



Southeastern Geology: Volume 41, No. 3 November 2002

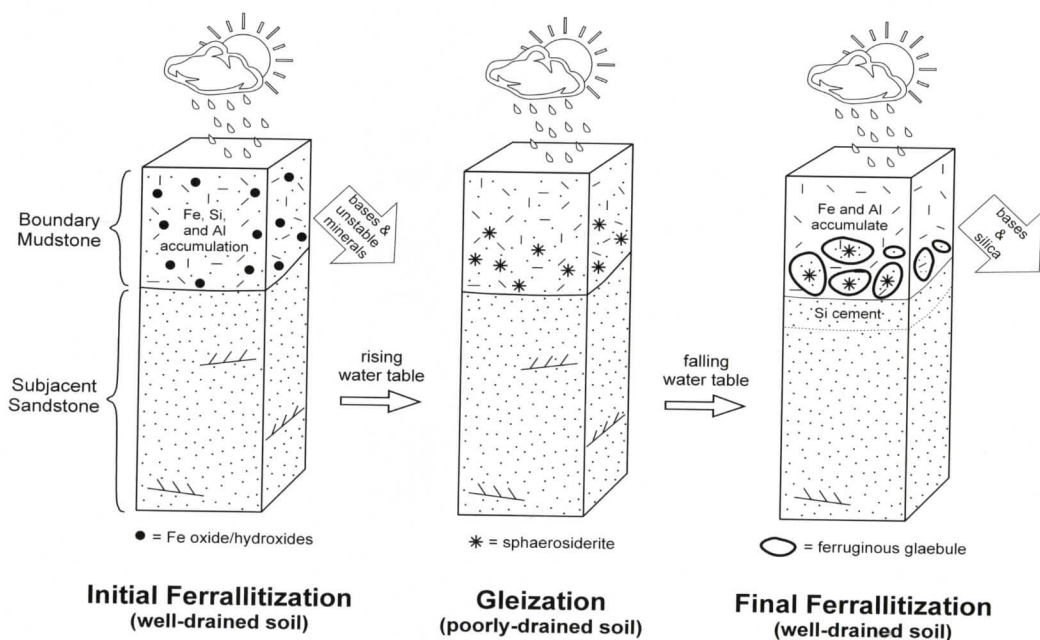
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Abstract

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SOUTHEASTERN GEOLOGY



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PALEOPEDOLOGY OF A RESIDUAL CLAY ASSOCIATED WITH THE MISSISSIPPIAN-PENNSYLVANIAN (MID-CARBONIFEROUS) UNCONFORMITY, SOUTHWESTERN PENNSYLVANIA

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ABSTRACT

The Mississippian-Pennsylvanian boundary near Johnstown, Pennsylvania is marked by an iron-rich mudstone ("boundary mudstone") that grades downward into a quartzose sandstone ("subjacent sandstone") in the Mauch Chunk Formation (Mississippian). Field, micromorphological, geochemical, and mineralogical characteristics indicate that these strata constitute a well-developed paleosol (herein referred to as the "Route 56 paleosol") with a complex pedogenic history. Specifically, the boundary mudstone comprises ferruginous glauclites that bear evidence of both ferrallitic/oxilic weathering (abundant kaolinite, ferric iron, and runiquartz) and gleization (sphaerosiderite). Such mineralogical and micromorphological disparities suggest that the Route 56 paleosol has a polygenetic origin involving contrasting soil-forming regimes. Although the Route 56 paleosol is lithologically similar to a gleyed underclay, micromorphological and geochemical analyses suggest that it formed under well-drained conditions on an interfluvial of the Mississippian-Pennsylvanian unconformity. Basic paleosol characteristics are consistent with pedogenesis under a tropical rainy climate, as existed during early Pennsylvanian time. This study demonstrates how paleopedology, particularly micromorphology, can be used to provide sedimentological evidence for an otherwise cryptic unconformity surface.

INTRODUCTION

The Mississippian-Pennsylvanian unconformity in the Appalachian foreland basin is a regional, Vail-type sequence boundary that apparently records both a mid-Carboniferous eustatic event (Saunders and Ramsbottom, 1986; Beuthin, 1997) and tectonic activity within the basin (Quinlan and Beaumont, 1984; Ettensohn and Chesnut, 1989; Chesnut, 1990; Ettensohn, 1994). Characteristically, the Mississippian-Pennsylvanian unconformity is marked by pronounced erosional relief associated with paleovalleys (Rice, 1984, 1985; Rice and Schwietering, 1988; Chesnut, 1988, 1992; Beuthin, 1994; Yang, 1998). Although well-developed paleosols would be expected to occur on interfluvial along the unconformity, these features have been reported at only a few locations throughout the basin (Smyth, 1984; Williams and Bragonier, 1985; Cecil and others, 1992; Chesnut and others, 1992; Beuthin, 1997; Humbert, 2001). Interfluvial paleosols can be easily overlooked because they are often subtle features that are difficult to identify using field evidence alone. Recently, McCarthy and Plint (1998) showed that paleopedology can be used to recognize interfluvial paleosols associated with sequence boundaries. In particular, they used micromorphology to show that sequence-bounding paleosols typically have a complex, polygenetic origin. This approach to unconformity analysis has the potential to yield new insights into the nature and origin of the Mississippian-Pennsylvanian boundary in the central Appalachian basin.

The Mississippian-Pennsylvanian boundary and associated strata are well exposed in a road-cut along Pennsylvania Route 56 near the city of Johnstown (Figures 1 and 2). The boundary is marked by an iron-rich mudstone with field characteristics of a semi-flint to plastic clay. This mudstone, which is sharply overlain by strata of the Pottsville Formation (Pennsylvanian), grades down into a subjacent sandstone of the Mauch Chunk Formation (Mississippian). Mudstones in the Pennsylvanian System with similar lithologic characteristics were interpreted by Williams and others (1965) as residual clays formed on well-drained interfluvial disconformities. In the present study we use a paleopedological approach to test the hypothesis that the "boundary mudstone" at the study site is a residual soil formed on an interfluvial associated with the Mississippian-Pennsylvanian unconformity.

STRATIGRAPHIC SETTING

Strata exposed at the study site belong to the upper part of the Mauch Chunk Formation (upper Mississippian) and the lower part of the Pottsville Formation (middle Pennsylvanian) (Figure 3). In this paper, the boundary mudstone and underlying Mauch Chunk sandstone are collectively termed the "Route 56 paleosol." Below the Route 56 paleosol about 30 m of the Mauch Chunk Formation are exposed, comprising interbedded red mudstone, green siltstone, and gray sandstone. The Wymys Gap Member of the Mauch Chunk Formation is recognized at the base of the section as a 30-cm-thick fossiliferous mudstone. This middle Chesterian unit records a regional transgressive event in southwestern Pennsylvania (Horowitz and Rexroad, 1972; Brezinski, 1989; Edmunds, 1993; Stamm, 1997). Above the Route 56 paleosol about 4 m of the Pottsville Formation are exposed, comprising carbonaceous mudstone and gray sandstone. These strata appear to be part of the Mercer shale member of Phalen (1910). Although direct biostratigraphic control on the age of the basal Pottsville strata at the Route 56 outcrop is lacking, palynomorphs and megafossil remains from the Mercer interval elsewhere in

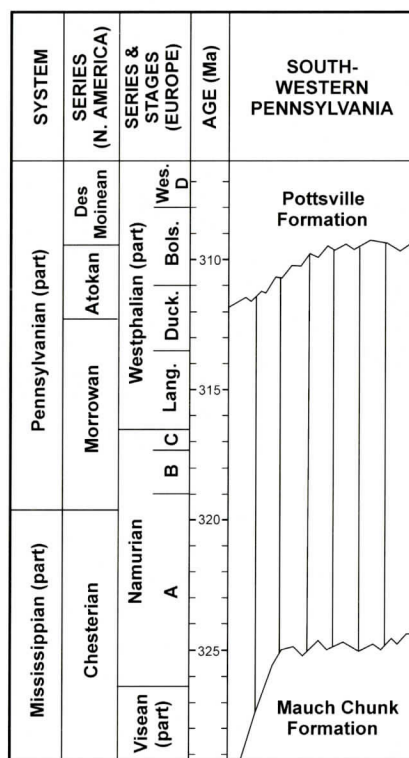


Figure 3. Upper Mississippian-lower Pennsylvanian stratigraphy of southwestern Pennsylvania showing hiatus associated with the Mississippian-Pennsylvanian unconformity. Chronostratigraphy after Edmunds (1993); numerical dates from Menning and others (2000). European stage names are abbreviated as follows: Langsettian (Lang.), Duckmantian (Duck.), Bolsovian (Bols.), and Westphalian D (Wes. D).

the basin indicate a Bolsovian (Atokan) age (Peppers, 1996; Blake and others, 2002). Plant debris is abundant in the Pottsville Formation, but uncommon in the Mauch Chunk Formation. The formation contact is an irregular erosional surface with about 1.0 m of relief across the outcrop face.

The Mississippian-Pennsylvanian unconformity in southwestern Pennsylvania apparently omits strata of upper Mississippian-lower Pennsylvanian age (upper Chesterian-Morrowan or Namurian-lower Westphalian) (Edmunds, 1993). The Carboniferous time scale of Menning and others (2000) suggests that the time gap may be as great as 13-16 million years (Fig-

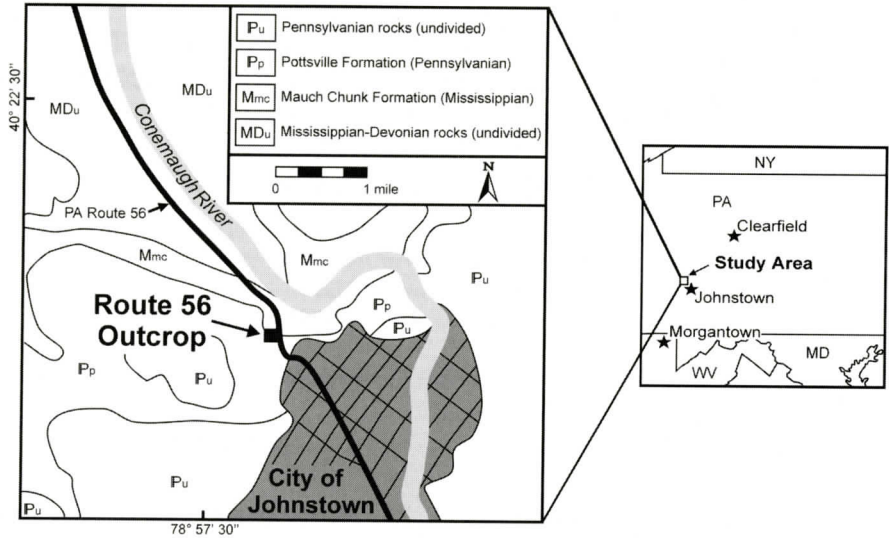


Figure 1. Location of the Route 56 paleosol near Johnstown, Pennsylvania (bedrock geology after Glover and Edmunds, 1981). Other paleosols associated with the Mississippian-Pennsylvanian boundary have been reported near Clearfield, Pennsylvania (Williams and Bragonier, 1985) and Morgantown, West Virginia (Cecil and others, 1992).

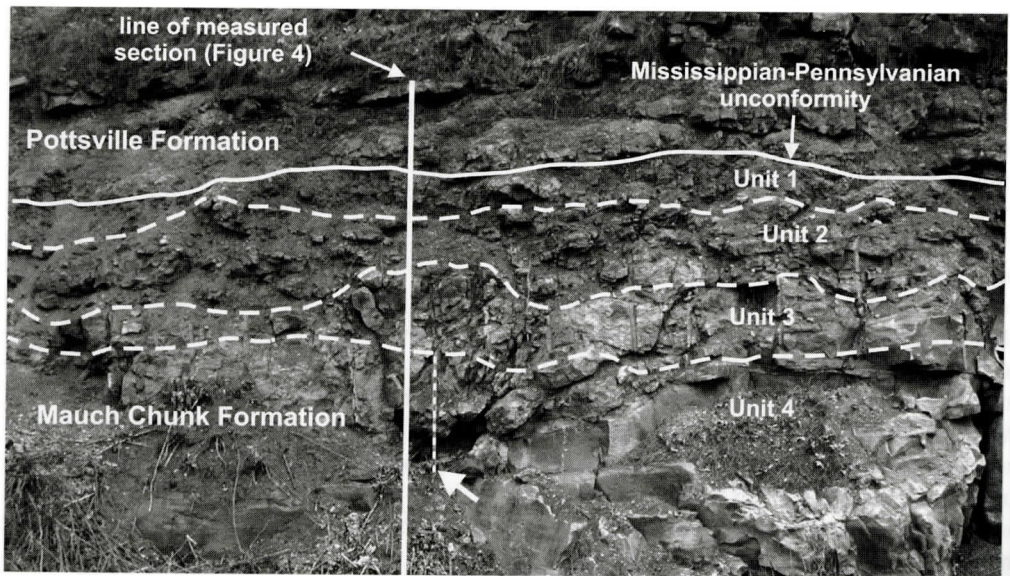


Figure 2. Outcrop photograph and interpretive stratigraphy of the Route 56 paleosol. Position of the Mississippian-Pennsylvanian unconformity (Mauch Chunk-Pottsville contact) is indicated by the bold solid horizontal line; contacts between units within the paleosol shown with dashed lines. Units 1 and 2 comprise the boundary mudstone; Units 3 and 4 comprise the subjacent sandstone. Exact position of the measured section shown is with thick vertical line. Note that unit thicknesses vary across the outcrop. Photoscale (measuring staff indicated by bold arrow) is 1.7 m tall with 0.1 m gradations.

Anatomy of a glaebule

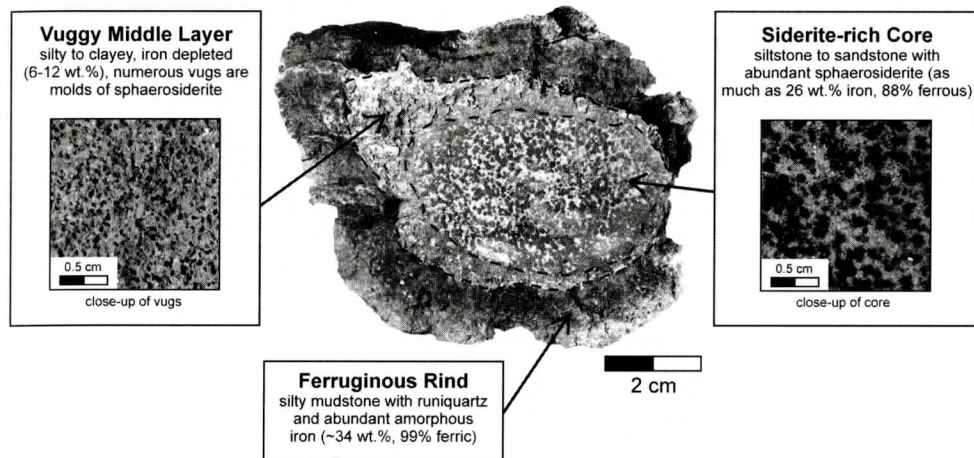


Figure 5. Tripartite concentric structure of a glaebule from Unit 2 of the boundary mudstone.

because of differential development. Across the outcrop the boundary mudstone is locally truncated at the base of the Pottsville Formation. Unit descriptions are as follows:

Unit 1: 50 cm thick; light-olive-gray; silty mudstone with isolated lenses of muddy, very fine- to fine-grained sand; internally massive; slight vuggy texture with yellowish-orange iron stain common around vug margins; basal contact wavy and gradational.

Unit 2: 100 cm thick; light-gray to reddish-yellow; silty mudstone; massive; numerous iron-rich glaebules ranging from 5 to 26 cm in diameter; glaebules have a tripartite concentric structure consisting of a siderite-rich core, iron-depleted middle layer, and a ferruginous outer rind (Figure 5); silty to sandy cores have abundant reddish-brown sphaerosiderite rosettes (0.1-1.0 mm in diameter); clayey middle layers have a well-developed vuggy texture (vugs ranging from 0.1-1.0 mm in diameter); reddish-brown rinds are silty; small glaebules lack siderite-rich cores; basal contact wavy and gradational.

Unit 3: 100 cm; white to very-light-gray; medium-grained sandstone; quartzose and silica-cemented; internally massive, lacks primary sedimentary structures and labile clasts; basal contact wavy and gradational.

Unit 4: 450 cm; light-gray; fine-grained sandstone; argillaceous; scattered siderite cement; very thick- to massive-bedded; planar-laminated to cross-bedded; tabular cross-strata indicate a southeasterly paleoflow direction; contains pebble-sized rip-up clasts of gray-green mudstone; contact with subjacent red beds is sharp and flat.

Micromorphology and Petrography

Boundary Mudstone (Units 1 and 2)

The relative proportion of skeleton grains (silt and sand) and plasma (clay) ranges from skeleton grains floating in plasma to skeleton grains that are in contact or are linked by intergranular braces. Variations in the distribution pattern occur at thin-section scale. The plasma fabric is mostly aseptic (weakly/randomly oriented clay, dull plasma) with scattered patches of sepic fabric (oriented, bright clay) (micromorphological terminology given in parentheses is after Brewer (1976)). In patches of sepic fabric, bright clay is aligned in isolated to interconnected streaks (insepic and mosepic fabrics, respectively). Where plasma forms intergranular braces, bright clay jackets many skeleton grains (skelsepic fabric). In contrast to the rest of the boundary mudstone, the cores and vuggy

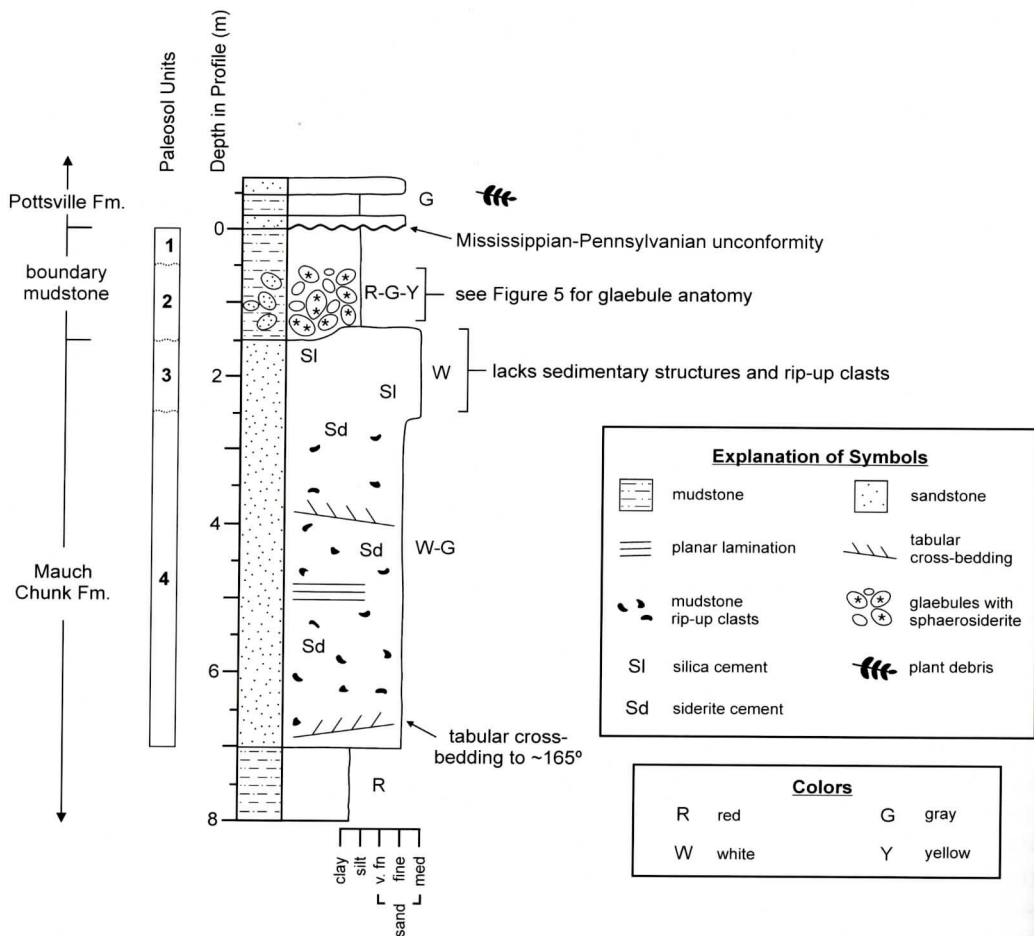


Figure 4. Sedimentological log of the Route 56 paleosol and associated strata.

ure 3). In addition to recording a prolonged base-level lowstand, the Mississippian-Pennsylvanian boundary in the Appalachian basin also marks a major climate change from wet-dry/monsoonal (Mississippian) to tropical rainy (Pennsylvanian) conditions (Cecil and others, 1985; Cecil, 1990). Throughout the basin, this climate change is evident in the paleosol record. Vertic and calcic paleosols are ubiquitous in the Upper Mississippian succession (Cecil and others, 1985; Cecil, 1990; Cecil and Dulong, 1992; Caudill and others, 1992; Caudill and others 1996; Beuthin, 1997; Greb and Caudill, 1998; Miller and Eriksson, 1999, 2000). Although thin, impure coals are present locally, these beds are probably the products of topographic control, rather than climate control (Beuthin

and Blake, 2002). In contrast, the lower Pennsylvanian succession comprises thick coals (paleo-Histosols), but lacks vertic and calcic paleosols (Donaldson and others, 1985; Cecil and others, 1985; Cecil, 1990; Cecil and Dulong, 1992).

THE ROUTE 56 PALEOSOL

Outcrop Description

Where measured, the boundary mudstone is 1.5 m thick and the subjacent sandstone is 5.5 m thick (Figure 4). Four horizon-like units are recognized in the paleosol, two in the mudstone (Units 1 and 2) and two in the sandstone (Units 3 and 4). Laterally, the units vary in thickness

iron. Scattered silt- and sand-sized quartz grains in the rinds are fractured and corroded, with microveins of ferric material frequently lining the fractures (Figure 6B). These quartz grains are nearly identical to runiquartz as described by Stoops (1983). In addition, numerous quartz grains in the cores of the glaeboles are etched and embayed by siderite (Figure 6C). Cutans are uncommon, but where present they typically are stained by iron (ferri-argillans) or comprise densely packed plasma (embedded grain argillans). Some ferri-argillans appear to coat collapsed voids. The iron-rich glaeboles generally lack cutans. In glaebole cores, siderite occurs as spherical aggregates with a radiating crystal habit (Figures 6D and 6E). Individual sphaerosiderite rosettes range in diameter from 0.6–1.0 mm. In places, rosettes have coalesced into larger, intergrown masses up to 1 cm in size. The radial morphology of many vugs indicates that they are molds of sphaerosiderite rosettes (Figure 6F). A few segregations of material resemble voids filled with skeleton grains (granotubules), some of which are lined with streaks of plasma (isotubules).

Subjacent Sandstone (Units 3 and 4)

Typically, skeleton grains are in contact with one another and there is little or no interstitial plasma. Where present, plasma forms scattered intergranular braces. Skeleton grains comprise fine- to medium-grained, subangular to subrounded sand. In Unit 3, all the sand is quartz, ~90% of which is monocrystalline. Grain contacts in Unit 3 are commonly flat or concave-convex. In Unit 4, polycrystalline quartz is common and increases in abundance downward in the profile. Mudstone rip-up clasts are scattered throughout the lower part of Unit 4. Many of these clasts are plastically deformed, probably from compaction. Quartz overgrowths are evident on several sand grains in Unit 3. Micro-morphologically, these overgrowths can be regarded as embedded grain quartzans. In Unit 3 and the top part of Unit 4, cherty silica fills many interstitial voids. In the upper part of Unit 4, a few plane argillans and embedded grain argillans are present. Siderite occurs as subhedral to euhedral interstitial cement, but not as

sphaerosiderite crystallaria.

Bulk Mineralogical and Geochemical Features of the Boundary Mudstone

Bulk XRD analysis showed that the clay content of the boundary mudstone ranges from 25.7 to 70.6% (Table 1). Kaolinite and illite were the only detectable clays, with kaolinite being the dominant species (kaolinite:illite ratios range from 1.6 to 3.5). Quartz was the only other abundant silicate mineral, with values ranging from 24.2 to 57.3%. A single sample of a siderite-rich glaebole core showed a siderite content of 15.0%; no other carbonate phases were detected. XRD analysis failed to detect iron-bearing minerals in the glaebole rinds, a result consistent with the observation of amorphous iron in thin section.

Silica, iron, and alumina are the primary chemical constituents of the boundary mudstone (Table 2). Silica is the most abundant constituent, with values from 41.95 to 88.84% by weight. Total iron ($\text{FeO} + \text{Fe}_2\text{O}_3$) reaches a maximum value of 33.78% in a glaebole rind from Unit 2. Ferric iron predominates in the rinds (99% of total iron), whereas ferrous iron is most abundant in the cores (88%). Alumina reaches a maximum value of 24.40% in Unit 1. Total abundance of leachable bases ($\text{MgO} + \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}$) ranges from 0.35 to 2.80%. Molecular weathering ratios of alumina to bases ($\text{Al} / \text{MgO} + \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}$) can be used to measure the degree to which soluble materials have been leached from a paleosol (Retallack, 2001). In the boundary mudstone, calculated alumina/bases ratios range between 13 and 23.

DISCUSSION

Evidence of Intense Pedogenesis

The four-unit profile recognized in the boundary mudstone and subjacent sandstone supports our interpretation of these rocks as a well-developed paleosol with a complex pedogenic history. Although the units have horizon-like properties, they are difficult to characterize using standard nomenclature for

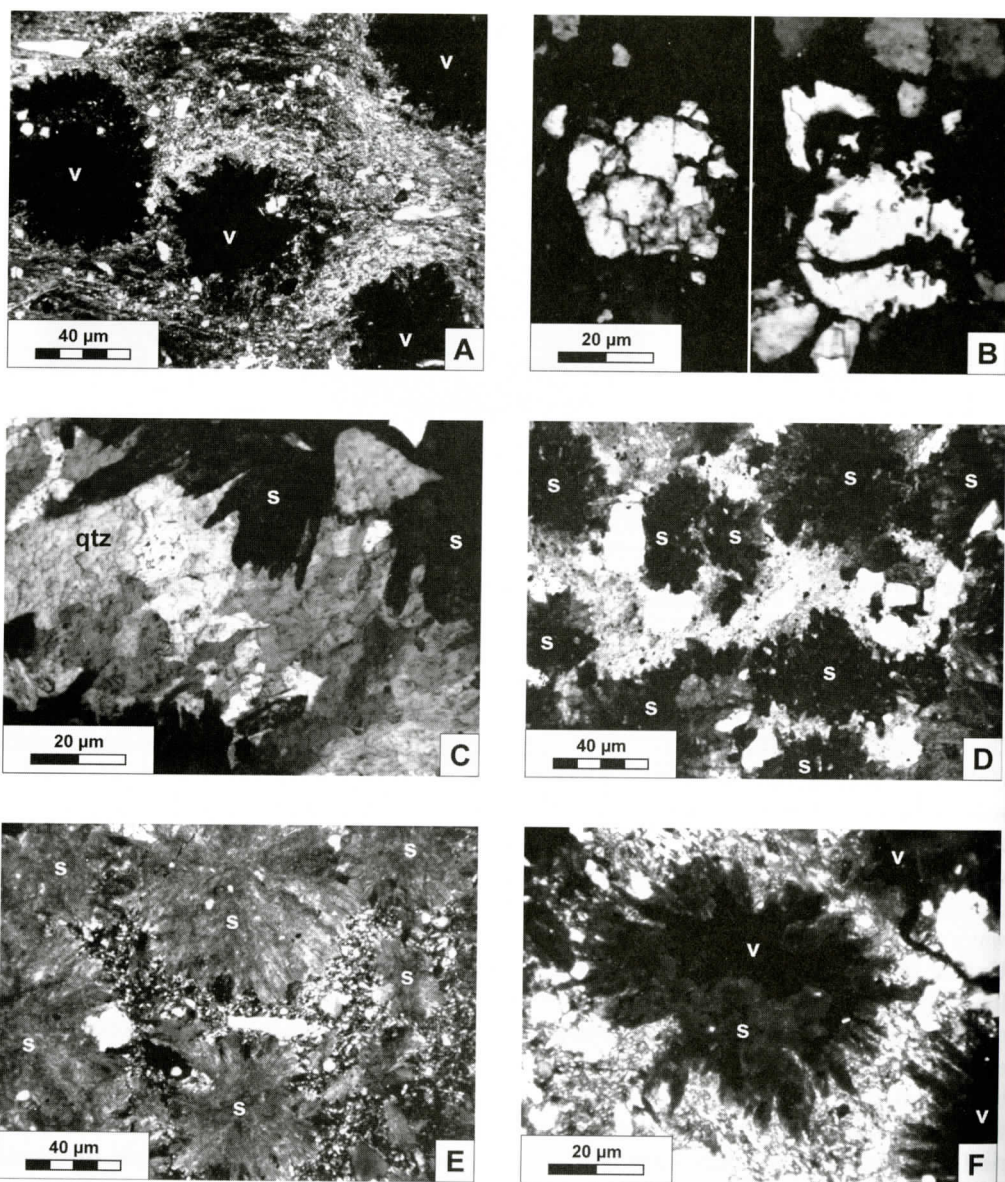


Figure 6. Selected photomicrographs from the Route 56 paleosol. (A) Sepsic plasmic fabric surrounding vugs (v) in sample 0.75V (middle layer of a glaebule; cross-polarized light). (B) Fractured and corroded quartz grains (runiquartz) from sample 1.00R (glaebule rind; plane-polarized light). (C) Sphaerosiderite (s) embaying quartz (qtz) in sample 1.00C (glaebule core; cross-polarized light). (D) Sphaerosiderite rosettes (s) in muddy matrix from sample 1.00C (glaebule core; plane-polarized light). (E) Sphaerosiderite rosettes (s) in silty-sandy matrix of sample 1.00C (glaebule core; cross-polarized light). (F) Vug (v) with radial margin and relict sphaerosiderite (s) in sample 1.00V (middle layer of a glaebule; circular-polarized light).

middle layers of the glaebules commonly exhibit sepic fabric with patches of densely woven bright clay (omnisepic fabric) (Figure 6A). Skeleton grains range in size from silt to medi-

um sand. Sand grains range from subangular to subrounded. Virtually all skeleton grains are quartz. Glaebule rinds are clayey; any clay fabric present is obscured by extensive amorphous

Table 2. Whole-rock geochemical data for the Route 56 paleosol (given in weight percent). Multiple samples from the same depth in the profile are identified using the nomenclature described in Table 1. Alumina/bases ratios were calculated using the methodology of Retallack (2001). Samples were prepared and analyzed (using X-ray fluorescence) by a commercial laboratory.

Unit	Depth in profile (m)	SiO ₂ %	TiO ₂ %	Al ₂ O ₃ %	Total Fe %	MnO %	MgO %	CaO %	Na ₂ O %	K ₂ O %	P ₂ O ₅ %	Cr ₂ O ₃ %	LOI %	Total %	Al/bases ratio
1	0.00	62.07	1.35	23.27	1.49	0.01	0.37	0.08	0.16	2.19	0.05	0.03	7.89	98.93	21
	0.25	80.15	0.97	10.74	1.17	0.01	0.17	0.05	0.08	1.04	0.03	0.03	3.53	97.94	20
	0.50-1	82.69	0.88	9.11	0.66	0.01	0.15	0.05	0.07	0.86	0.03	0.05	2.98	97.47	20
	0.50-2	59.89	1.24	24.40	1.18	0.02	0.36	0.01	0.16	2.15	0.07	0.03	8.31	97.89	23
2	0.75 V	78.92	0.70	6.97	6.12	0.04	0.10	0.06	0.05	0.57	0.06	0.08	3.73	97.40	22
	0.75 C	50.55	0.45	4.64	25.79	0.29	0.20	0.29	0.03	0.38	0.22	0.04	14.20	97.09	13
	1.00 R	41.95	0.57	9.84	33.78	0.42	0.15	0.04	0.06	0.95	0.30	0.03	9.23	97.32	20
	1.00 V	67.24	0.84	9.42	11.52	0.11	0.17	0.12	0.07	0.90	0.10	0.06	7.47	98.00	19
	1.00 C	70.46	0.46	4.22	13.90	0.15	0.12	0.14	0.03	0.39	0.15	0.06	7.29	97.37	15
	1.25	88.84	0.34	2.10	3.49	0.05	0.08	0.04	0.02	0.21	0.03	0.07	1.98	97.22	15
3	1.50	89.82	0.58	3.15	1.85	0.03	0.06	0.02	0.02	0.24	0.03	0.08	1.63	97.47	23
	1.75	91.37	0.54	2.78	1.29	0.01	0.03	0.01	0.02	0.22	0.03	0.10	0.89	97.25	25
	2.00	89.44	0.44	5.05	0.87	0.01	0.03	0.01	0.01	0.14	0.03	0.08	1.87	97.97	66
	2.50	90.90	0.58	4.10	0.63	0.01	0.04	0.01	0.02	0.33	0.03	0.06	1.24	97.92	25
4	3.00	90.81	0.67	4.66	0.48	0.01	0.07	0.02	0.03	0.53	0.03	0.04	1.34	98.66	18
	3.50	90.57	0.69	5.67	0.60	0.01	0.07	0.03	0.04	0.62	0.03	0.06	1.69	100.07	18
	4.00	91.23	0.67	5.34	0.47	0.01	0.08	0.02	0.05	0.72	0.03	0.04	1.62	100.23	15
	4.50	91.62	0.45	5.42	0.64	0.01	0.07	0.01	0.04	0.57	0.03	0.06	1.69	100.56	19
	5.00	90.57	0.44	5.17	1.21	0.04	0.09	0.02	0.04	0.56	0.03	0.04	1.97	100.17	18
	5.50	85.97	0.48	5.60	3.99	0.15	0.18	0.07	0.04	0.65	0.06	0.06	3.35	100.59	15
	6.00	90.24	0.43	4.38	2.31	0.09	0.13	0.02	0.03	0.48	0.03	0.05	2.27	100.45	16
	6.50	87.60	0.24	3.28	4.39	0.17	0.25	0.04	0.02	0.32	0.03	0.03	3.34	99.68	13
	7.00	90.70	0.52	4.39	1.45	0.04	0.10	0.06	0.03	0.49	0.08	0.05	1.88	99.79	16
	Detection Limit (%)	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.05	0.03	0.01	0.05	0.01	

Table 1. Whole-rock XRD data for the Route 56 paleosol (data are normalized). Common minor minerals include rutile, plagioclase, K-feldspar, diaspore, anatase, and hematite. Multiple samples from 0.50 m represent the sandy (0.50-1) and clayey (0.50-2) lithologies present at this depth. Glaebular samples are identified as having come from either the siderite-rich core (C), the middle vuggy layer (V), or the ferruginous rind (R). XRD samples were analyzed using a Scintag X-1 diffractometer; diffractogram peaks were analyzed using lab software available at the U.S. Geological Survey in Reston, VA

Unit		Depth in Profile (m)	Quartz (%)	Siderite (%)	Illite (%)	Kaolinite (%)	Minor Minerals (%)
Boundary mudstone	1	0.00	35.5	0.0	16.9	45.4	1.6
		0.25	52.8	0.0	12.8	28.8	5.6
		0.50-1	67.8	0.0	9.4	19.1	1.8
		0.50-2	24.2	0.0	16.5	57.1	2.1
	2	0.75 V	85.3	0.0	3.6	9.5	1.7
		0.75 C	57.3	15.0	6.9	18.8	2.1
		1.00 R	24.8	0.0	21.6	49.0	1.3
		1.00 V	69.2	0.0	8.5	19.7	2.7
		1.00 C	87.3	0.4	2.0	6.9	3.4
		1.25	95.7	1.1	0.0	3.0	0.2
Subjacent sandstone	3	1.50	95.5	1.5	0.0	1.9	0.0
		1.75	89.5	0.0	4.5	4.6	1.4
		2.00	92.5	0.0	0.0	6.6	0.9
		2.50	90.6	0.0	3.7	4.8	1.0
	4	3.00	84.0	0.0	6.1	6.2	3.3
		3.50	87.1	0.0	2.6	9.2	1.1
		4.00	84.9	0.0	6.5	7.1	1.5
		4.50	82.0	0.0	5.8	11.0	1.2
		5.00	80.8	0.0	5.8	10.5	2.8
		5.50	76.5	0.7	5.0	11.6	6.1
		6.00	83.4	0.9	3.2	7.9	4.7
		6.50	81.3	1.4	3.0	5.6	8.7
		7.00	83.7	0.2	4.6	8.0	2.5

modern soils (i.e., shorthand labels utilized by the Soil Survey Staff, 1998). It is possible, however, to interpret the pedogenic processes that created each unit.

The subjacent sandstone exhibits varying degrees of pedogenic alteration. The existence of bedding and labile rip-up clasts in the lower part of the sandstone (Unit 4) implies that it consists of relatively unaltered parent material. In contrast, the upper part of the sandstone (Unit 3) is pedogenically overprinted. Unit 3 possesses several characteristics of a well-leached, pedogenic ganister including quartz-rich sand, silica cement, absence of sedimentary structures, and a downward transition into a compositionally less mature sandstone (Percival, 1983).

Several aspects of Units 1 and 2 indicate that boundary mudstone is a residuum composed of strongly altered parent materials. Abundant ka-

olinite and quartz suggest destruction and removal of unstable constituents by intense hydrolysis and leaching. In modern soils, this degree of weathering is found in Oxisols (Soil Survey Staff, 1998; Birkeland, 1999; Retallack, 2001) or ferruginous soils (Duchaufour, 1982). Intense weathering is further supported by the presence of runiquartz, a common feature of modern Oxisols (Stoops, 1983). Sepic fabrics observed within the glaebules of Unit 2 are probably relict features inherited from an earlier stage of pedogenesis (pedorelicts). The oxidized rinds and reduced cores of these iron-rich glaebules suggest a complex history of iron mobilization and redoximorphism (Vepraskas, 1999). Although it is difficult to infer the nature of the parent material of the boundary mudstone, common coarse silt and sand suggest that at least some of the parent material was sandy

ly, the sphaerosiderite is embedded in a kaolinitic matrix, indicating that a phase of hydrolysis preceded the phase of gleization. Given the evidence for this initial phase of intense hydrolysis, it seems likely that iron oxides/hydroxides also accumulated as a residual product of deep weathering. During this phase, the paleosol may have approached a state of development comparable to that of a modern Oxisol.

Phase 2 (Gleization)

During this second phase, the water table was relatively high and drainage conditions deteriorated. Gleying by alkaline, reducing water transformed iron oxides/hydroxides accumulated during Phase 1 into sphaerosiderite (Curtis and others, 1975; Ludvigson and others, 1998; Ufnar and others, 2001). Most of the sphaerosiderite accumulated in the lower part of the boundary mudstone (Unit 2). Pore water was probably meteoric rather than marine, as marine water would have added sulfide to the system and favored the formation of pyrite over sphaerosiderite. Chemical destruction of quartz is suggested by embayment of quartz by sphaerosiderite in Unit 2. Hydrolysis, however, was suppressed because of poor drainage conditions.

Phase 3 (Final Ferrallitization)

During this phase, oxidizing waters again penetrated the profile during a period of lowered or seasonally fluctuating water levels, creating the glaebules in Unit 2 of the boundary mudstone. Oxidation of sphaerosiderite created the ferruginous materials in the glaebule rinds. Much of the iron in the rinds appears to be derived from the iron-depleted, vuggy middle layer of the glaebules. The siderite-rich cores of the glaebules represent unoxidized pedorelicts of the gleization phase. Runiquartz in the rinds of the glaebules suggests intense physical and chemical destruction of quartz during this phase. The solubility of quartz in the boundary mudstone may have increased significantly by redoximorphic reactions involving iron (Morris and Fletcher, 1987). Much of the silica cement in Unit 3 may have accumulated during this

time. This last recorded phase of pedogenesis was followed by deposition of Pottsville sediments during onlap of the mid-Carboniferous land surface.

Relationship to the Mississippian-Pennsylvanian Unconformity

The Route 56 paleosol appears to be the interfluvial expression of the Mississippian-Pennsylvanian unconformity. In contrast to paleosols developed on active floodplains, paleosols developed on interfluvies typically are polygenetic and have complex drainage histories related to changes in base level (Gardner and others, 1988; McCarthy and Plint, 1998; McCarthy and others, 1999; Plint and others, 2001). Micro-morphological features of the Route 56 paleosol clearly show evidence of polygenesis and drainage evolution, consistent with an interfluvial interpretation. However, it is difficult to unequivocally attribute polygenesis of the Route 56 paleosol to a single base-level event. Although a mid-Carboniferous eustatic event affected development of the Mississippian-Pennsylvanian unconformity throughout the basin (Saunders and Ramsbottom, 1986; Beuthin, 1997; Driese and others, 1998; Blake and others, 2002), the hiatus in western Pennsylvania is large enough that the development of the Route 56 paleosol may also have been profoundly impacted by tectonic activity, climate change, and landscape evolution.

Even though the Route 56 paleosol records protracted weathering and pedogenesis, it probably does not record the entire weathering history of the unconformity. The deeply leached character of the paleosol suggests development during the tropical rainy conditions of the early Pennsylvanian Period (Cecil and others, 1985; Cecil, 1990). In contrast, paleosols formed under the seasonal climate of late Mississippian time in the Appalachian basin are typically vertic, calcareous, or both (Cecil and others, 1985; Cecil, 1990; Cecil and others, 1992; Caudill and others, 1992, 1996; Beuthin, 1997; Beuthin and Blake, 2002). In terms of when it formed, the Route 56 paleosol may be analogous to the Olive Hill clay, which is a well-developed mid-

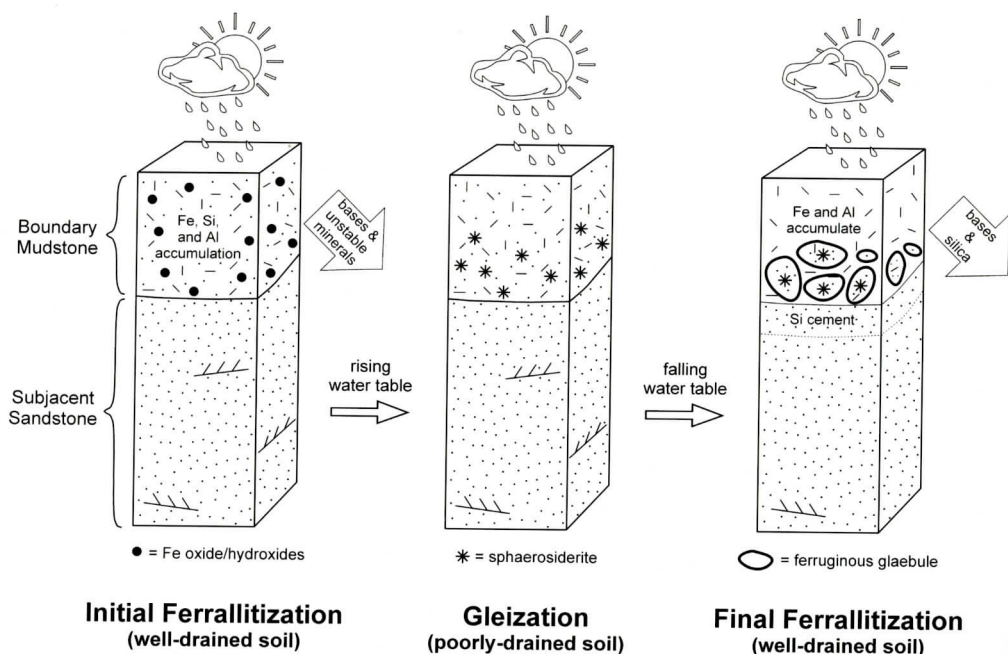


Figure 7. Three-phase pedogenic model for the Route 56 paleosol illustrating development of ferruginous glaeboles in the boundary mudstone.

(Loughnan and Bayliss, 1961).

Geochemical data also suggest that the boundary mudstone is a deeply weathered residuum. Alumina/bases ratios, which range from 13 to 23, indicate strong base depletion. This degree of hydrolysis and leaching occurs in modern Oxisols which characteristically yield alumina/bases ratios greater than 10 (Retallack, 2001). Even though iron is in both a ferrous and a ferric state, total iron in the Route 56 paleosol (up to 33.78%) compares well with that of some modern Oxisols (Birkeland, 1999). Also, paleosols enriched in resistate elements (iron, aluminum, silica, etc.) are common features of major geological unconformities (Retallack, 2001). The absence of root traces and general lack of clay coatings is consistent with evidence for extreme weathering. Organic matter has low preservation potential in the strongly oxidizing conditions typical of oxic soils. The rarity of clay coatings (i.e., illuvial clay), suggests that this material was itself weathered and assimilated into the soil matrix as the soil aged. Modern Oxisols are, in part, distinguished by a general absence of illuvial clay.

A Polygenetic History

Mineralogical and micromorphological disparities in the paleosol suggest a polygenetic origin involving contrasting soil-forming regimes. For example, abundant kaolinite and rare illuvial clay in the boundary mudstone indicate hydrolysis associated with a well-drained land surface, whereas sphaerosiderite in the same part of the paleosol suggests gleying associated with a poorly-drained land surface. Altogether, the contrasting features of the paleosol appear to record three phases of development, as outlined below and in Figure 7.

Phase 1 (Initial Ferrallitization)

This phase was characterized by a low water table and intense hydrolysis. The soil was well-drained and flushed by slightly acidic, oxidizing meteoric water. Over time, bases were leached and unstable minerals were destroyed by chemical weathering, leaving a residuum of kaolinite and quartz (the boundary mudstone). The only record of this phase is contained in the pedorelict interiors of the glaeboles. Specific-

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Carboniferous paleosol in eastern Kentucky (Smyth, 1984). Although this paleosol locally occurs at the Mississippian-Pennsylvanian boundary, it also occurs as much 30 m (100 feet) above the base of the Pennsylvanian (Chesnut and others, 1992). Like localities where the Olive Hill clay occurs at the Mississippian-Pennsylvanian boundary, the Route 56 paleosol probably records Pennsylvanian weathering of exposed Mississippian strata, rather than a "Mississippian-Pennsylvanian" event (eustatic or otherwise). Regardless of exactly when weathering occurred, the Route 56 paleosol provides sedimentological evidence for a pronounced unconformity between the Mississippian and Pennsylvanian systems at the study site.

The Route 56 paleosol provides critical data for assessing regional variations in pedogenesis associated with the Mississippian-Pennsylvanian unconformity. The Mercer clay near Clearfield, Pennsylvania (Figure 1) is also interpreted as a residual clay (Williams and Bragonier, 1985). In contrast to the ferruginous Route 56 paleosol, the Mercer clay is a more deeply weathered aluminous residuum that contains diaspor, boehmite, and gibbsite. Just east of Morgantown, West Virginia (Figure 1), the systemic unconformity is marked by a leached, clayey paleosol that appears to be a paleo-Ultisol (Cecil and others, 1992). These three paleosols indicate a south-to-north increase of paleopedogenesis across the northern flank of the basin. This "pedogenic gradient" possibly resulted from increased duration of subaerial exposure related to paleotopography on the unconformity.

CONCLUSIONS

Identification of residual clays related to the Mississippian-Pennsylvanian unconformity in the central Appalachian basin requires more than just field observation. In the case of the Route 56 paleosol, lithology alone (i.e., abundant siderite) might lead one to infer that it is a gleyed underclay formed in a poorly-drained environment. However, micromorphology combined with geochemical data indicates a

polygenetic origin involving intense leaching as well as gleying. In essence, this study further demonstrates the utility of the micromorphological approach to the identification of interfluvial paleosols developed by McCarthy and Plint (1998). However, the pedogenic history of the Route 56 paleosol is not a simple product of base-level change because of the large time gap represented by the Mississippian-Pennsylvanian unconformity.

Despite its fundamental stratigraphic importance, the Mississippian-Pennsylvanian unconformity has remained a cryptic feature at many localities in the central Appalachian basin (Brezinski, 1985; Edmunds, 1993). Although much focus has been placed on erosional features of the unconformity, this study adds to the growing body of evidence that paleosols can be equally diagnostic features (Cecil and others, 1992; Beuthin, 1997). In particular, it seems likely that there are more residual clays and maturely developed paleosols associated with interfluvial paleosols on the unconformity, especially on the northern and western flanks of the basin where prolonged exposure occurred. Continued analysis of clayey paleosols associated with the Mississippian-Pennsylvanian boundary should yield better evidence for the unconformity as well as reveal details of pedogenic history in relation to unconformity development.

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EVIDENCE FOR LARGE-SCALE REWORKING OF CAMPANIAN SEDIMENTS INTO THE UPPER MAASTRICHTIAN PEEDEE FORMATION AT BURCHES FERRY, SOUTH CAROLINA

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ABSTRACT

A 44-ft-deep corehole (FLO-311) was drilled at Burches Ferry, Florence County, S.C., in order to document the lithologic and paleontologic characteristics of the boundary between the upper Campanian Donoho Creek Formation and the upper Maastrichtian Peedee Formation. Palynomorph and calcareous nannofossil data provide detailed age control for these sediments. Examination of calcareous nannofossil assemblages shows that sediments from the type locality of the Peedee Formation consist largely of reworked Campanian materials. Robust Campanian species such as *Reinhardtites anthophorus*, *Reinhardtites levis*, and *Stoverius asymmetricus* commonly are found reworked into the Maastrichtian Peedee Formation in its entirety. Therefore, identification of lower upper Maastrichtian sediments is based on the presence of background assemblages rather than on the more traditional marker species. The presence of species restricted to the latest Campanian at the Donoho Creek-Peedee contact in outcrop indicates that the entire sedimen-

tary package represented by calcareous nannofossil Zones CC23 and CC24 has been removed.

INTRODUCTION

Exposures of Upper Cretaceous sediments on the Great Pee Dee River, Florence County, South Carolina, and particularly at Burches Ferry on the west bank of the river near Pamlico (Figure 1), have been investigated for more than 150 years. First described as the "Peedee beds" by Ruffin (1843), the Peedee Formation was formally described by Stephenson (1923) who designated Burches Ferry as the type locality.

The nature of the contact between the Peedee and the underlying Donoho Creek Formation of the Black Creek Group (Sohl and Owens, 1991) at Burches Ferry has been the subject of numerous studies. Benson (1969) and Van Nieuwenhuise and Kanies (1976) concluded that there was no obvious sedimentary or structural evidence of an unconformity at Burches Ferry. Hattner and Wise (1980) did not report an unconformity at the base of the Peedee Formation at Burches Ferry but did note the high abundance of *Reinhardtites anthophorus*, a Campa-

samples were later processed and examined for calcareous nannofossils and palynomorphs, but the results were inconsistent with those of earlier reports and geologic observations at the outcrop.

In an effort to improve fossil recovery and examine more of the stratigraphic section at the Peedee type locality, a short core was drilled approximately 25 feet from the edge of the right bank of the river bluff by the U.S. Geological Survey in August 1999 (Figure 1). The hole is located at 34°03'52"N latitude and 79°31'43"W longitude; it terminated at a depth of 44 ft. The land surface elevation is 44.5 ft as determined from a local river level gauge and measurement from water level at the outcrop. The hole was designated as Florence (FLO)-311 by the S.C. Department of Natural Resources. The resulting core contains sediments representative of both the Donoho Creek and Peedee Formations.

STRATIGRAPHY

The FLO-311 core encountered 23.7 ft of Donoho Creek Formation, overlain by 15 ft of Peedee Formation and 5.3 ft of Quaternary sand. The face of the bluff at Burches Ferry exposes only the lower 8-ft of the Peedee and the uppermost 3 to 6 ft of the Donoho Creek, depending on water level in the river (Figure 1). Therefore, the following discussion encompasses more section than is exposed in the bluff.

Donoho Creek Formation

The FLO-311 core encountered the Donoho Creek Formation from a depth of 20.3 ft to 44.0 ft below land surface (TD of the well). The Donoho Creek Formation consists predominantly of very fine- to fine-grained, moderately sorted, micaceous, calcareous clayey sands. The beds are thin and parallel and are accentuated by small changes in clay content. Clay-rich layers are dark gray to black and commonly contain lignitic debris. Shell fragments are common throughout the cored interval, and the sediments are moderately to heavily bioturbated. Benthic and planktic foraminifera, ostracodes, palynomorphs, and calcareous

nannofossils are common to abundant. The presence of phosphate granules and a distinct color change suggests a hiatus within the Donoho Creek Formation at 30.1 ft depth, but this interpretation is not supported by a change in the calcareous nannofossil or palynomorph assemblages (Figure 2).

Peedee Formation

The FLO-311 core encountered 15 ft of the Peedee Formation between 5.3 and 20.3 ft (Figure 2). The contact between the Donoho Creek and the Peedee in FLO-311 is a sharp, disconformable surface. The basal 2.5 feet of the Peedee Formation is clayey sand containing rounded phosphate pebbles and clasts up to 0.5 inches in length. Clay clasts up to 3 inches in length are scattered throughout this basal interval. Sand-sized phosphate occurs up to three ft above the contact in concentrations up to five percent. Vertebrate remains are concentrated in this lag deposit, as are pieces of permineralized wood up to 0.5 inches in length. Fragments of *Belemnetella americanus* are also common. Re-working of calcareous nannofossils is extensive throughout this basal interval. The remainder of the Peedee Formation consists of very fine- to coarse-grained, poorly sorted, massive, calcareous clayey sands rich in shell fragments with trace amounts of mica. Bedding is generally absent except for two indurated zones in the core between 10 and 11.5 ft, and 12.9 and 13.1 ft. The lower zone occurs in the outcrop just below the vegetation line. The uppermost 2.5 ft of the formation is noncalcareous with mottled coloring, varying between light olive gray and moderate yellowish brown. This interval most likely represents weathering and leaching at the top of the formation. Benthic and planktic foraminifera, ostracodes, calcareous nannofossils and palynomorphs are common.

Quaternary Deposits

The Peedee Formation is unconformably overlain by 5.3 ft of weathered, clayey, noncalcareous sand that has been described as Pleistocene (Stephenson, 1923), "post Cretaceous"

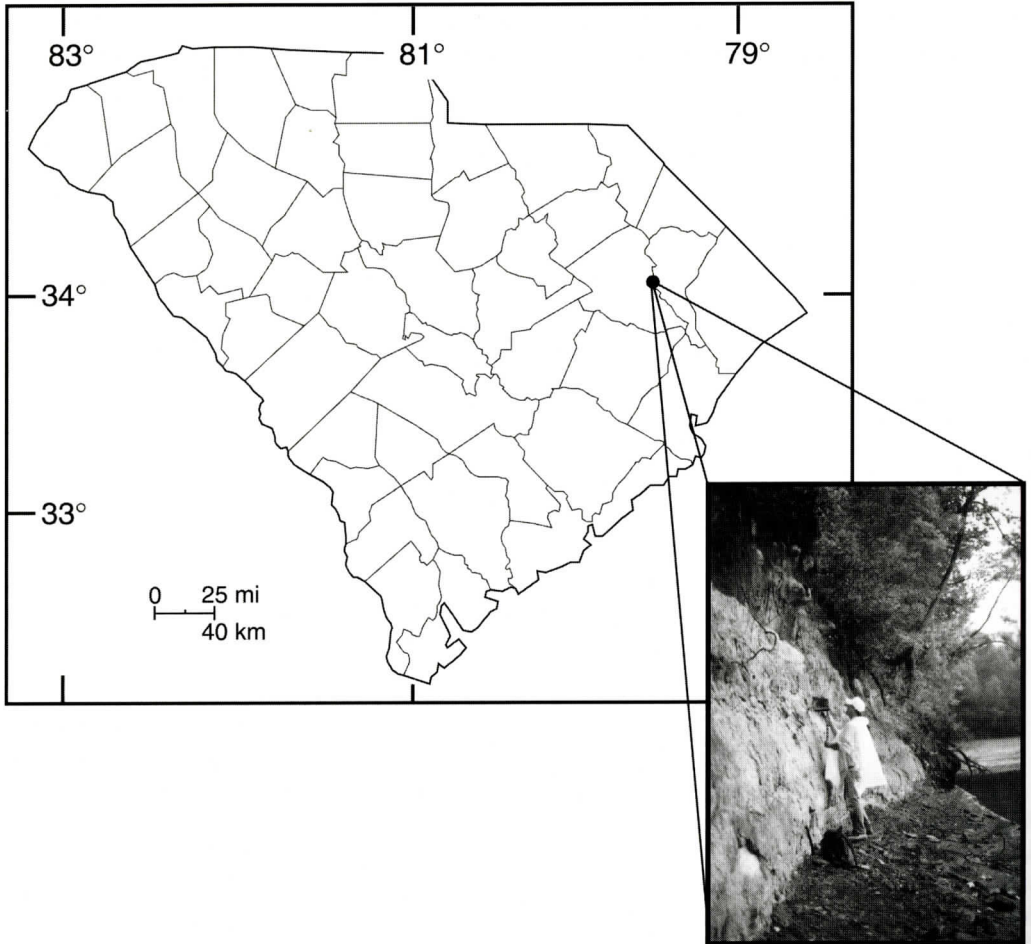


Figure 1. Map showing the location of the FLO-311 corehole drilled by the U.S. Geological Survey in South Carolina at Burches Ferry, type locality of the Pee Dee Formation. The inset picture is the bluff at Burches Ferry on the Great Pee Dee River. The bluff is the type locality of the Pee Dee Formation in its entirety and the resistant ledge just above water level is the uppermost Donoho Creek Formation.

nian calcareous nannofossil, in the Maastrichtian sediments of the Pee Dee (Hattner and Wise, 1979). Parr and Lawrence (1985) and Lawrence and Hall (1987), on the other hand, concluded that there was ample evidence to support a disconformity at the base of the Pee Dee Formation, a conclusion supported by Sohl and Owens (1991).

The present study uses outcrop and core data to address the nature of the formational contact at Burches Ferry and the changes in paleontological characteristics across this boundary. This study also provides detailed biostrati-

graphic correlation between calcareous nannofossils and palynomorphs for the late Maastrichtian of the region. Hence, these data facilitate correlation between updip palynomorph-bearing, nonmarine sediments and downdip, calcareous marine sediments of the Coastal Plain.

In the summer of 1997, the authors visited the Burches Ferry outcrop on the Great Pee Dee River in Florence County, South Carolina. Systematic samples were collected from the top of the exposure (Pee Dee Formation) down to the water level (Donoho Creek Formation). These

(Benson, 1969), and Pliocene (Van Nieuwenhuise and Kanesh, 1976).

SAMPLE PREPARATION

Calcareous Nannofossils

Calcareous nannofossil samples were extracted from the central portions of freshly broken cores in order to avoid contamination. Samples were dried in a convection oven in order to remove any residual water and placed in vials for long-term storage. A small portion of sample material was placed in a beaker and allowed to settle for one minute through 20 ml of water in order to remove the coarse size fraction. A second settling time of 10 minutes was used to concentrate the silt size fraction. Smear slides were prepared from the resultant slurry. Cover slips were adhered to 1 x 3 inch glass slides using Norland Optical Adhesive (NOA-65), a clear adhesive that bonds glass to glass and cures when exposed to ultraviolet light. Approximately one sample was taken every foot for a total of 38 samples. All slides were examined using a Zeiss Photomicroscope III that is housed at the U.S. Geological Survey National Center, Reston, Va. Slides are curated at the same location.

Palynomorphs

Four palynologic samples extracted from the central portions of freshly broken cores from the FLO-311 well, and five samples from the outcropping units at Burches Ferry, were processed using the palynological techniques discussed by Wood and others (1996). Approximately 20 g of each sample were physically disaggregated, placed in an acid-resistant beaker, and covered with 35 percent hydrochloric acid (HCl) until reaction ceased (usually 6 to 24 hours). The residue was centrifuged in de-ionized water for 10 minutes at 2000 rpm; this process was repeated three times, and the supernate was decanted after each centrifugation. The residue was covered with 70 percent hydrofluoric acid (HF) and left to react for 2 to 24 hours, at which time it was centrifuged three

times as described above. Following the final centrifugation, the sample was suspended in a zinc bromide (ZnBr) solution (specific gravity = 1.9) and centrifuged for 10 minutes at 2000 rpm. The float resulting from this centrifugation was removed from the centrifuge tube with a pipette and placed in a test tube. The sink was visually examined to determine if any palynomorphs were present; if none were observed, the sink was discarded. The test tube containing the heavy liquid float was filled with de-ionized water and subjected to ultrasonic vibrations for a period that ranged from 30 seconds to 5 minutes. Periodic visual inspections were made to ensure that the palynomorphs were not affected by the ultrasound. The residue was washed through a 5 micrometer nylon screen, and the material that remained on the screen was washed into a watch glass; the discard that passed through the screen was visually examined to ensure that it contained no palynomorphs. The organic material in the watch glass was panned, and the palynomorphs were removed with a pipette and placed in a 4 dram vial. A drop of the residue in the vial was placed on a 23 x 30 mm glass cover slip where it was thoroughly mixed with a drop of Natrosol and spread evenly across the surface of the cover slip. The cover slip containing the residue-Natrosol mixture was placed on a warming tray, covered, and left to dry for 4 hours. When dried, the cover slip was adhered to a 1 x 3 inch glass slide with Elvasite and placed under an ultraviolet light source to cure. All palynological slides were examined using an Aus Jena JENAVAL contrast GF-Papo research microscope housed in the Department of Geological Sciences, Clemson University. All palynologic slides are curated at the same location.

BIOSTRATIGRAPHIC FRAMEWORK

The calcareous nannofossil zones used in this report reflect the current understanding of their chronology (Burnett 1998; Self-Trail, 2001). Calcareous nannofossil biostratigraphy is based on the zonations of Sissingh (1977) and Perch-Nielsen (1985). A palynologic zonation for the latest Cretaceous sections of the Caroli-

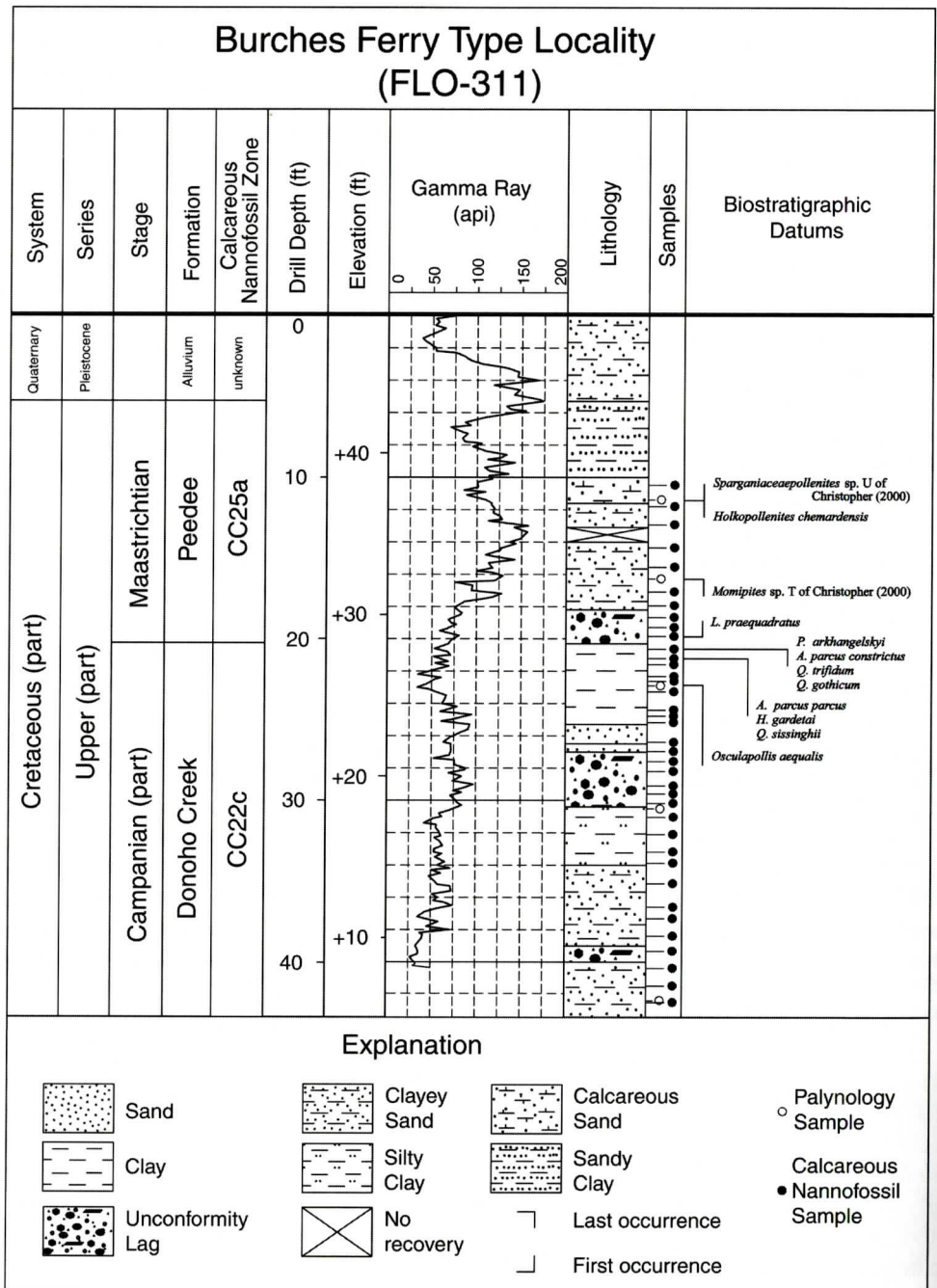


Figure 2. Stratigraphy and geophysical logs for the FLO-311 core.

nas did not exist prior to the work by Christopher and Sohl (1983). An informal and unpublished biostratigraphic framework based on their work was established in the 1980's to assist with mapping efforts then underway in the Carolinas (Owens, 1988; Owens and Sohl, 1989; Sohl and Owens, 1991). Since the palynologic framework was developed, interdisciplinary studies that utilized both calcareous nannofossil and palynologic biostratigraphy have been the basis for refining and formalizing the palynologic zonation (Self-Trail and Gohn, 1996; Christopher and others, 1999; Christopher and Prowell, in press). The palynologic zonation has proven particularly useful in correlating and determining chronostratigraphic relationships among the marine, down-dip sediments and the more terrestrial, noncalcareous, up-dip sediments of the region.

Biostratigraphic Results

Donoho Creek Formation

Twenty-eight calcareous nannofossil samples were collected from the Donoho Creek Formation in FLO-311 (Figure 2). In these samples, the co-occurrence of calcareous nannofossil species *Aspidolithus parvus constrictus*, *A. parvus parvus*, *Quadrum sissinghii*, *Q. trifidum*, *Reinhardtites anthophorus*, and *R. levis* is indicative of a latest Campanian age (Figure 3). Absence of the Maastrichtian species *Lithraphidites quadratus*, *Micula praemurus*, and *Lithraphidites grossopectinatus* further corroborates this age. The Donoho Creek strata at Burches Ferry are therefore equivalent to calcareous nannofossil Zone CC22c.

Calcareous nannofossil samples were similarly collected from the Donoho Creek strata exposed above water level in the outcrop. One sample, collected 0.5 ft (20.8 ft level in FLO-311) below the base of the Pee Dee strata, contains *Cylindralithus biarcus*, *Aspidolithus parvus constrictus* and *Reinhardtites levis*, but does not have *R. anthophorus*, indicating a Zone CC23 designation. This sample most likely represents a burrow filled with the transgressive lag deposits at the base of the Pee Dee Formation. Samples indicative of Zone CC24 were not

identified from either outcrop or core.

Pollen samples collected at 22.9 ft, 30.5 ft and 42.7 ft in the FLO-311 core contain the palynomorph species *Osculapollis aequalis* and *?Holkopollenites* sp. I (Figure 4). These species are indicative of pollen Zone B (see Sohl and Owens, 1991; their Figure 11-7), which is equivalent to calcareous nannofossil Zones CC22 and CC23 (Figure 2). All of the palynomorph samples contain moderate numbers of dinoflagellates.

Calcareous nannofossil and palynologic data support a late Campanian age for the Donoho Creek Formation. These data indicate that the Donoho Creek beds at Burches Ferry are within Zone CC22c, and perhaps older zones exist at lower stratigraphic levels and that younger beds (Zone CC23) must have been present prior to the Maastrichtian transgression. The presence of calcareous nannofossils and dinoflagellates indicates deposition of the Donoho Creek sediment occurred in an open marine setting. In addition, the abundance of pollen and other forms of land-derived organic matter indicates a significant terrestrial influence. Hence, a near-shore-marine paleoenvironment of deposition is inferred for the Donoho Creek section.

Pee Dee Formation

Eleven calcareous nannofossil samples were collected from the Pee Dee Formation in FLO-311 (Figure 2) and in outcrop above the basal contact (equals 20.3 ft in FLO-311). Calcareous nannofossil species from the core samples include *Markalius inversus*, *Lithraphidites praequadratus*, and *Prediscosphaera grandis*, which are indicative of nannofossil Zone CC25a. *Reinhardtites levis*, *R. anthophorus*, and *Stoverius asymmetricus* are also common, indicating reworking of Campanian Zone CC22c sediments into the Maastrichtian Pee Dee Formation (Figure 5). *Lithraphidites quadratus*, the basal marker species for Zone CC25b, is not recorded from any of the samples; thus, the Pee Dee Formation at Burches Ferry is probably restricted to late Maastrichtian Zone CC25a (Figure 2).

Two palynomorph samples were collected

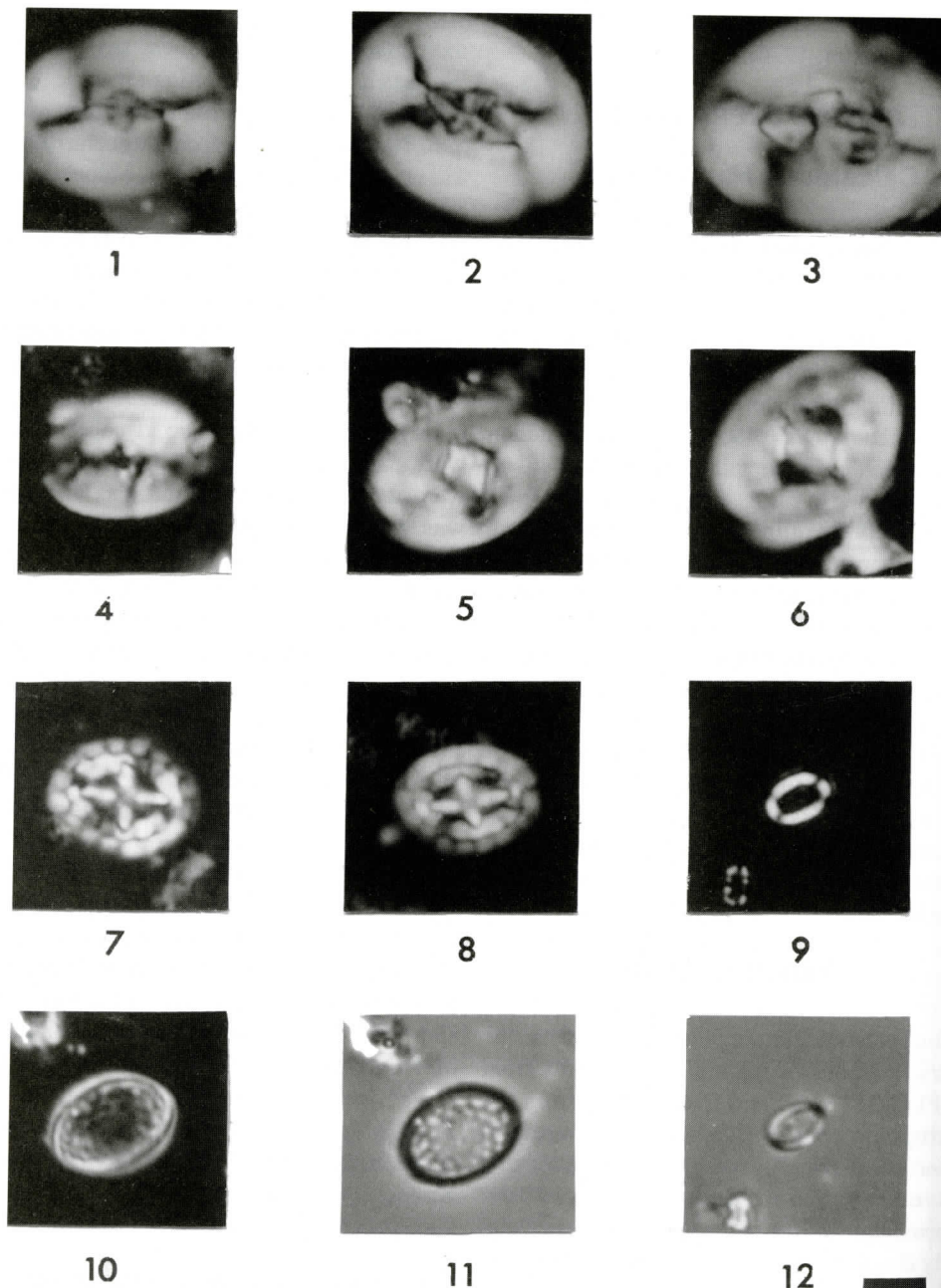


Figure 3. Calcareous nannofossils from the FLO-311 core, Donoho Creek Formation, 27.5 ft, Subzone CC22c, South Carolina. All specimens are in-place Campanian fossils. 1) *Aspidolithus parvus constrictus* (Hattner and others 1980) Perch-Nielsen 1984, crossed polars (XP). 2) *Aspidolithus parvus parvus* (Stradner 1963) Noel 1969, XP. 3) *Aspidolithus parvus expansus* Wise and Watkins in Wise (1983), XP. 4-5) *Reinhardtites levis* Prins and Sissingh in Sissingh 1977, XP. 6) *Reinhardtites anthophorus* (Deflandre 1959) Perch-Nielsen 1968, XP. 7-8) *Prediscosphaera arkhangelskyi* (Reinhardt 1965) Perch-Nielsen 1984, XP. 9, 12) *Amphizygus minimus* Bukry 1969, XP; 12, Same specimen, phase contrast (PC). 10-11) *Angulofenestrellithus snyderi* Bukry 1969, XP; 11, Same specimen, PC. Scale bar represents 5µm.

PEEDEE FORMATION AT BURCHES FERRY, SOUTH CAROLINA

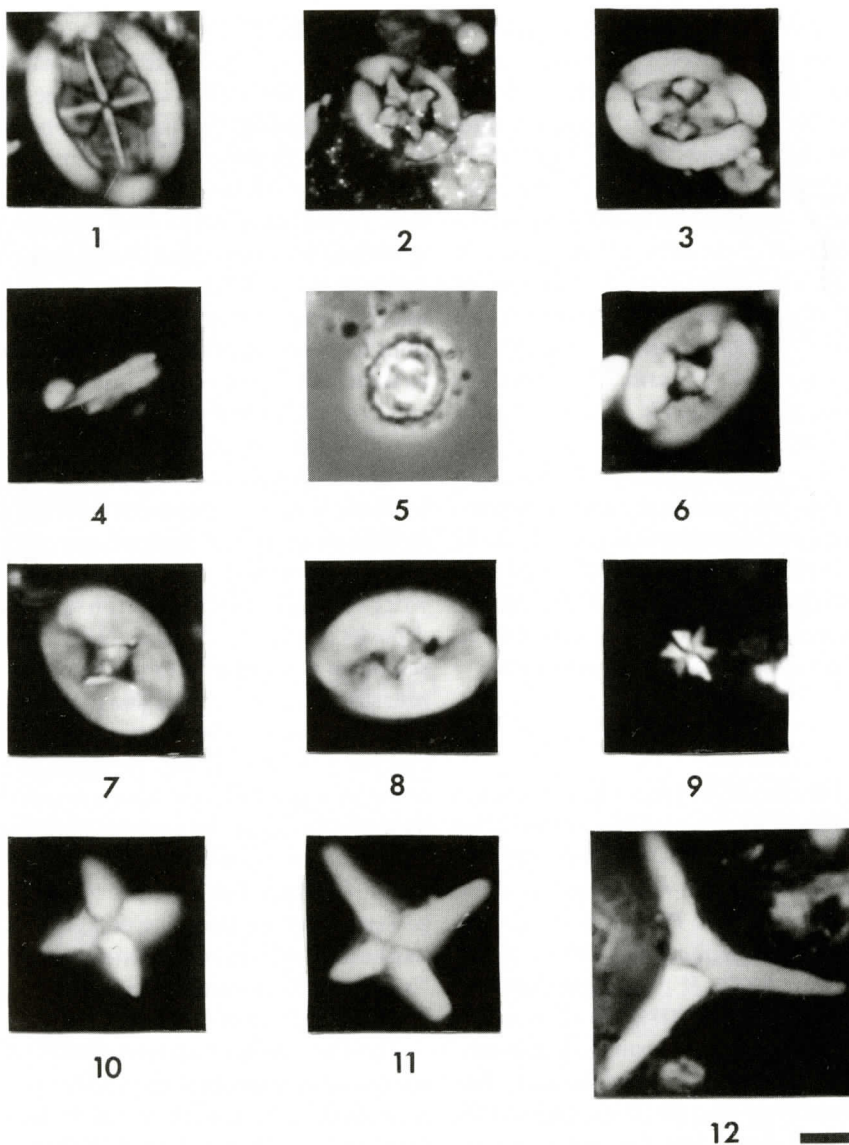


Figure 5. Calcareous nannofossils from the FLO-311 core. 1) *Arkhangelskiella cymbiformis* Vekshina 1959, crossed polars (XP), 14.2 ft, Subzone CC25a, in-place Maastrichtian fossil. 2) *Eiffellithus parallelus* Perch-Nielsen 1973, XP, 16.6 ft, Subzone CC25a, PDF, in-place Maastrichtian fossil. 3) *Arkhangelskiella cymbiformis* Vekshina 1959, XP, 15.5 ft, Subzone CC25a, PDF, in-place Maastrichtian fossil. 4) *Lithraphidites praequadratus* Roth 1978, XP, 15.5 ft, Subzone CC25a, PDF, in-place Maastrichtian fossil. 5) *Stoverius asymmetricus* (Bukry 1969) Perch-Nielsen 1984, phase contrast (PC), 16.6 ft, Subzone CC25a, PDF, reworked Campanian fossil. 6-7) *Reinhardtites anthophorus* (Deflandre 1959) Perch-Nielsen 1968, XP, 14.2 ft, Subzone CC25a, PDF, reworked Campanian fossils. 8) *Reinhardtites levis* Prins and Sissingh in Sissingh 1977, XP, 14.2 ft, Subzone CC25a, PDF, reworked Campanian fossil. 9) *Hexalithus gardetae* Bukry 1969, XP, 27.5 ft, Subzone CC22c, DCF, in-place Campanian fossil. 10-11) *Quadrum sissinghii* Perch-Nielsen 1986, XP, 27.5 ft, Subzone CC22c, DCF, in-place Campanian fossils. 12) *Quadrum trifidum* (Stradner in Stradner and Papp, 1961) Prins and Perch-Nielsen in Manivit and others 1977, XP, 27.5 ft, Subzone CC22c, DCF, in-place Campanian fossil. DCF, Donoho Creek Formation; PDF, Peedee Formation. Scale bar represents 5µm.

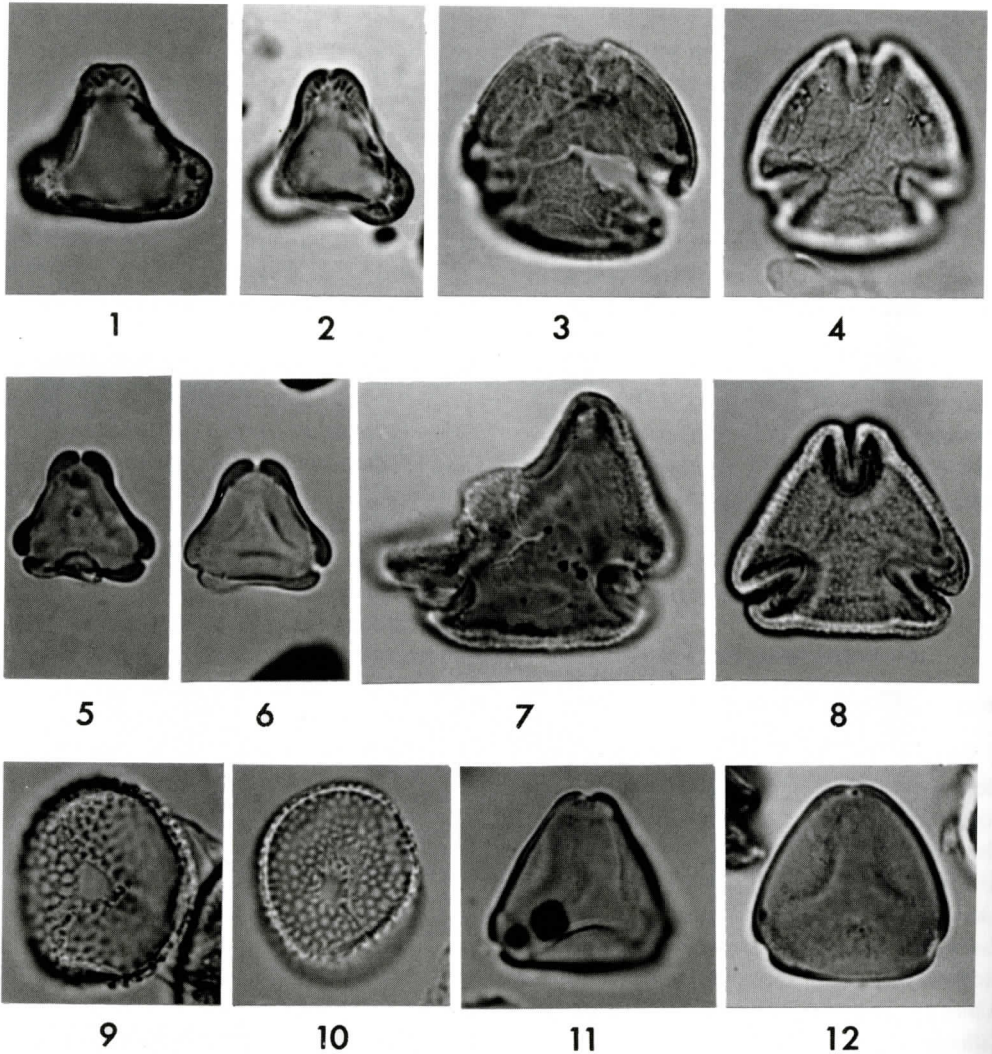


Figure 4. Palynomorphs from the palynological biozones represented in the FLO-311 core. All photomicrographs X1,000; all photomicrographs in brightfield illumination. 1-2) *Osculapollis aequalis* Tschudy 1975 (= NO-1 of Wolfe, 1976); pre-Maastrichtian; specimen 1 from the FLO-311 core at 22.9 ft, specimen 2 from the P21TA core (Aiken Co., SC) at 660 ft. 3-4) ?*Holkopollenites* sp. I (= CP3E-1 of Wolfe, 1976); pre-Maastrichtian; specimen 3 from the outcropping Donoho Creek Formation at Burches Ferry, 3 feet below the contact with the Peedee Formation, specimen 4 from the ORG-393 core (Orangeburg Co., SC) at 741.1 ft. 5-6) New Genus C sp. T; Ct Zone; specimen 5 from the outcropping Peedee Formation at Burches Ferry, 1 foot above the contact with the Donoho Creek Formation, specimen 6 from the ORG-393 core (Orangeburg Co., SC) at 446.0 ft. 7-8) *Holkopollenites chemardensis* Fairchild, in Stover, Elsik and Fairchild 1966; LAD at the top of the Hc Zone; specimen 7 from the outcropping Donoho Creek Formation at Burches Ferry, 3 feet below the contact with the Peedee Formation, specimen 8 from the ORG-393 core (Orangeburg Co., SC) at 1344 ft. 9-10) *Sparganiaceapollenites* sp. U; LAD at the top of the Su Zone; specimen 9 from the FLO-311 core at 11.4 ft, specimen 10 from the C-15 core (Jasper Co., SC) at 1120 ft. 11-12) *Plicatopollis cretacea* Frederiksen and Christopher 1978 (= NN-2 of Wolfe, 1976); range restricted to the Maastrichtian of the Atlantic Coastal Plain Province; specimen 11 from the outcropping Peedee Formation at Burches Ferry, 1 foot above the contact with the Donoho Creek Formation, specimen 12 from the MAN-1A core (Sumter Co., SC) at 150 ft.

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Figure 6. Calcareous nannofossil occurrences from core FLO-311. Sample abundance: C, common, or 1 to 10 specimens per field of view at X1,200; F, frequent, or 1 specimen per 1 to 10 fields of view at X1,200; R, rare, or 1 specimen per 11-100 fields of view; B, barren of calcareous nannofossils. Preservation: G, good; M, moderate; P, poor. Species abundance: C, common or 1 specimen per 1 to 10 fields of view at X1,200; F, frequent or 1 specimen per 11-100 fields of view at X1,200; R, rare or 1 specimen per >100 fields of view at X1,200. Other symbols: ? = questionable occurrence; rw = reworked specimen.

late Campanian										Maastrichtian	Age
Donoho Creek Fm.										Peedee Fm.	Formation
CC22c										CC25a	Nannofossil Zones (Perch-Nielsen, 1985)
42.8	42.7	42.6	42.5	42.4	42.3	42.2	42.1	42.0	41.9	41.8	Depth (ft)
											<i>Acuturris scotus</i>
											<i>Ahmuelerella octoradiata</i>
											<i>Ahmuelerella regularis</i>
											<i>Amphizygus minimus</i>
											<i>Amphizygus</i> sp.
											<i>Angulofenestrellithus snyderi</i>
											<i>Arkhangelskiella cymbiformis</i>
											<i>Arkhangelskiella specillata</i>
											<i>Aspidolithus parvus constrictus</i>
											<i>Aspidolithus parvus expansus</i>
											<i>Aspidolithus parvus parvus</i>
											<i>Biantholithus sparsus</i>
											<i>Biscutum constans</i>
											<i>Biscutum coronum</i>
											<i>Biscutum ellipticum</i>
											<i>Biscutum zulloi</i>
											<i>Biscutum</i> sp. A
											<i>Biscutum</i> sp. B
											<i>Braarudosphaera bigelowii</i>
											<i>Broinsonia bevieri</i>
											<i>Broinsonia dentata</i>
											<i>Broinsonia? verecundia</i>
											<i>Bukryaster</i> sp.
											<i>Calcutites obscurus</i>
											<i>Ceratolithoides aculeus</i>
											<i>Ceratolithoides arcuatus</i>
											<i>Chiasiozygus amphipons</i>
											<i>Chiasiozygus fessus</i>
											<i>Chiasiozygus litterarius</i>
											<i>Chiasiozygus trabalis</i>
											<i>Chiasiozygus</i> spp.
											<i>Corollithion exiguum</i>
											<i>Corollithion madagaskarensis</i>
											<i>Corollithion signum</i>
											<i>Cretarhabdus conicus</i>
											<i>Cretarhabdus multicaus</i>
											<i>Cretarhabdus schizobrachiatus</i>
											<i>Cribrosphaera circula</i>
											<i>Cribrosphaeraella ehrenbergii</i>
											<i>Cribrosphaeraella</i> sp.
											<i>Cyclagelosphaera margerelii</i>
											<i>Cylindralithus nudus</i>
											<i>Cylindralithus sculptus</i>
											<i>Cylindralithus serratus</i>
											<i>Discorhabdus ignotus</i>
											<i>Dodekapodorhabdus noeliae</i>
											<i>Eiffellithus gorkae</i>
											<i>Eiffellithus parallelus</i>
											<i>Eiffellithus turiseiffelii</i>
											<i>Eiffellithus</i> sp. D
											<i>Eiffellithus</i> sp. S
											<i>Flabellites</i> sp.
											<i>Gartnerago obliquum</i>
											<i>Glaukolithus bicrescenticus</i>
											<i>Glaukolithus diplogrammus</i>
											<i>Goniolithus fluckigeri</i>
											<i>Helicolithus trabeculatus</i>
											<i>Hexalithus gardetae</i>
											<i>Kamptnerius magnificus</i>
											<i>Kamptnerius punctatus</i>
											<i>Lilrhapidites carniolensis</i>
											<i>Lilrhapidites? charactozomo s.l.</i>
											<i>Lilrhapidites praequadratus</i>
											<i>Lordia</i> sp.
											<i>Lucianorhabdus cayeuxii</i>
											<i>Manivitella pemmatoides</i>
											<i>Markalius inversus</i>
											<i>Marthasterites inconspicuus</i>
											<i>Microrhabdulus attenuatus</i>
											<i>Microrhabdulus belgicus</i>
											<i>Microrhabdulus decoratus</i>
											<i>Microrhabdulus undosus</i>
											<i>Micula concava</i>
											<i>Micula cubiformis</i>
											<i>Micula decussata</i>
											<i>Micula</i> sp.
											<i>Monomarginatus</i> sp.

from the FLO-311 core at 16.3 and 11.4 ft. The sample at 16.3 ft contains the single occurrence of New Genus C sp. T and *Holkopollenites chemardensis*. New Genus C sp. T is the marker species for the Ct Zone of Christopher and Prowell (in press). Pollen markers in this sample were extremely rare due to the overwhelming dominance of dinoflagellate cysts. A sample at 11.4 ft yielded a single specimen of *Sparganiaceapollenites* sp. U, along with rare occurrences of *Plicatopollis cretacea*, and *Holkopollenites chemardensis*. This sample is tentatively assigned to the Hc Zone of Christopher and Prowell (in press; Figure 2). The presence of these two palynological zones in conjunction with a prominent increase in gamma radiation on the geophysical log (Figure 2) suggests a disconformity at a depth of 14 ft in FLO-311. Pollen Zones Ct and Hc are equivalent to calcareous nannofossil Subzone CC25a and basal Subzone CC25b elsewhere in S.C.

DISCUSSION

Sohl and Owens (1991) reported the first appearance datum of the planktic foraminifer *Gansserina gansseri* at Burches Ferry. In conjunction with macrofossil data from this same locality, they placed the Peedee Formation in that part of the *Exogyra costata* Zone that lies above their Campanian *E. cancellata* Subzone. As proposed by Odin (2001), the *G. gansseri* zone encompasses the Campanian-Maastrichtian boundary and has its top in the early late Maastrichtian. Similarly, the calcareous nannofossil and palynologic data reported in this report support a late Maastrichtian age for the Peedee Formation at Burches Ferry.

Detailed examination of both outcrop and corehole material has lead the authors to conclude that the Peedee Formation exposed at its type locality, Burches Ferry, consists of massively reworked upper Campanian sediments deposited during the first of several late Maastrichtian transgressions. The concentration of phosphate pebbles, bone fragments, rip-up clasts and shell fragments in the basal Peedee Formation at Burches Ferry strongly suggests that the contact between the Peedee and the un-

derlying Donoho Creek Formation is unconformable. Stephenson (1923), Cooke (1936), Swift and Heron (1969) and Lawrence and Hall (1987) recognized that the contact between the Peedee and Donoho Creek Formations is unconformable, but they did not document the extent to which Campanian sediments are reworked into the upper Maastrichtian sediments at Burches Ferry. They also did not note the significant removal of Donoho Creek strata prior to or during the first Peedee transgression as indicated by the removal of all Zone CC23 and Zone CC24 beds.

Calcareous nannofossils in the Peedee Formation at Burches Ferry are massively reworked. Robust Campanian species such as *Reinhardtites levis*, *R. anthophorus*, and *Stoverius asymmetricus* are commonly found reworked well above the transgressive lag deposits (Figure 6). At first, the Campanian assemblage was difficult to distinguish from the Maastrichtian assemblage due to the propensity of *R. levis* and *R. anthophorus*, both marker species for Subzone CC22c. However, the occurrence of *Cylindralithus biarcus*, a calcareous nannofossil commonly associated with Zone CC23, at the base of the Peedee Formation meant that any Zone CC22c marker species present above the basal lag deposit must be reworked. Therefore, a detailed study of the "background" assemblage in Peedee and Donoho Creek sediments was made.

The Campanian background assemblage is composed of less robust species that are not hardy enough to be reworked consistently; therefore, in-place Campanian sediments were distinguished from Maastrichtian sediments based on secondary assemblages rather than on marker species. In FLO-311, *Acuturris scotus*, *Aspidolithus parvus parvus*, *A. parvus constrictus*, *Hexalithus gardetae*, *Prediscosphaera arkhangel'skyi*, *Quadrum sissinghii*, and *Q. trifidum* all have their last occurrence at 20.6 ft (Figure 6). This coincides with a lithologic break between the clays of the Donoho Creek Formation and the clayey sands of the Peedee Formation.

Figure 6 shows that *Reinhardtites anthophorus* and *Stoverius asymmetricus*, which have

their last occurrences in the late Campanian, are reworked up into the Maastrichtian sediments to 11.7 ft, for a total reworked thickness of approximately 9 feet. Additionally, *Reinhardtites levis*, which goes extinct in the early Maastrichtian (Zone CC24), is reworked to the top of the calcareous Peedee Formation at 10.4 ft. These three species are very robust forms that do not easily dissolve and are therefore subject to reworking.

Evidence for reworking of palynomorph marker taxa was not observed in preparations from either the Burches Ferry outcrop or FLO-311 core samples. This apparent lack of reworking may reflect the extremely low abundance of palynomorph marker taxa more than it does a discrepancy between the palynological and calcareous nannofossil data. Typically, the palynomorph taxa constitute less than one percent of an assemblage. In some preparations, only a single specimen is present on an entire slide, and under these circumstances, it is highly unlikely that reworked specimens would be detected. Because of the low abundance of marker taxa in the Burches Ferry samples, it is difficult to identify specimens that display significant morphological details. For this reason, photomicrographs of marker taxa on Figure 4 from the Burches Ferry samples are supplemented with more photogenic specimens from the same stratigraphic intervals but different geographic locations.

CONCLUSIONS

Evaluation of calcareous nannofossils shows that the type section of the Peedee Formation at Burches Ferry consists of reworked Donoho Creek sediments. The presence of a Zone CC23 nannofossil assemblage at the basal contact of the Peedee but not in the underlying Donoho Creek strata indicates that the entire volume of sedimentary section deposited during Zone CC23 time has been removed. In addition to the removal of this large portion of the Donoho Creek Formation, the nannofossil evidence indicates that these strata and their fossil suite have been incorporated well up into the Peedee Formation.

The lithostratigraphic scenario documented at Burches Ferry has been observed in various other coreholes in the South Carolina Coastal Plain (Edwards and others, 1999; Self-Trail, 2001) and is often accompanied by reworking of calcareous microfossils of Campanian age into the overlying Maastrichtian sediments. Future researchers are well advised to study background assemblages in addition to marker species to avoid incorrect age assignments and stratigraphic miscorrelation.

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JEAN M. SELF-TRAIL AND OTHERS

late Campanian	Maastrichtian	Age
Donoho Creek Fm.	Peedee Fm.	Formation
CC22c	CC25a	Nannofossil Zones (Perch-Nielsen, 1985)
206 211 217 222 225 226 227 228 229 230 231 232 233 234 235 236 237 238 239 240 241 242 243 244 245 246 247 248 249 250 251 252 253 254 255 256 257 258 259 260 261 262 263 264 265 266 267 268 269 270 271 272 273 274 275 276 277 278 279 280 281 282 283 284 285 286 287 288 289 290 291 292 293 294 295 296 297 298 299 300 301 302 303 304 305 306 307 308 309 310 311 312 313 314 315 316 317 318 319 320 321 322 323 324 325 326 327 328 329 330 331 332 333 334 335 336 337 338 339 340 341 342 343 344 345 346 347 348 349 350 351 352 353 354 355 356 357 358 359 360 361 362 363 364 365 366 367 368 369 370 371 372 373 374 375 376 377 378 379 380 381 382 383 384 385 386 387 388 389 390 391 392 393 394 395 396 397 398 399 400 401 402 403 404 405 406 407 408 409 410 411 412 413 414 415 416 417 418 419 420 421 422 423 424 425 426 427 428 429 430 431 432 433 434 435 436 437 438 439 440 441 442 443 444 445 446 447 448 449 450 451 452 453 454 455 456 457 458 459 460 461 462 463 464 465 466 467 468 469 470 471 472 473 474 475 476 477 478 479 480 481 482 483 484 485 486 487 488 489 490 491 492 493 494 495 496 497 498 499 500 501 502 503 504 505 506 507 508 509 510 511 512 513 514 515 516 517 518 519 520 521 522 523 524 525 526 527 528 529 530 531 532 533 534 535 536 537 538 539 540 541 542 543 544 545 546 547 548 549 550 551 552 553 554 555 556 557 558 559 560 561 562 563 564 565 566 567 568 569 570 571 572 573 574 575 576 577 578 579 580 581 582 583 584 585 586 587 588 589 590 591 592 593 594 595 596 597 598 599 600 601 602 603 604 605 606 607 608 609 610 611 612 613 614 615 616 617 618 619 620 621 622 623 624 625 626 627 628 629 630 631 632 633 634 635 636 637 638 639 640 641 642 643 644 645 646 647 648 649 650 651 652 653 654 655 656 657 658 659 660 661 662 663 664 665 666 667 668 669 670 671 672 673 674 675 676 677 678 679 680 681 682 683 684 685 686 687 688 689 690 691 692 693 694 695 696 697 698 699 700 701 702 703 704 705 706 707 708 709 710 711 712 713 714 715 716 717 718 719 720 721 722 723 724 725 726 727 728 729 730 731 732 733 734 735 736 737 738 739 740 741 742 743 744 745 746 747 748 749 750 751 752 753 754 755 756 757 758 759 760 761 762 763 764 765 766 767 768 769 770 771 772 773 774 775 776 777 778 779 780 781 782 783 784 785 786 787 788 789 790 791 792 793 794 795 796 797 798 799 800 801 802 803 804 805 806 807 808 809 810 811 812 813 814 815 816 817 818 819 820 821 822 823 824 825 826 827 828 829 830 831 832 833 834 835 836 837 838 839 840 841 842 843 844 845 846 847 848 849 850 851 852 853 854 855 856 857 858 859 860 861 862 863 864 865 866 867 868 869 870 871 872 873 874 875 876 877 878 879 880 881 882 883 884 885 886 887 888 889 890 891 892 893 894 895 896 897 898 899 900 901 902 903 904 905 906 907 908 909 910 911 912 913 914 915 916 917 918 919 920 921 922 923 924 925 926 927 928 929 930 931 932 933 934 935 936 937 938 939 940 941 942 943 944 945 946 947 948 949 950 951 952 953 954 955 956 957 958 959 960 961 962 963 964 965 966 967 968 969 970 971 972 973 974 975 976 977 978 979 980 981 982 983 984 985 986 987 988 989 990 991 992 993 994 995 996 997 998 999 1000	Depth (ft)	
		<i>Nannoconus</i> spp.
		<i>Neobiscutum</i> sp.
		<i>Neocrepidolithus cohenii</i>
		<i>Neocrepidolithus neocrassus</i>
		<i>Percivalia porosa</i>
		<i>Petrobrasiella bowenii</i>
		<i>Placozygus ponticulus</i>
		<i>Placozygus sigmoides</i>
		<i>Placozygus spiralis</i>
		<i>Podorhabdus? elkefensis</i>
		<i>Prediscosphaera arhangelskyi</i>
		<i>Prediscosphaera cretacea</i>
		<i>Prediscosphaera grandis</i>
		<i>Prediscosphaera intercisa</i>
		<i>Prediscosphaera spinosa</i>
		<i>Prediscosphaera stoveri</i>
		<i>Quadrum gartneri</i>
		<i>Quadrum gothicum</i>
		<i>Quadrum sissinghii</i>
		<i>Quadrum trifidum</i>
		<i>Reinhardtites anthophorus</i>
		<i>Reinhardtites levis</i>
		<i>Repagulum parvidentatum</i>
		<i>Retacapsa angustiflora</i>
		<i>Retemediiformis teneretis</i>
		<i>Retemediiformis</i> sp.
		<i>Rhagodiscus angustus</i>
		<i>Rhagodiscus indistinctus</i>
		<i>Rhagodiscus plebeius</i>
		<i>Rhagodiscus reniformis</i>
		<i>Rhagodiscus splendens</i>
		<i>Rhombolothion rhombicum</i>
		<i>Rotelapillus crenulatus</i>
		<i>Rotelapillus munitus</i>
		<i>Rotelapillus octoradiatus</i>
		<i>Rucinolithus oriens</i>
		<i>Scapholithus fossilis</i>
		<i>Sollasites barringtonensis</i>
		<i>Sollasites lowei</i> s.l.
		<i>Sollasites</i> sp.
		<i>Staurolithes imbricatus</i>
		<i>Staurolithes minutus</i>
		<i>Stovarius asymmetricus</i>
		<i>Stradneria crenulata</i>
		<i>Tegumentum stradneri</i>
		<i>Tetralithus quadratus</i>
		<i>Tetrapodorhabdus decorus</i>
		<i>Thoracosphaera</i> spp.
		<i>Tortololithus hallii</i>
		<i>Tranolithus macleodae</i>
		<i>Tranolithus manifestus</i>
		<i>Tranolithus minimus</i>
		<i>Vagalapilla synquadriperforatus</i>
		<i>Vagalapilla</i> sp.
		<i>Vekshinella aachena</i>
		<i>Vekshinella stradneri</i>
		<i>Vekshinella</i> sp.
		<i>Watznaueria barnesae</i>
		<i>Watznaueria biporta</i>
		<i>Watznaueria bybelliae</i>
		<i>Watznaueria supracretacea</i>
		<i>Zeughrabdotus nanus</i>
		<i>Zeughrabdotus obliqueclausus</i>
		<i>Zeughrabdotus pomeroli</i>
		<i>Zeughrabdotus pseudanthophorus</i>
		<i>Zeughrabdotus</i> sp. A
		<i>Zeughrabdotus</i> sp. X
		<i>Zeughrabdotus small</i> sp.
		<i>Zeughrabdotus large</i> sp.
		Abundance
		Preservation

Figure 6. Continued from previous page.

MAMMALIAN TAPHONOMY OF THE EARLY IRVINGTONIAN (LATE PLIOCENE) INGLIS 1C FAUNA (CITRUS COUNTY, FLORIDA)

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ABSTRACT

Inglis 1C is an early Irvingtonian (late Pliocene) fossil vertebrate locality within a paleosinkhole. Taphonomic analyses of the mammalian remains reveal the depositional agents and biases in accumulation. The fossils are typically in excellent condition, although articulated elements do not occur. A few bones have a soft, chalky surface caused by post-depositional dissolution. The relative paucity of large carnivorans implies that the sinkhole did not serve as a den. Substantial water abrasion is rare, and evidence of predation is limited to a single tooth puncture on a deer phalanx and partial digestion of a few rodent teeth. The majority of clastic sediments were likely transported only short distances by water or possibly wind. Skeletons of the large mammals became disarticulated within the sinkhole, and small mammals accumulated as the result of predation by owls. After disarticulation, the very small elements were washed out of the deposit.

INTRODUCTION

Sinkholes are a natural and common geological feature in areas underlain by limestone. The abundant sinkholes of Florida form by solution of near surface limestone or by collapse of overburden into solution cavities. This karstification has been ongoing in Florida since a sea-level low stand in the late Oligocene (Randazzo, 1997), and has provided Florida with some of its most diverse and abundant vertebrate fossil collections. There are, however, unique prob-

lems associated with interpreting vertebrate fossil localities contained within paleokarst features. Chronological control of these deposits is limited to biostratigraphy due to an absence of physical stratigraphic controls and material suitable for radiometric dating. Application of stable isotopes to the chronology of karst features of Florida has not yet been attempted and efforts to incorporate paleomagnetic data were not successful (Ruez, 2001). A second problem in interpreting fossil faunas from sinkhole localities is the various potential methods of bone deposition. Vertebrate remains can accumulate from carcasses brought in by water, wind, or animals, from live animals physically trapped, or from the death of animals that lived in the sinkhole/cave.

Here I report the results of a taphonomic study of mammals found in the sedimentary filling of the late Pliocene Inglis 1C locality (Citrus Co., Florida). Based on these results, I propose the agents of accumulation and describe the biases introduced by the deposition method. Two approaches were used in taphonomic quantification of the mammalian remains: community-level and species-level analyses. The community-level approach evaluates the relative frequencies of taxa at Inglis 1C, while the species study shifts to abundances of elements within a taxon.

GEOLOGY

Inglis 1C (Figure 1; 29° 00' 43"N, 82° 40' 05"W) is a paleosinkhole within the Upper Eocene Ocala Limestone. Locally, the limestone is a buff-colored, foraminiferal and mol-

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MAMMALIAN TAPHONOMY OF INGLIS 1C

Table 1. Absolute and relative abundances of Inglis 1C mammals. Abbreviations: NISP, number of identified specimens; MNI, minimum number of individuals; Indet., indeterminate.

					without <i>Sylvilagus</i>	
	NISP	% total	MNI	% total	% total NISP	% total MNI
Xenarthra						
<i>Dasypus bellus</i>	293	4.93%	1	0.43%	20.07%	0.73
<i>Holmesina floridanus</i>	190	3.19%	1	0.43%	13.01%	0.73
<i>Paramylodon harlani</i>	154	2.59%	1	0.43%	10.55%	0.73
Insectivora						
<i>Cryptotis parva</i>	25	0.42%	8	3.45%	1.71%	5.74
<i>Blarina carolinensis</i>	1	0.02%	1	0.43%	0.07%	0.73
<i>Scalopus aquaticus</i>	5	0.08%	1	0.43%	0.34%	0.73
Chiroptera						
<i>Myotis</i> sp.	24	0.40%	6	2.59%	1.64%	4.38
Indet. Vespertilionidae	1	0.02%	1	0.43%	0.07%	0.73
Rodentia						
<i>Sciurus</i> sp.	65	1.09%	5	2.16%	4.45	3.65
<i>Glaucomys</i> sp.	3	0.05%	1	0.43%	0.21	0.73
<i>Orthogeomys pinetis</i>	91	1.53%	10	4.31%	6.23	7.30
<i>Sigmodon curtisi</i>	163	2.74%	25	10.78%	11.16	18.25
<i>Peromyscus sarmocophinus</i>	202	3.40%	35	15.09%	13.84	25.55
<i>Peromyscus polionotus</i>	8	0.13%	3	1.29%	0.55	2.19
<i>Reithrodontomys wetmorei</i> ¹	28	0.47%	9	3.88%	1.92	6.57
<i>Reithrodontomys humulis</i> ¹	1	0.02%	1	0.43%	0.07	0.73
<i>Baiomys</i> sp.	3	0.05%	1	0.43%	0.21	0.73
<i>Atopomys texensis</i>	6	0.10%	2	0.86%	0.41	1.46
<i>Ondatra idahoensis</i>	1	0.02%	1	0.43%	0.07	0.73
Lagomorpha						
<i>Sylvilagus webbi</i>	4489	75.46%	95	40.95%	-	-
Carnivora						
<i>Spilogale putorius</i>	104	1.75%	11	4.74%	7.12	8.03
<i>Lontra canadensis</i>	1	0.02%	1	0.43%	0.07	0.73
<i>Urocyon</i> sp.	1	0.02%	1	0.43%	0.07	0.73
<i>Smilodon gracilis</i>	3	0.05%	1	0.43%	0.21	0.73
<i>Homotherium</i> sp.	1	0.02%	1	0.43%	0.07	0.73
Artiodactyla						
<i>Platygonus bicalcaratus</i>	10	0.17%	1	0.43%	0.68	0.73
<i>Odocoileus virginianus</i>	43	0.72%	2	0.86%	2.95	1.46
<i>Palaeolama mirifica</i> ²	3	0.05%	1	0.43%	0.21	0.73
<i>Hemiauchenia macrocephala</i> ²	27	0.45%	2	0.86%	1.85	1.46
Perissodactyla						
<i>Equus</i> sp.	1	0.02%	1	0.43%	0.07	0.73
Indet. Proboscidea	1	0.02%	1	0.43%	0.07	0.73
Indet. Cetacea	1	0.02%	1	0.43%	0.07	0.73
Total	5949	100.00%	232	100.00%		

¹ Does not include eighteen upper incisors identified as *Reithrodontomys*, due to the presence of two included species; ² does not include 33 specimens identified only as Lamini (either *Palaeolama* or *Hemiauchenia*)

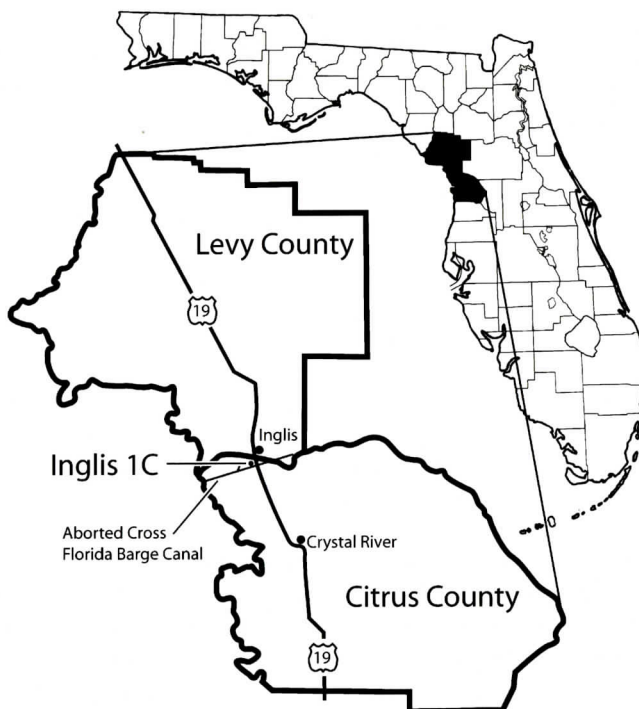


Figure 1. Location of Inglis 1C.

luscan grainstone. This site and others were intersected by the U. S. Army Corps of Engineers during the construction of the now-aborted Cross Florida Barge Canal. In 1967 Jean Klein and Robert Martin discovered vertebrate fossils in one of the paleosinkholes on the north bank of the canal. Fossils from that locality, Inglis 1A, were exhumed until 1974, when the site was excavated to five meters below current sea level. The Florida Museum of Natural History currently holds fossil vertebrates from seven of the intersected paleosinkholes, including the Inglis 1C locality examined in this paper.

The Inglis 1C sinkhole is currently approximately 5 m in diameter and 3.5 m deep. It is unknown what volume was removed by construction of the canal and by erosion from above. Inglis 1C is a cover-collapse sinkhole, recognizable from the relatively deep shape and small size. The area around Inglis 1C today consists of bare to thinly covered limestone in which solution sinkholes dominate (Sinclair and others., 1985). Cover-collapse sinkholes are usually confined to Areas III and IV of Sin-

clair and Stewart (1985). Their Area III consists of limestone covered by 30-200 ft of cohesive clay and is marked by numerous sinkholes of various sizes, while Area IV consists of limestone covered by more than 200 ft of cohesive sediment interlayered with discontinuous carbonate beds and marked by rare, very large sinkholes. While it is unknown how thick and of what type the overlying sediments were in this area during the Pliocene, the dimensions of the sinkhole conforms more closely with that described as dominant in Area III.

The Irvingtonian (late Pliocene) vertebrate fauna recovered from Inglis 1C includes 33 species of mammals (Ruez, 2001) and 32 species of birds (Emslie, 1998). Fish and herpetological remains were also recovered, but this component has not been studied. Pale-yellow quartz sands dominate the clastic fraction of the Inglis 1C infillings, with small amounts of rose quartz (~5 per 1,000 grains) and smoky quartz (~1 per 1000) also occurring. The quartz grains are roughly equidimensional and sub-rounded to sub-angular. Small Eocene foraminifer fossils

more accurately compared to one another. The revised values show a reduction in relative abundance from NISP to MNI of xenarthrans, as well as two artiodactyls: *Odocoileus virginianus* (2.95% NISP; 1.46% MNI) and *Hemiauchenia macrocephala* (1.85% NISP; 1.46% MNI). No other taxon shows a reduction in relative abundance from NISP to MNI. While no articulated material was recovered, the high NISP compared to MNI suggests that carcasses of several large-bodied taxa were disarticulated after deposition in the pit (Binford and Bertram, 1977).

Species Analysis

The high NISP of several taxa (*Sylvilagus webbi*, *Sigmodon curtisi*, and *Peromyscus sar-mocophinus*) allows for a different scale of taphonomic study, at the species level. The sample of *Sylvilagus* includes all skeletal elements, while the latter two are restricted to dental material. Some murid rodent postcrania are pre-

served at Inglis 1C, but the numerous forms preclude accurate assignment to species.

Sylvilagus webbi

While postcrania of lagomorphs are not diagnostic to the species-level, they are all allocated to *Sylvilagus webbi* because the dental characters indicate the presence of only a single species of lagomorph at Inglis 1C. The MNI of 95 is based on right maxillae, but this number is nearly reached by the other tooth-bearing elements. No postcranial element indicates abundance greater than 68 individuals. These values are listed in Table 2, along with a value used here to evaluate the quality of preservation at the site coupled with the diagnostic attributes of a particular element: the fragmentation ratio. The fragmentation ratio is calculated by dividing the NISP by the MNE (Richardson, 1980). A higher ratio indicates the presence of more fragmentary material, while the minimum value of 1.00 will occur when all specimens of that element are preserved intact or all specimens con-

Table 2. Selected *Sylvilagus* element abundance. The fragmentation ratio is calculated by dividing the NISP by the minimum number of elements (MNE). The MNI here is the number of individuals necessary to account for a particular element. I could not distinguish whether phalanges belonged to the forelimb or hindlimb, so they are grouped together. Likewise, I could not precisely identify distal ends of metapodials, so they are grouped.

Element	NISP	MNE	MNI	fragmentation ratio
maxillae	226	181	95	1.25
mandibles	219	177	93	1.24
scapulae	83	83	45	1.00
humeri	158	108	61	1.46
ulnae	113	84	44	1.35
radii	82	59	30	1.39
innominates	218	126	66	1.73
femora	223	123	68	1.81
tibiofibulae	206	111	60	1.86
astragali	58	58	33	1.00
calcanea	118	113	57	1.04
naviculars	35	35	25	1.00
cuboids	19	19	12	1.00
ectocuneiforms	25	25	15	1.00
metapodials	361	314	22	1.15
proximal phalanges	210	192	11	1.09
medial phalanges	117	114	7	1.03
distal phalanges	102	102	6	1.00

eroded from the surrounding Inglis Formation account for about two percent by volume of the matrix. Also weathered from the limestone walls are tests of the large foraminifer *Lepidocyclus ocalina*, gastropod external molds, and a few teeth of the extinct tiger shark *Galeocerdo alabamensis*.

Excavated sediments from Inglis 1C were sieved through 0.64 cm, 0.16 cm, and 0.8 mm screens. The fossils are typically in excellent condition, although articulated elements do not occur. A few bones have a soft, chalky surface, presumably caused by post-depositional dissolution. All fossils are curated in the vertebrate paleontology collection of the Florida Museum of Natural History, Gainesville.

METHODS AND RESULTS

Community Analysis

Table 1 shows the mammalian species abundance of Inglis 1C. Values are reported as both the number of identified specimens per taxon (NISP) and the minimum number of individuals necessary to account for those specimens (MNI), because these are the most commonly encountered quantitative terms in taphonomic analyses (White, 1953a; Grayson, 1984; Klein and Cruz-Urbe, 1984; Lyman, 1985). Use of the terms specimen and element in this paper follow the definition of Grayson (1984:16): a specimen is "a bone of tooth or fragment thereof, while an element is a single complete bone or tooth in the skeleton of the animal." "Identified" in this paper means identified to both taxon and element, including the side of the body when appropriate.

While NISP is the basic observational unit (although determination of what is identifiable is subjective), other quantitative terms are derivatives of NISP and need to be explained explicitly to avoid confusion with other uses. Most derived terms begin with estimating the minimum number of elements necessary to account for a sample of fragmented bones (MNE). Methods of calculating MNE have recently been reviewed by Marean and others (2001) who also advocated a spatial analysis GIS ap-

proach. They digitized each fragment and overlapped all the images to find the maximum number of overlaps. While conceptually this minimizes the grid size used in the manual overlap approach (Münzel, 1988; Marean and Spencer, 1991) and potentially provides a lower MNE, the results of Marean and others. (2001) show that MNE values are usually based on those areas of the element that have diagnostic features. For the Inglis 1C fauna, MNE values were determined by the manual overlap method as described by Morlan (1994) and Bunn and Kroll (1986). With the Inglis 1C fauna, MNEs of more than one were based only on complete teeth, except for *Spilogale putorius* (teeth, complete petrosals, maxillae, and mandibles), *Odocoileus virginianus* (right scapulae and left ulnae), *Hemiauchenia macrocephala* (right metatarsals), and *Sylvilagus webbi* (many elements). Therefore, with the exception of the *Sylvilagus webbi* remains, there was no ambiguity in determining MNE for the Inglis 1C fauna. Because of the abundance of *Sylvilagus webbi* fossils, diagnostic features were tabulated as demonstrated by Watson (1979). With maxillae and mandibles the features were specific alveoli and foramina; on most postcranial elements the features were articular surfaces, although some areas along the shaft of larger limb bones also proved useful.

The NISP of some species exceeds others due to the high number of identifiable elements, not necessarily as a result of more individuals in the deposit. A large number of specimens, especially osteoderms, record the presence of the xenarthrans, but the number of specimens for each species may be accounted for by the presence of a single individual. This bias is corrected by using the MNI as an indicator of taxonomic diversity. MNI is the greatest MNE value (separated into left and right when appropriate) for the particular taxon. Only the three xenarthrans and *Sylvilagus webbi* show a reduction in relative abundance with the change from relative NISP to MNI.

Because *Sylvilagus* material dominates the fossil assemblage (over 75% of fossil specimens), I recomputed relative abundances omitting the lagomorph. The other taxa can then be

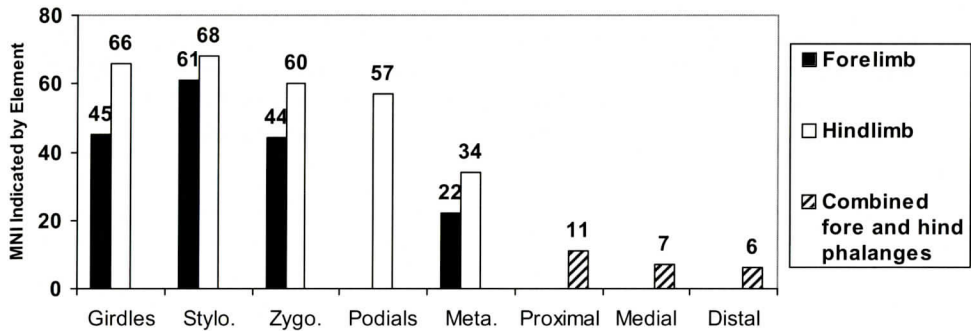


Figure 3. Element abundance along the appendicular skeleton. Phalanges are grouped together, as it is impossible to differentiate forelimb from hindlimb. MNI decreases away from the axial line of the skeleton. Abbreviations: Stylo., stylopodials (humeri and femora); Zygo., zygopodials (radii, ulnae, and tibiofibulae); Meta., metapodials (metacarpals and metatarsals); Proximal, proximal phalanges; Medial, medial phalanges; Distal, distal phalanges.

(1977), but is advantageous in not referring to "half animal units" and in recognizing possible fluctuations between sides of elements.

Modern avian and mammalian predators create bone assemblages with reproducible patterns in graphs of %MNI (Dodson and Wexlar, 1979; Andrews and Nesbit Evans, 1983; Denys, 1985, 1986; Andrews, 1990). While many taphonomic studies reference these works and attempt to find the best match between fossil and modern profiles, there is much variation within assemblages formed by a single predatory species, incomplete sampling of modern taxa, very little sampling of extinct taxa, no guarantee that fossil assemblages are derived from only a single source, and sometimes no assurances that predatory animals had any effect on bone accumulation. So, although the plot of %MNI in Figure 2 matches that of "the typical owl pattern" (Andrews, 1990: 48) any conclusions must be considered very tentative. The birds that produce this pattern include barn owls, long-eared owls, short-eared owls, and snowy owls (Andrews, 1990).

The trend of MNI for *Sylvilagus webbi* limb bones (Fig. 3) shows the taphonomic bias not only against the smaller elements, but more importantly, against more distal ones. The elements in the appendicular skeleton show a cline toward decreasing abundance away from the torso. This indicates that both the size and posi-

tion of the element determine preservation. The MNI of podials exceeds that of metapodials in spite of the much smaller size of the former. Carpals are not recognized in the collection, but are extremely small and difficult to identify.

Peromyscus sarmocophinus

During excavation of Inglis 1C, the sediment size distribution was not measured, nor is there any unsorted material left from the site. As proxy for such an analysis, it is possible to use the distribution of fossils of a known size. After *Sylvilagus*, the taxon at Inglis 1C with the highest MNI is *Peromyscus sarmocophinus* at 35. Most of the elements allocated to *P. sarmocophinus* are isolated molars, or maxillae and dentaries with associated teeth. Brachyodont teeth, such as those of *Peromyscus*, exhibit little ontogenetic variation in size and variation between individuals is much less than between different molars within the same individual. All this makes *P. sarmocophinus* the best choice to examine size bias at Inglis 1C.

For each of the six molars, Table 3 shows the total number, the number isolated, the percent isolated, and the average length. There is a strong correlation (significant at $p=0.01$) between the average length and the percent isolated (Pearson's product moment correlation coefficient of 0.9493). While the decrease is continuous across all teeth from largest to

tain the same portion of a skeletal element. An artifactual ratio of 1.00 can also occur when the element is not identifiable when broken. For example, identification of *S. webbi* scapulae was based on the morphology of the glenoid fossa. Although portions of the slender blade can be recognized as fragments of scapulae, they can not be confidently assigned to a species. The low values for podials, however, are real. Their high density and small size helps prevent breakage.

Many researchers exclude complete skeletal elements in their counts of NISP and MNE when calculating the fragmentation ratio (Lyman, 1994). This provides insight on how elements are broken when they are broken, but does not provide a representative view of the assemblage. If a collection contains 9 complete left humeri and one left humeri in 11 pieces, the fragmentation ratio would be 11 using Lyman's method. If another collection contains 10 left humeri, each in 3 pieces, the fragmentation ratio would be 3. Although the former example has a much great fragmentation ratio, I would argue that the latter collection is more fragmented because both the number of elements broken is greater and the total number of specimens is

greater even though the MNE of both is the same. Because of this, I have calculated the fragmentation ratio in Table 2 including complete specimens. If the first example above is calculated with the complete specimens included, the ratio is 2; the ratio in the second example would not change. This more accurately reflects the fact that the collection in the second example is more fragmented.

Normed MNI per skeletal portion (%MNI) is the minimum number of elements present in the deposit divided by the number of elements expected based on the MNI (White, 1953b, 1954, 1955, 1956). A graph of %MNI based on the MNI of 95 individuals of *S. webbi* (Figure 2) shows a significant bias against the presence of both the smaller and more fragile elements. The expected number of elements depends on the frequency within the skeleton. For example, a single individual will produce five lumbar vertebrae, two humeri, and only a single atlas. A %MNI of 100% would be assigned if the entire number of elements possible on an animal were preserved in all individuals represented in the deposit. This method is similar to the "normalized minimum animal units" used by Binford (1978, 1981, 1984) and Binford and Bertram

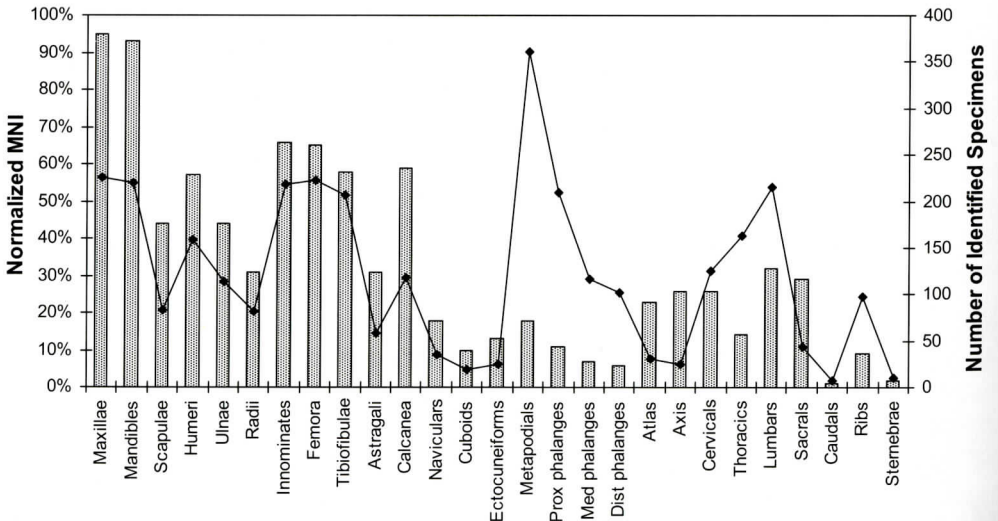


Figure 2. Normed MNI per skeletal portion (%MNI) of *Sylvilagus webbi* remains from Inglis 1C. The %MNI is calculated by dividing the minimum number of elements by the number expected based on a MNI of 95. The diamonds correspond to the NISP.

the average length. The extreme decrease in tooth abundance as isolated as the average length decreases to 1 mm is natural and not an artifact of excavation. Like the distal *Sylvilagus* elements, the smaller teeth were mostly removed from the deposit. The number of smaller rodents, insectivores, and bats may be significantly underestimated because of this bias against very small fossils at Inglis 1C. The relative abundances of lagomorph elements give a better impression of the physical depositional environment. Disarticulation occurred within the sinkhole, based on the high relative NISP to MNI ratio. The decrease in element abundance distally along the limbs is probably not due to predation; the lack of meat at the ends of limbs do not make them attractive as food, and the absence of evidence of dissolution on the distal elements present suggests the decrease is not due to complete digestion. Studies on the disarticulation of modern cow skeletons indicate that after separation of the mandible, disarticulation proceeds from the extremities (Weigelt, 1989). The relative absence of smaller elements, as well as clays and finer sands, implies subsequent removal by either wind or water. If the interpretation of the type of sinkhole is correct, there should be abundant clays in the area that would be expected in the deposit.

The small taxa may have fallen/climbed into sinkhole, but predation was suggested by Andrews (1990) as being the dominant mechanism by which small mammals are incorporated into any fossil deposit. Based on the little modification and little digestion of the Inglis 1C rodent material, the most likely predators would be those in Category I (sensu Andrews, 1990), which includes barn owls, snowy owls, and great grey owls.

The larger mammals either entered the sinkhole while alive, or were transported there; regardless, disarticulation of the carcasses occurred within the deposit. Small mammals may have entered the deposit by the action of owls. Fragmentation occurred within the deposit, but the broken elements were not removed. After disarticulation, fluid transport (presumably water) removed most of the elements about 1 mm or less in size. Because of the bias against

small fossils, the presence of chiropterans may be underrepresented; therefore the possibility of Inglis 1C serving as a roost for bats cannot be determined.

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Table 3. Abundance of *Peromyscus sarmocophinus* teeth. Total includes both isolated molars and those retained by the maxillae and dentaries. Broken molars were not included. Average lengths are taken from Ruez (2001). The Pearson's product moment correlation coefficient is 0.9493, a statistically significant correlation at the $p=0.01$ level. Abbreviations: M1, M2, M3, first, second, and third upper molars; m1, m2, m3, first, second, and third lower molars.

	total	isolated	% isolated	average length
M1	54	44	80.0	1.87
M2	24	15	62.5	1.43
M3	4	0	0.0	0.92
m1	25	20	80.0	1.74
m2	19	12	63.2	1.46
m3	18	1	5.6	1.13

smallest, the most dramatic decrease in percent isolated is from the second molars to the third molars. The decrease in average length from the first to second molars is just as large, but shows a much slighter decrease in percent isolated. Presumably there is a size threshold below which very little fossils are preserved at Inglis 1C; based on the size of the *Peromyscus* teeth used here, the threshold is slightly above 1 mm.

DISCUSSION AND SUMMARY

Of the 14 species of large mammals (carnivorans, ungulates, and xenarthrans) none is represented by more than two individuals. In spite of those limited numbers, xenarthrans, *Odocoileus*, and *Hemiauchenia* are "over"-represented in terms of the number of specimens (see discussion above). Such a high relative NISP compared to MNI suggests that these large mammals were disarticulated after deposition in the sinkhole (Binford and Bertram, 1977). Wind transport of such large carcasses is extremely unlikely. While water can transport large animals, the very low gradient rivers found in Florida, both today and in the Pliocene, would probably not be sufficient to move these animals. Hurricanes and other storm events could provide a mechanism by which either wind or water transport of carcasses is possible. However, such severe climatic events should leave recognizable debris (McFarlane and Lundberg, 1999).

Mammalian carnivores can drag carcasses into a den (eg. Meade, 1961), but evidence of predation at Inglis 1C is limited to a carnivoran

tooth puncture on a single *Odocoileus* phalanx. Even if evidence of predation was more prevalent, such activity could have taken place after deposition of the carcass. Carnivorans do occur in the deposit, but are rare (except for the small *Spilogale putorius*) and are probably not a result of denning which would typically leave many more carnivoran fossils. Deposition of the carcasses could also be the result of the animals falling or climbing into the sinkhole, where they then became trapped. In the absence of evidence for fluid or carnivoran transport, this method is accepted for the accumulation of the carcasses in the deposit. The single exception is the cetacean vertebra which is heavily worn and may have been washed into the site.

Among *Sylvilagus* postcranial elements, the ones with the four highest fragmentation ratios are the elements with the four highest MNI among postcrania (Table 2). This means that although the elements are broken, they are present in the deposit, hence the high MNI. There is no evidence of predation by mammalian carnivores or diurnal raptors. The single owl species present at Inglis 1C (*Otus asio*) is not large enough to take *Sylvilagus webbi* as prey (Emslie, 1998). Breakage therefore probably occurred within the sinkhole.

Examination of the sample of *Peromyscus* teeth suggests that the loss of *Sylvilagus* elements may be partly due to size as well as location. Although the sediment was screened at a mesh thickness finer than that of the minimum dimension of any *Peromyscus sarmocophinus* teeth, there is a significant correlation between the percent of teeth recovered as isolated and

TAPHONOMY OF LARGE ECHINOIDS: *MEOMA VENTRICOSA* (LAMARCK) FROM THE PLIOCENE OF SOUTH CAROLINA

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ABSTRACT

A test of the large spatangoid echinoid *Meoma ventricosa* (Lamarck, 1816), from the Middle Pliocene upper Goose Creek Limestone, shows an unusual and revealing pattern of preservation. A minimum estimate suggests it was resident on the seafloor for at least five months prior to entombment. Weight of overburden caused the specimen to start to collapse, but lithification "froze" it before it was completely disarticulated. Such preservation is certainly unusual and may be unique. It represents a frozen moment in the collapse of a large, thin-tested, high vaulted echinoid test, a morphology of irregular echinoids that is not well represented in the fossil record.

INTRODUCTION

The Cenozoic sedimentary successions of the southeastern United States and Caribbean preserve a diverse array of macrofossils, including common fossil echinoids. Many taphonomic patterns have been discerned by comparison between the Recent and fossil echinoids of the same region (Greenstein, 1993, 1995; and references therein), such as the extreme rarity of even partly disarticulated diadematooids (Greenstein, 1991; Donovan and Gordon, 1993; Donovan et al., 2001). These patterns can be related

to a few widely recognized parameters, such as environment of deposition, paleoecology and constructional morphology of the tests. One pattern that has not received definitive interpretation is the prevalence of smaller echinoids over larger individuals in the fossil record of the region. Although only a qualitative assessment, it is perhaps notable that, with the exception of the clypeasteroids, whose tests are particularly robust (Telford and Mooi, 1987), large (arbitrarily taken to be 100+ mm in maximum dimension herein) fossil echinoid tests are rare from these deposits. It may be that such large echinoids were not present as part of such faunas or were rarer than smaller tests, as might be expected, but fragments of tests of large echinoids form tantalizing components of some formations (e.g., Donovan, 1994, table 1). Further, the extant shallow-water echinoid fauna of the region includes some notably large taxa, particularly the spatangoids *Meoma ventricosa* (Lamarck, 1816) and *Plagiobrissus grandis* (Gmelin, 1788), which may both reach about 200 mm in length (Hendler et al., 1995).

Herein, we document the unusual preservation of a specimen of *Meoma ventricosa* (Lamarck) from the Pliocene of South Carolina (Figs. 1-3), which shows certain similarities to and differences from modern, dead tests of the same species found in shallow water environments (Donovan, 1991, fig. 11.3A, B). This specimen is deposited in the Department of Palaeontolo-

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Figure 2. *Meoma ventricosa* (Lamarck, 1816), RGM 212409, oral view. Cracks are both radial and following the lateral sutures of the plastron. The most obvious encruster is a balanid barnacle (upper left, arrowed). x 1. Scale bars in mm in all figures.

and crustaceans are well-preserved. Much of the section is filled with the large shells of *Carolinapecten eboreus walkerensis* (Tucker, 1934). Near the base of the section, just above the unconformable contact with the Maastrichtian Peedee Formation, about 1 m of upper Goose Creek Limestone is almost devoid of mollusc shells. It was in this less coarse-grained part of the succession that rare complete echi-

noids were found by the junior author.

Campbell and Campbell (1995, p. 100) listed the following echinoids from the Goose Creek Limestone: arbacioid *Arbacia improcera* (Conrad, 1843a); echinid *Psammechinus philanthropus* (Conrad, 1843a); clypeasteroids *Mellita caroliniana* (Ravenel, 1842), *Encope macrophora* (Ravenel, 1842) and *Clypeastrea* (sic) sp.; and spatangoids *Echinocardium orthono-*



Figure 1. *Meoma ventricosa* (Lamarck, 1816), RGM 212409, apical view. Note encrusting oysters, largest apically and generally smaller more adambitally, and the obvious radial cracks that more or less follow plate sutures. The arrow indicates an oyster that retains its free valve. x 1.

gy, Nationaal Natuurhistorisch Museum, Leiden (RGM). The terminology of the echinoid test follows Melville and Durham (1966) and that of encrusting organisms conforms to Taylor and Wilson (in press).

LOCALITY AND HORIZON

The echinoids discussed herein were collect-

ed from limestone quarries near Conway, Horry County, South Carolina, most notably those belonging to the Thompkins Mining Company, located about 4.8 km north of Conway on Route 905. About 18 m of leached Middle Pliocene upper Goose Creek Limestone are exposed in the Conway area (Campbell and Campbell, 1995). Aragonitic mollusc shells do not occur in the calcarenite, but calcitic molluscs, echinoids

tum (Conrad, 1843b) and *Spatangus glenni* Cooke, 1959 (for descriptions and illustrations of these species, see Cooke, 1959; Kier, 1963, 1972). The echinoids collected as part of the present study by the junior author (RGM 212409-212415) include *A. improcera*, *P. philanthropus*, *E. orthonotum*, *S. glenni* and three specimens of *Meoma ventricosa*, one complete test (RGM 212409), discussed below, and two "half tests" from different individuals (RGM 212410, 212411) that, in life, were larger than the complete specimen. However, neither of the "half tests" shows the same degree of radial disarticulation along sutures nor is as heavily encrusted as RGM 212409. Breakage of both "half tests" was undoubtedly post-exhumation.

DESCRIPTION

RGM 212409 is about 138 mm long, 110 mm wide and, less accurately, 55 mm high (estimated, but excluding encrusting oysters). The test is essentially complete apart from a broken portion in the anterior right ambital region (ambulacrum II, interambulacra 1 and, mainly, 2); this fracture cuts across plates and was undoubtedly post-exhumation in origin (Figs. 1, 2, 3A). Appendages (spines, pedicellariae) and delicate plate membranes (periproct, peristome) are lost, apart from rare spines that are disarticulated, but cemented to the test surface. The test is obscured, patchily, by coarse-grained carbonate sand, which includes some shell fragments, particularly adorally (Fig. 2). Although not obviously flattened, the test shows a pattern of radial cracks that are most prominent ambitally (Fig. 3) and correspond to sutures between columns of plates. There are 12 such radial cracks; the damage in the right ambital region is sufficient that, if any of the three plate sutures in this region had been cracked ambitally, the evidence has been lost. Other prominent cracks are along the sutures that flank the plastron (Fig. 2) and, rarely, occur between plates within columns (Fig. 3D, bottom left). Only one prominent crack is unrelated to plate sutures of the test and cuts across plates (Fig. 2, lower right, about one third of the test length from the posterior).

The test is encrusted, but not bored, by en-

dozozoans. Cementing organisms include small oysters that are closely clustered apically, and on adjacent interambulacral areas and the anterior ambulacrum; less commonly they occur in the ambulacral petals. At least one oyster retains the free valve (Fig. 1). There are also much rarer and smaller attached valves on the oral surface. The largest oyster has a maximum dimension of 15+ mm. Serpulid worm tubes are rarer and more dispersed. They are most common in the ambital region. Tubes are moderately elongate (<10 mm), sinuous but uncoiled, and preserved with the uncemented surface lost owing to breakage or abrasion. A prominent balanid barnacle is attached just subambitally and adjacent to the broken region; it does not overgrow the edge of the break (Figs 2, 3A). Small bryozoan colonies are rare on the oral surface; there is at least one such colony on the apical surface.

DISCUSSION

The most significant taphonomic observation concerning RGM 212409 is the evidence provided by the radial cracks, which indicate this test was almost certainly buried intact, but started to collapse under the weight of overburden after the collagenous ligaments holding the plates together had rotted (Smith, 1984, p. 17, 19). Most fractures of the test follow plate sutures. That the test did not collapse further was probably related to it being packed with sediment, the onset of lithification/cementation or both. Thus, this specimen preserves a frozen moment in its (incomplete) destruction by sedimentary processes. If RGM 212409 had been exhumed prior to lithification, then it would have left, at best, no more than so many unremarkable echinoid fragments. This pattern of cracking could not be confused with any of the several known forms of damage by predators and parasites that occur to this species (Kier and Grant, 1965, p. 49-50; Chesher, 1969, p. 93-95).

The spatangoids and clypeasteroids are the most commonly encountered and diverse groups of extant, shallow water irregular echinoids found in the tropical western Atlantic (Hendler et al., 1995) and southeastern U.S.A.,

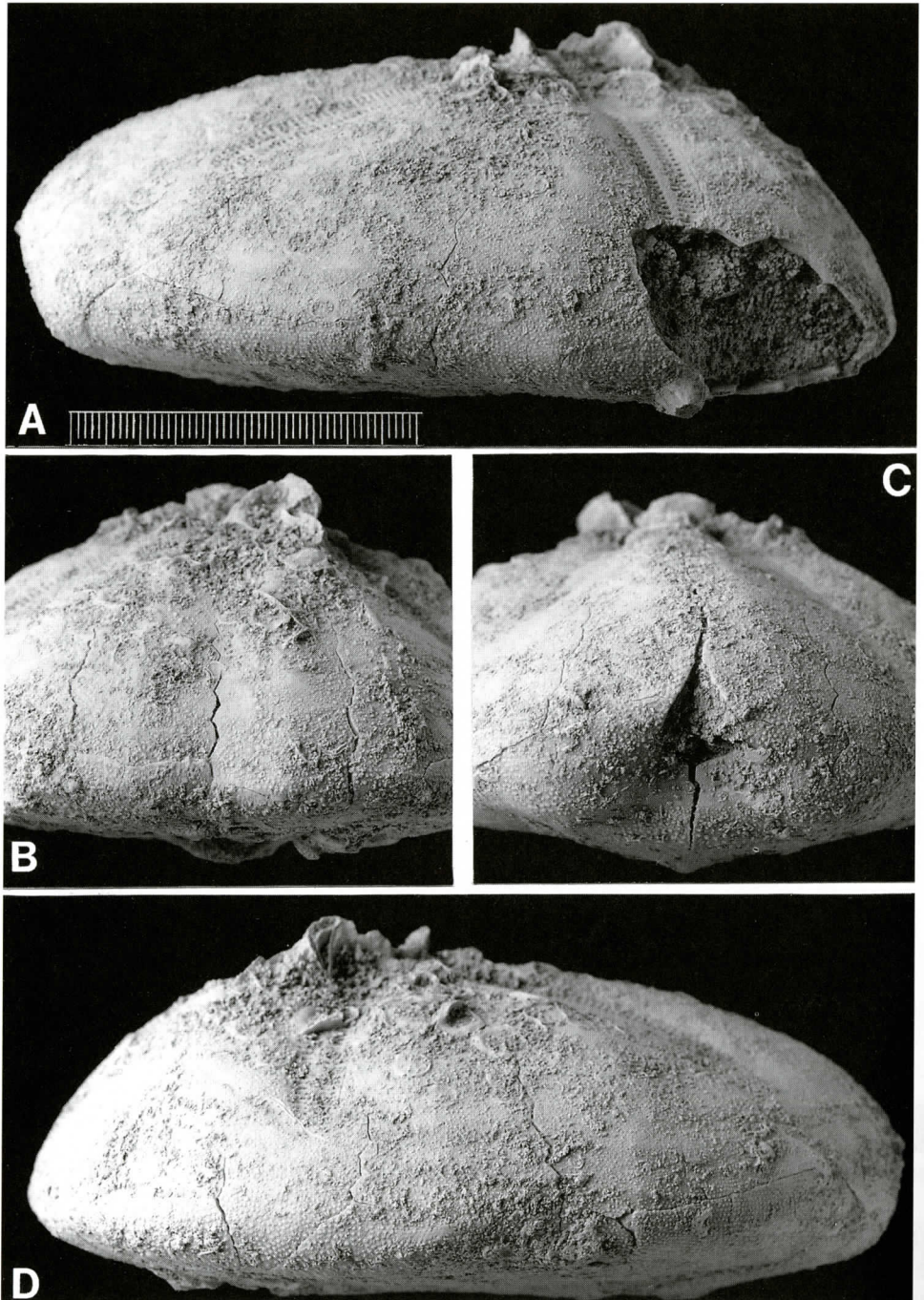


Figure 3. *Meoma ventricosa* (Lamarck, 1816), RGM 212409. (A) Right lateral view, showing anterior breakage of test, cluster of large oysters situated apically and the balanid barnacle just adoral of the ambitus. (B) Anterior view, showing oysters and radial cracks following radial plate sutures in anterior ambulacrum. (C) Posterior view, showing crack along suture through periproct. (D) Left lateral view. All x 1.

was "frozen" before it completely collapsed, producing a unique and informative specimen intermediate between complete and fragmentary preservation. Such preservation is unusual and may be unique. It represents a frozen moment in the collapse of a large, thin-tested, high vaulted test, the sort of irregular echinoid that does not have a good fossil record.

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Table 1. The fossil record of *Meoma ventricosa* (Lamarck).

Stratigraphy	Locality	Preservation	Reference
Upper Pleistocene	Barbados	internal mold	Donovan and Jones (1994)
Lower Pleistocene	Jamaica	test fragments	Donovan et al. (1994)
Pliocene	Aruba, Curaçao	incomplete tests	de Buissonjé (1974)
Middle Pliocene	South Carolina	one test, two "half tests"	herein
Lower Pliocene	Mexico	test	Buitrón-Sánchez et al. (1999)

yet the two groups have contrasting patterns of preservation (Smith, 1984). The specimen of *M. ventricosa* discussed herein is notable as an example of a taxon that, while locally common and widely distributed at the present day (Hendler et al., 1995, p. 244), has a patchy fossil record (Table 1). It is a fine specimen of a large, high-vaulted, non-clypeasteroid echinoid encrusted by endozoozoans and a particularly complete example of a large fossil spatangoid (compare with other occurrences in Table 1).

The cover of episkeletozoan encrusters (sensu Taylor and Wilson, in press) would be less unusual on a Recent dead test than on a fossil. All of the shells encrusting this test obviously did so after death of the echinoid, which first lost its spines. Extant *M. ventricosa* lives as a surface furrower and shallow burrower in and on sandy substrates (Smith, 1980, p. 40, 41; Hendler et al., 1995). It is not surprising, therefore, to find it encrusted as it did not necessarily have to be exhumed to reach the surface. Comparable fossil examples are rare, but Donovan et al. (1994) noted a partial specimen of *M. ventricosa* from the Pleistocene of southeast Jamaica that was sparsely encrusted by bryozoans and a serpulid around ambulacrum III. The apically positioned oysters are all about the same size and encrust both the echinoid and, laterally, each other, and thus represent a single spat fall.

Encrustation of the *Meoma* test by oysters and barnacles shows that the dead test was capable of surviving on the seafloor post-mortem for some time. Schäfer (1972, p. 104) noted that spines are lost in 7 days after death in at least one species of temperate, regular echinoid. Some tropical oysters may grow at about 0.1–0.5 mm day⁻¹ (Littlewood and Donovan, 1988, p. 1023); 0.1 mm day⁻¹ might be regarded as a reasonable rate in cooler waters (D. T. J. Little-

wood, pers. comm. to S. K. D.). As noted above, the largest oyster is 15+ mm in maximum dimension. Thus, a minimum residence time for this test on the seafloor might be regarded as approximately $(7 + [15/0.1])$ days, that is, about 160 days or 5 months. This would have been an ample period for decay of all collagenous ligaments within the stereom microstructure of the test, favoring breakage (before lithification) along plate boundaries (Smith, 1984), as is seen in this specimen, but not in similarly encrusted Recent tests resting on the sediment surface (Donovan, 1991, fig. 11.3B). The difference was produced by the burial and overburden loading of RGM 212409.

Among the extant shallow-water echinoids of the region, only the spatangoid *Plagiobrissus grandis* is consistently larger than *M. ventricosa*. It may be relevant that *P. grandis* has a poor fossil record and is only known from the Pliocene Tamiami Formation of Florida (= *Plagiobrissus* of Oyen and Portell, 2001, table 4).

CONCLUSIONS

Four points concerning this test are worth emphasizing:-

- This is an unusual specimen, representing a large, complete, non-clypeasteroid irregular echinoid test preserving a moderately diverse and locally dense endozoozoan fauna.
- This specimen could have been resident on the seafloor for at least five months prior to entombment, ample time for it to loose all of the collagenous ligamentation that bound the stereomic test plates together.
- The test was then buried and started to collapse under the weight of overburden, but the timing of lithification was such that it

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