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Abstract

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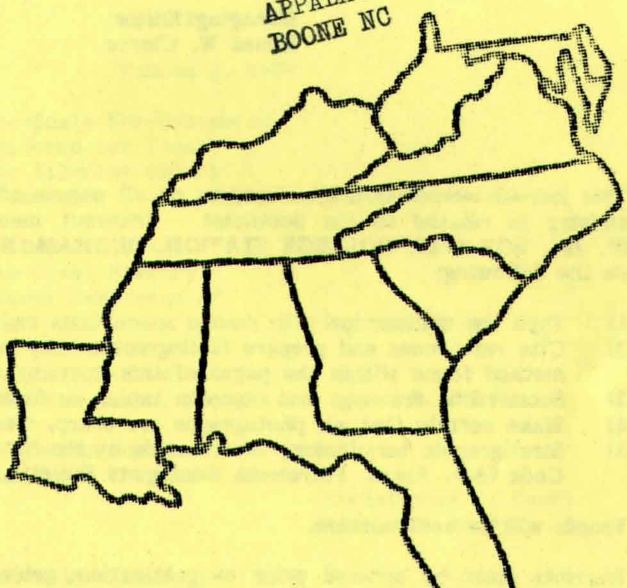

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SOUTHEASTERN GEOLOGY

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AN ANALYSIS OF LARGE-SCALE EBB-DOMINATED TIDAL BEDFORMS: EVIDENCE FOR TIDAL BUNDLES IN THE LOWER SILURIAN CLINCH SANDSTONE OF EAST TENNESSEE

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ABSTRACT

The Lower Silurian Clinch Sandstone of east Tennessee, previously interpreted as a shoreface to inner shelf sequence, contains large-scale (up to 1.5 m thick) sets of trough cross-strata that were deposited under strong tidal influence. Foresets are conspicuously bundled, and exhibit double-clay-drape sequences of the type: laminated sandstone, tens of cm thick → clay drape, mm thick → thin sandstone with ripple forms, <2 cm thick → clay drape, mm thick. These sequences are interpreted to represent, respectively: ebb-current deposition → slack-water after ebb current → flood current deposition → slack-water after flood current, and are thus inferred to equate with the tidal bundle sequences of Visser (1980). Deposition of thick bundles probably coincided with spring tides, whereas deposition of thin bundles and/or extensive bioturbation occurred during neap tides. Weak neap tides incapable of bedload transport of significant quantities of sand, and possible storm-wave reworking, may account for measured bundle periodicities (10-20 bundles) that are less than the theoretical 29.4 bundles for semi-diurnal tides. The bedforms that produced these bundled cross-strata were probably intermediate in characteristics (i.e., strength and degree of velocity asymmetry of governing currents) between Allen's (1980) Class III and Class IV bedforms. There was probably a significant stillstand period between threshold ebb and flood velocities necessary for bedload movement of sand, as evidenced by abundant clay drapes, and burrowed foreset and bottomset strata. In addition, abundant reactivation surfaces attest to bedform erosion during flow reversals and/or overtaking by smaller-scale superimposed bedforms.

INTRODUCTION

Tidal bundles represent increments of sediment (generally sand and mud) deposited on the slipface of a migrating bedform by tidal currents, in a regime in which ebb and flood currents differ considerably in strength (Boersma, 1969). It has been demonstrated by Visser (1980) that sand layers are deposited during active flow periods (ebb and flood), with the dominant current depositing thicker layers than the subordinate current; in addition, mud layers are deposited during the slack-water periods between ebb and flood flow. Visser (1980) also related the measured regular (periodic) variation in bundle thickness to the regular periodicity of the neap/spring tidal cycle; neap bundles are thinnest, whereas spring bundles are thickest. More detailed aspects of tidal bundle formation were later described by Boersma and Terwindt (1981), Terwindt (1981), Allen (1981 a,b; 1982), Van den Berg (1982), Siegenthaler (1982), De Mowbray and Visser (1984), and Yang and Nio (1985), among others.

Fossil tidal bundles have now been identified in a variety of Mesozoic to Cenozoic tidal deposits, including the Upper Jurassic Curtis Formation of the western U.S. (Kreisa and Moiola, 1986), the Lower Cretaceous Folkestone beds of southeastern England (Allen, 1981 a,b; 1982), the Lower Tertiary Roda Sandstone of the southern Pyrenees in northern Spain (Yang and Nio, 1985), the Middle Eocene of California (Clifton and Abbott, 1979), the Miocene Molasse in the vicinity of Fribourg, west Switzerland (Homewood and Allen, 1981; Allen and Homewood, 1984), and especially the Holocene of the Oosterschelde basin in the southwestern Netherlands (Visser, 1980; Van den

Berg, 1982; Siegenthaler, 1982; Nio and others, 1983; Yang and Nio, 1985). Surprisingly lacking are documented occurrences of Paleozoic and Precambrian sequences containing tidal bundles.

One purpose of this paper is to describe features observed in large-scale sets of trough cross-strata in the Lower Silurian Clinch Sandstone of east Tennessee (southeastern U.S.), and to present evidence for the interpretation that these features resemble tidal bundles produced by the migration of subtidal bedforms in response to strongly ebb-dominated tidal flow. In addition, this study will analyze the major differences between the features described in the Clinch Sandstone and "ideal" tidal bundles. It will be demonstrated that these differences are probably attributable to intense storm processes interacting with tidally produced bedforms and/or weak neap tides.

PREVIOUS WORK/BACKGROUND INFORMATION

Figure 1 summarizes the most recent research on the paleoenvironments of Lower Silurian (Llandoverian) strata deposited in the Appalachian miogeocline of the eastern U.S. Process-oriented sedimentologic research by Cotter (1983 a,b) established the existence of three major facies belts across Pennsylvania, namely alluvial fan, coastal alluvial (braided) plain, and marine shelf to shoreface. In east Tennessee, the work of Driese and others (Driese and others, 1984; Schoner, 1985 a,b; Schoner and Driese, 1985; Driese, 1985; 1986 a,b; Driese and others, 1986 a,b) serves to define three major facies belts, one of which is at least partly correlative (sedimentologically) with the western-most facies belt defined by Cotter (1983 a,b). Lower Silurian facies belts in east Tennessee include megarippled sand banks and shoals (inner shelf to shoreface), hummocky stratified storm-shelf, and carbonate (skeletal) sand bank to storm shelf (Figure 1). The deposits of the Clinch Sandstone discussed in this paper are part of the easternmost megarippled sand bank and shoal facies.

Figure 2 depicts, in greater detail, an interpretation of the paleoenvi-

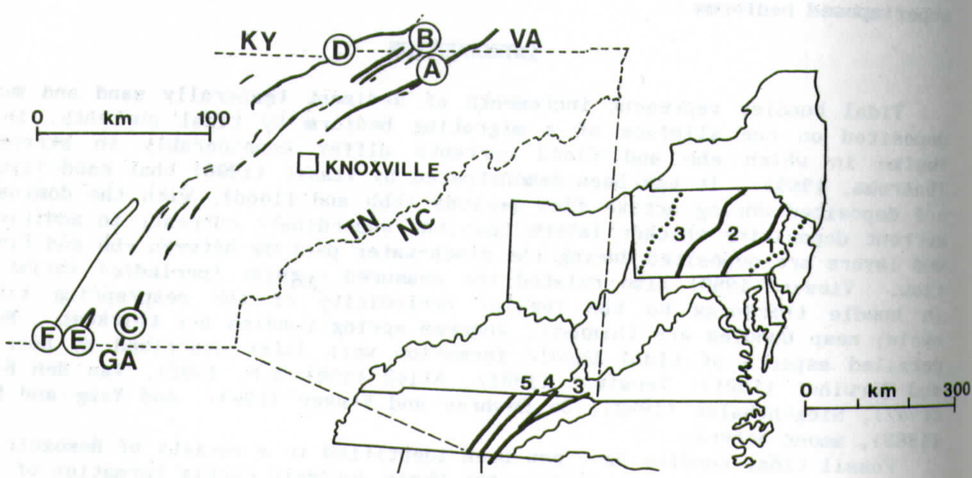

Figure 1. Correlation of Lower Silurian facies belts established by Cotter (1983 a,b) in northern Appalachian Basin, with those defined by Driese and others in the southern Appalachian basin (Driese and others, 1984; Schoner, 1985 a,b; Schoner and Driese, 1985; Driese, 1985; 1986 a,b; Driese and others, 1986 a,b). Clinch Sandstone (labelled A) represents easternmost belt of Lower Silurian rocks exposed in east Tennessee. 1 = Alluvial Fan; 2 = Coastal Alluvial Plain; 3 = Megarippled Shoreface/Inner Shelf; 4 = Hummocky Stratified Shelf; 5 = Carbonate Sand Bank/Shelf.

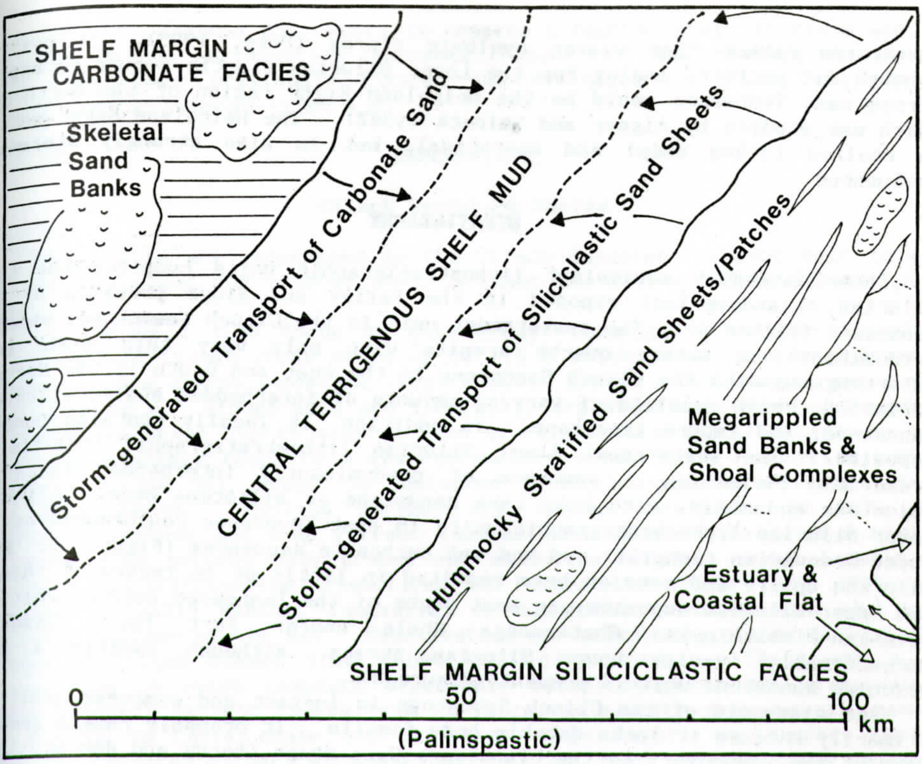

Figure 2. Paleoenvironmental reconstruction of east Tennessee during deposition of Lower Silurian strata (from Driese, manuscript in preparation). The model parallels, in many respects, one which has been proposed by Cotter (1983b) for slightly younger "Middle" Silurian strata exposed in Pennsylvania. Fair-weather wave currents, ebb-dominated tidal currents, and storm-generated geostrophic currents are all inferred to have influenced deposition. Facies 1 (discussed in this paper) probably was deposited in an estuarine setting characterized by strong ebb-dominated tidal flow.

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ronments of Lower Silurian strata, which are exposed in east Tennessee in a series of southeast-dipping imbricate thrust slices (Driese and others, 1986 a,b; Driese, 1986 a,b). This interpretation in many respects parallels one proposed by Cotter (1983 b) for "Middle" Silurian strata exposed in Pennsylvania. Siliciclastics were derived from a southeastern source (Taconic allochthons of Hatcher, 1972; 1978) and were transported westward by a combination of fairweather wave (trade winds) and tidal processes, with a northwest mode predominant. Major storm events also transported sand westward into deeper-water shelf environments, which were dominated (during fairweather periods) by terrigenous mud deposition and biogenic reworking (Driese and others, 1986 a,b; Driese, 1986 a,b). Storm-generated transporting agents probably included relaxation flows occurring after coastal set-up (c.f. Walker, 1984; Swift and Niedoroda, 1985), which developed as hurricane- generated waves approached from the west or northwest and piled a wedge of water along the coast. Geostrophic flows (i.e., currents flowing parallel to bathymetric contours) may have been generated both during storms and as southern Hemisphere Coriolis forces acted upon relaxation flows and deflected them to the southwest. The hurricane interpretation for storm processes is based on the paleolatitudinal/paleogeographic reconstructions of Ziegler and others (1977; 1979) placing east Tennessee about 30 degrees south of the paleoequator, a latitude dominated by hurricanes rather than winter cyclonic storms (Duke, 1985). A possible (though not perfect) analog for the Lower Silurian clastic sequences exposed across east Tennessee would be the Helgoland Bight region of the North Sea, which was studied by Aigner and Reineck (1982). The Helgoland Bight setting is shallow (3-50m deep) and macrotidal, but is also strongly storm-wave influenced.

STRATIGRAPHY

Three laterally equivalent lithostratigraphic units comprise the Lower Silurian (Llandoverian) exposed in the Valley and Ridge province of east Tennessee (Figure 3). The easternmost unit is the Clinch Sandstone, which is predominantly a mature quartz arenite with only very thin shale beds. Intertonguing with the Clinch Sandstone to the west and south is the Rockwood Formation, which consists of varying amounts of interbedded shale, siltstone, sandstone, and impure limestone. In addition, it locally contains hematite deposits. The westernmost Lower Silurian lithostratigraphic unit is the Brassfield Formation, a sequence of predominantly interbedded limestone, dolostone and shale, with only rare sandstone or siltstone beds. All three Lower Silurian lithostratigraphic units in east Tennessee conformably overlie Upper Ordovician (Ashgill) red bed and carbonate sequences (Figure 3). Post-Silurian uplift and erosion have resulted in little or no record of "Middle" and Upper Silurian deposits; in most parts of the Tennessee outcrop belt, the Devonian-Mississippian Chattanooga Shale and/or Fort Payne Limestone unconformably overlie Lower Silurian strata, although locally a Lower Devonian sandstone unit is present (Figure 3).

Age diagnosis of the Clinch Sandstone is inexact and somewhat arbitrary, primarily because it lacks datable body fossils. It probably ranges from the base of the Llandovery to the Llandovery C_{2-3} stage (Berry and Boucot, 1970). Overall it is a transgressive sequence which was deposited following a relatively rapid glaciogenic rise of sea level at the close of Ashgill time (Berry and Boucot, 1973; McKerrow, 1979). This coincided with a rejuvenation of Taconic highland areas to the east, thereby contributing sediment of greater size and quantity to the eastern margin of the Appalachian basin.

Northwest Georgia	East Tennessee (Sequatchie Valley)	East Tennessee (Westerly Belts)	East Tennessee (Easterly Belts)	Southwestern Virginia
Chattanooga Shale (Dev. - Miss.)	Chattanooga Shale (Dev. - Miss.)	Chattanooga Shale (Dev. - Miss.)	Wildcat Valley Ss. (Lower Devonian)	Rose Hill Fm. (Middle Silurian)
Red Mountain Fm. (Lower and Middle Silurian)	Brassfield Fm. (Lower Silurian)	Rockwood Fm. (Lower Silurian)	Poor Valley Ridge Mbr. Hagan Shale Mbr. CLINCH SS (Lower Silur.)	Tuscarora/ Clinch Sandstone (Lower Silurian)
Sequatchie Fm. (Upper Ordovician)	Sequatchie Fm. (Upper Ordovician)	Sequatchie Fm. (Upper Ordovician)	Juniata Formation (Upper Ordovician)	Juniata Formation (Upper Ordovician)

Figure 3. Stratigraphic nomenclature and approximate correlations for Upper Ordovician and Lower Silurian strata of the southern Appalachian Basin (northwest Georgia, east Tennessee, and southwestern Virginia). Sources include Berry and Boucot (1970), Chowns (1972), Dennison and Boucot (1974), Miller (1976), and Milici and Wedow (1977).

STUDY AREA

Large-scale sets of trough cross-strata displaying bundled foresets are exceptionally well-exposed in long roadcuts along the recently widened and improved U.S. Highway 25E, where it crosses Clinch Mountain at Beans Gap, in Grainger County, northeastern Tennessee (Figure 4). The outcrops utilized in this study were previously figured and discussed in Driese and others, (1984), and in Schoner (1985 a,b).

DESCRIPTION

Clinch Sandstone Facies

Four facies are recognized in the Clinch Sandstone outcrop belt in east Tennessee (Driese and others, 1984; Schoner and Driese, 1985; Schoner, 1985 a,b):

(1) Facies 1 (the focus of this study) consists of medium- to large-scale sets of trough cross-stata up to 1.5 m thick, developed in medium- to coarse-grained quartz sandstone (Figure 5a). Trough widths (axial exposures) are up to 10 m. Trough sets are internally organized into bundled foresets, in which sandstone (up to several tens of centimeters thick) alternates with shale (millimeters thick to a few centimeters thick) in a repetitive pattern (Figures 5b and 5c), except in instances in which complete megaripple forms are encountered. Maximum foreset dip angles rarely exceed 15°. Most bottomset laminae and some foreset laminae show some degree of bioturbation at interfaces between clay drapes and sandstone bundles; trace fossils are dominated by *Arthrophycus*, *Planolites* and *Paleophycus* (Figure 5d). Bottomset laminae typically exhibit tangential or sigmoidal contacts with the lower bounding surface of the set. Reactivation surfaces/pause planes lacking clay drapes are less common than those with clay drapes. There are a few occurrences in which complete megaripple (dune) bedforms are preserved in their entirety (height ranges from 25-75 cm, wavelength from 1-5 m), mantled by a 1-10 cm thick shale and siltstone drape layer; small-scale current ripples oriented either with dominant current flow, or oppositely directed

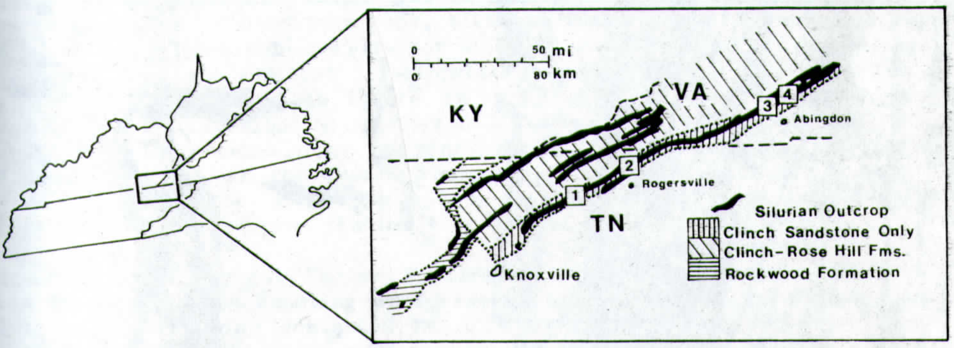

Figure 4. Location of Silurian outcrop belts in the Valley and Ridge Province of northeastern Tennessee and southwestern Virginia. Locality 1 (Beans Gap) is the focus of this study. Modified from Miller (1976).

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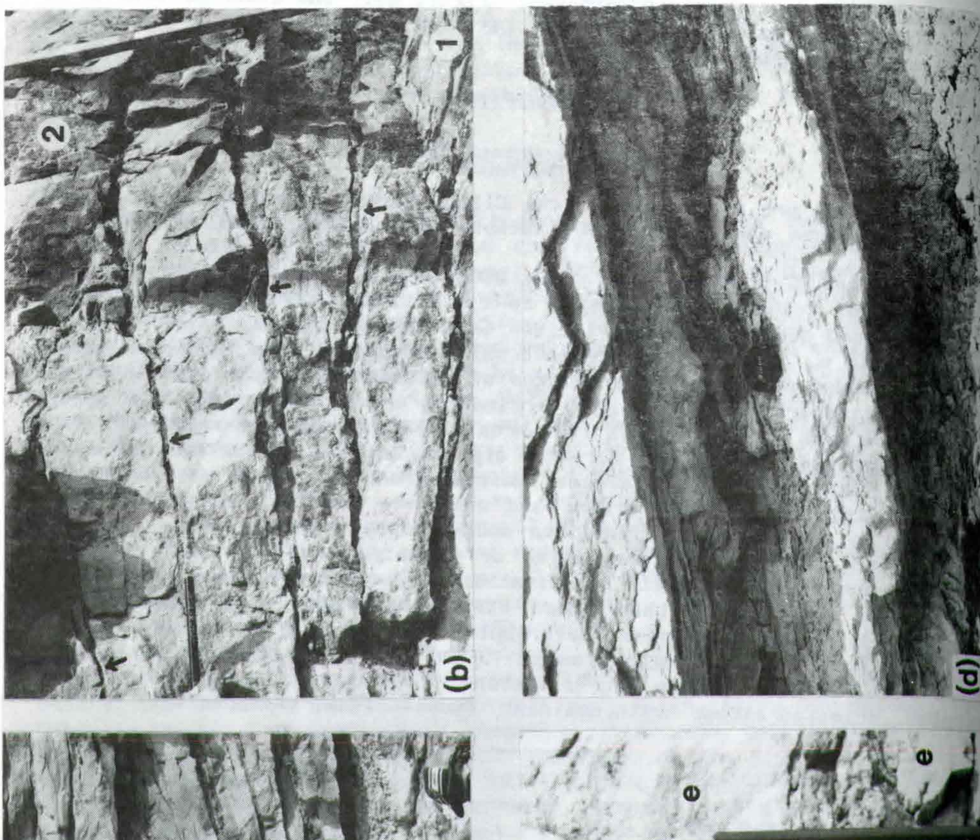
to it, occur on dune surfaces and along some bundled foresets. Topset laminae preservation was not observed except for the few complete bedforms described previously; most Facies 1 trough sets are truncated sharply along their upper bounding surfaces. In other words, the bulk of Facies 1 deposits consist of the lower parts of tidal bedforms which were incompletely preserved due to erosion of the top of the bedform. Individual cross-sets persist laterally over distances of up to 60 m.

(2) Facies 2 consists of medium- to very large-scale planar-tabular sets of cross-strata up to 2 m thick, developed in medium- to very coarse-grained quartz sandstone. Planar sets are comprised of centimeter-thick, normally graded, planar-abrupt foresets with abundant reactivation surfaces/pause


(a)
(b)
2
1
(c)
e
f
s
(d)

planes. Rare *Skolithos* and *Diplocraterion* traces descend from the upper surfaces of some sets or along reactivation surfaces between foresets. Shale drape layers are rare between foresets, although shale intraclasts are locally abundant. Winnowed granule and small pebble lag concentrations are common at the tops of some cross-sets. Foresets are generally inclined at or near the angle of repose (25-33°). Compound cross-sets are common; typically a subsidiary smaller-scale cross-stratification exists developed within thick, large-scale foresets with the dip direction of the smaller-scale cross-stratification more or less coincident with the larger-scale cross-stratification. Some thin (<5 m) vertical sequences exist in which a basal large-scale cross-set is overlain by progressively smaller-scale sets, and which may be accompanied by a slight upward-fining trend. Individual cross-sets persist laterally over distances of at least 10-20 m, and possibly more for the largest sets.

(3) Facies 3 consists of massive, densely bioturbated, fine- to coarse-grained quartz sandstone beds up to 3 m thick. Although some bedding planes are preserved, internal stratification features are rarely visible. Robust, centimeter-diameter *Diplocraterion* burrows with well-developed spreite, and millimeter-diameter *Skolithos* tubes, are the most abundant traces represented. Most horizontal bedding surfaces underlain by shale partings or layers (some up to 10-15 cm thick) exhibit abundant *Arthrophycus* traces. Rare cross-strata are preserved in the basal parts of some Facies 3 burrowed beds, with burrow density increasing progressively upward. Thick burrowed beds persist laterally for up to 150 m.

(4) Facies 4 consists of thin to medium beds of horizontally stratified and/or ripple cross-laminated, very fine- to medium-grained quartz sandstone, intercalated with shale lenses and layers up to a few tens of centimeters thick. The bases of most sandstone beds exhibit abundant *Arthophycus* and rare *Rusophycus* and *Cruziana* traces. Diminutive *Skolithos* and rare *Monocraterion* traces occur at the tops of some of the thickest sandstone beds. Both symmetrical (wave) and asymmetrical (current) ripples occur at the top surfaces of sandstone beds, although wave ripples are more abundant. Individual beds persist laterally for at least 10-20 m.

The lateral and vertical relationships between these four lithofacies are complex, as exemplified in Figure 6a. A previous study involving Markov Chain analysis revealed no preferred stacking patterns (Schoner, 1985a). Figure 6b illustrates, in greater detail, the complex intertonguing relationships between the four lithofacies. Burrowed units (Facies 3) may extend laterally for more than 150 m, grade downward into unbioturbated, cross-stratified sandstone (Facies 1 or 2) and are typically capped by thick

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Figure 5. (a) Typical Facies 1, large-scale set of trough cross-strata. Note the conspicuous bundling of foresets, as defined by clay drapes and/or zones of bioturbation dominated by *Arthrophycus* and *Paleophycus*. Jacob's staff is 1.5 m long, increments are 10 cm. (b) Several double-clay- drape sequences (arrows) observed in part of Facies 1 trough cross-set Unit A, which is depicted in Figure 7. Cross-set Unit A sharply overlies another Facies 1 sequence (1), and is overlain by a Facies 2 planar-tabular cross-stratified sequence (2). These bundle sequences greatly resemble those described by Visser (1980), which were attributed to deposition during each of the four phases of a complete tidal cycle. Scale bar increments are 10 cm. (c) Details of one complete double-clay-drape sequence from Facies 1 trough cross-set Unit A, depicted in Figure 5b. Interpreted are ebb-current-deposited sandstone layers (e), slack-water-deposited clay drapes (s), and a flood-current-deposited sandstone layer (f). Pencil is approximately 8 cm in length. (d) Concentrations of *Arthrophycus* burrows along foreset and bottomset planes in Facies 1 trough cross-set Unit B (shown in Figure 5a). Bottomset strata are generally more bioturbated than foreset strata. Some bioturbated zones extend all of the way up to the upper bounding surface of the cross-set. Lens cap is 5.5 cm in diameter.

(10-30 cm) shale layers continuous across the outcrop face. Most cross-stratified units (Facies 1 and 2) are limited in lateral extent to only 25 - 50 m, and erosively truncate underlying facies. Horizontally stratified and/or ripple cross-laminated, fine-grained sandstone sequences interbedded with shale are variable, averaging about 20 m in lateral extent. These Facies 4 sequences commonly appear as lateral equivalents of Facies 1 and 2 cross-strata.

Bundled Cross-Strata

General: Only in Facies 1 of the Clinch Sandstone were sedimentary structures suggestive of tidal bundles identified. Bundle-like features are especially apparent in the outcrop depicted in Figures 6a and 6b, where two large-scale sets of trough cross-strata were well-exposed and easily accessible for study. Figure 7 includes line drawings showing details of cross-stratification in these two cross-sets. In both drawings shale or silty shale strata are shaded dark, whereas sandstone strata are unshaded (white). All foresets depicted on Figure 7 were defined by either clay drapes, partings along texturally contrasting boundaries, shale intraclast concentrations, burrows descending from foresets, or reactivation surfaces/

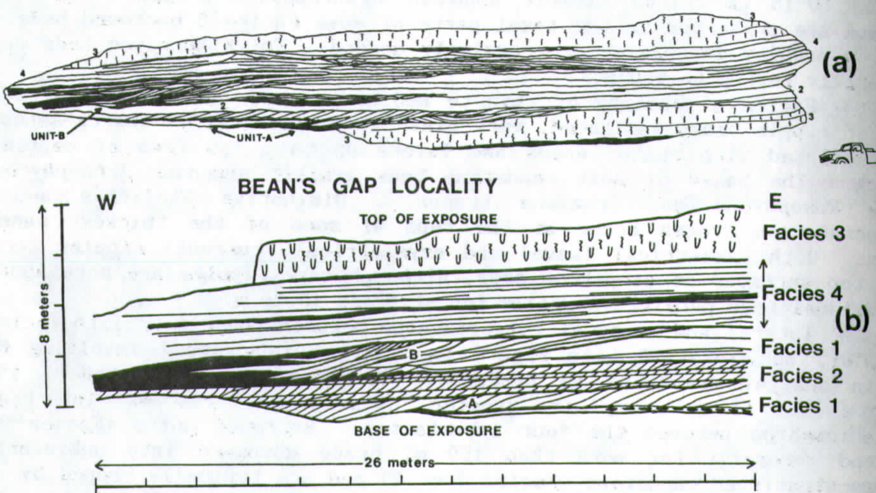

Figure 6 (a). Outcrop sketch showing lateral and vertical relationships between the four Clinch Sandstone facies defined and described in text. Note locations of two laterally persistent Facies 1 trough cross-set units, labelled A and B, which are discussed in detail in this study. Length of outcrop is 75 m, height is 10 m. (b) Detailed sketch of western half of outcrop depicted in sketch in Figure 6a, showing lateral and vertical relationships between the four Clinch Sandstone facies. Facies units are defined and described in text. Note two laterally persistent Facies 1 trough cross-set units, labelled A and B, which are discussed in detail in this study and are shown in Figure 7.

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pause planes discordant to foreset bedding.

Unit A: Cross-set A (Figure 7) was the most laterally continuous trough set studied, and is traceable for over 40 m across the outcrop face. It sharply overlies Facies 3 burrowed strata (Figure 6a). Sequential bundle thickness measurements were collected (bundle thicknesses measured at half of the set height) and are summarized in Figure 8. Two types of measurements were taken:

Method a - thickness of bundles bounded both above and below by well-developed clay drapes that extended all of the way from the bottom to the top of the cross set (Figure 8), and

Method b - thickness of bundles bounded above and below by any type of discontinuity feature (i.e., clay drapes, partings along texturally contrasting boundaries, shale intraclast concentrations, burrows descending from foresets, or reactivation surfaces/ pause planes discordant to foreset bedding) (Figure 8).

For method (a) there seems to be a general asymmetric pattern of gradually increasing thickness of bundles, then very abrupt thinning, followed by gradually increasing thickness (Figure 8). Bundle thickness ranges from 2-75 cm, with an average of about 30 cm. Periodicity between either two successive thickness maxima or two successive thickness minima seems to average about 15 bundles.

For method (b) a less well-developed pattern is indicated than for method (a), although there is some suggestion of slightly asymmetric to nearly symmetric bundle thickening and thinning trends (Figure 8). Bundle thickness ranges from 2-28 cm, with an average of 9.3 cm. Periodicity between either two successive thickness maxima or two successive thickness minima averages about 20 bundles.

Maximum foreset dip angles range from 12^{o} to 30^{o}, and average about 16^{o}. Whereas Unit A shows no periodic variation of dip angles, Unit B shows a

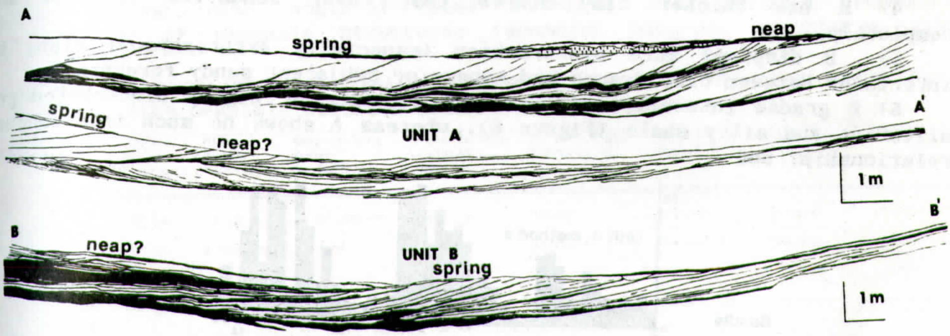

Figure 7. Detailed sketches of Facies 1 trough cross-set units A and B, depicted in Figure 6. The mode of formation of these cross-sets is discussed in the text. Note the conspicuous bundling of foresets and bottomsets; bundles are defined by either clay drapes, concentrations of shale intraclasts, reactivation surfaces/pause planes, or zones containing biogenic structures. Bar scales are shown for each cross-set.

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progressive decline in dip-angle as the bedform migrated. Sandstone foreset bundles toe into burrowed (mostly *Arthrophycus* and *Paleophycus/Planolites*) silty shale bottomset strata, some of which exhibit evidence of soft-sediment deformation. Double clay drapes are especially common in one part of the exposure of this cross-set (Figure 5b), in which the following sequence is repeated numerous times: clay drape, mm thick, → laminated thick sandstone, tens of cm thick → clay drape, mm thick, → thin sandstone with rippled upper surface (ripples oriented opposite or transverse to major foresets), <2 cm thick → clay drape, mm thick (Figure 5c). The best preservation of the above-described sequence occurs only in the middle and toe regions of each foreset. Topset strata were never observed, and the whole cross-set is capped by either Facies 2 or Facies 4 strata (Figure 6).

Unit B: Cross-set B (Figure 7) is laterally continuous for over 25 m across the outcrop face, and sharply overlies Facies 2 strata (Figure 6). Sequential bundle thickness measurements were collected and are summarized in Figure 8. Again, bundle thicknesses were measured in two different ways, as was the case for cross-set A.

For method (a) (i.e., bundles bounded above and below by clay drapes extending across the entire foreset to the upper truncation surface), there is a nearly symmetrical pattern of bundle thickness increase and decrease, although a longer record would be desireable (Figure 8). Bundle thickness ranges from 2-28 cm, with an average of about 13 cm. The periodicity between either two successive thickness maxima or two successive thickness minima averages around 10 bundles (?).

For method (b) (bundles bounded by any sort of discontinuity) the bundle thickness pattern is much less apparent (Fig. 8). Bundle thickness ranges from 1-11 cm, with an average of 6.3 cm. Periodicity between either two successive thickness maxima or two successive thickness minima are difficult to assess, mostly because the lateral continuity of this unit is not extensive enough. There is a slight suggestion of a 15-20 bundle periodicity.

Maximum foreset dip angles range from 11^o to 16^o; and average about 12^o. Cross-set unit B differs from unit A in the following ways:

1) B has a more symmetrical pattern of bundle thickness variation, whereas A is distinctly asymmetric,

2) B has less variation, in terms of range of bundles thicknesses, than does A,

3) B has overall lower foreset dip angles than A,

4) B has thicker clay drapes that cover sandstone foresets more completely,

5) B displays more bioturbation (especially *Arthrophycus*) along the interfaces between clay drapes and super- or subjacent sandy foresets,

6) B grades laterally into a thick sequence of highly bioturbated gray siltstone and silty shale (Figure 6), whereas A shows no such intertonguing relationship, and

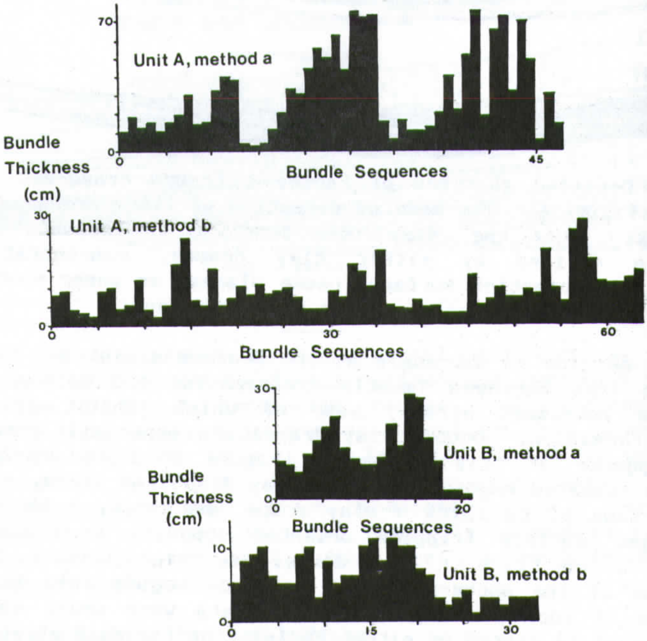

Figure 8. (a) Histograms for cross-set Unit A, showing relationship between bundle thickness and bundle sequences: (method a) for bundles bounded above and below by continuous clay drap layers, and (method b) for bundles bounded by any type of discontinuity, as discussed in text. (b) Histograms for cross-set Unit B, showing relationship between bundle thickness and bundle sequence: (method a) for bundles bounded above and below by continuous clay drape layers, and (method b) for bundles bounded by any type of discontinuity, as discussed in text.

7) the foresets of B display a progressive flattening of dip-angle with bedform migration, unlike A, which shows no pattern.
Double clay drapes, identical morphologically to those observed in cross-set A, were also observed in B, and again, optimum preservation was in the toe regions of foresets, extending only part of the way up the foresets.

INTERPRETATION

Comparisons Between Clinch Sandstone "Bundles" and Holocene to Recent Tidal Bundles

General: The bundle-like features reported here in the Clinch Sandstone compare favorably with tidal bundles reported in the literature. Table 1 lists some physical characteristics of Holocene tidal bundle sequences. Similar attributes include scale (cross-set thickness), maximum and minimum bundle thicknesses, neap-to-neap lateral spacings, and occurrences of double clay drapes. Dissimilar attributes include bundle periodicity (average periodicity for Clinch Sandstone bundles of about 15, versus ideal periodicity of 29.4 bundles for semi-diurnal tides (see Table 2)), angle-of-repose foresets (average foreset dips of 12-16^{o} for Clinch Sandstone cross-sets, versus near angle-of-repose foresets for Holocene deposits), occurrences of biogenic structures (abundant biogenic structures along

Table 1. Comparison of physical characteristics of Holocene tidal bundle sequences with possible Silurian tidal bundle sequences.

Cross-set Thickness	Maximum Bundle Thickness	Minimum Bundle Thickness	Neap-to-Neap Bundle Periodicity	Neap-to-Neap Lateral Spacing	Burrows Along Foresets (?)	Double Clay Drapes (?)	Angle-of-Repose Foresets (?)	Reference
0.2-2.5 m	80 cm	3 cm	26-30 bundles	10-11 m	Absent	Present	Present	Visser (1980)
0.25-1.0 m	115 cm	a few cm	No data	No data	Absent	Present (?)	Present	Boersma and Terwindt (1981)
0.2-1.5 m (Phillipsdam)	50 cm	a few mm	26-30 bundles	8-13.5 m	Absent	Present	Present	van den Berg (1982)
2.5 m (Schaar)	50 cm	a few mm	26-30 bundles	5-15 m	Absent	Present	Present	van den Berg (1982)
0.2-2.5 m	125 cm	5 cm	26-30 bundles	No data	Absent	Present	Present	de Mowbray and Visser (1984)
Unit A, Clinch Sandstone, 1.5 m	75 cm	2 cm	15-20 bundles	7-12 m	Present, <u>Arthrophycus</u>, <u>Planolites</u>, <u>Paleophycus</u>, <u>Skolithos</u>	Present locally	Rare, in general not present	(This study)
Unit B, Clinch Sandstone, 1.0 m	28 cm	2 cm	10-15 bundles (?)	10-12 m (?)	Present, <u>Arthrophycus</u>, <u>Planolites</u>, <u>Paleophycus</u>	Present locally	Not present	(This study)

interfaces between sandstone foresets and bounding clay drapes in Clinch Sandstone, versus near absence of biogenic structures in the Holocene deposits), and lack of identifiable structures in the Clinch Sandstone reflecting acceleration, full stage and deceleration of flow respectively during the dominant tide (as was reported by Boersma and Terwindt [1981] and Kreisa and Moiola [1986]).

Bundle Periodicity: The discrepancy between average bundle periodicity observed in the Clinch Sandstone (approximately 10-15 bundles) and the expected or ideal periodicity of 29.4 bundles can be explained in several ways:

1) Possibly the features observed in the Clinch Sandstone and described in this report do not represent tidal bundles but were produced by some other type of repetitive process characterized by alternating deposition of thick layers of sand and thin layers of mud on the slip face of a migrating sub-

Table 2. Comparison of astronomical parameters for Recent and Silurian.

Age	N_1 *	N_2 *	N_3 *	N_4 *	Reference
Recent	365.25	29.54	12.3	28.5-28.6	Lambeck(1978);Scrutton(1978)
Silurian	400-405	30.35-30.40	12.5-13.0	29.4	Lambeck(1978);Scrutton (1978)

*N_1 = number of solar days/year, N_2 = number of solar days/synodic month, N_3 = number of synodic months/year, N_4 = theoretical number of bundles in a neap/spring/neap sequence for semidiurnal tides

aqueous dune bedform. However, the author rejects this hypothesis because there are too many other similarities, cited in Table 1, which support a tidal bundle interpretation. Furthermore, there are a host of other sedimentary structures present in the Clinch Sandstone that attest to a high-energy regime that was undoubtedly shallow-marine (Driese and others, 1984; Schoner, 1985 a,b; Schoner and Driese 1985; Driese and others, 1986 a,b), which include:

(a) very large-scale (up to 2 m thick) sets of planar-tabular and wedge-planar cross-strata containing multiple reactivation surfaces and/or exhibiting compound cross-stratification,

(b) complete megaripple bedforms preserved and mantled with smaller-scale current ripples oriented oblique or opposite to dominant current flow.

(c) sequences of cross-sets arranged into thinning- and fining-upward sequences,

(d) deeply incised channel surfaces lined with concentrations of shale intraclasts,

(e) winnowed granule and small-pebble lags concentrated at the tops of some cross-sets,

(f) a shallow-marine trace assemblage dominated by high-energy suspension-feeders (*Skolithos* assemblage) interbedded with lower-energy deposit-feeders (*Cruziana* assemblage), and

(g) a unimodal paleocurrent pattern suggesting offshore or ebb-dominated (northwesterly) flow.

Lastly, the intimate association of the bundled trough cross-sets with marine trace fossils precludes a fluvial origin for the repetitive sandstone-shale alternations; within the marine realm, a tidal environment seems most favorable for the development of these features because of the characteristic alternation of bedload and suspension transport processes (Klein, 1977).

2) It is probably more realistic to consider the possibility of either incomplete sequences of bundles due to weak neap currents that left little or no record, and/or erosion and reworking of some bundles by episodic storm events. Boersma and Terwindt (1981) documented that the degree and sense of tidal dominancy varied with time and place across a carefully studied intertidal sand bank. Relevant were their observations showing that during neap tides the large-scale sand bedforms may become inactive and replaced by small-scale current ripples. The incomplete neap-spring-neap sequences was therefore produced in this type of situation. Visser (1980) observed a slight discrepancy between his observed bundle sequence periodicities (ranging from 26-30 bundles) and the average theoretical value of 28.5. He reasoned that the differences might be due to storms or irregular movement of the bedform during one dominant current stage. Visser (1980) felt that this could be especially important during a neap tide period, when the thickness of adjacent bundles would be very small. The hypothesis of episodic storm events modifying tidal bedforms in the Clinch Sandstone is supported by the documentation of abundant storm-wave features in the more offshore and laterally equivalent Rockwood Formation (Driese, 1985; 1986 a,b; Driese and

others, 1986 a,b). Allen (1981 a; 1982) discussed the possibility that storm-generated currents may inhibit, if not totally prevent mud drape deposition. The effect on the bundle sequence record would be to lower the number of drape-sandy foreset pairs which might otherwise have formed during the neap-spring cycle. There is also evidence for megaripple overtaking (cross-set A, Figure 7), as has been documented by De Mowbray and Visser (1984). Megaripple overtaking could have resulted in limited lengths of bundle sequences.

3) Thus far the author has only entertained the hypothesis of ideal semi-diurnal (twice daily) tides. If the Clinch tidal regime was characterized by mixed or diurnal tides, this would reduce the number of bundles in a neap-spring-neap sequence. A present-day diurnal tide has a period of approximately 25 hours, and is characterized by only one high and one low water daily; thus only 14 bundles would be produced over the neap- spring-neap time interval. The average bundle periodicity in the Clinch Sandstone of 10-15 bundles is therefore not in conflict with a diurnal tide interpretation. The marked asymmetry of Clinch Sandstone bundle sequences will be discussed in a subsequent section.

4) A fourth possible explanation for the lower number of bundles in Clinch bundle sequences is a preservational bias due to variable quality of outcrop exposures. Clinch Sandstone strata have experienced pervasive silica cementation (as syntaxial overgrowths), pressure solution (probably related to overburden), and structural deformation (Schoner, 1985 a). Consequently it is not inconceivable that some Clinch Sandstone bundles were undetected because diagenesis and tectonic overprinting masked some depositional features. Nevertheless, explanations #2 or #3 are still favored by the author for the reasons outlined previously.

High degree of bundle asymmetry: As noted earlier, Clinch Sandstone bundles do not exhibit a symmetrical (or even near-symmetrical) pattern of thickening and thinning; instead, the general pattern is of gradually increasing thickness, followed by very abrupt thinning (Figure 8). Possible explanations have been offered by Allen (1981 a; 1982) based on his study of tidal bundles in the Lower Cretaceous Folkestone Beds of England. However, the asymmetry noted by Allen (1981 a; 1982) and P. L. de Boer (Yang, pers. comm., 1986) is of the reverse type, that is, with bundle thickness increasing more quickly than it decreases. One explanation was that of bed-hardening (bedform armoring by mud) consequent on extensive mud deposition and ageing, as the peak spring tidal currents declined towards the neaps. Alternatively, it may have been that during neap periods currents were generally too weak to deposit sand layers, or that the threshold for dune migration was only reached during a part of the tidal cycle characterized by very strong spring tides; for example, Allen and Friend (1976) documented that at the Lifeboat Station Bank along the coast of England, dunes migrated under the influence of tidal flow only during 4 to 6 tides in each spring-neap cycle, because the minimum threshold for dune migration was not reached at all other times. A third explanation is that the bundle thickness apparently reflects an asymmetric curve of Umax in the neap-spring cycle. For example, Boersma and Terwindt (1981) observed on an estuarine shoal that a flood dominance which was present towards spring tide changed into an ebb dominance toward neap, leading to a relatively rapid decrease in Umax after spring tide.

Less-than-angle-of-repose foresets: The discrepancy between the low to intermediate dip-angles of Clinch Sandstone foresets (12-16^{o}) versus the common occurrence of near-angle-of-repose foresets in Holocene to Recent bundled cross-sets also requires further explanation:

1) One possibility is that these large-scale trough cross-sets in the Clinch Sandstone represent only the toe and possibly middle regions of what were once considerably larger bedforms, which were incompletely preserved- that is, the dune bedforms that produced the trough cross-sets could have been considerably higher (perhaps 3-4 m ?) than the thickness of the

preserved cross-set would indicate (<1.5 m). If angle-of-repose foreset bedding existed, it was removed by erosion, prior to deposition of the overlying stata. The fact that no topset strata were ever observed in conjunction with these trough cross-sets, and that the sets are always sharply truncated along their upper bounding surfaces, together support the hypothesis of extensive erosional truncation.

2) A somewhat related hypothesis is that Clinch Sandstone trough sets were produced by the migration of very large-scale, subaqueous tidal bedforms that possessed only moderately dipping lee slopes, such as are discussed by Allen (1980). Of the six structural classes of sand bedforms considered by Allen, Class III and Class IV deposits most resemble the Clinch Sandstone deposits (Facies 1) described in this report. Table 3 lists some of the important characteristics of Class III and IV structures. Class III has a significant stillstand period between threshold ebb and flood velocities for bedload movement of sand. Class IV shows major reactivation surfaces which are related to bedform erosion during flow reversals or overtaking by superimposed bedforms. Note that the large-scale trough cross-sets of the Clinch Sandstone display both Class III characteristics (abundant clay drapes and biogenic structures along foreset and bottomset surfaces) and Class IV characteristics (bundled foresets, shale clasts concentrated along foresets and bottomsets, abundant reactivation surfaces, etc.). Thus, one might conclude that the flow charactersitics of the Clinch paleoenvironment that produced Facies 1 might have been intermediate between Class III and IV

Table 3. Comparisons between Allen (1980) sandwave and dune models and Facies 1 large-scale trough cross-sets in the Clinch Sandstone.

Structure	Bedform Type	Flow Characterization	Bedform Morphology	Associated Structures
Allen (1980) Class III	sand wave or dune	1.) Dominated by unidirectional flow or reversing flow 2.) Bedload transport during dominant current 3.) Velocity periodically declines below threshold for movement of sand, and clay is deposited due to velocity reduction and/or high suspended sediment concentrations 4.) Flow separation on lee side of bedform	1.) Overlies laterally extensive first-order erosion surface 2.) Moderately steep foreset dip-angles ($\beta = 20^{o}$) 3.) + rippled bottomsets 4.) Possible smaller-scale superimposed bedforms	1.) Mud drapes cover some foresets, and especially bottomsets 2.) Weakly developed reactivation surfaces due to erosion by subordinate current 3.) Biogenic structures descending from foreset and bottomset surfaces
Allen (1980) Class IV	sand wave	1.) Reversing flow, with dominant and subordinant flows clearly developed 2.) Direction of bedload transport reserves w/each reversal of tidal currents 3.) Flow separation on lee side of bedform	1.) Overlies laterally extensive first-order erosion surface 2.) Moderate foreset dip-angles ($\beta = 12.5^{o}$) 3.) + rippled bottomsets 4.) Superimposed smaller-scale bedforms	1.) Foresets are divided up into bundles bounded by two types of second-order erosion surface: E_{2a} = gently inclinded sigmoidal discontinuities emerging from bottomsets; E_{2b} = steep, short surfaces that do not extend downward into bottomsets 2.) Rare mud drapes 3.) Shale clasts along foresets
Facies 1 large-scale trough cross-set, Clinch Sandstone (this report)	principally dunes but locally intergrading with sand waves	1.) Reversing flow, with dominant and subordinant flows clearly developed 2.) Bedload transport principally during dominant current stage, with rare subordinant current deposition 3.) Velocity periodically declines below threshold for movement of sand, and clay is deposited due to velocity reduction and/or high suspended sediment concentration 4.) Flow separation on lee side of bedform	1.) Overlies laterally extensive first-order erosion surface 2.) Moderate foreset dip-angles ($\beta = 12\text{-}16^{o}$) 3.) $\pm$ rippled bottomsets and/or foresets 4.) Possible smaller-scale superimposed bedforms	1.) Foresets are divided up into bundles 2.) Bundles are defined by clay drapes that extend from bottomsets up to upper bounding surface where they are truncated 3.) Biogenic structures descend from, or occur along foreset and bottomset boundaries 4.) Shale clasts dispersed within foresets 5.) Steep, short reactivation surfaces do not extend downward into bottomsets, $\pm$ clay drapes

conditions. However, caution should be exercised in such interpretations because Allen's model was originally developed for very large-scale bedforms (height = 4.25 m, wavelength = 210 m), and whether or not the original dune bedforms that produced the 1.0 - 1.5 m thick trough cross-sets in the Clinch Sandstone were also this large is certainly debatable.

DeMowbray and Visser (1984) have demonstrated that the slopes of bundle sequences are a function of degree of tidal asymmetry, such that low-angle sequences are formed by stronger subordinate flows and high- angle sequences are formed by weaker subordinate flows (assuming a constant dominant-current flow character).

Burrowed Foreset Strata: The common occurrence of horizontal burrows in Clinch Sandstone foreset and bottomset strata (Figure 5d) appears inconsistent with the Holocene tidal bundles described in the literature (Table 1), which lack biogenic structures. If the repetitive sandstone-shale alternations in the Clinch Sandstone do indeed represent daily increments of strata associated with tidal rhythms, then what part of the tidal phase do the burrowed horizons represent? Would it have been possible for benthonic organisms inhabiting lower-energy areas (such as the low areas between bedforms) to have migrated up onto the bedforms during periods of bedform inactivity associated with flood or slack phases, and, in the matter of a few hours or so, produce the bioturbation observed along some Facies 1 bottomsets and foresets? Or do these burrowed zones represent a period of days or months during which the bedform was inactive and became extensively colonized by benthonic organisms, in which case the sandstone-shale alternations are not at all related to tidal rhythms? A perusal of the literature existing on modern large- scale tidal sand bedforms shows that many of these questions have not been adequately addressed. In the North Sea region, bioturbation in the sand sheet facies is generally associated with a more distal tidal transport path position, whereas bioturbated sand banks are typically those existing in an inactive or moribund state (Stride and others, 1982). Only in the literature on inferred ancient tidal sands does one encounter reports of burrowing along foresets and bottomsets of presumed large- to very large-scale tide-dominated bedforms (Table 4). In most of these examples, the biogenic structures occur in association with obvious diastem or reactivation surfaces, or are interpreted to represent a reduced sedimentation rate caused by lower- energy conditions. In no case does an author specify that the traces were produced during a single tidal cycle-in fact, no attempts were made to estimate the amount of time represented by a burrowed foreset plane.

Another approach to questions concerning the timing of formation of biogenic structures along Clinch Sandstone foresets lies in paleoecological analysis of the organism(s) that produced *Arthophycus*, the most commonly observed ichnogenus. This trace fossil, first interpreted by Sarle (1906) as a burrow structure, has been the subject of much controversy with respect to interpretation of the organism(s) responsible for its formation. It has been typically attributed to sedentary polychaete worms, although a more recent interpretation suggests anthoptiloid sea pens as the trace-makers (Bradley, 1980; 1981). The concentration of these traces along clay drapes between sandy foresets in the Clinch Sandstone (Figure 5d) clearly indicates a preference for a finer-grained substrate and the nutrients contained within it. It seems entirely possible that instead of bioturbation occurring sequentially as each clay drape was added to the slipface of the migrating bedform, the organisms mined from the upper surface of the stabilized bedform downward, following each clay drape horizon to the base of each cross- set (E. Cotter, pers. comm., 1986). Such a process could explain the rather uniform distribution of *Arthrophycus* from the toeset to foreset on each bundle sequence containing a clay drape layer in excess of 0.5 cm.

Nevertheless, the interpretation that the burrowed foreset and bottomset planes represent reworking during the period: ebb slack → flood → flood slack, or perhaps during several days of exceptionally weak neap tides seems likely considering the similarities between Clinch Sandstone foreset bundles and Holocene tidal bundle sequences (Tables 1, 3), as well as the fact that greater concentrations of biogenic structures occur in what are interpreted as neap bundle sequences. In addition, the research of Rhoads (1967) on measuring the actual rate of bioturbation in Recent intertidal sediments showed that the upper 10 cm of sediment can be completely reworked in a period of one or two months; ultimately, the reworking zone could extend downward 30 cm, although bioturbation to this depth might require several

Table 4. Examples of bioturbation in large- to very large-scale sets of cross-strata interpreted to be tidal.

Stratigraphic Unit	Age	Locality	Trace Fossil Occurrences	Reference(s)
Rancho Rojo Mbr. of Schnebly Hill Fm.	Permian	Arizona	Both vertical burrows (Skolithos) and horizontal traces (Planolites, possible Agrichnium, Paleophycus, Scalarituba) occur in association with trough erosion surfaces and/or second-order bounding surfaces.	Blakey (1984)
Sandstone Sequence (Un-named)	Lower Jurassic	Bornholm, Denmark	Skolithos burrows descending from reactivation surfaces of megaripple strata and from individual clay-draped foresets	Sellwood (1972, 1975)
Osmington Oolite Series of the Corallian Beds	Upper Jurassic	England	Vertical burrows on topset and foreset planes (Arenicolites and Diplocraterion) and horizontal burrow galleries in bottomset strata (Thalassinoides and Rhizocorallium).	Wilson (1968, 1975)
Lower Greensand	Lower Cretaceous	England, France	Large cylindrical burrows, straight to slightly curved, ranging from 2 mm to 4 cm in diameter and 2 mm to 1 m long, occur concentrated in zones that parallel foresets, in cross-sets up to 5 m.	Middlemiss (1962); Narayan (1971); Nio (1976); Allen (1981 a,b; 1982); Walker (1984)
Roda Sandstone	Eocene	Spain	Simple vertical and horizontal burrows in association with major erosional surfaces and clay drapes, within the proximal slope and distal slope subfacies.	Nio (1976); Johnson (1979); Yang and Nio (1985)

years. Thus the 0.5 - 5 cm thick zones of bioturbation (of varying intensities) which occur along the shale-sandstone contacts that define foreset planes (Figure 5d), possibly represent a period of low bedform activity of at least a day, or possibly several days duration, perhaps in association with weak neap currents.

SUMMARY AND CONCLUSIONS

An analysis of several large-scale sets of trough cross-strata suggests the former existence of large- to very large-scale, ebb-dominated bedforms in a shallow subtidal, tide-dominated environment. The bedforms were probably large dunes (megaripples) that were intermediate in characteristics between Allen's (1980) bedforms Classes III and IV. Moderately inclined (12- 16°), trough-tangential to sigmoidal foresets are arranged into 2-75 cm thick bundles that are clearly defined by repetitive alternations of sandstone and shale. The physical characteristics of these bundles compare favorably with the tidal bundles reported recently from Holocene to Recent tidal sands, and the complete double-clay-drape sequence: thick sandstone layer (inferred to represent deposition during ebb flow) → thin clay drape (slack after ebb) → thin sandstone layer, commonly rippled (flood flow) → thin clay drape (slack after flood), was observed in several instances. An analysis of bundle thickness as a function of bundle sequence reveals a periodicity of 10-20 bundles, much less than the ideal 29.4 bundles. This discrepancy is probably related to either incomplete sequences of bundles due to weak neap currents (that left little or no depositional record) and/or erosion and reworking of previously deposited bundles by episodic storm events. Mixed or diurnal tides (instead of the hypothesized ideal semi-diurnal tides) would also reduce the number of bundles in a neap-spring- neap sequence. The occurrences of biogenic structures along some foreset and bottomset planes represent periods of low bedform activity of at least a day, or possibly several days duration, perhaps in association with weak neap currents.

During periods of stronger spring tides, large-scale dune bedforms migrated actively and deposition of double-clay-drape sequences occurred. Thick increments of cross-stratified sand were added to the bedform slipface during ebb flow by migration of smaller-scale megaripples or dunes across the surface of the larger bedform. During weaker flood flow, thin increments of sand with poorly defined current ripple forms were deposited; ripple orientations reflect flow reversal. Slack-water phases were dominated by suspension deposition of mud drapes over the surface of the large bedform. Some erosion of mud drapes and ebb current deposits may have occurred during subsequent flood flow, as evidenced by the more common preservation of double clay drapes towards the lower parts of foresets and in bottomset strata.

Approaching neap tides, bedform activity (migration) was greatly diminished - thinner bundles were deposited, and double-clay-drape sequences were not preserved. Less sand was deposited by ebb flow on the bedform slipface. Little or no sand was deposited during flood flow; instead, bioturbation and suspension deposition of mud drape laminae prevailed, as well as mud drape deposition during slack water periods. During periods of weaker neap tides, bedforms commonly ceased migration and no bundles were deposited. Bioturbation probably resulted in selective destruction of formerly deposited thin bundles, therefore, most of the bundles preserved are probably spring bundles.

This documented occurrence of tidal bundles in the Lower Silurian Clinch Sandstone in east Tennessee represents the oldest report of these features in the literature (i.e., the first Paleozoic occurrence), and the first in the southeastern United States. Careful scrutiny of clay-draped sets of cross-strata will likely yield more examples in the future.

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SEDIMENTARY FACIES, DEPOSITIONAL ENVIRONMENTS, AND SEA-LEVEL HISTORY - MOOREVILLE CHALK, LOWER CAMPANIAN OF EAST-CENTRAL ALABAMA

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ABSTRACT

The Upper Cretaceous Mooreville Chalk (Lower Campanian), a 160 m-thick formation in the inner Coastal Plain of east-central Alabama, contains ten shelf facies. In stratigraphic order these facies are: 1) interbedded glauconitic sandstone, sandy foraminiferal chalk, sandy marl, and calcareous clay (interpreted as innermost shelf), 2) sandy calcareous clay and silty marl with storm-scour features (open shelf at or just above storm wave base), 3) thoroughly bioturbated marl (open shelf below storm wave base), 4) grey laminated silty calcareous clay with inoceramids (open shelf, poorly oxygenated and below storm wave base), 5) thoroughly bioturbated silty marl and chalk (open shelf below storm wave base), 6) sandy marl and calcareous sandstone containing coarsening-upward sand bodies capped by hummocky cross-bedded and megarippled sands (open shelf above storm wave base with shelf bar development), 7) silty and non-silty calcareous clay and silty marl with thin scour-surface sand laminae (open shelf at or just above storm wave base), 8) calcareous sandstone, clayey micaceous sand, and sandy calcareous clay - all thoroughly bioturbated (open-shelf sands, probably above storm wave base), 9) micaceous marl and silty micaceous calcareous clay containing detrital *Exogyra* (open shelf well above storm wave base), and 10) algal calcisphere limestone beds (brecciated and non-brecciated) and intercalated hard, silty and micaceous chalk and soft, silty marl and calcareous clay (open shelf, well above storm wave base). Based on the vertical arrangement of these facies and to some extent on their lateral continuity, a relative sea-level curve for the Early Campanian on this part of the northern Gulf rim has been inferred. The curve shows three main cycles of relative sea-level rise and fall spanning the 5 Ma of the Early Campanian. This curve does not match well with the curve for the Lower Campanian in the Western Interior Seaway, suggesting a non-eustatic, local control on sedimentary cycles of this order on the northern Gulf rim.

INTRODUCTION

The purpose of this paper is to describe the sedimentary facies, facies relationships, and sea-level history of the Mooreville Chalk in east-central Alabama. The Mooreville is the lowermost unit in the Upper Cretaceous Selma Group of central Alabama. The Mooreville crops out in the inner Gulf Coastal Plain from northern Mississippi, through western Alabama, and into east-central Alabama where it grades laterally into the age-equivalent clastic Blufftown Formation of eastern Alabama and western Georgia (Figure 1). The sandy faces of the Mooreville are detrital tongues of the Blufftown or are genetically related to the Blufftown, the linear shoreline equivalent of the Mooreville (Skotnicki and King, this volume).

The Mooreville Chalk, including its upper member, the Arcola Limestone, is approximately 160 m thick and dips due south at 7.5 m/km in the study area. The resulting outcrop belt is approximately 19 km wide. The Mooreville can be traced in the subsurface many tens of kilometers south of the outcrop belt (Moore and Joiner, 1969).

Even though the Mooreville is referred to as a "chalk," the unit consists of several lithologies including calcareous clay, marl, impure chalk, limestone, and sandstone. In addition, virtually all gradations between these lithologies are present. Volumetrically, the most common lithology is a poorly indurated lutite that stradles the line between a calcareous clay and a very argillaceous chalk with the calcareous component deriving mainly from

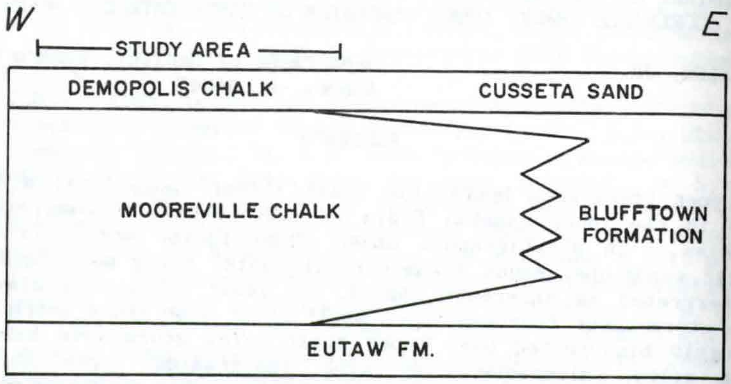

Figure 1. Schmatic of part of the Upper Cretaceous stratigraphy in the study area showing the Mooreville-Blufftown relationship across the region.

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calcareous nannofossils. This lithotype has been referred to as a marl by some workers (e.g. Monroe, 1941; Russell and Keady, 1983) and a marly chalk by others (e.g., Copeland, 1968).

According to Berger (1974) and Scholle and others (1983), marl and chalk are hemipelagic sediment types that are classified by percentage of calcium carbonate. In their scheme, marl contains 30 to 70% calcareous material; chalk contains over 70%. Calcareous clay has less carbonate than a marl. Their classification is used in this paper.

Previous work on the Mooreville Chalk in east-central Alabama has centered on physical relationships of formal lithostratigraphic units (Monroe, 1941 and 1947; Eargle, 1950; Copeland, 1968 and 1974) and on biostratigraphic relations (Cepek and others, 1968; Sohl and Smith, 1980; Smith and Mancini, 1983). The biostratigraphic studies have assigned an Early to Late Early Campanian age to the Mooreville based on calcareous nannoplankton and planktonic foraminiferal occurrences.

METHODS

One hundred outcrops were measured, described, and sampled in this study (Figure 2). Outcrops, mainly roadcuts, were used because they yielded good information on sedimentary structures, facies relations, and the relatively sparse macrofauna. The outcrops were located mainly along north-south trending roads. Only fresh outcrops were used for study; many of them were created in a relatively recent road-building and road-imporvement phase in the western part of the study area (Montgomery County). Measured sections were positioned stratigraphically according to elevation and north-south composites were used as single stratigraphic sections (numbers 1 - 10, Figure 2) in an east-west correlation panel diagram (Figure 3). Petrology of the shallow-subsurface section was taken from chip logs of five water wells described by Scott (1962) and Knowles and others (1963). The formational boundaries in the wells were picked according to geophysical log responses. No shallow subsurface core is available for the study area.

Laboratory methods on samples included binocular-microscope examination, thin-section study of indurated samples, acid digestion to determine carbonate content, and oil treatment to bring out trace fossils (technique of Bromley, 1980).

RESULTS

Ten sedimentary facies have been defined in the Mooreville. These facies are distinguished on the basis of their lithology, megafossils, sedimentary structures including trace fossils, and relative stratigraphic position.

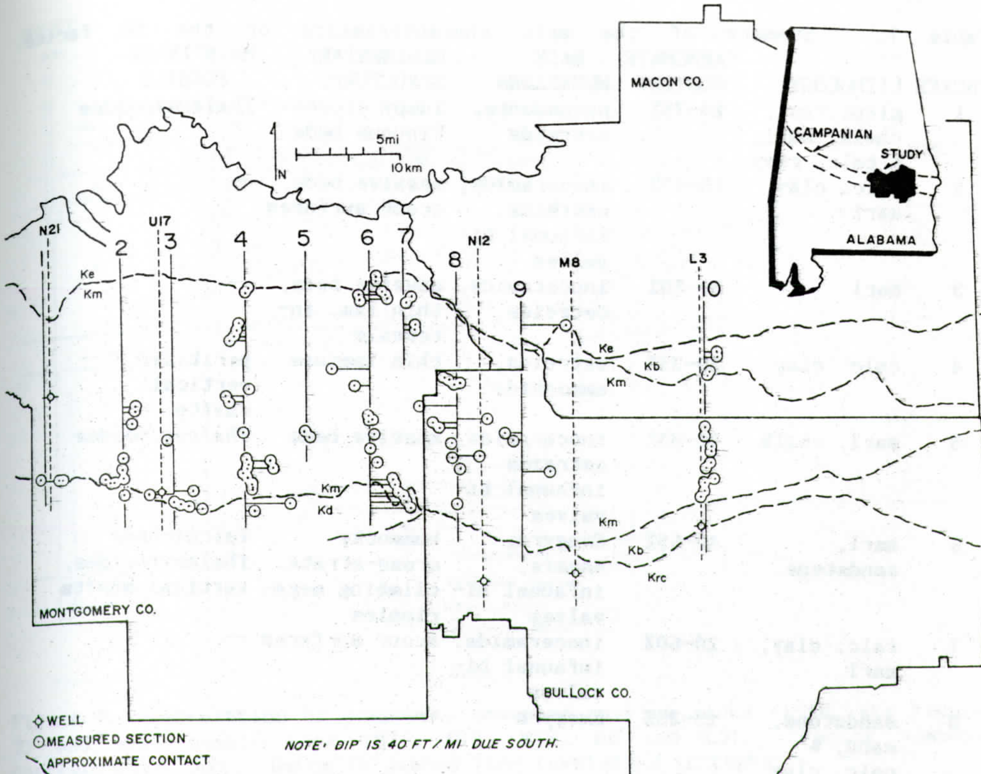

Figure 2. Study area showing surface outcrop of the Mooreville Chalk (Km) and Blufftown Formation (Kb). Underlying unit is Eutaw Formation (Ke) and overlying units are Demopolis Chalk (Kd) and Cusseta Sand (Krc). Composite section lines (1-10) are shown with measured-section locations projected onto them. Position and dashed projection lines for wells are also shown. Well data are from Knowles and others (1963, wells N21 and U17) and Scott (1962, wells N12, M8 and L3).

Table 1 shows the main characteristics of these facies. The ten facies will be correlated, described in detail and interpreted in the next sections.

STRATIGRAPHIC CORRELATION

Figure 3 shows the results of correlation among the measured sections (projected into their porper stratigraphic position) and the well sample logs. The facies numbers shown in Figure 3 correspond to the facies descriptions in the next section. The base of the section is the Eutaw-Mooreville conformable contact. The top of the section is a scour-surface discontinuity which separates the sandy marl of the overlying Demopolis Chalk (Kd) and Cusseta Sand (Krc) from the Mooreville facies. This scour-surface discontinuity, marked by a layer rich in shark teeth, fish bones, phosphatic shell molds and casts, and phospatic marl and limestone clasts, coincides with a faunal break (see Smith and Mancini, 1983; Carter and Wheeler, 1983). The datum is the top of the most continuous limestone bed in the Arcola Limestone Member (or Facies 10) of the Mooreville. In the eastern sections, the estimated projection of the limestone level and the base of the Facies 6 sands are used as datum levels.

Table 1. Summary of the main characteristics of the 10 facies.

FACIES	LITHOLOGY	CARBONATE CONTENT	MAIN MEGAFAUNA	SEDIMENTARY STRUCTURES	MAIN TRACE FOSSILS
1	glauc. ss., chalk, marl, & calc. clay	20-75%	pycnodonts, ostreids	lumpy discontinuous beds	*Thalassinoides*
2	calc. clay, marl	15-45%	inoceramids, ostreids, infaunal bivalves	massive beds, scour surfaces	--
3	marl	30-50%	inoceramids, ostreids	massive beds, thin lam. intervals	--
4	calc. clay	20-35%	ostreids, ammonoids	thin laminae	pyritized vertical shafts
5	marl, chalk	40-80%	inoceramids, ostreids infaunal bivalves	massive beds	*Thalassinoides*
6	marl, sandstone	40-65%	*Exogyra*, *Anomia*, infaunal bivalves	hummocky cross-strata, climbing megaripples	*Teichichnus* *Thalassinoides*, vertical shafts
7	calc. clay, marl	20-60%	inoceramids, infaunal bivalves	scour surfaces	--
8	sandstone, sand, & calc. clay	15-35%	*Exogyra*	--	--
9	marl, calc. clay	15-55%	inoceramids, *Exogyra*, ostreids	massive beds	
10	limestone, chalk, marl, & calc. clay	15-90%	*Exogyra*, bivalves, gastropods	thick beds, clast & shell imbrication	*Thalassinoides*

SEDIMENTARY FACIES

In this section, the facies are described in approximate stratigraphic order; Facies 1 is the lowermost in the study area (Figure 3).

Facies 1: This facies consists of roughly equal proportions of 1) tan calcareous, glauconitic sandstone, 2) tan to yellow sandy foraminiferal chalk, 3) tan sandy marl, and 4) tan to grey calcareous clay. Beds of sandstone and sandy chalk, 1 to 3 m thick, are generally more common near the base. The sandstone and sandy chalk contain intraclasts, phosphatic clasts, and phosphatized shell casts. These coarser beds contain small inoceramid impressions and casts of gastropods, ammonites, and *Baculites*. Further, abundant pycnodonts, including *Gryphaea*, and ostreids (attached to clasts and bedding surfaces) occur. The sandstone and sandy chalk display thoroughly bioturbated textures. Beds, where present, are 5 to 10 cm thick and are lumpy or discontinuous. *Thalassinoides* is a common trace fossil in the sandy beds. The sandy marls and calcareous clays contain large inoceramids with attached ostreids and display a general bioturbation and rare, vague horizontal lamination. Facies 1 ranges in thickness from 6 to 30 m and is thickest in the eastern part of the study area (Figure 3).

Facies 1 is laterally adjacent to the lower tongue of the Blufftown Formation. The lower tongue is composed mainly of lower-shoreface sands (Skotnicki, 1985; Skotnicki and King, this volume).

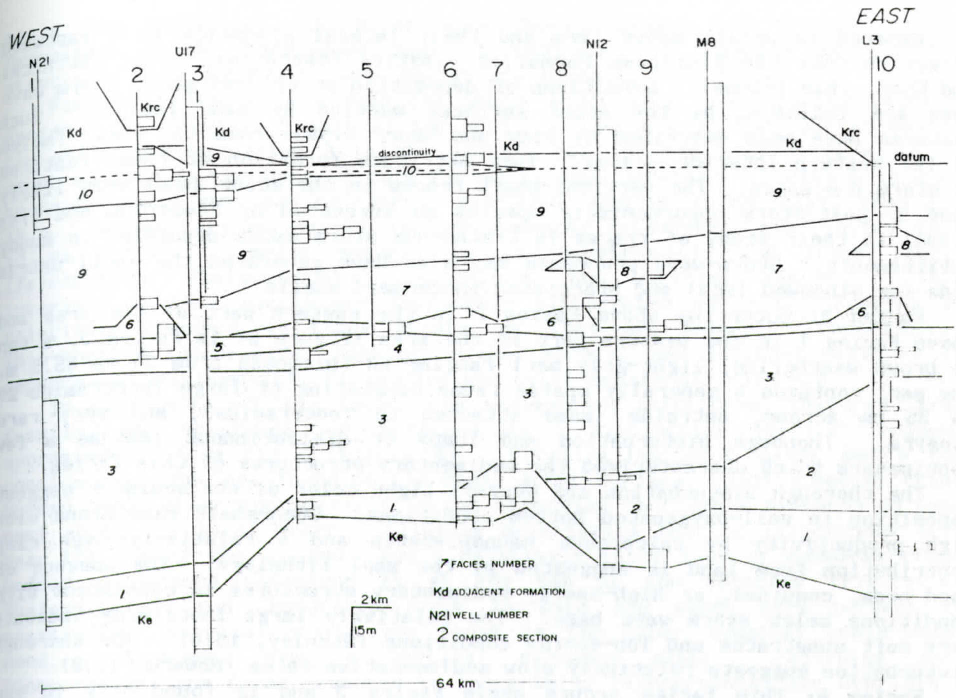

Figure 3. Correlation of composite measured section lines (1-10, see Figure 2) and well sample logs (N21, U17, N12, M8 and L3). Facies numbers correspond to text. Datum is dashed line (explained in text).

Facies 1 is interpreted as a shallow shelf, at or just below normal wave base. The glauconitic sandstone beds in this facies probably derived from a proximal nearshore source, whereas the sandy foraminiferal chalk suggests a shelf environment (Scholle, 1977). Intraclasts in the sandstone and chalk beds show that wave energy acted on the sea bottom at times. The glauconitic sandstone with phosphatic clasts corresponds to Jenkyns' (1980) innermost shelf facies in his Cretaceous transgressive shelf model. The transgressive nature of Facies 1 is suggested by the upward decrease in sandstone and sandy chalk beds. The marl and calcareous clay are interpreted as hemipelagic deposits of the inner shelf.

Facies 2: Occurring stratigraphically above Facies 1 in the western part of the study area (Figure 3) is Facies 2, a tan to grey sandy calcareous clay to silty marl with layers of dark-tan calcareous clay up to 6 m thick. Facies 2 ranges in thickness from 12 to 52 m in the study area. The main sedimentary structure is massive bedding with thorough bioturbation. However, some outcrops show minor intervals of thin, parallel lamination. In addition, rare, glauconitic sand laminae a few millimeters thick rest on shallow scour surfaces, 1 to 3 m wide. These thin sands are penetrated by vertical shaft burrows 1 cm in diameter. Inoceramid shell fragments oriented at high angles to bedding and thin discontinuous layers of fine shell (mainly ostreid) debris up to 1 cm thick are seen on a few outcrops. The fauna are relatively sparse, consisting of inoceramids (some bearing clinoid-sponge borings), small ostreids (as detached single values and attached to each other), and impressions and hematitic films of small infaunal bivalves leached of their shells.

This facies is interpreted as an open-shelf deposit having formed at or just above storm wave base but relatively close to the shoreline. Proximity to shore is confirmed by the greater amount of sand and silt in these strata as compared to strata above them and their lateral gradation in a eastward direction into the Blufftown Formation clastics (Skotnicki, 1985; Skotnicki and King, this volume). Conditions of deposition at or just above storm wave base are indicated by the scour surfaces mantled by sand laminae. Such features have been described by Rice and Shurr (1983) from the shelf facies of the Western Interior Seaway. They attribute formation of these features to storm processes. The vertical-shaft traces in the scour sands were likely made by post-storm opportunistic species as suggested by Pemberton and Frey (1984) in their study of traces in Cretaceous storm sands deposited in muddy environments. Storm-wave processes may also have generated the shell debris beds (as winnowed lags) and brecciated inoceramid shells.

Facies 3: Occurring above Facies 2 in the eastern part of the area and above Facies 1 in the western part of the area (Figure 3) is Facies 3, a tan to brown weathering, light-grey marl ranging in thickness from 20 to 45.5 m. The marl contains a generally sparse fauna consisting of large inoceramids 20 to 35 cm across, ostreids (some attached to inoceramids), and very rare *Exogyra*. Thorough bioturbation and lumpy or discontinuous laminae a few centimeters thick characterized the sedimentary structures of this facies.

The thorough bioturbation and overall light color of the sediment suggest deposition in well-oxygenated bottom conditions. Open-shelf conditions with high productivity by calcareous nannoplankton and a relatively high clay contribution from land is suggested by the marl lithology. The absence of sand beds, coquinas, or high-energy sedimentary structures is consistent with conditions below storm wave base. The relatively large inocermids indicate very soft substrates and low-energy conditions (Stanley, 1970). The thorough bioturbation suggests relatively slow sedimentation rates (Howard, 1978).

Facies 4: This facies occurs above Facies 3 and is found only in the western part of the study area (Figure 3). Facies 4 is a grey to olive, fissile silty calcareous clay that ranges in thickness from 0 to 10 m. The main sedimentary structure is thin parallel lamination. The facies show no good evidence of bioturbation except for some small shaft burrows, 5 to 10 millimeters in diameter and 2 to 3 cm long, that are mineralized by FeS_2. Very large inoceramids, over 30 cm across are numerous in this facies. Small ostreids and impressions of ammonoids (*Scaphites*) occur also. Nodules of FeS_2 are found on some outcrops.

This facies, unlike others below or above, shows evidence of deposition under relatively poorly oxygenated conditions. This evidence includes lamination with little bioturbation suggesting a very low density of benthic macro-organisms to churn the sediment (Reiskind, 1983). In addition, the very limited fauna (mainly inoceramids), grey color of the calcareous clay, and FeS_2 mineralization of burrows and in nodules suggest low-oxygen conditions at the sediment-water interface and below.

A likely explanation is that the oxygen-minimum zone was higher, perhaps in or near the study area, during deposition of Facies 4. Frush and Eicher (1975) postulated such an event to explain the dearth of benthonic foraminifera in the Upper Cenomanian transgressive sequence in Texas. Reiskind (1983) used a similar explanation to interpret laminated *versus* bioturbated units in the Lower Campanian of North Dakota. The lack of sedimentary structures, other than parallel laminations, suggests conditions below storm wave base.

Facies 5: This facies occurs above Facies 4 and is restricted to the western part of the study area (Figure 3). Facies 5 is a tan to cream-colored silty marl and chalk, ranging from 0 to 7.5 m thick. The megafauna are inoceramids, ostreids, *Hamulus*, and infaunal bivalve impressions that remain as hematitic films on broken surfaces of the rock. Thorough bioturbation with some *Thalassinoides*, more common in the chalk, characterizes this facies. Bedding is very vague and the sediment has a blocky nature due to many small joints in the outcrops.

This facies is interpreted as a relatively distal, open-shelf deposit. The *Thalassinoides* in well-churned sediment is characteristic of shelf-sea marls and chalks of 100 m or more depth (Ekdale and Bromley, 1984). Deposition below storm wave base is indicated by negative evidence; no structures attributable to wave energy are found in this facies.

Facies 6: Traceable across the whole study area, this facies rests on Facies 5 in the western part of the area, on Facies 4 in the central part of the area, and on Facies 3 in the eastern part of the area (Figure 3). A scour surface is the base of this facies in some sections. Facies 6 is a tan to brown sandy marl and calcareous sandstone ranging in thickness from 3 to 15 m. Sedimentary structures include hummocky cross-stratified layers (3 to 10 cm thick), climbing megaripples (in sandstones), and trace fossils (*Teichichnus*, *Thalassinoides*, *Chondrites*, and vertical shafts). Thick intervals of this facies are thoroughly bioturbated as well. *Exogyra* are relatively abundant in these strata and they are commonly clionid-bored. Thin-shelled inoceramids, *Hamulus*, *Pinna*, *Anomia*, and infaunal bivalves (as impressions and hematitic films) are the other megafaunal elements.

In most parts of this facies the sedimentary structures, textures, trace fossils, and megafauna are arranged in a distinctive vertical sequence approximately 6 m thick (Figure 4). Parallel-laminated sands and a thin conglomerate zone are found on the basal scour surface. Thin hummocky cross-bedded sands occur sporadically in the otherwise thoroughly bioturbated sandy marl above the basal scour. Climbing megaripples bearing *Teichichnus* occur near the top of the sequence and a relatively thick zone of hummocky cross-bedded calcareous sandstone is at the top of the sequence (Figure 4). *Thalassinoides* and *Chondrites* occur in sand and sandy marl between the megaripples; vertical shafts are found in the upper hummocky cross-bedded sand. Megafauna show a vertical sequence also: reclining forms (inoceramids) and forms cemented to inoceramids (*Alectryonia*) are overlain by strata with bysally-attached forms (*Pinna*, *Anomia*) and calcareous worms (*Hamulus*). *Exogyra* are found only at the top of the sequence in the most sandy strata. The sand content increases from approximately 15% just above the base to 75% in the upper hummocky cross-bedded zone; size increases from fine to medium sand from basal marl to sandy top.

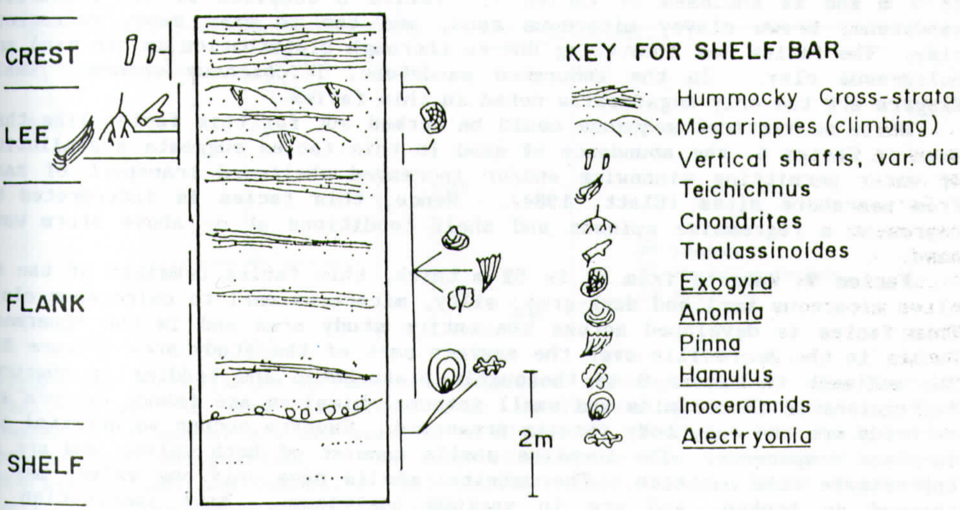

Figure 4. Generalized vertical sequence of interpreted shelf-bar sands. See text for explanation and interpretation.

The vertical sequence above suggests three phases of development: initial scour, hydrodynamic piling with occasional winnowing, and a storm-winnowing phase. The latter phase was likely punctuated with long intervals of quiet conditions as indicated by the distinct, low-energy trace fauna, especially *Teichichnus* (Fursich, 1975). Similar coarsening-upward sequences in fine-grained rocks of the Western Interior Seaway have been interpreted as shelf sand bars or sand lobes (Boyles and Scott, 1982). The distribution of physical and biogenic structures suggests that the lower part of the sand body represents shelf-bar flanks; the megarippled and *Teichichnus*-bearing zone, lee of crest; and the upper hummocky cross-bedded zone, active crest (subdivisions indicated in Figure 4).

Deposition of the shelf sands occurred in the water well above storm wave base, as indicated by the hummocky cross-stratification and megaripples. Conditions below normal wave base are indicated because much of the sediment is not well winnowed and thorough bioturbation is common. These points suggest a shallowing of water during deposition of Facies 6. In the Western Interior Seaway, shelf sand-body development was closely associated with regressive espisodes (Slatt, 1984).

Facies 7: This facies, ranging in thickness from 0 to 36 m, consists of dark tan to grey, silty and non-silty calcareous clay and tan silty marl. The facies has generally thorough bioturbation and a mottled appearance. Rare, thin sand laminae and vague parallel lamination are present in parts of some outcrops. The sand laminae, a few millimeters thick, rest on shallow concave-up scour surfaces and are penetrated by isolated vertical-shaft burrows, 1 cm across. The macrofauna are very sparse, consisting of impressions and hematitic films of thin-shelled inoceramids and small infaunal bivalves.

This facies exhibits some of the open-shelf characteristics discussed previously under Facies 2 and 4, namely dark colors, storm scour-produced sand laminae, parallel laminations, and a sparse fauna including inoceramids. This facies is interpreted as having formed at or near storm wave base. The presence of this facies above Facies 6 suggests a deepening of water in the area and development of less well-oxygenated conditions.

Facies 8: This facies is not continuous across the area; it occurs in two lenses in the study area (Figure 3). The facies ranges in thickness from 0 to 9 m and is enclosed by Facies 7. Facies 8 consists of tan calcareous sandstone, brown clayey micaceous sand, and tan to grey sandy calcareous clay. The facies has a mottling due to thorough bioturbation in the sand and calcareous clay. In the indurated sandstone, *Teichichnus* occurs. Small *Exogyra* are the only megafossils noted in this facies.

While no vertical sequence could be worked out for this facies like that seen in Facies 6, the abundance of sand in this facies suggests a shallowing of water permitting winnowing and/or increased shelfward transport of sand from nearshore sites (Slatt, 1984). Hence, this facies is interpreted to represent a regressive episode and shelf conditions at or above storm wave base.

Facies 9: Ranging from 21 to 52 m thick, this facies consists of tan to olive micaceous marl and dark-grey, silty, micaceous marl to calcareous clay. This facies is developed across the entire study area and is the uppermost facies in the Mooreville over the eastern part of the study area (Figure 3). The sediment in Facies 9 is thoroughly bioturbated and bedding is obscure. Impressions of inoceramids and small infaunal bivalves are found; *Exogyra* and ostreids are the only body fossils preserved. *Exogyra* occurs as detrital and in-place components. The in-place shells consist of both valves and are in approximate life position. The detrital shells have only one valve, may be chipped or broken, and are in various positions. Rare imbrication of detrital *Exogyra* valves also occurs in this facies.

This facies is interpreted as a shallow-shelf deposit having formed well above storm wave base. Wave energy acting on the bottom is indicated by the transport, comminution, and rare imbrication of *Exogyra*. An upward increase in the frequency of occurrence of these detrital shells along with an increase in the general sand content suggests a shallowing-upward sequence over the span of this facies.

Facies 10: Facies 10 corresponds approximately to the Arcola Limestone interval of the uppermost Mooreville and is the lateral equivalent of Facies 9 in the western part of the area (Figure 3). Facies 10 consists of 1 to 4 hard, white limestone beds and intercalated hard, tan silty and micaceous chalk, and soft, dark-tan silty marl and micaceous, calcareous clay. The interval is 0 to 22 m thick.

The limestones are 90% calcium carbonate and are composed mainly of benthic-algal calcispheres and calcite cement. The beds range in thickness from 3 to 35 cm. *Thalassinoides* is common in the limestones, and the burrows are filled with the overlying marl or chalk.

The limestone beds in some outcrops consist of imbricated angular rubble in the size range of 5 to 15 cm. Some clasts display conchoidal fracture. Isolated angular limestone clasts occur along a bedding plane in lieu of a continuous limestone bed in some outcrops.

The spacing between limestone beds is not regular. In the western part of the area, the lowest limestone is 6.3 m below the second limestone, whereas 1.4 m to 16 cm of marl separates the second and third limestones. Figure 5 shows the spacing of limestone beds in Facies 10. The correlation shows that only one limestone bed persists as far east as the central part of the study area (Figure 5).

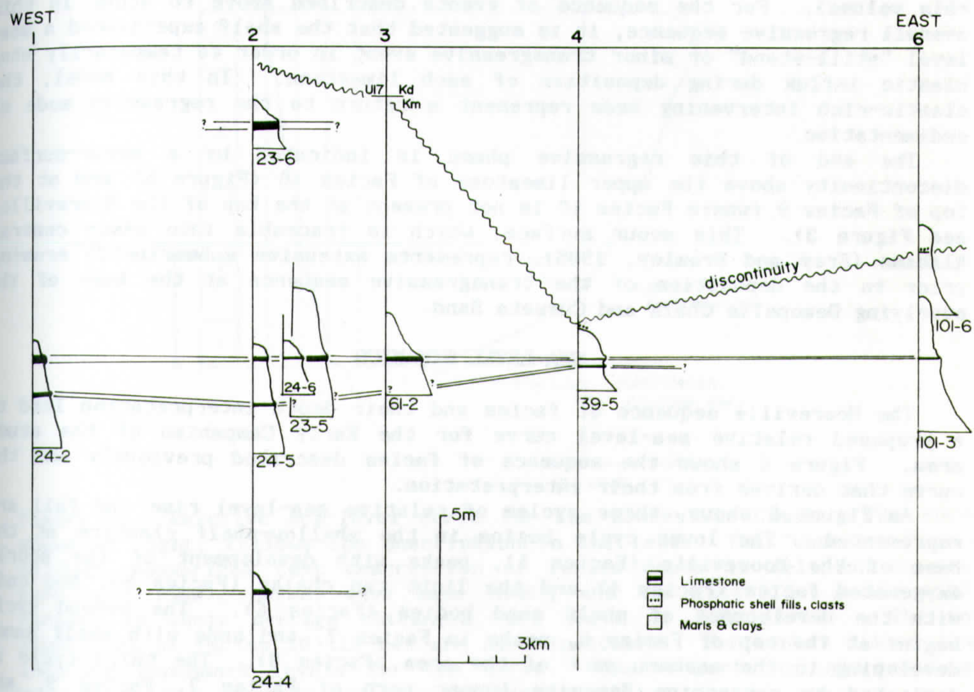

Figure 5. Correlation of measured sections showing relative stratigraphic position of the four limestone beds in Facies 10. Composite section numbers (1-4 and 6) are the same as in Figure 3. Numbers below each section are outcrop designations. The wavy line shows the discontinuity of the Mooreville (Km) and Demopolis (Kd) boundary (see text). Km/Kd boundary in well U17 is projected on section line 4.

Megafauna in this facies occur in the chalk, marl, and calcareous clay between limestones. The fauna are detrital *Exogyra* (in layers), infaunal bivalve impressions, *Hamulus*, and internal molds of *Baculites*, *Dentalium*, and various gastropods. *Thalassinoides*, some containing *Chondrites* in their internal fills, is moderately common in the chalk, marl, and calcareous clay of this facies.

Silt and mica content of the intervening marl increases from less than 10% near the base of the facies to 30% near the top. The limestone beds are without silt and have only a trace of mica.

Facies 10, like its lateral equivalent Facies 9, is interpreted as a regressive shelf deposit representing sedimentation well above storm wave base. The main evidence of depth in relation to storm wave base is the presence of detrital limestone clasts and detrital *Exogyra* beds. Abundant *Thalassinoides* in the limestone, marl, and calcareous clay suggests a general shallow shelf-sea setting (Ekdale and Bromley, 1984). The vertical trend of increased silt and mica content is evidence of the regressive nature of the interval.

The limestone beds of this facies formed in clear, relatively shallow water as indicated by the prolific benthic-algal remains (Bottjer, 1980) which would require much light to flourish. Further, these beds, composed of 90% calcium carbonate, became cemented (as a non-mineralized "hardground") at or near the sediment-water interface where storms could dislodge and comminute pieces of the layers.

Facies analysis of the coeval interval in the Blufftown Formation shows a regressive sequence of clastic facies (Skotnicki, 1985; Skotnicki and King, this volume). For the sequence of events described above to occur in this overall regressive sequence, it is suggested that the shelf experienced a sea-level "still-stand" or minor transgressive event in order to temporarily stop clastic influx during deposition of each limestone. In this model, the clastic-rich intervening beds represent a return to the regressive mode of sedimentation.

The end of this regressive phase is indicated by a scour-surface discontinuity above the upper limestone of Facies 10 (Figure 5) and at the top of Facies 9 (where Facies 10 is not present at the top of the Mooreville, see Figure 3). This scour surface, which is traceable into west- central Alabama (Frey and Bromley, 1985), represents extensive submarine(?) erosion prior to the deposition of the transgressive sequence at the base of the overlying Demopolis Chalk and Cusseta Sand.

SEA-LEVEL DYNAMICS

The Mooreville sequence of facies and their depth interpretation lead to a proposed relative sea-level curve for the Early Campanian of the study area. Figure 6 shows the sequence of facies described previously and the curve that derives from their interpretation.

As figure 6 shows, three cycles of relative sea-level rise and fall are represented. The lower cycle begins in the shallow-shelf clastics at the base of the Mooreville (Facies 1), peaks with development of the poorly oxygenated facies (Facies 4) and the light tan chalks (Facies 5), and ends with the development of shelf sand bodies (Facies 6). The second cycle begins at the top of Facies 6, peaks in Facies 7, and ends with shelf sands developing in the eastern part of the area (Facies 8). The third cycle is dominated by regressive deposits (upper part of Facies 7, Facies 9, and Facies 10), including the stepwise regressive sequence suggested by the occurrence of the algal-calcisphere limestones. A scour-surface discontinuity (Monroe, 1941; Frey and Bromley, 1985) marks the end of the cycle (a faunal hiatus at this level is indicated by some workers, e.g., Smith and Mancini, 1983; Carter and Wheeler, 1983).

DISCUSSION

Information on Campanian cyclic sea-level changes on other parts of the globe fall into two categories; these categories correspond to the second-order (10 to 80 Ma) and third-order (1 to 10 Ma) sea-level cycles of Vail and others (1977). Studies describing second-order sea-level curves do not show enough detail for good comparison with figure 6 in this report. Figure 6 shows those interpreted changes only for the Early Campanian, a span of 5 Ma (78 to 83 according to Harland and others, 1982). It is interesting to note that the second-order sea-level curves for the Campanian of southern England, the Arabian Penninsula, and western Africa (Seigle and Baker, 1984) show the sea-level rise after an end-Santonian regression as noted in this study (Figure 6). A similar pattern occurs in the Santonian-Campanian curve for northern Europe (Hancock and Kauffman, 1979).

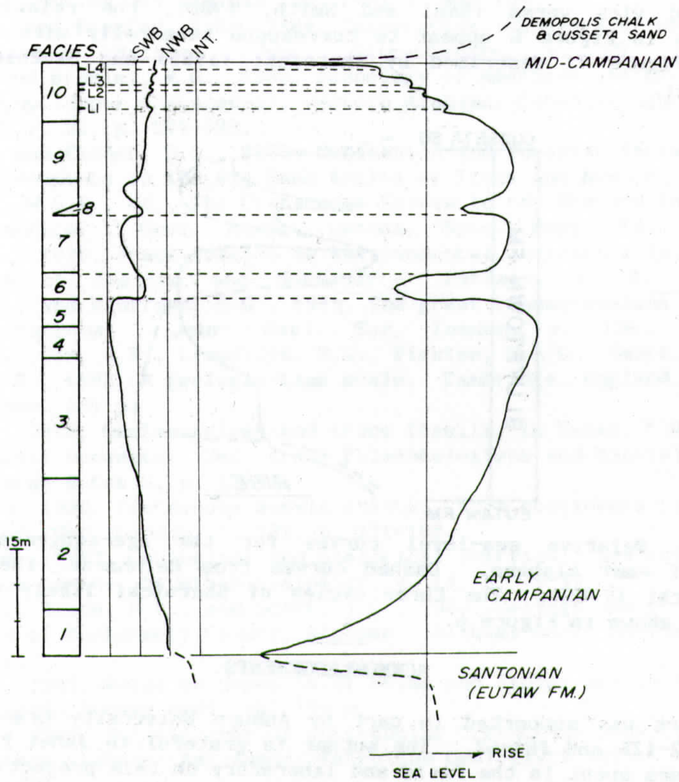

Figure 6. Relative sea-level curve for the Mooreville Chalk. The curve, shown at right, follows the description in the text. The "wave-base" curve at left summarizes text description of facies depth in relation to storm wave base (SWB), normal wave base (NWB), and the intertidal realm (INT). Facies, in their average thickness, are shown at left (1-10). The four limestones of Facies 10 (L1-L4) are also indicated. Horizontal dashed lines indicate benchmark levels for the curve. Dashed parts of sea-level and wave-base curve are inferred from Reinhardt (1980).

Almost no information on the detail of Cretaceous third-order cycles is available in the literature (see Vail and others, 1977; Harland and others, 1982; Ziegler, 1982; Seigle and Baker, 1984). Only in the Western Interior Seaway is sufficient data available to resolve shifts in sea level corresponding to third-order cycles. Kauffman (1977, his Figure 7) presents these data and describes a single transgressive-regressive cycle in the Lower Campanian of the Western Interior.

Because three cycles are evident in the study area (Figure 6), a history of relative sea-level change more complex than in the Western Interior is suggested for the northern Gulf rim. This may indicate local rather than eustatic controls on the relative sea-level changes of Figure 6.

Reinhardt (1980) has suggested two cycles of relative sea-level rise and fall (Figure 7) in the age-equivalent Blufftown Formation clastics of the Chattahoochee River valley area (70 km east of the study area). Skotnicki (1985), studying additional outcrops, has suggested that Reinhardt's upper cycle is divisible into two distinct cycles, making a total of three cycles of relative sea-level change in the Blufftown Formation (Figure 7). While exact correlations are not possible because the biostratigraphy of the chalks in western and central Alabama has not yet been carried into the Blufftown clastics and vice versa (Sohl and Smith, 1980), the relative sea-level cycles shown in Figure 6 appear to correspond temporally with those of the Blufftown (Figure 7) described by Skotnicki (1985) and Skotnicki and King (this volume).

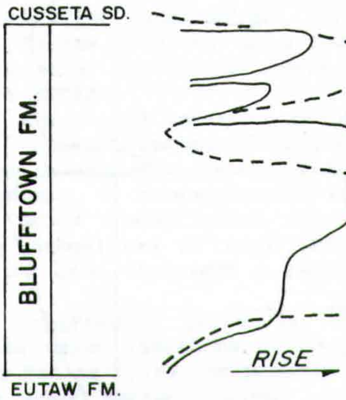

Figure 7. Relative sea-level curves for the age-equivalent Blufftown Formation of east Alabama. Dashed curves from Reinhardt (1980); solid is from Skotnicki (1985). The three cycles of Skotnicki likely correspond to the changes shown in Figure 6.

ACKNOWLEDGEMENTS

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DISTRIBUTION OF ARK SHELLS (BIVALVIA: *ANADARA*), CABRETTA ISLAND BEACH, GEORGIA[1]

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ABSTRACT

Distributions of *Anadara brasiliana* and *A. ovalis* valves, the predominant arks on Cabretta Island Beach, are governed by several interrelated variables: (1) proximity of bivalve life sites, (2) valve symmetry and ornamentation, which affect hydraulic behavior, (3) distances and processes of shoreward transport, (4) beach configurations, (5) interactions between waves and tidal currents, and their influences upon valve transport, and (6) prevailing patterns of longshore and/or ebb-current drift. Because of slight differences in shell morphology, however, detrital valves of these two species respond differently within the same hydrologic regime. In a general way, right valves of *A. brasiliana* tend to accompany left valves of *A. ovalis*, and vice versa.

INTRODUCTION

Recent studies have stressed the importance of shell accumulations along Georgia beaches (Dörjes and others, 1986; Frey and Henderson, 1987) and on the contiguous shelf (Henderson and Frey, 1986; Frey and Howard, 1986). However, dominant mollusks in these detrital assemblages consist of small bivalves such as *Abra aequalis*, *Donax variabilis*, *Mulinia lateralis*, and *Tellina versicolor*, hence these species have received major emphasis during previous investigations. Present work, in contrast, is concerned mainly with the characteristics of selected larger species, of which the arks, *Anadara*, are typical (Abbott, 1974).

Beach shell accumulations may be evaluated with respect to depositional dip (perpendicular to shore) or depositional strike (parallel to shore) (Dörjes and others, 1986). As reiterated by present work, the wrack line (previous high-tide line) is a convenient, replicable datum for establishing shell displacements along depositional strike. There, valve distributions are mainly a function of alongshore hydrography and local geomorphic features.

Principal objectives of this study, conducted on Cabretta Island Beach (Figure 1), include the characterization not only of overall size trends and wrack-line distributions of ark shells but also the wave-swash sorting and longshore transport of left and right valves.

The results should be useful in taphonomic interpretations of ancient shell accumulations (Kidwell and Jablonski, 1983), especially those observed in beach sequences (McCarthy, 1977; Howard and Scott, 1983). Particularly important in this regard is the concept of shell budgets and shell provenance, in addition to overall distributional patterns within essentially uniform regimes (Boucot, 1953; Frey and Henderson, 1987).

METHODS AND SETTING

Georgia beach sediments consist primarily of fine-grained, well-sorted, angular quartz sand (Howard and Frey, 1980). Beach slopes are 1^{o} to 2^{o}. Mean tidal range is about 2.4 m, yet mean nonstorm wave amplitude is only about 0.25 m (Kuroda and Marland, 1973). Thus, these beaches are part of a mesotidal, low- to moderate-energy system (Hubbard and others, 1979; Oertel, 1985).

[1]Contribution number 562, University of Georgia Marine Institute, Sapelo Island, Georgia

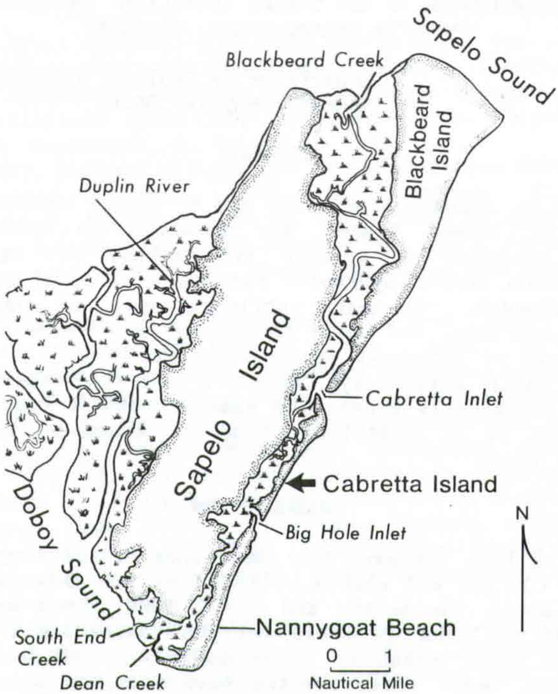

Figure 1. Index map of study area: Cabretta Island, Georgia.

Beach shell sampling was conducted along the length of Cabretta Island (Figure 1). Physiographically, this island is contiguous with Nannygoat Beach, where most previous work was undertaken. However, Cabretta Beach is slightly broader and gentler of slope than Nannygoat Beach (Frey and Mayou, 1971), and its northern part is shielded by a predominantly subtidal shoal-spit system extending southward from Blackbeard Island. A decade or so ago Cabretta Beach suffered severe erosion (Howard and others, 1972, p. 7) but has essentially recovered (Frey and Basan, 1981, Pl. 9, figs. 43-45). Present samples are judged to be representative of normal, broad, low-energy beaches of this type.

Samples were obtained from the wrack line (landward extent of high-tide wave swash) during typical low-energy conditions of summer. Sample plots consisted of 11 linear sites along the wrack line (parallel to shore), each 1 m wide and 150 m long; with one exception these sites were spaced 150 m apart, from Big Hole Inlet through Cabretta Inlet (Figure 1). [One potential site was omitted (between Sites 6 and 7, Figure 2) because of local disturbances along the wrack line there.] Each whole, undamaged ark shell observed within these linear plots was collected, measured, and identified. Because all shells were disarticulated, individual valves were tallied as separate specimens.

CHARACTERISTIC ARK SHELLS

The predominant ark on Cabretta Beach, *Anadara brasiliana* (Figure 2A), is nearly spherical in shape; valves are highly inflated, and shell height is approximately the same as shell length. This ark is slightly inequivalved, the left valve overlapping the right valve toward the postero-ventral margin (Abbott, 1974). Valves are typified by beaded ribs (costae), which are somewhat more prominent on the larger left valve than on the smaller right valve; posterior ribs on the right valve may be rather smooth. All ribs are

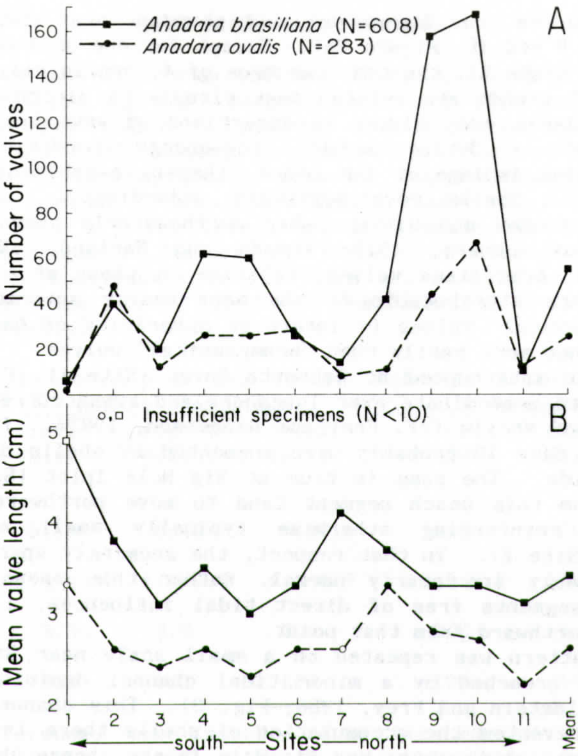

Figure 2. Distribution of ark shells, Cabretta Beach. A, Total valve abundance. B, Mean size.

rounded in cross section.

Next most abundant is A. *ovalis*, which is smaller than A. *brasiliana* (Figure 2B). A. *ovalis* is equivalved and somewhat oval in shape (Abbott, 1974), but shell height is slightly less than shell length. Both valves bear prominent, unornamented ribs that are rather rectangular in transverse section. Whereas umbones of A. *brasiliana* are nearly orthogyrate and centralized, those of A. *ovalis* are prosogyrate and displaced anteriorly.

These collective differences in shell dimensions and ornamentation cause corresponding differences in the hydraulic properties of A. *brasiliana* and A. *ovalis* valves, both of which are moderately common in local intertidal and subtidal shell assemblages. [Live individuals of both species bear a periostracum, which probably modifies the hydraulic behavior of freshly disarticulated valves (Bottjer and Carter, 1980); however, the periostracum was missing from all detrital valves evaluated here.]

The only other arks observed within sample plots on Cabretta Beach were A. *transversa* (1 valve) and *Noetia ponderosa* (5 valves). These species (Abbott, 1974) are comparably rare in other local shell assemblages (Dörjes and others, 1986; Henderson and Frey, 1986).

WRACK-LINE DISTRIBUTION OF ARK SHELLS

During normal low-energy conditions of summer, large mollusk valves are comparatively rare on the broad, gently sloping foreshore of Cabretta Island. Instead of lodging there, most such shells are transported rather rapidly from the shoreface directly to the wrack line. Shell accumulations observed along this wrack line thus are representative of valves newly introduced to the overall beach.

Valves of *Anadara brasiliana* were most abundant near the northern end of the beach (Sites 9 and 10, Figure 2A). Except for one area near the southern end of the beach (Site 2), the same was true of *A. ovalis* valves.

These overall trends are related most closely to directions of longshore drift which, in turn, are related to directions of wave incidence (Frey and Henderson, 1987). During normal low-energy conditions of summer, southeasterly waves impinge on the beach, thereby establishing a pattern of northward drift. Shells move northward accordingly. [The pattern is reversed during autumn and winter, when northeasterly waves impinge on the beach (Howard and others, 1972; Kuroda and Marland, 1973).] Increased proportions of *A. brasiliana* valves, relative to those of *A. ovalis* (Figure 2A), evidently are attributable to the more nearly spherical shape of *A. brasiliana*; "spherical" valves (= *intercept sphericity* of Menard and Boucot, 1951) are entrained more easily than "nonspherical" valves.

This trend is interrupted at Cabretta Inlet (Site 11, Figure 2A), where ebb-tidal currents predominate over longshore and swash currents, and tend to clear the beach of shells (cf. Frey and Henderson, 1987). Indeed, the large accumulations at Site 10 probably were augmented by shells derived from Site 11 during ebb tide. The same is true at Big Hole Inlet (Site 1); however, shells swept from this beach segment tend to move northward with longshore drift, thereby reinforcing otherwise typically small accumulations in adjacent areas (Site 2). In that respect, the seemingly sparse accumulations at Site 3 probably are nearly normal, rather than anomalously low, for southern beach segments free of direct tidal influences, i.e., most valves are in transit northward from that point.

The inlet pattern was repeated on a small scale near Site 7, where the beach ridge was breached by a minor tidal channel draining a small back-ridge lagoon (Goldstein and Frey, 1986, Fig. 3). This channel (ca. 15 m wide and 1 m deep) prevented the accumulation of shells there in a manner analogous to the regimes at Cabretta and Big Hole Inlets, hence the reduced number of valves at Sites 6 to 8. The remaining shells tended to be larger in mean size (Figure 2B), indicating a lag-residue effect and winnowing of smaller valves. (The same seems to have been true of valve sizes at Site 1; but tidal currents evidently were much stronger at Site 11 and removed most valves.)

Thus, except for irregularities in beach configurations and corresponding tidal effects, shell accumulations on Cabretta Beach became progressively greater from south to north, corresponding to prevailing directions of longshore drift during the summer regime. To that extent, the pattern is similar to ones discerned during previous work (Dörjes and others, 1986; Frey and Henderson, 1987).

However, previous results revealed minimal shell accumulations along beach segments that projected seaward, whereas Cabretta arks were most abundant along an arcuate, convex-seaward beach segment (Figures 1, 2A). Along open-ocean beaches, shell accumulations are related not only to longshore drift but also to patterns of wave refraction around headlands or within embayments (cf. Ross, 1982, Fig. 9-21); even on a small scale, therefore, beach protrusions tend to be swept clean of valves, and beach reentrants tend to be sediment traps for valves. The principal difference here is that, unlike "sinuous" open-ocean beach segments examined previously, the arcuate northern end of Cabretta Beach is protected by a sizeable spit extending southward from Blackbeard Island (not shown in Figure 1). South of the spit influence, the beach is essentially straight or only very gently arcuate.

An additional aspect of valve transport stressed by Kornicker and Armstrong (1959) is the effect of boreholes drilled by predatory gastropods; valves with holes are less buoyant and therefore tend to lag behind unbored valves. Although not investigated in detail on Cabretta Beach, this effect is not considered sufficient to mask major trends in valve distributions because both left and right valves are bored.

LEFT-RIGHT VALVE SORTING

In addition to overall trends outlined above, Cabretta ark shells exhibit considerable variation in the characteristics of left and right valves. Discrepancies in size (Table 1) indicate that unpaired valves, or unequal numbers of left and right valves, were delivered from the sea to the wrack line. The disparity is considerably less for *Anadara brasiliana* than for *A. ovalis*, evidently because *A. brasiliana* dwells closer to shore (Henderson and Frey, 1986, Appendix 1); its valves are in transit for shorter distances before reaching the beach. The ecologic range of *A. ovalis* overlaps that of *A. brasiliana*, yet individuals of *A. ovalis* are more broadly distributed in the nearshore-estuarine zone. Additional ramifications of this disparity are discussed subsequently.

Table 1. Density and size of ark shells, Cabretta Island Beach, Georgia.

	Anadara brasiliana			*Anadara ovalis*		
	left valves	right valves	total valves	left valves	right valves	total valves
Mean density (valves/m^2)	0.19	0.18	0.37	0.07	0.10	0.17
Range in valve length (cm)	1.4-6.7	1.4-6.6	1.4-6.7	0.9-4.9	1.3-5.5	0.9-5.5
Mean valve length (cm)	3.3	3.5	3.4	2.6	2.7	2.6
Number of observations	306	302	608	123	160	283

Further differences are imparted once the shells reach the beach. Although general distributional patterns remained similar to ones described previously (Figure 2A), individual valves behaved quite differently. For *A. brasiliana*, right valves predominated through most of the beach area but were subordinate to left valves at sites of maximum shell accumulation (Figure 3A). The opposite tended to be true of *A. ovalis* valves (Figure 3B).

Differences in local shell sizes were even more pronounced. For *A. brasiliana* (Figure 4A), and to some extent for *A. ovalis* (Figure 4B), the

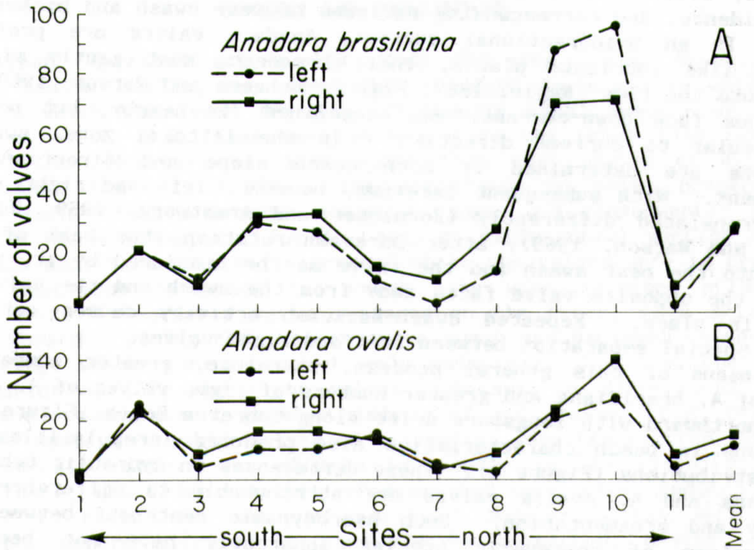

Figure 3. Distribution of left and right valves. A, *Anadara brasiliana*. B, *Anadara ovalis*.

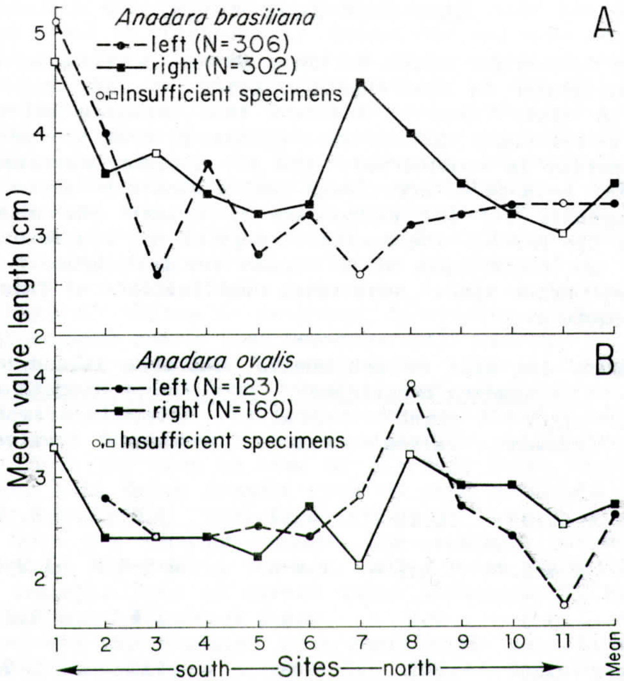

Figure 4. Mean size of left and right valves. A, *Anadara brasiliana*. B, *Anadara ovalis*.

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size distribution of left valves tended to be inversely related to that of right valves. The greatest discrepancies occurred at sites of minimal shell accumulation (and may be partly an artifact of small sample sizes).

These aspects of shell sorting, commonly termed "left-right phenomena" (see summaries by Lever and Thijssen, 1968; Lawrence, 1979; Frey and Henderson, 1987), are attributable mainly to valve symmetry, directions of wave incidence, and corresponding patterns of wave swash and backwash (Nagle, 1964). In an unidirectional current, *Anadara* valves are preferentially oriented like imbricate plates, their broadest, most gently sloping edge facing into the flow (Nagle, 1967, Fig. 2; Behrens and Watson, 1969, Fig. 3); beaks thus face down-current, and hingelines lie nearly, but not exactly, perpendicular to current direction. In the littoral zone, swash-current directions are determined by both beach slope and directions of wave impingement. With subsequent backwash, however, left and right valves tend to be translated differently (Kornicker and Armstrong, 1959; Nagle, 1964; Behrens and Watson, 1969); after backwash rotation, the beak of one valve faces into the next swash and the valve may be displaced by it, whereas the beak of the opposite valve faces away from the swash and the valve tends to remain in place. Repeated swash-backwash activity causes corresponingly greater spatial separation between left and right valves.

By means of this general process, therefore, greater numbers of left valves of *A. brasiliana* and greater numbers of right valves of *A. ovalis* were driven northward with longshore drift along Cabretta Beach (Figure 5). Local variations in beach characteristics also produced irregularities in valve-size distributions (Figure 4). These differences in hydraulic behavior of *A. brasiliana* and *A. ovalis* valves are attributable to differences in shell symmetry and ornamentation. Such hydrodynamic contrasts between left and right valves of congeneric species apparently have not been reported previously.

More difficult to evaluate is the amount of left-right valve sorting that

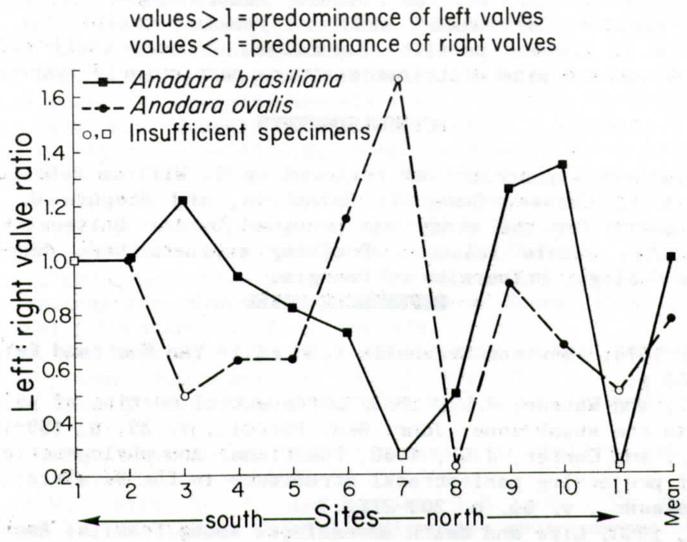

Figure 5. Ratios of left to right valves. Ratios for *Anadara brasiliana* and *A. ovalis* tend to be inversely related.

occurred subaqueously, before the valves reached the beach. Although Behrens and Watson (1969) considered such sorting to be strictly an intertidal phenomenon, the disparity in numbers and sizes of left and right valves of *A. ovalis* (Table 1, Figure 5) is evidence to the contrary. Subtidal left-right valve sorting also is well documented for *Donax variabilis* and *Abra aequalis* in Sapelo Island waters (Frey and Henderson, 1987), even though the mechanism involved remains poorly understood. One can only speculate that, in general, the observed deficit in left valves of *A. ovalis* is due to differential transport by currents originating in the shoreface (cf. Niedoroda and others, 1985).

SUMMARY AND CONCLUSIONS

Predominant detrital ark shells along Cabretta Island Beach, *Anadara brasiliana* and *A. ovalis*, are comparably abundant in nearshore shelf assemblages from which they were derived. Virtually equal numbers of left and right valves of *A. brasiliana* were delivered to the wrack line (left:right ratio = 1.01). However, far more right than left valves of *A. ovalis* migrated there (ratio = 0.77). The difference in valve proportions is attributed mainly to the broader ecologic range, and consequently increased distances of valve transport, for *A. ovalis*; differential sorting of valves in the beach shoreface evidently occurred while the valves were in transit.

Valve distributions along the wrack line were governed primarily by longshore drift. Southerly waves impinged on the beach, thus driving most valves northward. During the process, wave swash and backwash caused further shell sorting, which influenced not only valve-size distributions but also the separation of left and right valves. Because of differences in valve symmetry and ornamentation, greater numbers of left valves of *A. brasiliana*, and greater numbers of right valves of *A. ovalis*, were transported northward.

This overall pattern was altered profoundly in the vicinity of tidal inlets. There, ebb-tidal currents predominated over wave- induced currents and tended to clear the beach of shells. Many evidently were returned seaward; but others migrated to, and thereby augmented, nearby wrack-line accumulations otherwise generated by longshore drift.

Finally, even though wrack-line shell accumulations stand little chance for direct preservation in the fossil record, general processes and parameters revealed by such studies remain useful in taphonomic interpretations of ancient paralic assemblages. Local shell budgets (left-right valve ratios and size distributions) are particularly instructive.

ACKNOWLEDGMENTS

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AN ALLOCHTHONOUS PRESERVED WOODGROUND IN THE UPPER CRETACEOUS EUTAW FORMATION IN MISSISSIPPI

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ABSTRACT

A new occurrence of the little known Ichnogenus, *Teredolites* is described from the Upper Cretaceous, Eutaw Formation in northeast Mississippi. Clusters of *Teredolites clavatus* and a smaller trace fossil of indeterminate origin are found in an allochthonous, limonitized woodground. The woodground occurs in the sparingly fossiliferous lower Eutaw sands. A chemotactic response to adjacent domiciles is suggested to be responsible for the presence of severely distorted specimens. Although the xenoglyph and evidence of the original xylic substrate are preserved, there is no evidence of a bioglyph or the causative organism.

INTRODUCTION

This paper presents the first occurrence of a preserved, bored woodground in the Upper Cretaceous, Eutaw Formation in Mississippi. The woodground contains the ichnogenus *Teredolites* and is described from a single railroad cut exposure (Figure 1) near Quincy, Monroe County, northeast Mississippi (SW 1/4, SW 1/4, Sec 29, T13S, R17W, Greenwood Springs 7.5 minute Quadrangle). The Eutaw Formation (Figure 2) is defined as a fine to medium grained, micaceous, glauconitic sand. The formation was originally defined by Hilgard (1860) for all sediments lying between the fossiliferous Tombigbee Sand and the Palaeozoic rocks. The Eutaw is, however, presently restricted to the glauconitic sands that occur below the marls of the Mooreville Formation and which overlie the McShan Formation (Monroe and others, 1946). The sparingly fossiliferous, lower Eutaw sands are interpreted to be a shallow marine unit (Smith and Johnson, 1887; Stephenson, 1914; Russell, and others, 1983). The Tombigbee Sand Member of the upper Eutaw Formation is a finer-grained richly fossiliferous lithology of deeper water origin. The only macrofossils thus far described from the lower Eutaw Sands in Mississippi are the shrimp burrow *Ophiomorpha nodosa* Lundgren (=*Halymenites major* Lesquereux), and plant remains (Berry, 1919; Stephenson and Monroe, 1940; Vestal, 1943, 1947). An upper Cretaceous age can be assigned to the Eutaw Formation based on the following evidence. Palynological analysis of the basal Eutaw indicates a possible late Santonian or early Campanian age; the nannofossil and planktonic foraminiferal biostratigraphy of the Tombigbee Sand Member is indicative of an early Campanian age and palynological analysis of the McShan Formation suggests a late Coniacian to early Santonian age (Russell, and others, 1983).

LITHOLOGIC ASSOCIATION

The woodground occurs at a single horizon in a six meter railway cutting (Figure 3, near Quincy, Monroe County, Mississippi.

At the base of the section, inter-laminated blue/grey clays and orange sands of the McShan Formation are exposed. Immediately above the McShan unit is a discontinuous, clay clast (>5 cm) conglomerate which is overlain by a fine to medium grained glauconitic sand of the Eutaw Formation. The sand exhibits both trough and herring-bone cross bedding. Low angle foresets range from 30-80 mm thick are often accentuated by clay drapes.

Incised into this unit is a channel containing a similar sand with larger foresets. Near the base of the channel is a large population of clay clasts. On the west side of the channel are rafts of limonitised, bored wood. The variable orientation of the borings in the outcrop and nature of the section suggest that the woodground is allochthonous.

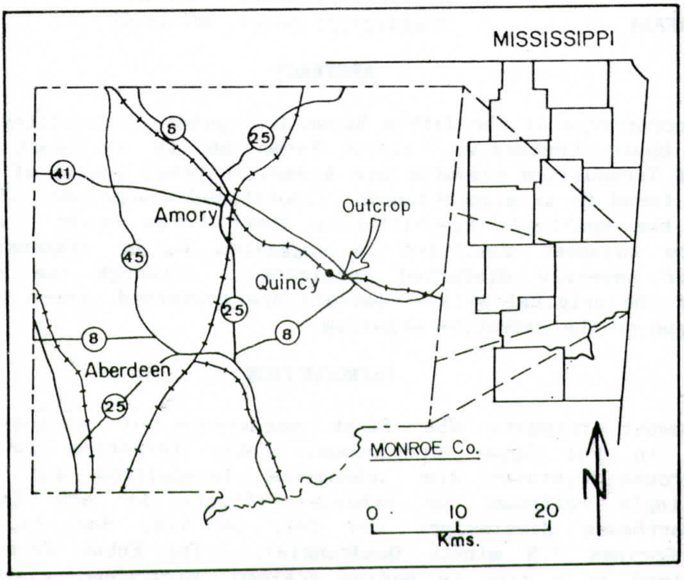

Figure 1. Location Map of *Teredolites* Outcrop.

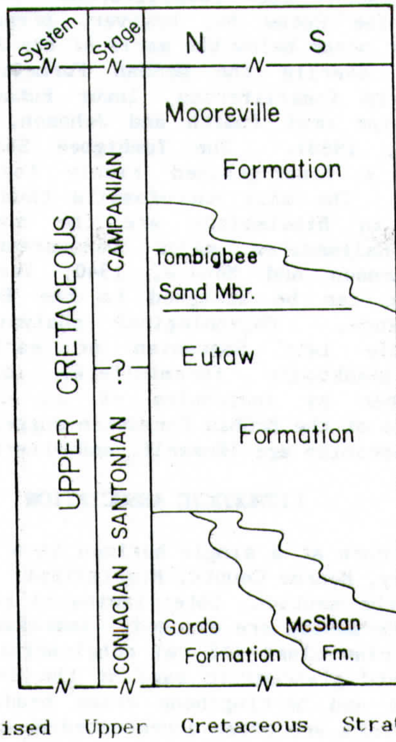

Figure 2. Generalised Upper Cretaceous Stratigraphy of Northeast Mississippi.

Above the channel is an ironstone layer and about 2 m of deeply weathered, red, medium grained, glauconitic sand. This lithology is "typical" of the basal Eutaw in this area.

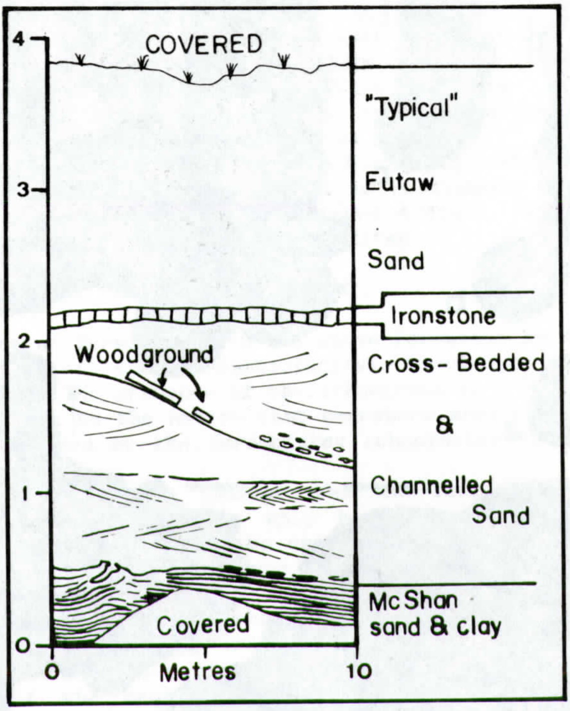

Figure 3. Simplified Diagram of Railway Cutting Outcrop.

DEPOSITIONAL ENVIRONMENTS

The interlaminated sands and clays of the McShan, which occur at the base of the section, are considered to be equivalent to Lithofacies 2 of Cook (1986) who suggests an estuarine origin for the McShan as a whole. Cook, (1986) further suggests that Lithofacies 2 was deposited in the estuarine channels.

The lower Eutaw Formation is a transgressive sand, which, at this locality shows evidence of having reworked the McShan, and shows both tidal influences and channelling. It is therefore considered to be a shallow, nearshore high energy deposit which may represent the infilling of the McShan estuarine channel. In its updip facies, the Eutaw truncates the McShan, in northeast Mississippi, and directly overlies the fluvial facies of the Gordo Formation.

It is likely therefore, that the woodground was transported into a channel which was cut into the earliest Eutaw sands during the transgression of the Eutaw over the McShan.

ICHNOFOSSIL DESCRIPTION

The traces referable to *Teredolites clavatus* Leymerie, are found as variably clustered groups in limonitised wood. The traces are sand filled casts that possess a tubular, J-shaped morphology with a closed bulbous lower end. Specimens range from 3-4 cms in diameter and can be as much as 15 cms in length (Keady and Dewey, 1986). Although the tubes are subparallel and mostly perpendicular to the woodgrain, in some cases the tubes are severly

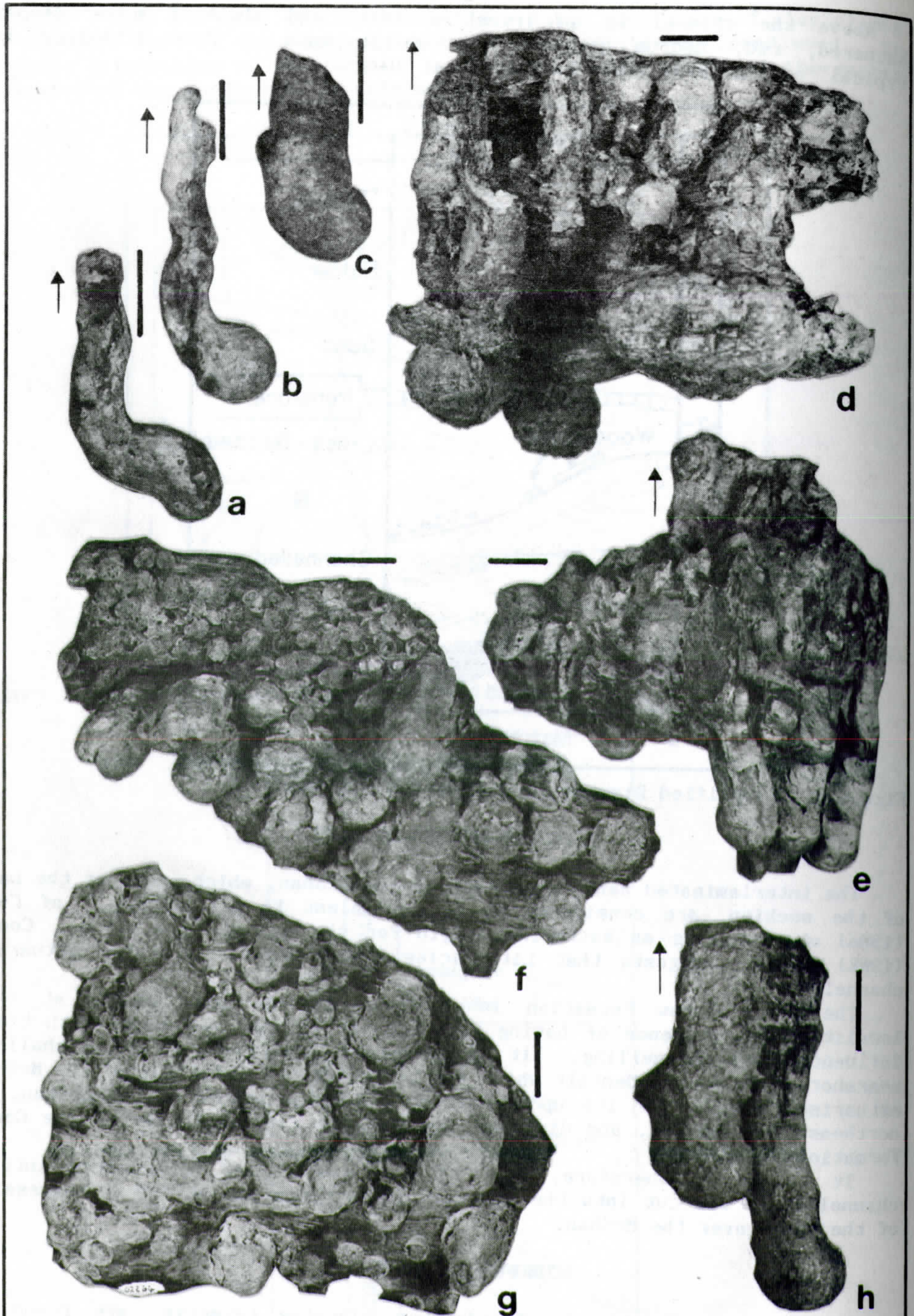

Figure 4. (a,b) Distorted *Teredolites* tubes in side view. (c) Distal end of *Teredolites* tube. (d,e,h) *Teredolites* clusters in side view. (f,g) *Teredolites* clusters in basal plan view. Note: Scale Bar = 2 cm., arrows indicate "up".

distorted. Since a chemotactic response would prevent the boring organism from drilling into an adjacent domicile, the distortion of tubes (Figure 4 a,b) can be accounted for by late stage colonisers of the crowded substrate.

Where preserved, the walls of the traces exhibit a linear sculpture imposed by the substrate. This feature has been called the xenoglyph by Bromley, and others (1984). The traces show no preserved markings by the causitive organism (bioglyph), probably because the traces were infilled with sand during the transportation and redeposition processes.

In regions of the woodground where the *Teredolites* specimens are widely spaced, the original grain of the wood can be observed.

A second, smaller type of cylindrical boring of indeterminate origin is also present. The borings are less than 1 cm in diameter and have a similar orientation to the *Teredolites* traces (Figure 4 f,g). The borings are not club shaped and cannot be referred to *Teredolites*.

DISCUSSION

The occurrence of *Teredolites* in the Eutaw Formation in Mississippi is an important addition to the palaeobiological knowledge of this sparcely fossiliferous unit. The presence of the ichnogenus in a xylic substrate also gives further support to the notion that nearshore shallow marine woodgrounds should be characterized by the *Teredolites* ichnofacies (Bromley and others, 1984).

Bromley and others (1984), indicate that the *Teredolites* trace is formed by the pholadid bivalve *Martesia* sp.; however, material from the Eutaw Formation contains no remains of the causative organism. It is clear from both the orientation of the woodground in the outcrop and the sedimentological relationships, that it was not bored in the final depositional environment. It is suggested that the woodground was bored prior to transportation, and that during transportation and redeposition the borings were filled with sand. Loss of the body fossils may have occurred during transportation (if the neck of the boring was wide enough) or during subsequent diagenesis and limonitisation.

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FACIES FAUNAS OF THE SALEM LIMESTONE (MISSISSIPPIAN) IN SOUTHERN INDIANA AND CENTRAL KENTUCKY

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ABSTRACT

The Somerset Shale Member of the Salem Limestone (Meramecian, Mississippian) contains a rich fauna distinct from the better known fauna of the Salem carbonate facies. The Somerset is a thin (0-15 m.) discontinuous shale to shaley limestone at the base of the Salem Limestone in central Kentucky and southern Indiana. The shale was deposited below normal wave base in linear troughs between grainstone shoals of the Salem Limestone. The muddy substrate supported locally a diverse benthic community that included brachiopods, crinoids, blastoids, rugose and tabulate corals, bryozoans and sponges.

Cluster analysis distinguished three distinct assemblages that correspond to three distinct facies in the lower Salem in southern Indiana and central Kentucky. The three facies are the open tidal embayment with shifting carbonate sand, the sand bar shoal with stable carbonate shoals separated by deeper water troughs and the Somerset Shale. The Somerset fauna lacks the diverse mollusks that are characteristic of the two assemblages in the carbonate facies in Indiana. Differences between the two assemblages in the carbonate facies are the result of the ability of species to survive on a shifting substrate. Differences among the three communities are attributable to differences in substrate type and stability and the presence of the other organisms.

INTRODUCTION

The Salem Limestone is part of a carbonate shelf that developed on top of the Lower Mississippian Borden Delta as clastic deposition waned. The Salem is composed principally of cross-bedded pack- and grainstones that were deposited in clear, shallow, well-agitated, tidally influenced marine water to the north and east of the Illinois basin (Sedimentation Seminar, 1966; Benson, 1976; Pryor and Sable, 1974).

The fauna from the carbonate facies of the Salem Limestone in Indiana has been well studied (Cumings and others, 1906; Hall, 1856, 1883; Whitfield, 1882; Miller, 1891, 1892). Cumings and others (1906) collected extensively from nine sites in Indiana, including the well-known Spergen Hill railroad cut, and described and illustrated over 180 invertebrate species, however, there are only a few brief descriptions of the Somerset fauna (Butts, 1922; Taylor, 1968; Feldman and others, 1982). The purpose of this study was to determine the composition and distribution of the Somerset fauna and compare it with the fauna of the Salem in Indiana.

The Somerset Shale Member is a thin shale at the base of the Salem Limestone (Mississippian, Meramecian) extending from south-central Kentucky to southern Indiana (Nicoll and Rexroad, 1975; Stockdale, 1939). Although Butts (1922) designated the type section of the Somerset Shale in Somerset, Kentucky, he noted that the best exposures are in glades on hilltops two miles west of Colesburg (localities 5 and 6). The nearby section at Tunnel Hill (locality 2, Figure 1) was used as a reference section in this study because it is well-exposed and was illustrated and described by Butts (1922).

The Somerset is primarily a dark gray dolomitic or calcareous shale with interbedded fossiliferous packstones. Where there is little carbonate, the shale weathers and erodes quickly leaving a residue of fossils. The shale grades from almost unfossiliferous to highly fossiliferous. Geodes are common at some localities and are usually less than five centimeters in diameter.

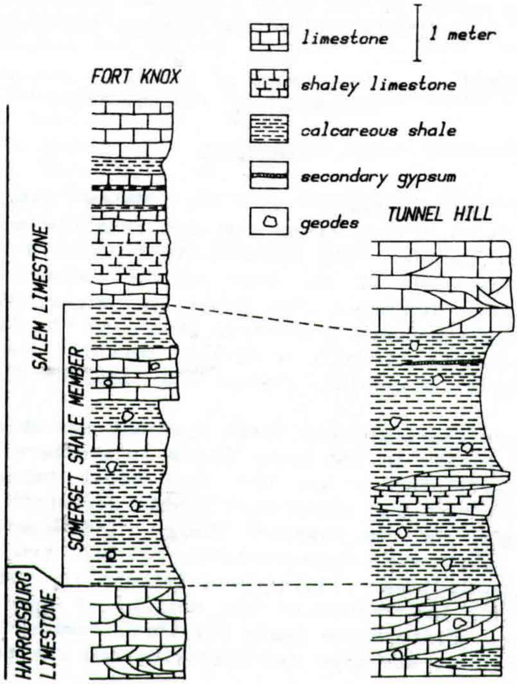

Figure 1. Measured sections of the Somerset Shale at localities 1 (Fort Knox) and 3 (Tunnel Hill).

The lower contact with carbonate rocks is sharp and marks the Salem-Harrodsburg boundary (Figure 1). The upper contact of the shale is either sharp or gradational. Limestone interbeds may become more common upwards grading into massive cross-bedded carbonate sands typical of the Salem Limestone.

METHODS OF INVESTIGATION

Eighteen exposures of the Somerset Shale in southern-most Indiana and central Kentucky were measured, described and collected (Figure 2). Surface collections were used because 1 to 3 kg bulk rock samples yielded small numbers of fossils. All available complete fossils were collected from small exposures. At larger exposures only parts of the exposure were collected. The surface residue of areas about 1 square meter were collected so that relative abundances of macrofossils could be assessed.

SEDIMENTOLOGY OF THE SOMERSET

In south-central Kentucky the stratigraphic equivalent of the Somerset Shale is the Science Hill Sandstone Member of the Warsaw Formation (Figure 3). Based on an isopach map of the Science Hill Sandstone (Figure 3) and sedimentological data, Lewis and Taylor (1979) concluded that the Science Hill Sandstone was deposited as a constructive, lobate delta in a shallow marine environment near the eastern edge of the Illinois Basin. They also suggested that the Somerset Shale was deposited as prodeltaic mud of the Science Hill delta. The distribution of the shale, mostly north of the delta, indicates that the currents flowed generally northward.

The Somerset Shale is preserved as discontinuous lenses between carbonate beds. U.S.G.S. workers (e.g. Moore, 1970: Taylor and others, 1968; Taylor,

1965) mapped the shale (they did not name it) in facies relationship with carbonates of the Salem or Salem-Warsaw Formation. Many of the shale lenses are elongate with a northeast-southwest axis suggesting that the shale was deposited in linear troughs between elongate carbonate sand bars.

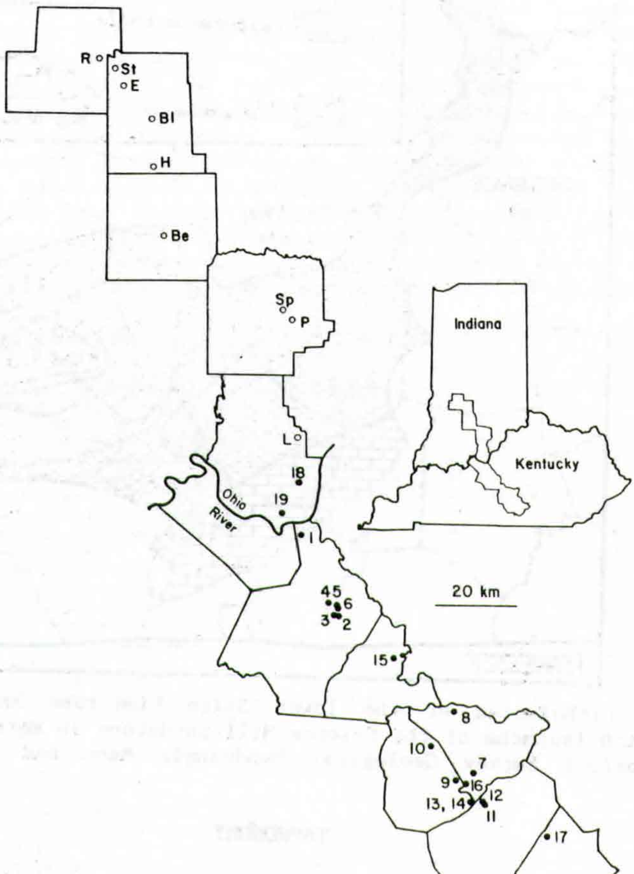

Figure 2. Map of localities used in this study. Open circles represent the limestone localities collected by Cumings and others (1906) and closed circles represent shale localities collected for this study. R, Ramona; St, Stinesville; E, Elletsville; Bl, Bloomington; H, Harrodsburg; Be, Bedford; Sp, Spergen Hill; P, Paynter's Hill; L, Lanesville.

Mud was deposited below normal wave base but above storm wave base. The influence of storms is indicated by common shell beds and by two local erosional unconformities (localities 7 and 12). Currents must have been considerable to produce scour surfaces with over a meter of relief. Sessile benthic organisms, such as corals and brachiopods, indicate that mud deposition was slow enough for these animals to become established and live to maturity. The presence of sponges that typically could not tolerate turbidity (as demonstrated by Rigby and Ausich, 1981) indicates that the water was clear most of the time. The abundance of filter feeders indicates that plankton were abundant. When terrigenous deposition ceased, carbonate sediments gradually spread over the mud.

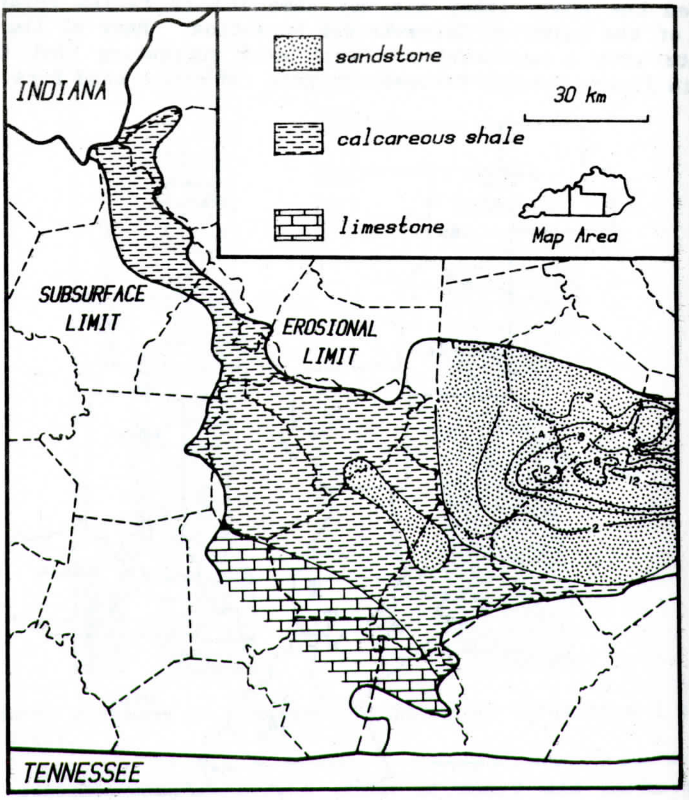

Figure 3. Lithofacies of the lower Salem Limestone and equivalents in Kentucky with isopachs of the Science Hill Sandstone in meters. Compiled from U.S. Geological Survey Geological Quadrangle Maps and Lewis and Taylor (1979).

TAPHONOMY

Almost all fossils in the Somerset Shale were crushed during compaction of the shale but otherwise are well-preserved. Only fossils from shell hash layers show signs of abrasion. At some localities brachiopods and bivalves are articulated and some delicate bryozoan fronds of *Cystodictya lineata* Ulrich over 6 cm long are preserved. Locality 7 yielded articulated specimens of the echinoid *Archaeocidaris norwoodi* Hall. The abundance of isolated echinoderm (mostly crinoid) plates indicates that half or more of the echinoderm calyces were disarticulated. Although most of the disartculation probably occurred shortly after death while the organisms were lying on the seafloor (Liddell, 1975; Meyer, 1971), burrowing organisms also helped distribute echinoderm ossicles. Trace fossils, especially *Helminthoidea* and to a lesser extent *Chondrites*, are abundant throughout the fossiliferous parts of the Somerset.

The overall nature of fossil preservation suggests that, except for shell hash layers, the fossils were not transported any significant distance. Thus the assemblage may be considered the fossilized portion of a paleocommunity.

PALEOECOLOGY

The Somerset Shale Community

Most localities from which a large number of specimens were collected show a moderate degree of similarity in faunal composition. Collections from localities 5-7, 11, 13, 14 and 16-19 are judged to represent the preserved assemblage of the dominant high diversity community in the Somerset Shale (Table 1).

All the common species are suspension feeders and are classified by tropic group in Table 2. Interspecific competition was minimized by niche partitioning through vertical tiering and differences in food particle sizes. Such partitioning is similar to the model of crinoid niche differentiation proposed by Ausich (1980). The lowest tier, which includes all brachiopods and the coral *Hapsiphyllum cassedayi* (Milne-Edwards), filtered water just above the sediment-water interface. The coral probably fed on larger food particles than did the brachiopods and so did not compete with them for food. The intermediate tier consists of organisms that filtered water from 2 cm to about 15 cm above the sediment-water interface, and includes blastoids, *Dichocrinus simplex* Shumard and other small (mostly rare) crinoids, all the bryozoans except those encrusting brachiopods and rugose corals, and probably the tabulate corals. The high level suspension feeders fed more than 15 cm

Table 1. Species occurrence data for the Somerset Shale [1,2].

Species	Localities													
	4	5	6	7	8	10	11	12	13	14	16	17	18	19
Brachiopods														
Crania? sp.	+													
Cleiothyridina hirsuta (Hall)	+	+	+	+		+	+		+	+		+	+	+
Camerotoechia mutata (Hall)		+	+	+			+	+			+	+	+	+
Composita trinuclea (Hall)	+	+	+	+		+	+	+	+	+	+	+	+	+
Dimegalasma sp.		+	+	+	+		+	+	+	+	+	+	+	
Echinoconchus biseriatus (Hall)													+	
Eumetria verneuliana (Hall)		+	+	+	+		+	+	+		+	+	+	+
Girtyella? sp.	+	+	+	+		+	+		+	+	+		+	+
Orthotetes sp.		+	+			+						+		
Perditocardinia dubia (Hall)				+					+				+	+
Reticulariina salemensis (Weller)	+		+			+			+	+		+	+	+
Setigerites setiger (Hall)		+	+	+		+	+	+	+	+	+	+	+	+
Spirifer bifurcatus Hall	+	+	+	+		+	+	+			+	+	+	+
S. Subaequalis Hall				+										
S. tenuicostatus Hall	+	+	+	+	+		+		+			+		
S. washingtonensis Weller	+	+	+	+	+		+		+	+	+			
Torynifer pseudolineatus (Hall)	+	+	+	+					+	+		+	+	+
Bryozoans														
Cystodictya lineata Ulrich		+	+	+	+	+	+	+	+	+	+	+	+	+
Fenestella spp.	+	+	+	+	+	+	+	+	+	+		+	+	+
Fenestralia sp.													+	
Fistulipora spergenensis Rominger	+	+	+	+		+	+		+	+		+	+	+
Glyptopora sp.									+	+			+	
Hemitrypa spp.	?	+	+	+	+	+	+	+	+	+		+	+	+
Polypora spp.	+	+	+			+	+			+		+	+	+
Tabulipora sp.		+	+	+		+	+		+	+		+	+	
Worthenopora sp.		+											+	+
Corals														
Cladochonus beecheri (Grabau)		+	+	+										
Hapsiphyllum cassedayi (Milne-Edwards)	+	+	+	+	+	+	+	+	+	+	+	+	+	+
Palaeacis cuniformis (Haime)			+											
Syringopora monroensis (Beede)	+	+					+					+	+	
Crinoids														
Batocrinus calyculus (Hall)				+		+			+			+		
Batocrinus icosodactylus Casseday			+										+	
Batocrinus spergenesis Miller			+	+		+								
Barycrinus sp. aff. *B. asteriscus* Van Sant	?		+				+		+			+	+	
Camptocrinus sp.	?		+				+		+			+		
Cyathocrinites multibrachiatus (Lyon & Casseday)				+										
C. parvibrachiatus (Hall)			+						+					
Dichocrinus simplex Shumard		+	+	+			+		+	+		+	+	+
Dizygocrinus whitei (Wachsmuth & Springer)			+							+	+			
Dizygocrinus sp. A			+											
Eretmocrinus sp.			+	+			+							
Forbesiocrinus multibrachiatus Lyon & Casseday	?	?	?	?		?		?	?	+		?		?
Platycrinites bonoensis (White)		+	+	+			+		+			+	+	+
Synbathocrinus swallovi Hall			+						+			+		+

Table 1. Continued.

Species	4	5	6	7	8	10	11	12	13	14	16	17	18	19
							Localities							
Blastoids														
Diploblastus sp.			+											
Metablastus bipyramidalis (Hall)												+		
Metablastus sp. aff. *M. wortheni* (Hall)			+						?			?		
Pentremites conoideus Hall		+	+	+			+		+	+		+	+	+
Sponges														
Belemnospongia parmula Rigby, Keys & Horowitz				+					+		+	+	+	
Haplistion armstrongi Young & Young									+					
Molluscs														
Bellerophon (Bellerophon) sp.												+		
Cypricardinia sp.			+											
Euconospira conula (Hall)						+			+					
Platyceras (Orthonychia) acutirostre (Hall)			+			+			+	+		+		
Mayalinid									+					
Gastropod A									+					
Miscellaneous														
Archaeocidaris norwoodi Hall				+										
Griffithides bufo Meek & Worthen									+				+	
Paraconularia sp.													+	

1 Macrofossils were not found at localities that are not listed.
2 + present; ? questionably identified parts.

Table 2. Trophic classification of suspension feeding fauna from the Somerset shale.

Tropic group	Taxa	Most Common species
Low level suspension feeders, large food particle size	rugose corals, tabulate corals?	*Hapsiphyllum cassedayi*
low level suspension feeders, small food particle size	brachiopods, some encrusting bryozoans *Spirorbis*	*Cleiothyridina hirsuta* *Composita trinuclea*
intermediate level suspension feeders	fenestrate, ribbon-like and ramose bryozoans, *Fenestella* spp., blastoids, some crinoids, tabulate corals and *Spirorbis*	*Dichocrinus simplex* *Pentremites conoideus*
high level suspension feeders	all other crinoids	all are rare

above the sediment-water interface and include all other crinoids. Competition for food between crinoid species can also lead to niche partitioning based on food particle size and height above the substrate (Ausich, 1980). However, in the Somerset assemblage all crinoids except *D. simplex* are so rare that interspecific competition was probably not a major factor.

Species of the lowest tier display a variety of adaptations for life on a soft substrate. The most common brachiopids, *Cleiothyridina hirsuta* (Hall) and *Composita trinuclea* (Hall), had functional pedicles that were tethered to shell debris or anchored in the mud. Some low and broad species of brachiopods used the snow-shoe effect and others had spines. Adult *Orthotetes* sp. rested upon its large triangular interarea. About half of the *Orthotetes* sp. specimens collected were compressed anteroir-posteriorly indicating burial in life position.

The rugose coral *Hapsiphyllum cassedayi*, which is the most common solitary invertebrate other than the brachiopods, has a spinose epitheca that helped to stabilize it in the mud. A few well-preserved specimens have coiled protocoralla indicating that as juveniles they were attached (Ausich and Smith, 1982).

Bryozoans are the most abundant fossils from the second tier at most localities. Thickets of fenestrate bryozoans, among the most common fossils in the Somerset, may have baffled bottom currents and served to trap mud. Encrusting and ramose bryozoans also are common.

The sponge *Belemnospongia parmula* Rigby, Keys and Horowitz is present at most localities with high diversity and is common, along with *Haplistion armstrongi* Young and Young, at locality 13. Rigby and Ausich (1981) found that the abundance of sponges, including *Belemnospongia* and *Haplistion*, was negatively correlated with turbidity. The presence of these sponges in the Somerset indicates that turbidity generally was low during deposition.

The only evidence of predatory behavior is cylindrical or parabolic holes in the beaks of some brachiopods. Ausich and Gurrola (1979) interpreted similar borings to be the work of infaunal parasitic polychaete worms and predatory ?gastropods, respectively. Borings are most common on *Spirifer tenuicostatus* Hall and *Dimegalasma* sp.

Fossils are rare or absent at several localities which may, in some cases, be owing to the type of exposure and weathering. At other localities (e.g. 12 and 15) there are far fewer fossils than expected for the type of exposure. The high diversity assemblage was not present everywhere mud was being deposited. At some localities the assemblage is modified. For example *Hapsiphillum cassedayi* and the common pelmatozoans are absent from locality 11, *Cleiothyridina hirsuta* is absent from locality 17, and locality 14 contains an unusually high abundance of sponges. The scale of faunal patchiness was larger than outcrop size but smaller than the spacing between outcrops. Community composition probably was controlled by physical factors, such as turbidity (as related to rate of deposition) and depth, that changed laterally over kilometers. The diverse community probably developed in areas with low turbidity in shallow water below fair-weather wave-base.

Differences among Middle Mississippian crinoid communities were related to substrate and turbidity. Monocyclic camerates preferred carbonate substrates in shallow, clear water (Lane, 1972, 1978) whereas inadunates and flexibles preferred clastic substrates and perhaps more turbid water (Breimer and Lane, 1978, p. 344; Lane, 1973; Kammer, 1982). This environmental selectivity may be because the fine mesh of filtration fans formed by pinnulate camerate crinoid arms clogged easily in tubid waters, whereas the filtration fans of the inadunates could cope with larger quantities of suspended sediment.

The Somerset crinoid assemblage is dominated in numbers of both specimens and species by advanced monobathrid camerates. However, the most common crinoids include cyathocrine inadunates and flexibles as well as monobathrid camerates (although, with the exception of *Dichocrinus simplex*, all crinoids are generally rare).

Comparitively the Lower Mississippian Borden Delta platform crinoid-dominated assemblages have been classified into three facies controlled communities (Ausich, 1978; Ausich and others, 1979). Poteriocrine inadunates dominated the communities in high turbidity and low energy paleoenvironments of the interdistributary siltstones and distributary channels. Monobathrid camerates and disparid inadunates dominated the higher energy biohermal deposits composed of crinoidal sands. Cyathocrine inadunates, poteriocrine inadunates, flexibles and monobathrid camerates are more equably distributed in the interdistributary mudstones and siltstones. Bryozoans and brachipods also are commonly abundant in this paleocommunity which is associated with low energy, slow sedimentation and high food supply (Ausich, 1978).

The Somerset community is most similar to the equably distributed paleocommunity of the Borden Delta interdistributary mudstones. Similar paleoenvironmental conditions include soft substrate, low energy, and a low sedimentation rate. The abundance and high diversity of rheophilic (pinnulate) camerates probably is owing to proximity of mud deposition to the shallower carbonate shoals in which the camerates are common.

Comparison With Contemporaneous Fossil Assemblages

Comprehensive faunal studies of Salem limestone facies in Kentucky do not exist so the Somerset was compared to the Salem in Indiana (Cumings and others, 1906). Presence-absence data for genera from the Somerset localities and Salem Limestone localities were analyzed using a cluster analysis program developed at Indiana University. Generic names of brachiopods, echinoderms, gastropods, coelenterates and bryozoans were updated from Cumings and others (1906) so they could be compared with the Somerset collections. Genera with fewer than 3 occurrences were eliminated from the data set. Somerset localities from which fewer than 20 brachiopods were collected also were not used. A total of 68 genera were used to cluster 21 localities. Three statistically significant clusters are recognized (Figure 4) that correspond to distinct facies of the Salem Limestone: the Somerset Shale facies, the sand bar shoal and the open tidal embayment (Donahue, 1967).

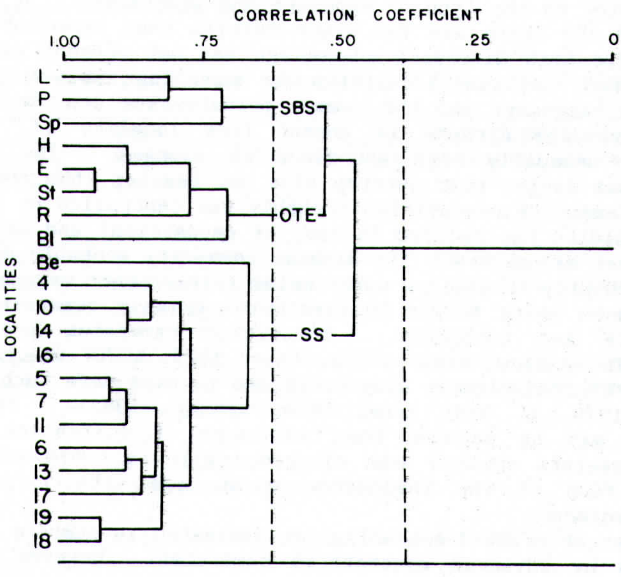

Figure 4. Q-mode cluster analysis of the Somerset and Salem localities in Kentucky and Indiana using the simple matching coefficient with unweighted pair group strategy. Those correlations that fall between the dashed lines are indistinguishable from random data, those correlations to the left of the left-most dashed line show a statistically significant correlation at the 95% confidence level. For locality abbreviations see Figure 3. SBS, sand bar shoal facies; OTE open tidal embayment facies; SS Somerset Shale.

In Indiana between Bedford and the Ohio River the Salem is comprised of the sand bar shoal facies (Donahue, 1967). This facies includes persistent shallow water shoals, preserved as algal-mollusk grainstones, separated by deeper water troughs in which carbonaceous dolosiltite was deposited. These shoals were stationary and did not migrate over finer sediment in the troughs. The open tidal embayment facies lies immediately to the north of the sand bar shoal facies (Donahue, 1967). Extensive shoaling in this facies resulted in deposition of widespread carbonate sands with a diverse fauna.

The dendrogram (Figure 4) indicates the distinctness of each facies fauna (Table 3, 4). The most diverse assemblage is from the sand bar shoal facies. High diversity and abundance of crinoids and corals indicates that the sand surface must have been stable over long periods of time to allow these organisms to grow. The abundance and diversity of crinoids and corals is considerably lower in the northern open tidal embayment facies owing to the intolerance of these taxa to less stable continously shifting sand. Fossils from the Bedford locality were probably collected from a fine- grained facies in the upper Salem which may be a carbonate equivalent to the shale facies; thus the Bedford locality clusters with the Somerset fauna.

Mollusks, particularly gastropods, are abundant and diverse in both carbonate facies but are rare in the shale. Gastropods and many bivalves are mobile and well-suited for life on a shifting substrate. Many modern gastropods make their living by scraping microepibionts from hard surfaces in shallow water. Many of the Salem gastropods may have done the same. Lack of abundant hard surfaces probably prevented them from invading the shale facies. The brachiopods and bryozoans show little preference of one facies over another except that there is a slight decrease in diversity of the brachiopods to the north in Indiana.

Table 3. Comparison of generic diversity of fauna from the Salem Limestone in Indiana (data from Cumings and others (1906) and the Somerset Shale Member in Kentucky.

	FACIES[1]		
	SBS	OTE	SS
Brachiopoda	16	11	14
Pelmatozoans	12	3	13
Coelenterata			
Rugosa	11	3	1
Tabulata	3		3
Mollusca			
Pelecypoda	13	10	2
Gastropoda	21	22	4
Cephalopoda	4	1	
Trilobita	1		1
Bryozoa	6	9	9
Grand Total	87	59	47

[1] SBS sand bar shoal; OTE open tidal embayment; SS Somerset Shale.

Community composition in the three facies can be explained using the component concept of paleoecology (Ausich, 1983). In this model the taxa are divided into components based on their response to various environmental factors. The distribution of the taxa in each component is controlled by a similar set of physical factors. Four components have been identified in this study: organisms that preferred carbonate sand, including most mollusks; organisms that preferred terrigenous mud, including inadunate crinoids; organisms that required stable substrates, including corals, crinoids and some brachiopods; and organisms that were dependent on other organisms, including platycerid gastropods and encrusting and boring organisms.

CONCLUSIONS

The Somerset Shale was deposited between carbonate bars of the Salem Limestone below normal wave base but above storm wave base. The muddy substrate supported a locally diverse benthic community typical for the upper Paleozoic. Competition between suspension feeders (the dominant trophic group) was reduced by tiering of the community.

Three distinct paleocommunities are recognized from the Somerset Shale and contemporaneous carbonate deposits of the Salem Limestone in Indiana.

The northern exposures of the Salem Limestone had shifting, shallow-water sands and a community dominated (in terms of diversity) by mollusks, brachiopods and bryozoans. The Salem Limestone of southern Indiana was deposited as stable sand bars with deeper water troughs in between and was dominated by a diverse assemblage of gastropids, brachiopods, corals, pelmatozoans and bryozoans. The Somerset community had an intermediate diversity of pelmatozoans, brachiopods, bryozoans and corals.

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Table 4. Characteristic genera for facies of the Salem Limestone in the study area.

	FACIES[1]		
	SBS	OTE	SS
Brachiopods			
Beecheria	+	+	
Camarotoechia	+	+	+
Cleiothyridina	+	+	+
Composita	+		+
Dimegalasm			+
Eumetria	+	+	+
Girtyella?	+	+	+
Perditacardinia	+	+	+
Reticulariina			+
Reticulariina?	+		
Setigerites			+
Spirifer	+	+	+
Streptorhynchus	+	+	
Tetracamera	+		
Torynifer			+
Bryozoans			
Cystodyctia	+	+	+
Dichotrypa	+		
Fenestella		+	+
Fistulipora	+		+
Hemitrypa	+		+
Polypora		+	+
Tabulipora			+
Corals			
Bordenia	+		
Cladochonus	+		+
Cystelasma	+		
Hapsiphyllum	+	+	+
Syringopora			+
Crinoids			
Batocrinus	+		+
Barycrinus			+
Cyathocrinites			+
Dichocrinus	+		+
Dizygocrinus	+		+
Forbesiocrinus	+		+
Platycrinites	+		+
Blastoids			
Metablastus	+		
Pentremites	+	+	+
Sponges			
Belemnospongia			+
Bivalves			
Conocardium		+	
Cypricardinia	+	+	
Edmondia?	+	+	
Eodon		+	
Nuculopis		+	

Table 4. continued.

	SBS	OTE	SS
Gastropods			
Bellerophon		+	
Cyclonema		+	
Dictiomaria		+	
Euconospira	+	+	
Glosseletina		+	
Hypergonia		+	
Ianthopsos		+	
Murchisonia		+	
Platyceras	+	+	+
Pleurotomaria		+	
Retispira		+	
Solenospira		+	
Stroporallus			
Strophostylus		+	

[1] SBS sand bar shoal facies; OTE open tidal embayment facies; SS Somerset Shale facies.

LOCALITY LIST

1. Fort Knox U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 6 & 15-R-42, Hardin County, Kentucky. Road cut on west side of U.S. Highway 31w, 2 km south of Hardin-Mead county line and about 4 km southeast of West Point.
2. Elizabethtown U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 5-O-45, Hardin County, Kentucky. Old railroad cut in Tunnel Hill near tunnel, IU16711.
3. Elizabethtown U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 5-O-45, Hardin County, Kentucky. New railroad cut in Tunnel Hill just south of old cut.
4. Colesburg U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 10-P-44, Hardin County, Kentucky. Roadcut on south side of State Highway 434 less than 0.25 km east of Cedar Creek, IU16712.
5. Colesburg U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 18-P-45, Hardin County, Kentucky. Glade 0.33 km south of State Highway 434, and 5.5 km east of the intersection of State Highways 251 and 434, IU16713.
6. Colesburg U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 17-P-45, Hardin County, Kentucky. Glade adjacent to, and west of locality 5, IU16714 and IU10887.
7. Campbellsville U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 12-J-50, Taylor County, Kentucky. Road cut about 100 m south of County Highway 1061 on the east side of State Highway 55, about 5 km south of Campbellsville, IU16715.
8. Saloma U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 16-L-50, Taylor County, Kentucky. Small road cut 0.75 km north of County Highway 1252 on State Highway 527, IU16716.
9. Greensburg U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 21-J-50, Green County, Kentucky. Small road cut on State Highway 208 about 0.5 km south of Meadow Creek, IU16717.
10. Greensburg U. S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 25-J-49, Green County, Kentucky. Glade on both sides of State Highway 323, 5.1 km east of Summersville (measured from the intersection with State Highway 61), IU16718.
11. Cane Valley U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 16-I-51, Adair County, Kentucky. Road cut on State Highway 55 about 0.5 km south of State Highway 633, IU16719.
12. Cane Valley U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 15-I-51, Adair County, Kentucky. Road cut on State Highway 55 about 50 m south of State Highway 682, IU16720.
13. Cane Valley U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates SE 1/4, SE 1/4 18-I-50, Green County, Kentucky.

Large glade near power lines, IU16721 and IU13058.

14. Cane Valley U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates NE 1/4, NE 1/4 23-I-50, Larue County, Kentucky. Small Glade in field, IU/16722 and IU13058.
15. Hodgenville U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 23-I-47, Taylor County, Kentucky. Road cut on U.S. Highway 31E, about 0.4 km northeast of intersection with State Highway 84, IU16723.
16. Gresham U.S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates 3-I-50, Taylor County, Kentucky. Small, low road cut on County Highway 1701, 1.6 km north of the Lemon Bend Church, IU/16724.
17. Russell Springs U. S. Geological Survey 7.5-minute Geologic Quadrangle, Carter Coordinates NE 1/4 20-J-53, Russell County, Kentucky. Glade near barn, IU10856 and IU13965.
18. Kosmosdale U.S. Geological Survey 7.5-minute Topographic Quadrangle, NE 1/4, SE 1/4 S. 20, T. 5 S., R. 5 E, Harrison County, Indiana. Glade on Buena Vista road next to St. Peters Church, IU/16869.
19. Laconia U.S. Geological Survey 7.5-minute Topographic Quadrangle, SW 1/4, SE 1/4 S. 35, T. 5 S., R. 4 E, Harrison County Indiana. Glade on the east side of a small country road approximately 150 m east of the West Branch of Mosquito Creek, IU16870.

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