the transition count matrix does show cyclicity. Calculation of chi-square statistics for individual positive cells in the difference matrix confirms the presence of cyclicity with a high degree of confidence (>90% to >99.5%) in six matrix cells and a lesser degree of confidence (80% to 89%) in two cells. Four positive cells with very low chi-square statistics (0.08 to 1.28 with one degree of freedom; transitions A-C, C-A, C-D, F-A on Table 3) in the difference matrix were rejected as insignificant or due to random chance.

A path diagram based on results found to be significant in the difference matrix graphically depicts preferred lithic transitions in upper Pocahontas strata (Figure 9). Four depositional successions are emphasized in the path diagram, including 1) fining-upward transitions, 2) a coarsening-upward transition, 3) a two-way transition, and 4) a uniform (constant

Table 3.- Difference matrix. Positive numbers indicate nonrandom transitions.

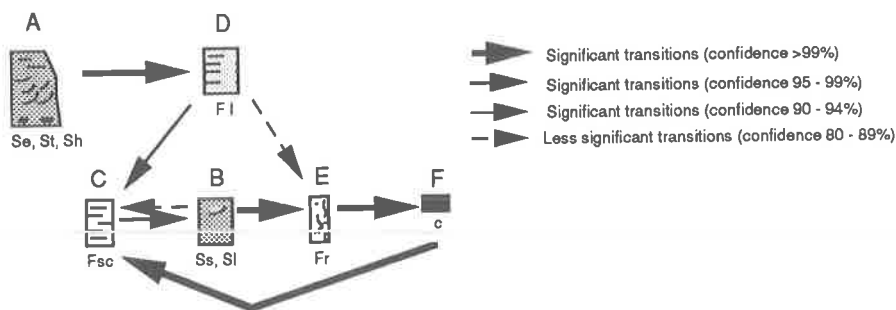


Figure 9. Path diagram of preferred lithic transitions.

grain size) transition. Each of these is described and discussed below.

Fining-upward transitions include couplets A-D, D-C, B-C, B-E, and D-E. The fining-upward transition is the most common type of transition present and is characteristic of both high-sinuosity meandering fluvial systems (Allen, 1970; Cant, 1982) and some types of cyclic, low-sinuosity braided systems (Saskatchewan-type rivers of Miall, 1978, 1982). Transition A-D, channel-belt sandstone overlain by levee-splay deposits, is characteristic of a fining-upward bar sequence where upper-bar deposits are overlain by finer levee and overbank deposits. Transitions D-C and B-C are similar in origin; splay and levee deposits overlain by finer floodplain and backswamp deposits. Gradual reduction in energy and fining of deposition above splay lobes and waning flood deposits (levees) has been reported by many workers (Galloway, 1981; Walker and Cant, 1984; Ghosh, 1987). Transitions B-E and D-E are also similar in origin. This association represents the development of a soil on top of abandoned crevasse lobes and levee deposits which would tend to form a local topographic high and offer drainage for plant growth.

A coarsening-upward sequence exists in couplet C-B. The transition from C to B records a progradation of coarser sediment represented by levee-overbank deposits and crevasse-splay complexes over a muddy backswamp or floodplain assemblage. A two-way transition also exists in couplet C-B and suggests interbedding of floodplain-backswamp deposits with overbank and crevasse-

splay deposits. This association is especially evident in outcrop in the northernmost part of the study area where thick sequences of lithofacies C are interrupted by thin prograding sandy deposits of lithofacies B and D.

A uniform transition is recorded by couplets E-F and F-C. In contrast to the other types of transition, the uniform transition has no significant change in depositional energy or grain size across each transition boundary. The boundary is primarily marked by a change in vegetation. Couplet E-F represents an obvious upward-transition from rooted soil horizon to a vegetated peat swamp. The presence of backswamp deposits over coal in couplet F-C is inferred to indicate drowning of a peat swamp.

The lithofacies sequence model depicted in Figure 9 is most valid only for the southern part of the cross-section A-A' (Figure 4) where lithofacies A is present. Because lithofacies A is absent from the northern area, the sequence is "short cut" and lithofacies A is entirely excluded from the path diagram. However, even in the southern area there is no preferred transition from any lithofacies upward to lithofacies A (*i.e.*, A apparently can be found at random above any other lithofacies).

DEPOSITIONAL MODEL AND DISCUSSION

Examination of the gross geometry of lithofacies suggests that the upper part of the Pocahontas Formation can be grouped into two general facies assemblages: a channel-belt sandstone and an overbank-floodplain facies

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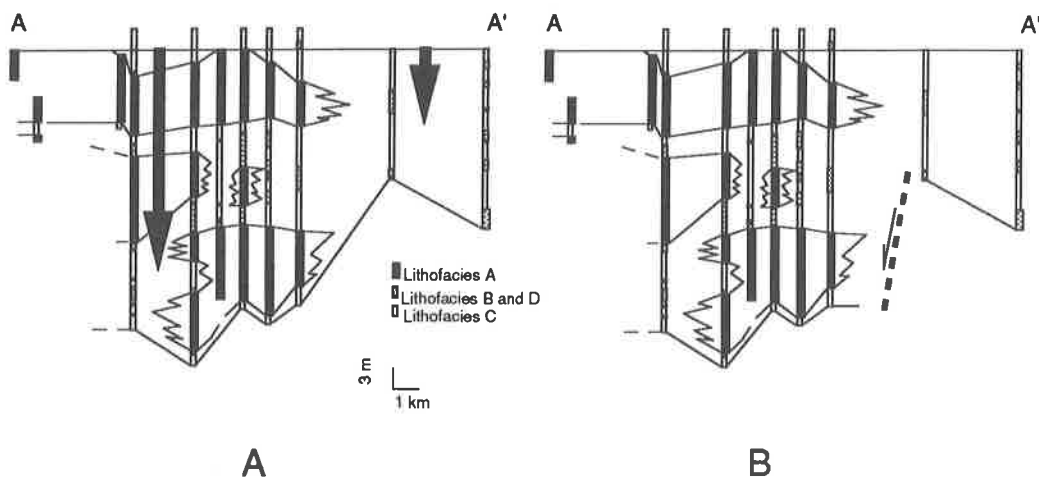


Figure 10. Depositional model for the upper part of the Pocahontas Formation. A) Simple differential subsidence. Arrows indicate relative subsidence required to explain north to south thickness increase. B) A modification of Figure 10A. Growth fault accommodating differential subsidence and influencing distribution of channel-belt sandstones.

assemblage. The channel-belt sandstone facies assemblage is represented by rocks of lithofacies A, and the overbank-floodplain facies assemblage consists of lithofacies B through F. A generalized depositional model capable of generating these major sedimentary features is a fluvial-deltaic model (e.g., Figure 1A) that is similar to models described by other workers (Ferm and Cavaroc, 1968; Haszeldine, 1983; Fielding, 1986) and consistent with regional interpretations (e.g., Englund, 1974; Ferm, 1974). However, certain modifications must be made to this general model in order to account for: 1) the reported presence of brackish and marine fauna in the roof mudstones of the No. 6 and No. 8 coal beds, and 2) the extreme increase (~20 m) in thickness of strata between the No. 6 and No. 8 coal beds from north to south (Figure 4).

The vertical distribution of fauna indicates that some freshwater to estuarine conditions existed in the region immediately following deposition of the No. 6 coal bed and marine to brackish conditions existed following deposition of the No. 8 coal. The majority of the interval between the roof mudstones of the No. 6 coal and the base of the No. 8 coal is considered to be freshwater because continental plant

fossils are ubiquitous and marine or brackish fauna have not been discovered. These environmental conditions place the upper Pocahontas delta in an intermediate state between an entirely continental upper-delta plain and a periodically estuarine-marine-influenced transitional upper- to lower-delta plain.

The 20 m increase in thickness of section from north to south could possibly be attributed to either simple differential compaction (e.g., Englund, 1974) or differential subsidence. For example: 1) the sandstone-dominated facies may have compacted less than the mudstone-dominated facies, or 2) the individual channel-belt sandstone complexes may have stacked in an *en echelon* arrangement similar to the model of Ferm and Cavaroc (1968) causing a major sandstone delta lobe to subside faster than flanking overbank-floodplain assemblages (Figure 10A). Differential compaction does not adequately explain thickness variations because the mean total thickness of all overbank-floodplain assemblages (mudstone-dominated facies) between the No. 6 and No. 8 coal beds in the southern sections is 16.6 m compared to 21.4 m in the north -- a difference of only 4.8 meters. Therefore, the northern sections probably do not contain an appreciably

greater thickness of mudstone-dominated facies in order to account for an additional 20 m of thinning solely by processes of differential compaction. Simple differential subsidence (Figure 10A) is also difficult to apply. If the more competent meander-belt sandstone facies had preferentially and contemporaneously loaded soft mudstone-dominated facies in the south, then a variety of soft-sediment deformation features would be expected beneath or proximal to meander-belt sandstone assemblages. However, few soft-sediment deformation features, such as convoluted bedding, and no major synsedimentary deformational features, such as mud lumps, were observed in the backswamp-floodplain assemblages. Additionally, the mudstones are not any more deformed in the south than in the north suggesting that contemporaneous loading of sandstone must have occurred after mudstone-dominated facies were adequately lithified or stiffened. An alternative conclusion that accounts for increased subsidence in the south without a greater degree of regional soft-sediment deformation in the mudstone-dominated facies can be derived based on comparative evidence.

Comparison of the synsedimentary fault-controlled depositional model developed for coal-bearing rocks in the Black Warrior basin (Figure 1B; Weisenfluh and Ferm, 1984) with the gross geometry of facies in Figure 4 reveals a striking similarity. The southern part of section A-A' is very similar to the *en echelon* model of Ferm and Cavaroc (1968; Figure 1A), but the distribution of facies along the entire line of section is more similar to the cross section of Weisenfluh and Ferm (1984), suggesting that a down-to-the-south contemporaneous normal fault may have been a critical factor in controlling the differential subsidence and spatial distribution of channel-belt sandstone facies (Figure 10B). This would have allowed for channel-belt facies to episodically subside and accumulate on a down-thrown fault block under conditions that would minimize the creation of regional soft-sediment deformation features in underlying mudstone-dominated units. For example, Fielding and Johnson

(1987) show that synsedimentary structures in coal-bearing strata are localized and found only in the immediate vicinity of growth faults that accommodated differential subsidence.

Poor exposure in the vicinity of the northern limit of the meander-belt sandstone facies precludes direct observation of the inferred fault depicted in Figure 10B. However, several studies have postulated that down-to-the-south growth faults may have influenced deposition of other strata in the central Appalachian basin (Horne and others, 1978; Padgett and Ehrlich, 1978; Horne, 1979; Keiser and others, 1981; Ferm and Weisenfluh, 1989; Liu and Ferm, 1992), although such faults have rarely been directly observed in surface exposures (Keiser and others, 1981). In contrast to the central Appalachian basin, the effects of synsedimentary faults on local stratigraphy have been well documented in many other coal-bearing basins, including the Black Warrior basin (Thomas, 1968, 1986, 1988; Weisenfluh and Ferm, 1984), the Denver basin (Weimer and Davis, 1977), and the northern England coal fields (Fielding and Johnson, 1987). In fact, growth faults have long been considered to be a primary factor accommodating differential subsidence in many sedimentary basins (Shelton, 1968). The results of this study suggest that synsedimentary faults may have been an important control on the deposition and local geometry of strata in the upper Pocahontas Formation and that differential compaction was not a primary control on regional thickness variations.

CONCLUSIONS

1) Six lithofacies have been identified and interpreted based on a local facies analysis. These include variably sinuous channel-belt sandstones, crevasse-splay sandstones and siltstones, backswamp-floodplain mudstones, levee-overbank and distal crevasse-splay sandstones and mudstones, rooted mudstone paleosol, and coal deposited in peat swamps.

2) Lithofacies are cyclic and primarily show preferred fining-upward transitions from chan-

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nel-belt sandstones to levee-overbank deposits to backswamp-floodplain deposits. The cyclicity and fining-upward sequences are characteristic of both classic meandering fluvial models (e.g., Allen, 1970) and sandy Saskatchewan-type braided fluvial models (e.g., Miall, 1978, 1982). Additionally, a coarsening-upward transition exists where crevasse-splay complexes have prograded across backswamp-floodplain environments. Channel-belt sandstones succeed other lithofacies at random and are not present at all in the northern part of the study area.

3) Sediments were deposited on a transitional mid- to upper-delta plain locally discharging to the west. The deposition and distribution of facies between the Pocahontas No. 6 and No. 8 coal beds was controlled by depositional autocyclicity, locally modified by an inferred down-to-the-south synsedimentary fault that influenced the distribution of channel-belt sandstone complexes.

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BIOSTRATIGRAPHIC RELATIONSHIP OF THE GOSPORT SAND AND THE MOODYS BRANCH FORMATION (EOCENE) AT LITTLE STAVE CREEK, ALABAMA

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ABSTRACT

For many years there has been confusion about the stratigraphic relationship of the well-known and very fossiliferous Gosport Sand of Alabama and the Moodys Branch Formation (Eocene) in Mississippi. At Little Stave Creek, Alabama, a 25-foot section of glauconitic sand is present that has been divided into the Gosport and Moodys Branch in various ways. The occurrences of ostracodes in the greensands indicate that bed 43 of the commonly used section of Toulmin (1962) correlates with the lower part of the Moodys Branch of Mississippi and with the Dellet Sand of the Claiborne Bridge section. There is no unconformity in the greensands, except at the base between the Gosport and the Lisbon Formation.

INTRODUCTION

Two of the most macrofossiliferous Cenozoic lithostratigraphic units in North America are the Eocene Gosport Sand and the Moodys Branch Formation (e.g., Toulmin, 1977). However, for many years there has been some confusion about the stratigraphic relationship of these two units. They are lithologically similar. The original concept of the Gosport includes a bed that is of lower Moodys Branch age (MacNeil, 1946; Stenzel, 1952). Further, in Alabama, beds that have been referred to as "lower Moodys Branch" are in part of Gosport age (for example, unpublished ostracode data indicate that the lower part of the lower Moodys Branch of Huddlestun and Toulmin, 1965, on the Conecuh River belongs in the *Coccolitha grigsbyi*

Zone of Hazel, 1990). In this paper the authors seek to clarify the biostratigraphic relationships using the Ostracoda, primarily from the well-known locality at Little Stave Creek, Clarke County, Alabama.

PREVIOUS WORK

The Moodys Branch was formally proposed by Lowe (1915) and the Gosport prior to that by Smith (1907). The Gosport was originally referred to as the "Claiborne sands" by Conrad (1847). The type locality of the Moodys Branch is at Jackson, Mississippi, and that of the Gosport is 150 miles to the southeast on the Alabama River four miles below the bridge at Claiborne (Figure. 1). Moodys Branch has been applied to rocks from central Louisiana to the Alabama-Georgia border. The Gosport, on the other hand, is only identified in southwestern Alabama where, traditionally, it has been considered the youngest unit of the Claiborne Group. Stenzel (1952) points out that in Alabama it is difficult to find the contact between the Gosport and Moodys Branch "by lithologic contrast alone."

Cooke (1939) believed that some of what had been traditionally assigned to the Gosport was equivalent to the lower part of the Moodys Branch of Mississippi. He recommended abandonment of the term Gosport and the inclusion of all of the Gosport in the Moodys Branch, therefore transferring the concept of the Gosport from the Claiborne to the Jackson Group. Harris (1940) immediately made a plea for retaining both units. MacNeil (1946) was also aware that the upper bed of the Gosport at Clai-



Figure 1.—Index map showing the location of Little Stave Creek and Claiborne Bridge localities.

borne Bluff (his bed 5) was faunally more related to the Moodys Branch than the rest of the Gosport. He felt that this bed might be slightly older than the basal Moodys Branch in the type area or was equivalent to the glauconitic sand-filled burrows penetrating the underlying Cockfield Formation. He retained this bed, which contains *Periarchus lyelli* (Conrad), *Chlamys deshayesi* (Lea), and *Gigantostrea trigonalis* (Conrad), in the Gosport.

Stenzel (1952) reviewed previous work on the problem and discussed the field observations of himself and others concerning the Gosport primarily in its type area. It was his interpretation that in the type area MacNeil's bed 5 is in sharp contact with the bed below and that there is a phosphatic pebble layer in the base of bed 5 at the contact. He considered

this to represent a regional unconformity. Stenzel removed MacNeil's bed 5, which is 2.2 feet thick, from the Gosport and proposed the name Dellet Sand for this unit. The type locality is at the then "new bridge" over the Alabama River at Claiborne. There is now a newer bridge at this locality. The concept of the Dellet is basically one of biostratigraphy rather than lithostratigraphy. Subsequent workers have not used the term Dellet Sand. At that point in their studies neither Stenzel nor MacNeil were aware of, or at least had not studied, what has subsequently become one of the more famous Paleogene localities in the Coastal Plain, the Little Stave Creek locality at Jackson, Alabama (Figure 1).

Little Stave Creek is a tributary of Stave Creek, which flows into the Tombigbee River at Jackson. It is a small meandering stream that heads about three miles north of Jackson. Paleogene rocks are exposed here, about 15 miles from the normal outcrop belt, because of the influence of the Hatchetigbee Anticline and the Jackson Fault and downcutting by Little Stave Creek. Although mapped by Hopkins (1917), the locality was not examined and collected for its rich fossil beds until the 1930s. Toulmin (1940) provided the first general description of the section. A team of Louisiana State University personnel mapped the section by plane table in the early 1940s and the LSU section was published by Smith and others (1944). The same map and sections can also be found in Rainwater (1955). H. V. Howe of LSU collected samples for microfossils in the creek in 1943. Independently, Orville Bandy mapped the creek in the late 1940s and published his section in his monograph of the smaller foraminifers (Bandy, 1949). He included macrofossil data extracted from Smith and others (1944) on his section. There were a few discrepancies in terms of thickness between the Bandy and Smith and others sections, and in 1959 LSU graduate students Lewis Nichols and Phil Deboo made another plane table traverse of the section. Nichols and Deboo made collections for microfossils from most of the distinct beds and these are in the Museum collections at

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LSU. The Nichols and Deboo section was not published in its entirety, although some of it is included in Deboo (1965).

In 1962, Toulmin presented yet another measured section for Little Stave Creek (in this paper Toulmin also discusses the history of the Little Stave Creek locality). Thicknesses of individual beds in the Nichols and Deboo section and Toulmin's section are very close. Toulmin numbered the distinct beds of the section and indicated where the microfossil samples of Bandy (1949) were collected. Little Stave Creek is an important collecting locality for Toulmin's (1977) monograph of the Paleocene and Eocene macrofossils of the Coastal Plain.

Earlier, in the mid 1930s, Winnie McGlamery of the Geological Survey of Ala-

bama had collected fossils from Little Stave Creek (Toulmin, 1962). In about 1938, paleontologists Julia Gardner and C. W. Cooke of the U. S. Geological Survey collected fossils from Little Stave Creek. Gardner (1939) compiled the Gosport Sand macrofossil fauna. In 1957 she discussed the paleoecology of the exposed formations in the creek. Her section and formation boundaries in the 1957 paper are adapted from Smith and others (1944).

THE GREENSANDS AT LITTLE STAVE CREEK

In its outcrop area the Gosport is distinguished "by the presence in the formation of carbonaceous leaf-bearing clays which under-

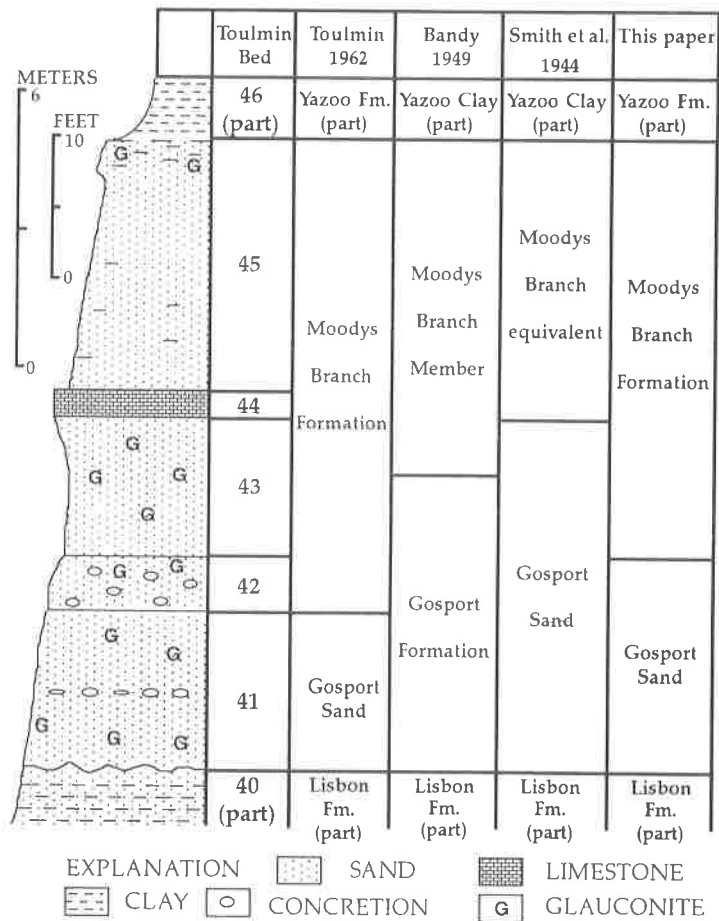


Figure 2.-Stratigraphic section of the Gosport Sand and Moodys Branch Formation interval at Little Stave Creek, Alabama. The nomenclature used by various workers is shown.

lie, overlies, or interfinger with the shell bed from place to place" (Toulmin, 1977, p. 115). However, at Little Stave Creek there are no leaf beds. There is 25 feet of glauconitic sand between the distinct erosional contact (hiatus of about 0.8 my, Hazel and others, 1984) between the Gosport and the underlying Lisbon Formation and the silty limestone ledge above (bed 44 of Toulmin, 1962) that causes the 12-foot waterfall (Figure 2). The Lisbon/Gosport contact, in sequence stratigraphic terms, is considered to be a Type 1 unconformity, and a sequence boundary (Baum and Vail, 1987; Mancini and Tew, 1991). The richly fossiliferous interval of the Gosport at Little Stave Creek is in the lower few feet of Toulmin's bed 41 of the formation. Macrofossils are comparatively rare above this interval. Because the lithologies are similar in the greensands, workers have had different opinions about which beds are equivalent to the Moodys Branch and which correlate with the Gosport at other localities (Figure 2). On lithological grounds alone there is no justification for placing a lithostratigraphic boundary anywhere in the 25 feet of glauconitic sand.

Smith and others (1944) and Gardner (1957) put all of the greensands in the Gosport, placing the boundary below the silty limestone ledge (bed 44). Bandy (1949), on the other hand, places this boundary in the middle of what came to be Toulmin's (1962) bed 43 because his sample from that level "...contains a few of the diagnostic Claiborne species..." However, the list of four "diagnostic" forms is not convincing. As pointed out above, important macrofossils are rare in the upper part of bed 41; however, Toulmin (1962) reports the biostratigraphically diagnostic bivalve *Venericardia alticostata* (Conrad) from the upper part of bed 41. He then placed the Gosport/Moodys Branch boundary between beds 41 and 42, but he found no macrofossils in bed 42. Bed 42 is concretionary and forms a small waterfall in the creek.

In this paper the lithostratigraphic scheme for Little Stave Creek of Toulmin (1962) is followed, except for the placement of his bed 42,

which is included in the Gosport rather than the Moodys Branch. The authors hasten to point out that they are aware that there are extensive problems of terminology and lithostratigraphic concepts in the interval under study that involve formations, groups, and provincial stages, and that these impact sequence stratigraphic interpretations. Such problems need to be addressed, but more research is needed and this is beyond the scope of the present study.

INTERPRETATION OF OSTRACODE DATA

Blake (1950) first reported on the ostracodes from the Gosport at Little Stave Creek, but it cannot be determined exactly where his samples came from. Not all samples from the greensand interval contain abundant microfossils. However, using samples from a trip to the locality in 1988 and samples collected from trips in the past and housed in the LSU collections, the authors have been able to document the ostracode fauna of the greensand interval at Little Stave Creek (Table 1). In addition, a sample from Stenzel's (1952) Dellet Sand at the Claiborne bridge section was also examined. Martin (1939), Stephenson (1942), and Blake (1950) have reported on the ostracodes from the Gosport on the Alabama River. The locality data for the samples used are given in Table 2.

There are numerous references to the ostracodes of the Moodys Branch, the most important of which are the monographs of Howe and Chambers (1935) and Huff (1970). However, none of these mention Little Stave Creek. Hazel (1990) recognized that the Gosport ostracode fauna was distinctive and assigned the Gosport to his *Coccolia grigsbyi* Zone and the Moodys Branch to the lower part of the *Actinocythereis montgomeryensis* Zone. At the time that paper was written, details of the distribution of the ostracodes at Little Stave Creek were not known.

At Little Stave Creek, bed 42 of Toulmin (1962) contains *Actinocythereis gosportensis* (Blake), *Cyamocytheridea goochi* (Stephen-

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Table 1. Occurrences of ostracodes at Little Stave Creek and the Dellet Sand at Claiborne Bridge. Description of the sample locations is Table 2.

Species	Sample								
	1	2	3	4	5	6	7	8	9
<i>Acanthocythereis</i> aff. <i>A. florienensis</i> (Howe & Chambers)									X
<i>Acanthocythereis florienensis</i> (Howe & Chambers)			X	X	X		X	X	X
<i>Actinocythereis</i> aff. <i>A. boldi</i> Huff				X	X		X		
<i>Actinocythereis</i> aff. <i>A. gosportensis</i> (Blake)	X								
<i>Actinocythereis</i> aff. <i>A. montgomeryensis</i> (Howe & Chambers)	X		X		X		X		
<i>Actinocythereis boldi</i> Huff			X				X	X	
<i>Actinocythereis davidwhitei</i> (Stadnichenko)						X			
<i>Actinocythereis gibsonensis</i> (Howe & Chambers)	X								
<i>Actinocythereis gosportensis</i> (Blake)			X	X	X	X	X		
<i>Actinocythereis montgomeryensis</i> (Howe & Chambers)								X	X
" <i>Acuticythereis</i> " <i>cocoaensis</i> Krutak				X			X		
<i>Alatacythere ivani</i> Howe									X
<i>Argilloecia</i> sp.								X	
<i>Bairdia gosportensis</i> Blake	X		X		X	X	X		X
<i>Bairdoppilata</i> sp. (Howe & Chambers)	X		X	X	X	X	X	X	X
<i>Buntonia shubutaensis</i> Howe								X	
<i>Bythocypris</i> sp.									X
<i>Clithrocytheridea</i> aff. <i>C. shubutaensis</i> (Stephenson)					X				
<i>Clithrocytheridea garretti</i> (Howe & Chambers) large form								X	X
<i>Clithrocytheridea garretti</i> (Howe & Chambers) small form	X	X	X	X	X	X	X		
<i>Cocoaia</i> aff. <i>C. smithwillensis</i> (Stepheson)	X				X		X		
<i>Cocoaia grigsbyi</i> (Howe & Chambers)	X	X	X	X	X	X	X		X
<i>Cyamocytheridea chambersi</i> (Stephenson)									X
<i>Cyamocytheridea goochi</i> (Stephenson)	X		X	X	X	X	X		
<i>Cyamocytheridea husseyi</i> (Stephenson)							X		
<i>Cyamocytheridea watervalleyensis</i> (Stephenson)									X
<i>Cytherella</i> spp.		X	X	X	X	X	X	X	X

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Table 1. Occurrences of ostracodes at Little Stave Creek and the Dellet Sand at Claiborne Bridge. Description of the sample locations is Table 2.

<i>Cytherelloidea montgomeryensis</i> Howe			X				X		X
<i>Cytheretta jacksonensis</i> (Meyer)	X	X	X	X	X	X	X		X
<i>Cytheromorpha</i> aff. <i>C. asperata</i> Huff.								X	
<i>Cytheromorpha</i> aff. <i>C. ouachitaensis</i> Howe & Chambers					X			X	
<i>Cytheropteron</i> aff. <i>C. montgomeryensis</i> Howe & Chambers								X	
<i>Cytheropteron montgomeryensis</i> Howe & Chambers								X	X
<i>Cytheropteron variosum</i> Martin			X	X	X	X	X	X	
<i>Cytherura moorei</i> Huff							X		
<i>Cytherura semireticulata</i> (Blake)			X			X	X		
<i>Digmycythere russelli</i> (Howe & Lea)								X	X
<i>Echinocythereis jacksonensis</i> (Howe & Pyeatt)								X	X
<i>Haplocytheridea montgomeryensis</i> (Howe & Chambers)	X		X	X	X	X	X	X	X
<i>Hazelina couleycreekensis</i> (Gooch)	X	X	X	X	X		X	X	
<i>Hemicythere bellula</i> Howe								X	
<i>Hermanites</i> aff. <i>H. moodysbranchensis</i> Huff							X		
<i>Hermanites claibornensis</i> (Gooch)		X							
<i>Hermanites collei</i> (Gooch)	X		X	X	X	X	X	X	X
<i>Hermanites melleni</i> Huff									X
<i>Hermanites moodysbranchensis</i> Huff								X	
<i>Konarocythere spurgeoniae</i> (Howe & Chambers)								X	X
<i>Loxoconcha</i> aff. <i>L. chamfera</i> Murray	X		X	X		X	X	X	
<i>Loxoconcha</i> aff. <i>L. creolensis</i> Howe & Chambers	X								
<i>Loxoconcha clarkensis</i> Blake			X	X	X	X	X		
<i>Loxoconcha stavensis</i> Blake		X	X	X	X	X	X	X	
<i>Monoceratina alexanderi</i> Howe & Chambers			X				X		X
<i>Monoceratina laevis</i> Stephenson							X		
<i>Monoceratina paucipunctata</i> Stephenson			X						
<i>Occultocythereis broussardi</i> (Howe & Chambers)								X	
<i>Opimocythere</i> cf. <i>O. watervalleyensis</i> (Howe & Chambers)	X	X		X		X			

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Table 1. Occurrences of ostracodes at Little Stave Creek and the Dellet Sand at Claiborne Bridge. Description of the sample locations is Table 2.

<i>Opimocythere martini</i> (Murray and Hussey)	X								
<i>Opimocythere mississippiensis</i> (Meyer)	X		X	X				X	X
<i>Opimocythere watervalleyensis</i> (Howe & Chambers)								X	X
<i>Ouachitaia caldwellensis</i> (Howe & Chambers)	X	X	X	X	X		X	X	X
<i>Ouachitaia gosportensis</i> (Stephenson)				X					
<i>Paracypris franquesi</i> Howe & Chambers				X	X			X	
<i>Paracytheridea belhavenensis</i> Howe & Chambers	X	X		X	X		X	X	X
<i>Pontocythere papula</i> (Krutak)							X		
<i>Pontocythere? alta</i> (Blake)			X	X			X		X
<i>Pontocythere? gosportensis</i> (Blake)	X	X	X	X		X	X		
<i>Pontocythere? perforata</i> (Blake)			X	X			X		
<i>Tringglymus hyperochus</i> Blake	X		X	X			X		
<i>Tringglymus? longicostata</i> (Blake)	X		X				X		
<i>Tropidocythere carinata</i> Huff								X	
<i>Xestoleberis sarsi</i> Howe & Chambers								X	
<i>Xestoleberis</i> sp. (of Blake)			X	X				X	

son), *C. husseyi* (Stephenson), *Cytheropteron variosum* Martin, and others that indicate it is faunally more related to bed 41 than bed 43. Forms that are typical of the Moodys Branch in Mississippi, such as *Actinocythereis montgomeryensis* (Howe and Chambers), *Tropidocythere carinata* Huff, *Konarocythere spugeonae* (Howe and Chambers), and others appear in bed 43. Thus, the boundary between the *Cocoaia grigsbyi* and *Actinocythereis montgomeryensis* Zones of Hazel (1990) is placed between beds 42 and 43. It should be pointed out that Edwards (1977), in an unpublished doctoral dissertation, using dinocysts and graphic correlation, makes the same correlation. There is no evidence of a hiatus in the section above the Lisbon/Gosport unconformity. In fact, the glauconitic sands grade to a bluish (when wet), clayey sand, which grades to a sandy clay, which grades to the bluish green clay of the North Twistwood Creek Clay.

The bluish, clayey sand and the sandy clay are the informal Blue Sand Member of the Moodys Branch of Hazel (1990). On their chart Baum and Vail (1987) place a sequence boundary between what they refer to as the Lower and Upper Moodys Branch. The authors find no physical or faunal evidence for this. Mancini and Tew (1991) place the Gosport and Moodys Branch in different sequence cycles with a Type 1 sequence boundary separating them.

Figure 3 is based on the ostracode data generated in the present study plus the available literature. The ranges of the most useful ostracode species for biostratigraphy are shown. Several species that heretofore had not been known to occur below the Moodys Branch and its equivalents, such as *Actinocythereis boldi* Huff, *A. gibsonensis* (Howe and Chambers), "*Acuticythereis*" *cocoaensis* Krutak, and *Opimocythere watervalleyensis* (Howe and Chambers) are found in the Gosport

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Table 2. Location of samples used in this study.

Sample number	Locality
1	Base of the Gosport Sand in Little Stave Creek, 3.5 miles north of Jackson, Alabama, Section 21, T7N, R2E, Clarke County. This is equivalent to Bandy's (1949) sample 34. Collector H.V. Howe, 1958. LSU H.V. Howe Collection slide number M1662.
2	Same location and stratigraphic position as sample 1. Collectors L.G. Nichols and P.B. Deboo, 1961. LSU slide number M426.
3	Same location. Gosport Sand. Lower bed 41 of Toulmin (1962). Approximately equivalent to sample 35 of Bandy (1949). Collector J.E. Hazel, 1988.
4	Same location. Middle of the Gosport Sand. Three feet below the top of bed 41 of Toulmin (1962). Approximately equivalent to sample 36 of Bandy (1949). Collector H.V. Howe, 1943. LSU slide number M1663.
5	Same location and stratigraphic position as sample 4. Collectors L.G. Nichols and P.B. Deboo, 1959. LSU slide number M414.
6	Same location. Gosport Sand. From the description, most probably bed 42 of Toulmin (1962). Collector H.V. Howe, 1943. LSU slide number 1824.
7	Same location. Gosport Sand. Middle of bed 42 of Toulmin (1962). Collector J.E. Hazel, 1988.
8	Same location. Moodys Branch Formation. Lower part of bed 43 of Toulmin (1962). Equivalent to sample 38 of Bandy (1949). Collectors L.G. Nichols and P.B. Deboo, 1961. LSU slide number M425.
9	Exposure at north side of the east end of the new bridge over the Alabama River near Claiborne, Monroe County. Dellet Sand of Stenzel (1952). Collector J.E. Hazel, 1988.

(beds 41 and 42 and equivalents). However, there are several forms that do not occur above the Gosport and another suite that have not been found below the Moodys Branch and its equivalents. The species *Hirsutocythere hornotina* Howe and *Pokornyella? bellula* (Howe) were originally described (Howe, 1951) from the Avon Park Limestone of central Florida. The Avon Park is thought to be correlative with the Gosport (e. g., Toulmin, 1977), but is faunally very different. By Moodys Branch time these taxa had migrated into the Coastal Plain.

Correlation to planktic zones in Figure 3 is based on incorporating the Eocene and early Oligocene at Little Stave Creek into a chronostratigraphic framework model for the Paleogene using the graphic correlation technique. Part of this model is described in Hazel (1989).

CONCLUSIONS

At Little Stave Creek at Jackson, Alabama, there is a 25-foot thick section of glauconitic sands (beds 41, 42, and 43 of Toulmin, 1962). Various authors have placed the Gosport/Moodys Branch Formation boundary at various places in the greensand section. These faunal discontinuities are actually biostratigraphic interpretations, as there is no justification for dividing the sands into lithostratigraphic units. The ostracodes indicate that the Moodys Branch Formation correlative assemblage (lower *Actinocythereis montgomeryensis* Zone of Hazel, 1990) begins at the base of Toulmin's bed 43. Bed 43 is the equivalent of the Dellet Sand of Stenzel, 1952) at Claiborne Bridge on the Alabama River and the lower part of the Moodys Branch of Mississippi. Beds 41, 42, and 43 at Little Stave Creek

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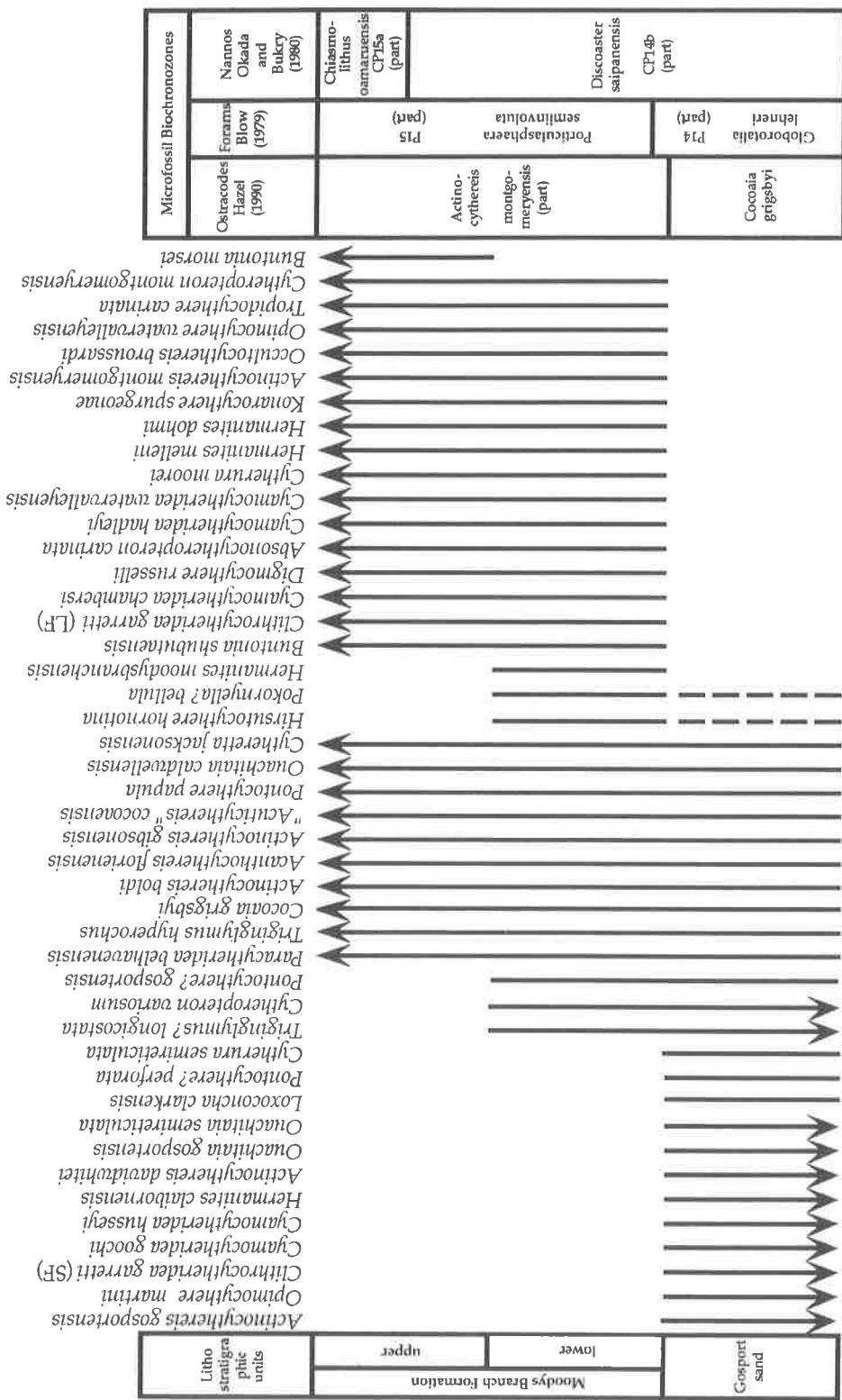


Figure 3.-Distribution of important ostracodes at Little Slave Creek. The lower Moodys Branch includes Beds 43 and 44 of Toulmin (1962). The upper Moodys Branch is Bed 45 of Toulmin and the so designated "Blue Sand Member" of Hazel (1990, Figure 2). The range determinations are from Table 1 and published and unpublished data. Correlation of zonal schemes from Hazel (1989;1990). LF=large form, SF=small form.

are part of the same transgressive interval. There is no evidence for hiatuses above the Gosport/Lisbon contact.

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