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Abstract

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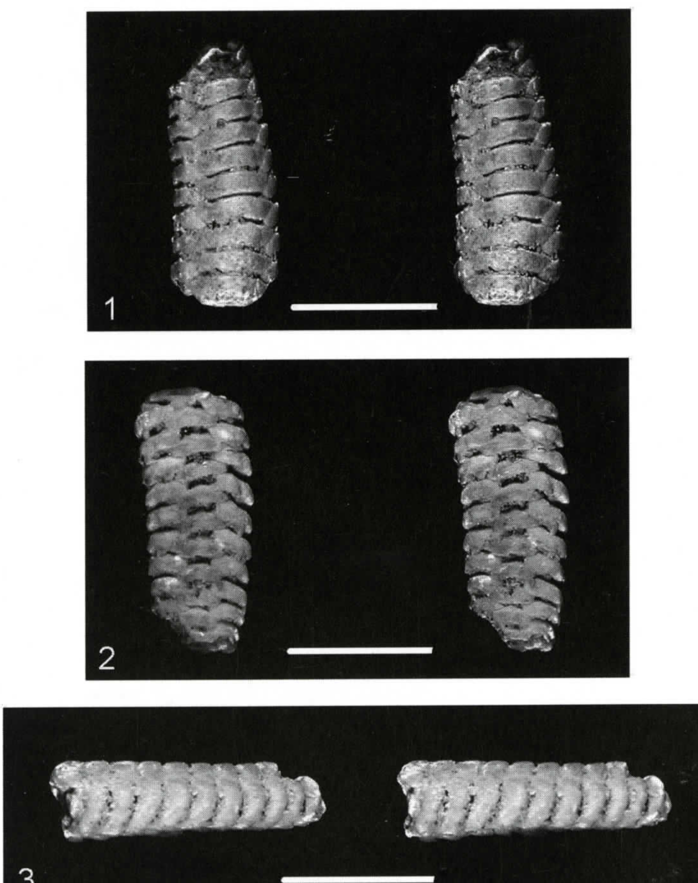

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PSEUDOBEDDING, PRIMARY STRUCTURES AND THRUST FAULTS IN THE GRANDFATHER MOUNTAIN FORMATION, NW NORTH CAROLINA, USA

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ABSTRACT

Primary structures, including beds and laminations, are overprinted by four sets of curviplanar structural-metamorphic S-surfaces in rocks of the Neoproterozoic Grandfather Mountain Formation of western North Carolina. The S-surfaces include a penetrative, but diffuse metamorphic foliation (S_1); diffuse, ductile deformation zones, in part characterized by thin structural laminations (S_2); discrete, ultramylonitic (ductile) thrust fault zones (S_3); and spaced joints (S_4). The structural laminations of diffuse, ductile deformation zones and some of the thin mylonitic fault zones mimic primary bedding, laminations, cross bedding, and cross laminations leading to errors in paleocurrent and structural analysis. Intense deformation along deformation zones, especially in the central part of the Grandfather Mountain Formation, has resulted in a stacked series of thrust faults and intervening, mesoscopic isoclinal folds that, combined with the metamorphic foliation, accommodate a significant amount of shortening.

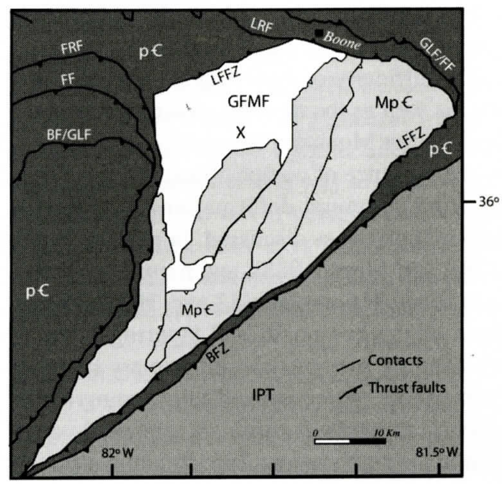


Figure 1. Map of the Grandfather Mountain Window and the surrounding geology (modified from Bryant and Reed, 1970a; Neton and Raymond, 1995; Adams and Su, 1996; Stewart, 2003). X-location of the study. GFMF - Grandfather Mountain Formation; IPT – Inner Piedmont Terranes; MpC - mesoproterozoic rocks; pC - undifferentiated precambrian rocks; Pz - Paleozoic rocks and underlaying Precambrian Rocks; BF - Burnsville Fault; FF - Fries Fault; FRF – Fork Ridge Fault (equivalent); GLF - Gossan-Lead Fault; LFFZ - Linville Falls Fault Zone; LRF - Long Ridge Fault

INTRODUCTION

The Grandfather Mountain Formation is a Neoproterozoic rift basin sequence of metasandstones, metasiltstones, metaconglomerates, and phyllites, with very thinly layered marble-like units and locally extensive metarhyolites and metabasalts (Bryant and Reed, 1970a; Schwab, 1977; Neton, 1992; Fetter and Goldberg, 1995). The formation appears uniquely within a structural window in the Blue Ridge Belt of the southern Appalachian Orogen (Fig. 1) (Bryant and Reed, 1970a; Boyer and Elliott, 1982). Primary and secondary features within the Grandfather Mountain Formation yield a valuable data set important for analysis of the sedimentary and structural history along part of the Neoproterozoic to Paleozoic continental margin of Laurentia.

The first modern analysis of the primary and secondary structures of the Grandfather Mountain Formation was completed by Bryant and

Reed (1970a). Yet, in spite of the fact that several additional studies were conducted since the publication of the detailed map of the Grandfather Mountain Window by Bryant and Reed (1970a), their observations (1970a, p. 73) — that (1) there is a general lack of distinctive marker horizons in the Grandfather Mountain Formation; (2) no major stratigraphic unit appears as a repeated marker unit in regional folds within the window; (3) there are significant variations in thickness and facies along strike within the formation; and (4) the structure is complex — remain arguably valid.

The difficulty of resolving the overall structure, as well as the detailed structure within the Grandfather Mountain Window, is exacerbated by the presence of compositional layering arising from structural deformation along ductile deformation zones discussed, in part, by Boyer (1984; and Mitra, 1988); and narrow mylonitic fault zones (Raymond and Love, 2005). In fact, some of the compositional layering is easily mistaken for metamorphosed primary layering. Furthermore, unrecognized thin ultramylonitic (ductile) thrust fault zones are apparently common in the southeastern, (presumably) structurally lower half of the formation. Both structural laminations and mylonitic thrust fault zones create pseudobedding and pseudo-cross bedding that confound attempts to resolve the folded structure; and the newly discovered thrust faults represent an unresolved amount of shortening of the lower part of the section. The new evidence of intrawindow deformation presented here, further demonstrates the poorly understood structural complexity of this region. The described characteristics have confounded attempts to detail the structural geology within the Grandfather Mountain Formation and the Window as a whole.

The purpose of this paper is to lay a foundation for re-evaluation and future studies of the sedimentological and structural history of the Grandfather Mountain Formation. We review existing primary and secondary (deformational, metamorphic) structures, including penetrative foliation, structural laminations, and discrete mylonitic fault zones (thrust faults) present in the central part of the Grandfather Mountain Formation as a step towards resolving details of the local and regional structure; and we expand on the brief descriptions of faults and diffuse deformation zones reported by Raymond and Love (2005). We also note the existence of secondary structural layers arising from the development of an S_2 fabric that many who have casually examined Grandfather Mountain rocks in southeastern exposures of the Formation, particularly along the Blue Ridge Parkway, considered to be primary beds. We concentrate our study between Rough Ridge and Dixon Creek on the southeastern flank of Grandfather Mountain, but note features at other localities to emphasize particular points.

In the Dixon Creek – Rough Ridge area, both structurally produced layering that mimics cross bedding and newly discovered thrust faults are locally well exposed. Because the structural layers superficially look like primary beds, they and the pseudo-cross beds may have been used in the past for structural and stratigraphic analyses, which, in light of this new evidence, become suspect. The thrust faults and some associated folds represent zones of structural shortening in the structurally lower part of the formation and reflect an heretofore unrecognized degree of intense ductile deformation. Recognition of widespread pseudobedding will require re-evaluation of the sedimentological and structural framework of intra-window rocks.

PREVIOUS WORK

The rocks of the Grandfather Mountain Formation were first described by Kerr (1875; in Bryant and Reed, 1970) and the description, structural character, and map distribution was greatly expanded by Bryant and Reed (Bryant, 1962; Bryant and Reed, 1970a,b). Aspects of the Grandfather Mountain Formation stratigraphy and sedimentology, petrology, and structure are described by Schwab (1977; 1986), Boyer (1978; 1984; 1992; Boyer and Mitra, 1988); Raymond et. al. (1992), Raymond and Pippin (1993), and Neton (1992; Neton and Raymond, 1995). Schwab (1977) used cross bedding to determine paleocurrent directions and a general paleogeography for the depositional setting of the Grandfather Mountain Formation. The details of the lithostratigraphy were expanded by Neton (1992; Neton and Raymond, 1995). Raymond and Pippin (1993) suggested that the provenance of Grandfather Mountain granitoid conglomerate clasts was somewhere other than the window basement or the Neoproterozoic Crossnore Volcanic-Plutonic Complex plutons of the surrounding (overthrust) rocks of Linville Falls Block.

The Grandfather Mountain Window is a duplex developed in Proterozoic rocks during Paleozoic orogenesis (Boyer and Elliot, 1982; Boyer and Mitra, 1988; Van Camp and Fullagar, 1982; Adams and Su, 1996). It is outlined by the surrounding Linville Falls Thrust Fault Zone (Boyer and Elliot, 1982; Adams and Su, 1996; Trupe, 1997), which represents the roof thrust of the duplex. Within the window, a "basement" of Meosproterozoic gneisses with Neoproterozoic granitoid intrusions provides the foundation upon which the Neoproterozoic Grandfather Mountain sediments were deposited (Bryant and Reed, 1970a; Fetter and Goldberg, 1995; Carrigan and others, 2003).

The Linville Falls Fault appears folded at the northeast and northwest corners of the window (e.g., Bryant and Reed, 1970a Neton, 1992; Raymond et al., 1992; Fig. 1), yet Bryant and Reed (1970a) were unable to recognize large regional, macroscopic scale folds *within* the Grandfather Mountain Formation. In contrast, they did map some smaller macroscopic folds that affected both underlying Mesoproterozoic gneisses and the overlying Grandfather Mountain rocks and they recognized "ubiquitous" mesoscopic (outcrop scale) folds within the formation. Mesoscopic folds were also recognized by Boyer (1984), especially near the Mesoproterozoic-Neoproterozoic contact, but like Bryant and Reed (1970a), Boyer did not show stratigraphic units repeated by macroscopic folding or describe regional macroscopic folds within the Window.

A comparison of published maps and cross sections clearly shows that there are different interpretations of the overall structure of the rocks inside the window (and below the Linville Falls Fault Zone). Although the stereographic structural data of Bryant and Reed (1970a) suggest the presence of regional folding in the northwestern part of the Window, their cross sections depicted a thick, generally northwest dipping section of rocks overprinted by a southeast dipping metamorphic cleavage. At the macroscopic (regional) scale, the rocks of the window were figured and described by Bryant and Reed (1970a) as a single overturned, northwest dipping limb of a complex synclinorium, whereas Boyer and Mitra (1988) show a regional anticline-syncline-anticline fold belt. The regional folding suggested by Boyer and Mitra (1988) is depicted in a general way in a cross section and shows tight to isoclinal folding on the southeast giving way to open to gentle folding in the northwestern part of the Window. Using fold axis symbols, Neton (1992), Raymond and others (1992), and Neton and Raymond (1995) depicted regional macroscopic folds on maps, but in the published work did not discuss details of the folding. Following Boyer's (1978) suggestion that Bryant and Reed's (1970) lower and middle siltstone units are correlative units on opposing fold limbs, Neton and Raymond (1995) argued that an anticline may be represented by similar conglomerates exposed to the northwest and southeast of the proposed anticlinal axis, but the conglomerates are not traceable around the fold. The different patterns of folding shown by these and other authors are based largely on regional relations. Analyses of folding using cross bedding to determine facing have proved difficult, because facing determinations, like the paleocurrent directions determined by Schwab (1977), seem to give confusing and conflicting results. No work describes significant large-scale, intraformational folds in detail.

No large scale faults are mapped that cut the informal members of the Grandfather Mountain Formation within the stratigraphic section above the southeastern structural base of the formation. Boyer (1978; 1984) and Boyer and Mitra (1988) proposed that the Mesoproterozoic-Neoproterozoic boundary, between the younger, 742 my old Grandfather Mountain Formation (Fetter and Goldberg, 1995) and the underlying 1.1 billion year old gneisses (Bryant and Reed, 1970; Carrigan and others, 2003) is largely faulted, in part along now deformed basin-forming normal faults and in part along Acadian-Alleghenian thrust faults, notably the intra-duplex thrust named the Goldmine Branch Fault. The latter defines a major part of the Mesoproterozoic-Neoproterozoic contact. Neither Boyer and Mitra (1988) nor Bryant and Reed (1970a) described faults *within* the Grandfather Mountain section. Raymond and Love (2005) provided the first report of significant thrust faulting in the structurally lower part of the Grandfather Mountain Formation. They report, however, that although numerous exposures of mylonitic thrust fault zones are present, continuous mapping of individual, closely spaced faults and calculation of the amounts of shortening represented by these faults has not been successful, because of the limited exposures.

Ductile deformation zones were recognized within the Grandfather Mountain Window by Boyer (1978; 1984) and Boyer and Mitra (1988). They report that these zones are particularly common in the southeastern part of the section and are absent to the northwest. The ductile deformation zones apparently cut a somewhat older metamorphic cleavage in the basements rocks (Boyer, 1984). In addition to these deformation zones, Raymond and Love (2005) report a pervasive foliation and mylonitic fault zones

PRIMARY STRUCTURES

The dominant primary structure in the Grandfather Mountain Formation is bedding. Beds vary from 1 cm to amalgamated metaconglomerate beds of more than 8 meters. Metasandstone beds typically occur in the range of 5 cm to 80 cm, whereas phyllite (metashale) beds are usually less than 50 cm thick (Fig. 2A, B). Metaconglomerate beds typically are 2m to 7m thick, but thinner and thicker beds occur locally (cf., Neton and Raymond, 1995). For structural purposes, beds are labeled S_0.

Laminations are the second most abundant primary structure. By definition, laminations range from <1mm to 10mm thick. Heavy mineral concentrations and, under current metamorphic conditions, chlorite concentrations, typically define the laminations that occur in metasandstones. Thin layers, commonly mistaken for laminations, are the structural (metamorphic), alternating quartz-white mica "compositional" layers described by Boyer (1984). These compositional layers represent diffuse deformation zones. In layered marble-phyllite units, questionable primary laminations are defined by alternating layers of phyllosilicate minerals and carbonate minerals. In some phyllites, alternating layers of phyllosilicate minerals and quartz-rich layers define a lamination. Some of these laminations may be formed by transposition.

Cross-bedding and cross-laminations are common in the less deformed parts of the Grandfather Mountain Formation. Cross-beds range from small troughs of less than 15 cm to large cross-bed sets that extend for several meters and consist of meter-scale, laminated subunits (Fig. 3). Cross laminations occur as climbing ripple, cross-laminations (Neton and Raymond, 1995) and as small scale laminations within cross-stratified beds.

Other primary structures include graded bedding and ripple marks. Graded bedding occurs primarily in metaconglomerate units, but occurs locally in some metasandstone beds. Both normal and reverse grading, generally coarse-tail grading, occurs in metaconglomerates. Because the rocks are extensively deformed and have a penetrative cleavage, ripple marks are rarely preserved in metasandstones (Fig. 4).

METAMORPHIC-STRUCTURAL S-SURFACES

Mesoscopic crudely planar, metamorphic structures in Grandfather Mountain rocks include an S_1 metamorphic foliation; S_2 structural laminations representing diffuse, ductile deformation zones [DDDZs]; S_3 discrete, mylonitic fault zones [DMFZs]; and vein-filled S_4 joints. Each S-surface type defines a fabric and the chronology indicated by the designations as S_1, S_2, and S_3 defines the general sequence of formation apparently developed during progres-

Figure 2. Bedding in the Grandfather Mountain Formation. A) Photograph of interbedded metasandstone (Mss), meta-conglomerate (Mcg), and phyllitic metasiltstone (Msh) of the Grandfather Mountain Formation, Highway 321, 1/2 mile south of Boone, North Carolina. A rock hammer and DNAG scale provide scale.

B) Photograph of metaconglomerate structurally overlying metasandstone along the Underwood Trail on Grandfather Mountain. The first author provides scale.

Figure 3. Photo of crossbedding and cross-lamination in the Grandfather Mountain Formation, Highway 184, 1/2 mile north of Linville Gap, North Carolina.

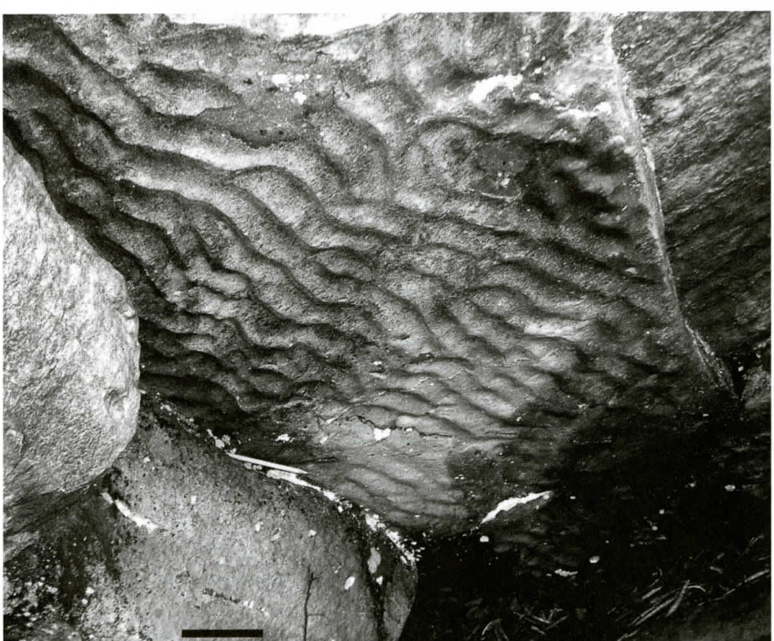


Figure 4. Overturned ripple marks in metasandstones of the Grandfather Mountain Formation, Boone Fork valley, Grandfather Mountain, NC. Ripples have wavelengths of about 4 cm. Note the pencil in the crack at lower center and the equivalent length bar scale = 16 cm.

Figure 5. Bedding (S_0) and metamorphic foliation (S_1) in metaconglomerate. Storyteller's Rock, Boone Fork valley, Grandfather Mountain, NC. Upper (light) pencil parallels S_0; lower (dark) pencil parallels S_1.

sive Alleghenian deformation, as we argue below. S_4 formed over a period of time that at least, in part, spans the same time interval as that during which S_2 and S_3 developed, as the observations presented below demonstrate. To date, S_2 and S_3 are only recognized in southeastern exposures of the formation, generally southeast of the crest of Grandfather Mountain. All S-surface types are not present in every outcrop in this southeastern part of the region. S_1 is not ubiquitous, being absent in some exposures containing S_0 and S_2. S_2 is widely distributed and is commonly the dominant fabric element in the southeastern exposures. S_3 is only present locally. S_4 is widely distributed throughout the formation.

Rocks of the Grandfather Mountain Formation are metamorphosed to greenschist facies grade (Bryant and Reed, 1970a), but no detailed study of metamorphism has been completed on these rocks. The metamorphic foliation (S1) is a spaced cleavage imparted to the rocks by the alignment of greenschist facies phyllosilicate minerals (Fig. 5). Thus, in the most common rocks (the metasandstones and metaconglomerates), thin, shape preferred orientation (SPO) zones dominated by phyllosilicate minerals define the foliation. In metasandstones, the SPO zones consist of one or more of the minerals white mica, biotite, and chlorite. Light green phengitic white mica is the dominant mineral in most rocks, but in metawackes and phyllites, chlorite is a major component of the SPO fabric. Biotite defines the foliation in some phyllites, metasandstones, and metaconglomerates, where the metamorphic grade may be slightly higher. In local exposures of the Linville Metadiabase southeast of the crest of Grandfather Mountain, white mica, chlorite, and green amphiboles are variably the dominant contributors to the SPO fabric.

Diffuse, ductile deformation zones [DDDZs] define a fabric here assigned the S_2 designation. DDDZs range in thickness from one quarter of a meter to more than 10 meters thick (Fig. 6). Adjoining exposures suggest that some DDDZs may be over 30 meters thick. Within these zones there are isolated areas in which relict primary clastic fabrics persist in weakly metamorphosed sandstone layers. Overall, however, where exposures allow a clear view of the fabric, such as at the exposures along Highway 221

Figure 6. Photo of diffuse, ductile deformation zone (DDDZ) at Pilot Knob on the eastern slope of Grandfather Mountain. Note the cross-cutting, structural laminations (Domain type 2) at the right center that yield pseudocross bedding; and the light colored layers of coarse metasandstone that have formed boudins and folded boudins in Domain types 2 and 3 (type 3 lacks structural laminations). Approximately two meters of section are shown. Access to this area is restricted by the Nature Conservancy.

Table 1. GPS coordinates of Important Field Localities.

Name	Coordinates (Lat)	Coordinates (Long.)
HWY 184 Outcrop	N 36º 07.50'	W 081º 50.67'
Shiprock (Rough Ridge)	N 36º 05.85'	W 081º 48.10'
Morphin-Endorphin	N 36º 05.84'	W 081º 46.62'
Pilot Knob	N 36º 05.90'	W 081º 47.15'
Story Tellers Rock	N 36º 07.18'	W 081º 47.69'

popularly known among rock climbers as "Morphin-Endorphin," and in exposures along the Blue Ridge Parkway, where it crosses Rough Ridge (Table 1), the fabric is pervasive and strongly overprints or completely replaces primary textures and structures. Locally, in some areas within S_2 domains and in locales where S_3 cuts S_2, there are apparent S-C fabrics of the C-type that contain discrete shear surfaces (C) cutting a schistosity (S) (Passchier and Trouw, 1996, p. 111).

In DDDZs there are several domains of various character. DDDZ domain type 1 consists of 15-25 cm thick quartz-dominated layers with thin (<1cm thick), somewhat continuous SPO, phyllosilicate-dominated interlayers (Fig. 7A). The latter are dark, thin, mylonitic, laminae-like layers. In many places, e.g., at Storyteller's Rock at the end of the Nuwati Trail on Boone Fork at the northeast end of Grandfather Mountain (Table 1), multiple sets of these dark thin laminae create pseudo-crossbedding (Fig. 8). Along the Blue Ridge Parkway, weathered exposures of type 1 domains give the appearance of sandstone beds with thin shale partings. Some type 1 sections consist of 1/4 meter thick packets of subtle to distinct compositional bands. The layers are here called structural layers or structural laminations, depending on their thickness. Type 2 DDDZ domains consist of "pinstripe" units. Some consist of closely spaced, alternating, mm scale, quartz-dominated and SPO, phyllosilicate-dominated layers (also called structural laminations) that give the appearance of metamorphosed, laminated, medium to fine-grained sandstones with thin shale interbeds (Fig. 7B). Others are phyllosilicate rich, but also exhibit thin, alternating quartz-

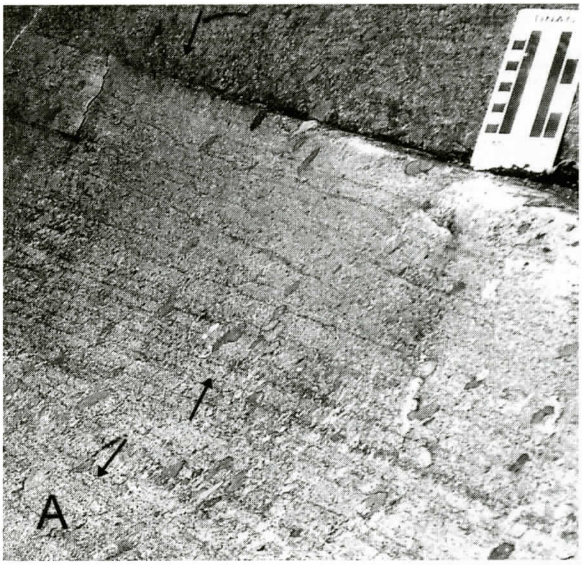
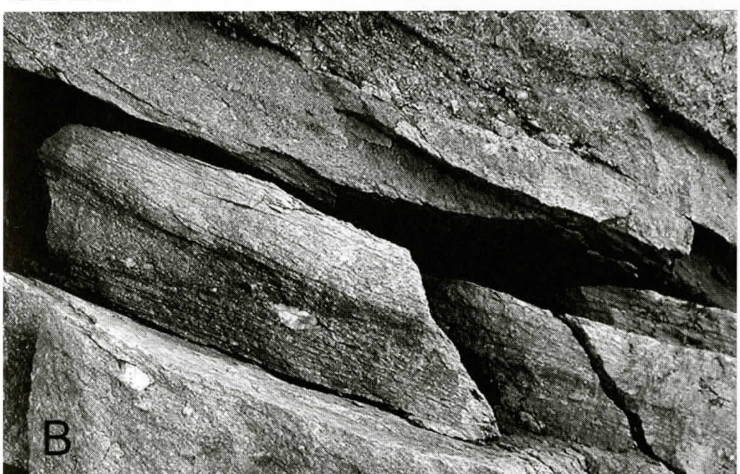
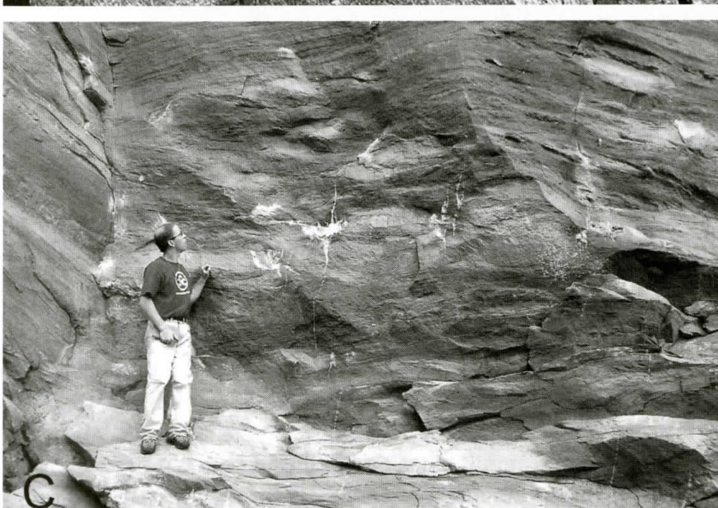


Figure 7. Diffuse, ductile deformation zones (DDDZs)(S_2). A) DDDZ with quartz-dominated layering (type 1 domain) at Rough Ridge along the Blue Ridge Parkway. Approximately 1/2 meter of section shown. Note pseudocross beds (arrows indicate apparent facing, i.e. bed tops). B) Pin-stripe structural laminations in DDDZ (type 2 domain), Swinging Bridge parking lot, Grandfather Mountain, NC. Approximately 1/3 m of section is shown. C) Boudins of epidote metaquartzite (light colored) in type 3 DDDZ domain. Boudins occur as light patches in a dark phyllosilicate matrix and form a ledge below the second author's feet. "Morphin-Endorphin" rock, Highway 221, Grandfather Mountain, NC. Second author (1.3 m) serves as scale.





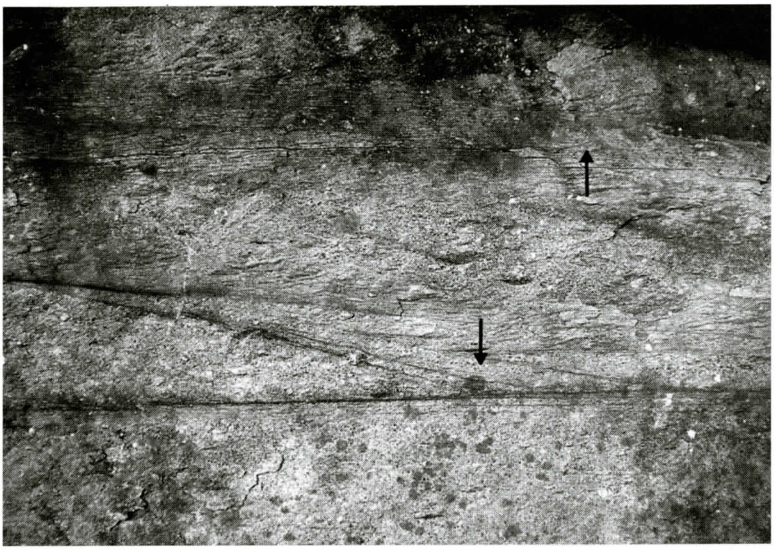


Figure 8. Pseudocrossbeds, Storyteller's Rock, Boone Fork valley, Grandfather Mountain, NC. Note conflicting facing indications indicated by arrows.

Figure 9. Photo of folded structural laminations (right center), cut by later-formed laminations near the round, gray lichen at left center and along the fold axis, Pilot Knob, Grandfather Mountain, NC. Such cut-offs locally give the appearance of cross-laminations. Small divisions on scale at right are in cm. Black line drawn to highlight fold.

richer and quartz-poorer laminations. These DDDZs include the type described by Boyer (1984), but we note that the scale and deformational histories of both type 1 and type 2 domains are larger and locally more complex, respectively, than are those described by Boyer (1984; and Boyer and Mitra, 1988). For example, in an ecologically fragile, restricted area of the northeastern part of Grandfather Mountain controlled by the Nature Conservancy, on a guided tour, we observed as many as three generations of cross-cutting structural laminations in a 30 cm wide band of a clean cliff exposure (Fig. 6). Similar relations exist at the easily ac-

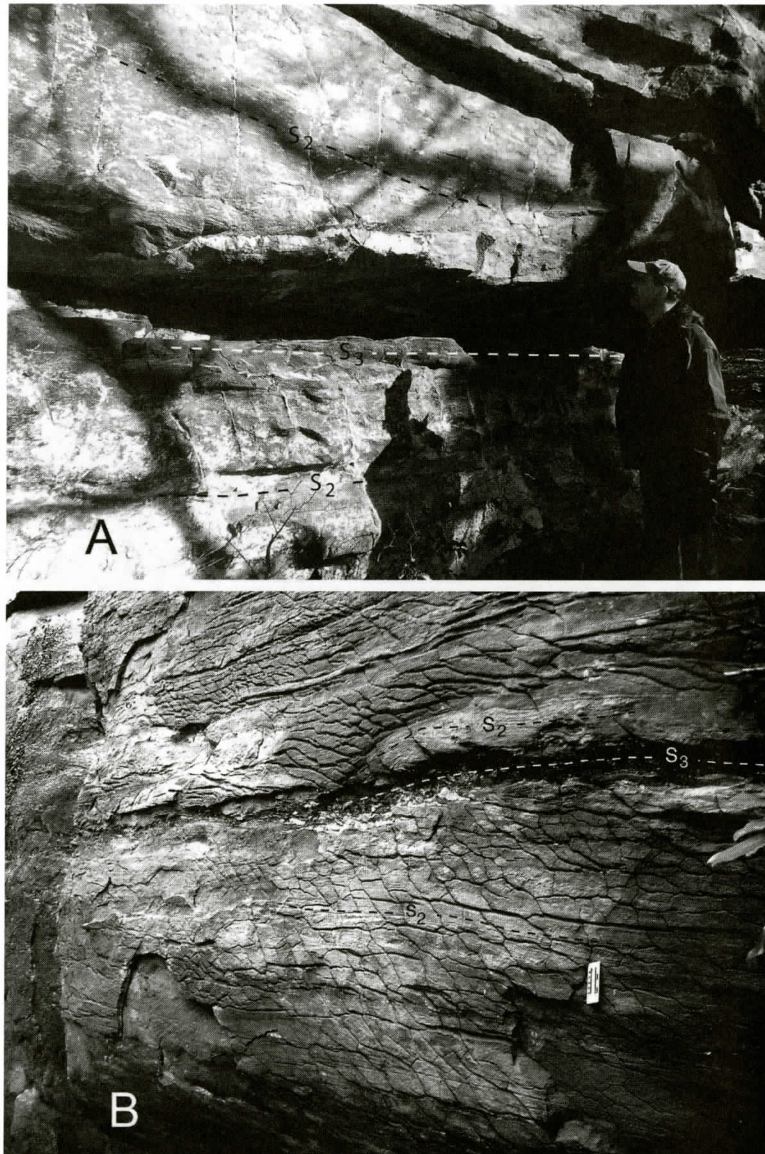


Figure 10. Photos of discrete mylonitic fault zones (DMFZs)(S_3). A) DMFZ at Ship Rock (Rough Ridge). Steve Hageman (1.8 meters) provides scale. Note nearly vertical S_4 quartz veins. B) DMFZ at "Morphin-Endorphin", cutting S_2 DDDZ. DNAG scale in lower right. Both localities are on the SE slope of Grandfather Mountain, NC.

cessible site where the Blue Ridge Parkway crosses Rough Ridge (Table 1; also Figure 12). Pseudo-cross beds are developed where folded or layered, earlier formed structural laminations are cut by a later set of laminations that truncate the earlier formed layers or limbs of folds (Fig. 6; Fig. 9). DDDZ domain type 3 consists of boudins of epidosite and epidote-quartz metasandstone in a matrix of deformed (mylonitic) finer-grained metasandstone, meltasiltstone, or metagabbro (Fig. 7C). Epidosite boudins are locally common and are the dominant types of boudins in the Linville Metadiabase (metagabbro) that intruded the Grandfather Mountain Formation. Southeast of the crest of Grandfather Mountain, the metadi-

Figure 11. Photo of Ship Rock (Rough Ridge) with multiple DMFZs (S_3), Blue Ridge Parkway, SE slope of Grandfather Mountain, NC.

abase displays some of the same fabric elements as the metaclastic rocks. Epidote-quartz boudins are also locally common in metasedimentary sections. Boudins are typically 4 to 25 cm thick and 7-30 cm long, but meter-scale, longer beds with pinched ends or boudins occur in some exposures. The longer beds show that in some cases S_0 and S_2 are parallel. Some thinner beds of only 2-5 centimeters have also developed the type 3 boudinaged fabric (Figs. 9, 13). Locally, the boudins are rotated or folded at the ends. Some thinner boudinaged beds are tightly to isoclinally folded within layers of domain type 2 or type 3 matrix (Fig. 9). Large cliff exposures of DDDZs may contain all domain types.

Discrete, mylonitic fault zones [DMFZs] form local S_3 fabrics (Fig. 10A, B). In general, these zones are less than one meter thick and may thin down to less than one millimeter or may disappear along strike. Some packets of DMFZs range up to five meters thick and consist of alternating more and less deformed zones. Stacked sets of DMFZs occur in a few major cliff exposures (Fig. 11). In some cases, stacked sets of DMFZs appear to form stacked, but *en echelon* sets of faults. Where we can determine the sense of shear, e.g. at Rough Ridge (the climber's Ship Rock) along the Blue Ridge Parkway (Table 1), the faults are top to the northwest thrust faults. At Rough Ridge, we have recognized eight DMFZs, two of which merge at the upper end of the exposure. The mylonite that distinguishes the DMFZs is commonly dark green to black, chloritic mylonite that is typically pervaded with milky quartz veins. Not uncommonly, the quartz veins are deformed and exhibit boudins or cm scale folds. Locally, cm scale apophyse-like dikes of the chlorite mylonite appear to be injected into surrounding layers or joints in less deformed metasedimentary rock. Because the chlorite mylonite locally contains pods of amphibole rich rock, we suggest that many DMFZs formed preferentially along the weak interfaces between Linville Metadiabase sills and the adjoin-

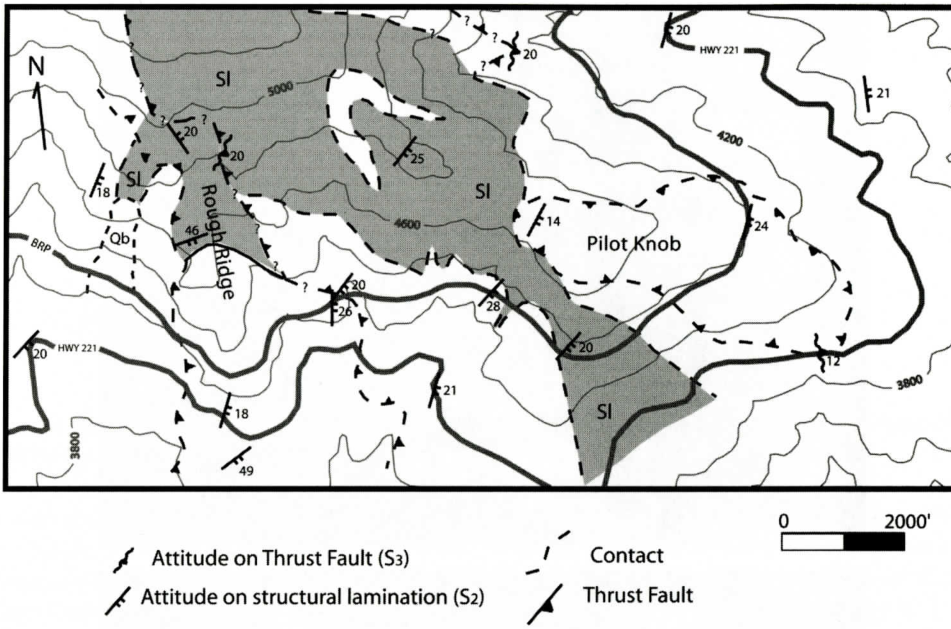


Figure 12. Geologic Map of part of the southeastern slope of Grandfather Mountain showing attitudes on DDDZs (S_2) and the locations of DMFZ thrust faults (modified from Bryant and Reed, 1970). Sl - Linville metadiabase; Qb - Quaternary blockfield.

ing quartz-rich country rock. At some locales, where the dark chloritic mylonite is missing, the faults are difficult to recognize or distinguish from similar appearing, thin bands of phyllosilicate in DDDZs.

In most cases, individual faults marked by DMFZs are not mapped across the southeastern flank of Grandfather Mountain for two reasons. First, multiple zones occur a few meters apart and spacing prohibits regional mapping of individual faults at any reasonable scale. Second, the DMFZs occur within cliff and rare outcrop exposures, but are obscured between exposures by soil covered areas tens to hundreds of meters across. Determining which of five or six faults in one exposure is represented by one fault in the next exposure along strike is simply not possible, especially considering (1) the similarities of the zones, (2) the observation that faults merge or die out, and (3) local arrangements in which one fault may be replaced by an *en echelon* companion up or down section. For this reason, we depict the localities of DMFZ exposure on Figure 12, and show projected traces of DMFZs, but the area as a whole appears to be pervasively deformed by DDDZs and DMFZs.

The fourth type of metamorphic/structural fabric element in the southeastern exposures of the Grandfather Mountain Formation is the joint. Many joints are filled by milky quartz veins (Fig. 13a, b). These veins form a widely spaced S_4 fabric. Examination of the relationships between quartz veins in S_2 and S_3 fabrics indicates that veins formed over an extended period of time during progressive deformation. The vein forming fluids likely derived their silica from the metaquartz arenites and quartz pebbles within metaconglomerates. In general, the veins are developed at a high angle to S_2 and S_3 fabric elements. That they formed over a period of time is evidenced by their variable relationships to other fabric elements. A few quartz veins cut across all preexisting S-surfaces without deflection (Fig. 13a), indicating their post-movement age of formation. In contrast, some veins show sigmoidal bends (Fig. 13b) where they extend from epidote-quartz boudins into the surrounding matrix, suggesting minor post-emplacement flow of the matrix. In other cases, the ends of veins end abruptly at the boudin

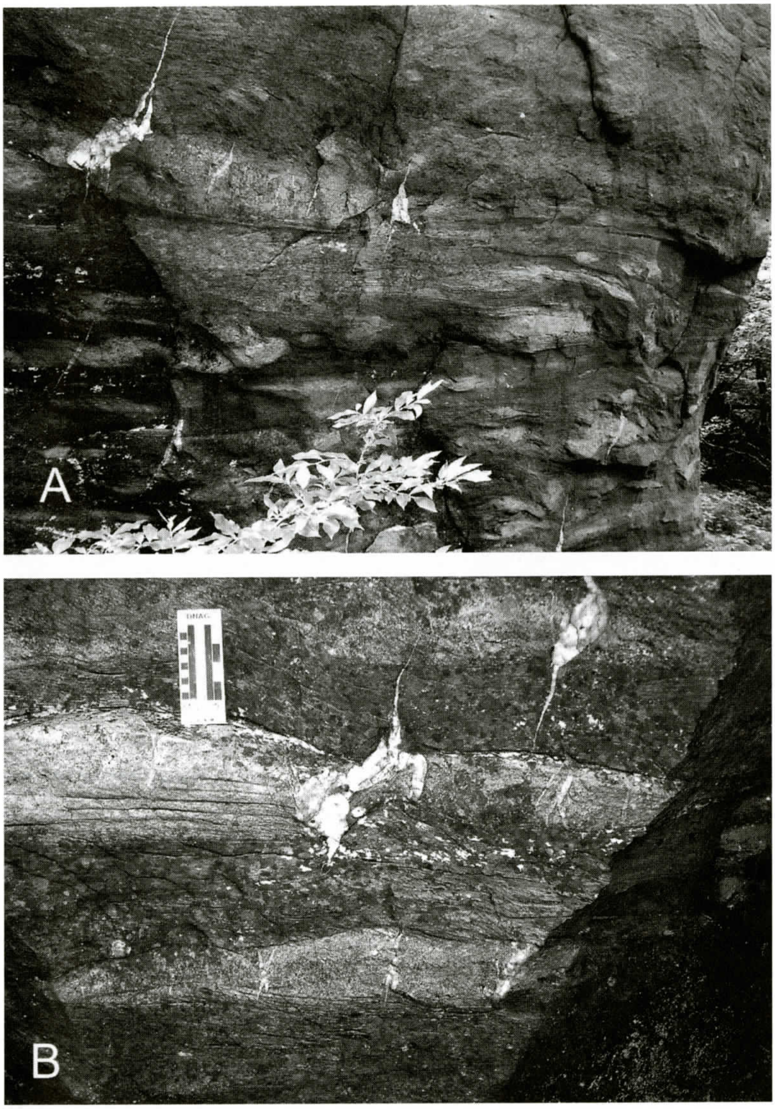


Figure 13. Quartz veins in jointed rock (S_4). a) Photo of south face of Pilot Knob exposure showing multiple quartz veins in Type 2 and 3 DDDZ. Approximately 3 m of exposure are shown. Note the straight vein at the far left and the slightly sigmoidal vein near the right end of the outcrop. b) Photo of boudin showing various relationships with quartz veins, Pilot Knob, Grandfather Mountain, NC. Note truncated and sigmoidal veins.

margin, indicating significant dislocation or flow along the boudin margin after the vein formed (Fig. 13b). The presence of folded and boudinaged quartz veins within DMFZ fault zones indicates significant post-formation movement within the zones. Thus, veins cut various fabric elements and formed during and after formation of S_2 and S_3.

We suggest that reconstruction of the relationships between the various post – S_0 fabric elements indicates that they are all Alleghenian in age and represent a single progressive deformation. Figure 14 depicts the various relationships. S_0 (primary) fabrics and structures are cut by the S_1 metamorphic foliation, which represents a widely distributed, very diffuse defor-

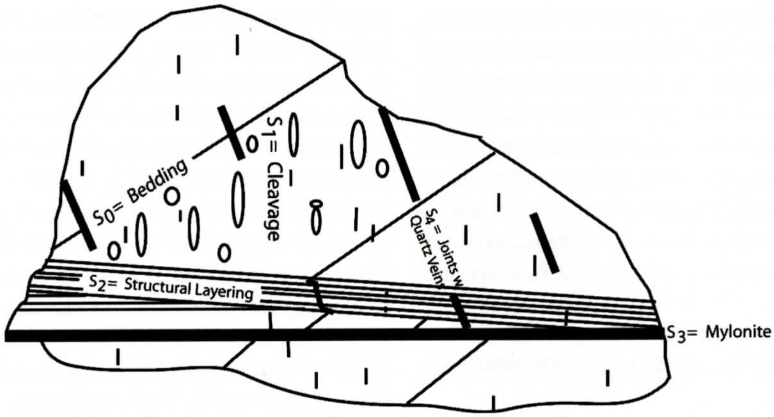


Figure 14. Sketch showing the relationships between S_0, S_1, S_2, S_3, and S_4 in a boulder of the Grandfather Mountain Formation.

mational and metamorphic feature formed early in the deformation of the Grandfather Mountain Formation. To our knowledge, no radiometric dating of this fabric element is available. Fabric element S_2 (the structural laminations, boudinage zones, and layered and laminated DDDZs) represent a style of deformation in which strain is partitioned into less diffuse, but still distributed zones of mylonitic deformation. Within the type 2 domains, we can recognize multiple periods of local deformation as indicated by cross-cutting packets of structural layers indicating shifting locations of flow (Fig. 7A). Locally, coarse-grained meta-quartz arenite beds were boudinaged and folded, cut by a first set of structural laminations, and successively formed into boudins by a second and perhaps third set of structural laminations, formed along different zones of flow. The latter laminations cut earlier formed laminations and folded laminations. S_3 formed as strain was more highly partitioned into the very discrete DMFZs. Continued shortening was facilitated by thrust faulting along these zones.

DISCUSSION

Pseudobeds and Cross Beds

Deformational features in the Grandfather Mountain Formation, as noted above, mimic primary sedimentary structures. Even knowing that the features exist, we still experience difficulty, locally, in determining whether a feature is an S_0, S_1, S_2, or S_3 feature. There are significant consequences of the discovery of diverse structural layers that mimic primary layering. First, we suspect that previous workers, as did we initially, mixed primary and secondary layers in their analyses. In particular, DDDZ domain type 1 and type 2 structures were likely measured as beds to determine the overall structure of areas within the Grandfather Mountain Formation. We do note that on their map, Bryant and Reed (1970a), perceptively, distinguished between bedding and "compositional layers." The latter layers are visible because of differences in both composition and grain size. Nevertheless, inasmuch as they are pseudobeds, any structural analysis based on measurements that assume that the layers are primary is flawed. Second, most of the apparent cross bedding and cross lamination features in the southeastern area of Grandfather Mountain Formation are likely pseudo-cross beds and pseudo-cross laminations representing multiple periods of (progressive) structural lamination development. As a consequence both paleocurrent analyses and structural analyses that are based on apparent cross beds are suspect.

In outcrop, the *appearance* of the pseudo-cross beds is decidedly primary. Thin section analyses do not always help distinguish deformed from undeformed rock. As noted by Boyer (1992), some of the rocks in the region have undergone significant recrystallization (recovery) and do not appear to be mylonitic. Were it not that opposing facing directions are evident in the same outcrop or same exposure face (Figs. 7A, 8), many geologists would likely not be convinced that the structures are, in fact, secondary. Thus, we offer a cautionary note to those using apparent cross beds for structural and paleocurrent analyses in metamorphic terranes, that such features can be, in fact, secondary.

Single or Multiple Orogenies?

We suggested above that the various deformational features represent progressive deformation during a single, protracted Alleghenian deformation. Boyer (1992) argued, instead, that the structural features of the Grandfather Mountain Window represent repeated deformation extending over the Ordovician (Taconic) to Pennsylvanian (Alleghenian) age range of Southern Appalachian orogenesis. Boyer based this conclusion on two data sets. First, as do we, Boyer recognized that there is a succession of structural features, but he attributed some of these to different orogenies. Second, he linked the structural features in the Window to the metamorphic events recognized in overlying thrust blocks (Dallmeyer, 1975; but also see McSween and others, 1989; Adams, and others, 1995; Abbott and Raymond, 1997).

A poly-orogenic hypothesis, however, is not consistent with the bulk of the data now available. First, the likely large magnitude of pre-Alleghenian separations of Blue Ridge thrust fault blocks (Hatcher, 1989), means that at the time of metamorphism of rocks such as the Ashe Metamorphic Suite, which occur in a thrust block structurally overlying the Grandfather Mountain Formation (above the Linville Falls Fault), Grandfather Mountain rocks and these overlying rocks were widely separated geographically. Second, the 302 m.y. age on Linville Falls Fault Zone rocks determined by Van Camp and Fullagar (1982; also see Adams and Su, 1996; Schedl and others, 1997) and the associated petrologic evidence, indicates that movement of rocks above the Linville Falls Fault, over the Grandfather Mountain Window, occurred during the Alleghenian Orogeny under greenschist facies conditions (Adams and Su, 1996). Significantly, "diffuse phyllonites" in basement rocks beneath the Grandfather Mountain Formation also formed at this time (Schedl and others, 1997). Third, petrographic evidence indicates greenschist facies metamorphism along faults separating the Grandfather Mountain Window rocks from those of the overlying fault blocks, and similar, but retrograde metamorphism, along faults separating some of those overlying fault blocks (e.g., Abbott and Raymond, 1984; Raymond and others, 1992). Fourth, if the Grandfather Mountain rocks were linked to rocks in the overlying thrust blocks during earlier Taconic (Ordovician) and Acadian (Devonian) metamorphic events, the metamorphic pattern (with greenschist facies rocks at depth and amphibolite to eclogite facies rocks at the highest structural levels) would be an "upside-down" metamorphic pattern that would require an explanation not provided by the Boyer hypothesis.

Boyer (1992) recognized recovery in some rocks and perhaps considered that retrogressive, greenschist facies recrystallization was pervasive in the Grandfather Mountain Formation rocks, explaining some of the features listed above. Yet, there is no compelling evidence for multiple metamorphic events, and by implication, multiple deformational events affecting the Grandfather Mountain rocks. We note that the Silurian Linville Metadiabase (Fetter and Goldberg, 1993) locally displays each of the S-surface fabric elements, so that, in fact, deformation must be post-Silurian in age. Furthermore, there is no specific evidence of any Ordovician metamorphism or deformation within the Grandfather Mountain Window. The Linville Falls Fault Zone was clearly active during the Alleghenian Orogeny and cuts some S1 fabric elements, but as noted by Bryant and Reed (1970a) and Boyer (1992), the fabrics in the Window appear to be synmetamorphic. Rocks in the Grandfather Mountain Formation are all of greenschist facies grade. We know of no evidence of retrograde metamorphism or multiple metamorphic events in these rocks. Boyer (1992) did not address the overall lack of higher grade enclaves within rocks of the Grandfather Mountain Formation, which, if present, would justify a history involving multiple orogenies. In their absence, there is only evidence of one metamorphic event, and that is associated with the Alleghenian Orogeny. We know of no cases where S_2 and S_3 fabrics cut across the Linville Falls Fault, suggesting that all S surfaces are pre- to syn-faulting features. As a consequence, we interpret the S_1 to S_4 sets of structures to represent one protracted, progressive Alleghenian deformation during which strain was progressively partitioned into successively tighter zones of deformation. Late stage movement along duplex faults would produce local cross-cutting structural relationships. Clearly, detailed metamorphic and additional radiometric dating analyses are needed to confirm our hypothesis.

CONCLUSIONS

In metamorphic terranes, partitioned strain may be revealed by a variety of structural features. Some of these features mimic primary structures — most notably bedding, laminations, cross bedding, and cross laminations. Failure to distinguish pseudo-cross beds from their primary analogs clearly will lead to erroneous structural and sedimentological analyses.

In southeastern exposures of the Grandfather Mountain Formation, pseudo-cross beds and false bedding are common features among the array of secondary structures. Four secondary metamorphic structures — a metamorphic foliation S_1, structural laminations and related features forming diffuse ductile deformation zones (S_2), discrete mylonitic fault zones (S_3), and spaced fracture-filling quartz veins (S_4) characterize exposures of greenschist facies metasedimentary and metaigneous rocks. The structural laminations and layers in diffuse ductile deformation zones mimic primary bedding and cross laminations, but are pseudo-beds and pseudo-cross laminations. Since they were likely considered to be primary features in some previous analyses, the conclusions of those studies are suspect. Distinguishing these secondary structures from primary structures will be a critical element of future structural analyses.

Previously unrecognized folds and thrust faults, the later represented by discrete mylonitic fault zones (S_3), represent a significant, but undocumented amount of shortening in the southeastern part of the Grandfather Mountain section. Recognition of the various unrecognized and previously misinterpreted structural features described here is a first step in developing a better understanding of the overall structure of the Grandfather Mountain Formation and the window in which it is exposed.

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SOUTHEASTERN GEOLOGY
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ADDITIONAL ECHINODERMS FROM THE PCS (LEE CREEK) PHOSPHATE MINE, NEAR AURORA, BEAUFORT COUNTY, NORTH CAROLINA

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ABSTRACT

Three recently discovered echinoids, *Arbacia* sp. cf. *A. sloani* (Clark in Clark and Twitchell), *Encope macrophora* (Ravenel), and *Agassizia scrobiculata* Valenciennes, and one partial ophiuroid arm, family Ophiuridae, from the PCS Phosphate (Lee Creek) Mine, near Aurora, Beaufort County, North Carolina, are herein reported. The occurrence of these echinoderms supplement our knowledge of the taxa found in the mine.

INTRODUCTION

Kier (1983) identified five species of echinoids from the Lee Creek (= PCS) Phosphate Mine: *Abertella aberti* (Conrad, 1842) from the Middle Miocene Pungo River Formation; *Echinocardium kelloggi* Kier, 1983, *Psammechinus philanthropus* (Conrad, 1843), and *Arbacia improcera* (Conrad, 1843) from the Lower Pliocene part of the Yorktown Formation; and *Mellita* sp. cf. *M. aclinensis* Kier, 1963 and *A. improcera* from the Pliocene-Pleistocene Croatan (equals James City Formation used herein). Echinoids (spines and pieces of tests) are quite common in the mine spoil piles but the vast majority of these are too fragmentary for identification. The most easily identifiable remains are of the sand dollar *Mellita* sp. cf. *M. aclinensis*, but even for this species near-complete specimens are scarce. Recently, collectors found and donated to the North Carolina Museum of Natural Sciences, three species of echinoids not discussed by Kier (1983): *Arbacia* sp. cf. *A. sloani, Encope macrophora* (Ravenel, 1842), and *Agassizia scrobiculata* Valenciennes, 1846. These specimens, along with a partial ophiuroid arm, representing the first Ophiuridae from this locality, are reported here. Together, with the echinoids documented by Kier (1983), these specimens give a more accurate account of the echinoderms found in the mine.

GEOLOGICAL SETTING

The PCS Phosphate (Lee Creek) Mine is located in Beaufort County, North Carolina, about eight km north of the community of Aurora. To gain access to the ore (which lies approximately 30-35 m below sea level) the top 10 m or so are removed using a bucket-wheel excavator and transported on a conveyor belt to a previously mined pit. The next 20-25 m are removed by large electric draglines and piled in the immediately previous cut (McLellan, 1983). This "topside-down" process results in the older sediments generally being placed on top of younger sediments in spoil piles; however mixing inevitably occurs. Once the equipment is removed to safe distances, fossil collectors are allowed to hunt the spoil piles under tightly regulated conditions. The 25 m of spoil contains the James City Formation (formerly known as the Croatan Formation), various levels of the Yorktown Formation, and the top level of the Pungo River Formation (Gibson, 1983) (Figure 1).

The Late Pliocene–Early Pleistocene James City sediments contain primarily invertebrate remains, including corals (e.g., *Septastrea*), mollusks (mostly extant subtropical species) (Ward and Blackwelder, 1987), arthropods, and bryozoans. Echinoid spines and test fragments are very common, including *Mellita* sp. cf. *M. aclinensis* and *Arbacia improcera* (Kier, 1983).

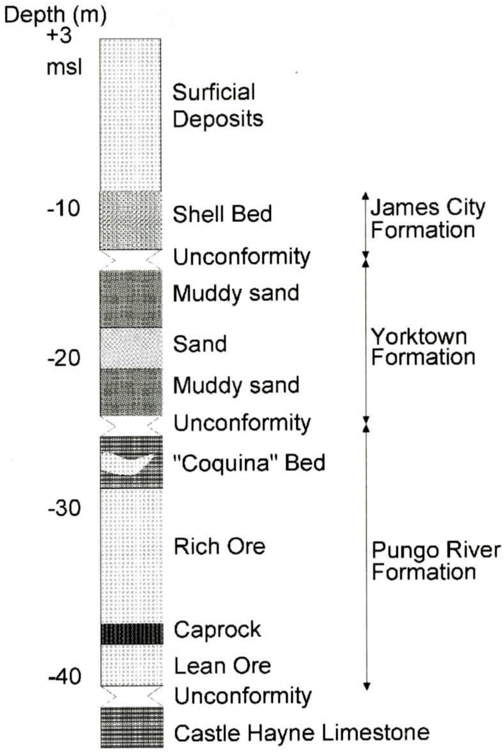

Figure 1: Simplified stratigraphic column of the PCS Phosphate (Lee Creek) Mine, modified from Riggs et al., 2000.

Vertebrate remains are rarely encountered but include shark teeth from species such as *Carcharodon carcharias* and *Galeocerdo cuvier* (Purdy et al., 2001).

The lower unit of the Yorktown Formation occurs at the Miocene–Pliocene boundary. In the mine, the Yorktown is particularly rich in vertebrate remains, including fish, marine mammals, and birds. Characteristic mollusks are *Chesapecten* and *Ecphora* (Gibson, 1987; Wilson, 1987). Echinoids include *Echinocardium kelloggi*, *Psammechinus philanthropus*, and *Arbacia improcera* (Kier, 1983).

The oldest unit commonly encountered in the mine is the Middle Miocene Pungo River Formation. This unit is also rich in vertebrate remains, especially shark teeth. Mollusks are mainly moldic. Kier (1983) attributed remains of *Abertella aberti* to the Pungo River Formation at Lee Creek (PCS) Phosphate Mine but provided no illustration and implied that only fragments were found. To our knowledge, no complete or near-complete echinoid remains have been found in Pungo River spoils.

SYSTEMATIC PALEONTOLOGY

All figured specimens are housed at the North Carolina Museum of Natural Sciences (NCSM), Raleigh, North Carolina.

Class **STELLEROIDEA Lamarck, 1816**

Subclass **OPHIUROIDEA Gray, 1840**

Order **OPHIURIDA Müller and TROSCHEL, 1840**

Family **?OPHIURIDAE Lyman, 1865**

Material — One partial arm (NCSM 9811: Figure 2.1 - 2.3).
Measurements — Length 18.2 mm, width 7.8 – 6.5 mm tapered end to end, height 5.4 – 4.7 mm tapered end to end.
Description — Arm fragment triangular in cross section. Dorsal shield plates single, moderately curved, parallelogram shaped, 5.7 mm long, 1.8 mm wide. Lateral shield plate single, strongly curved, 4.1 mm high, 1.8 mm wide with curved protuberance on ventral end in contact with ventral shield plate. Ventral shield plate flat, 3.5 mm long, 1.2 mm wide, tapered on both ends and adjoining lateral plate protuberance. Dorsal shield forms vertebrae with neural hole in cross section.
Occurrence — Spoil piles, Yorktown Formation (Lower Pliocene), PCS Phosphate (Lee Creek) Mine, near Aurora, Beaufort County, North Carolina.
Discussion — Records of Pliocene asteroids or ophiuroids from the southeastern United States have been very limited (Portell and Oyen, 2001). Jones and Portell (1988) reported *Heliaster microbrachius* Xantus, 1860 from the Tamiami Formation of southeastern Florida. To this, Oyen and Portell (2001), added *Luidia* sp. and an unknown ophiuroid from the same deposit. The ophiuroid, although complete, was

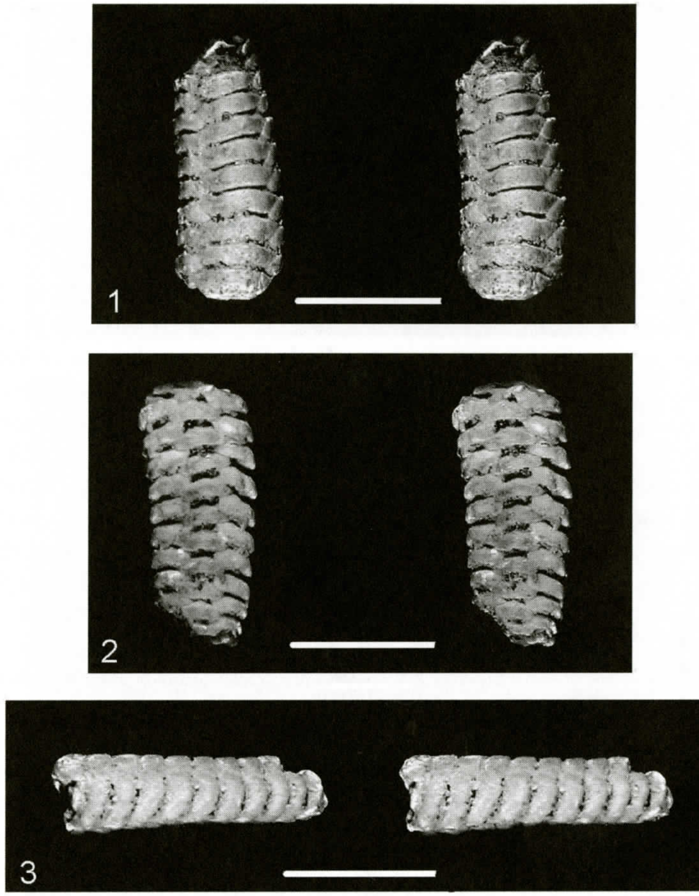

Figure 2: Stereo pairs of NCSM 9811,Ophiuridae (partial arm); 2.1 dorsal view, 2.2 ventral view, 2.3 lateral view. Scale bars = 10 mm.

highly recrystallized thus precluded identification. The discovery of a partial ophiuroid arm from the Yorktown Formation in the PCS Phosphate Mine is a welcome addition to the poorly known asteroid and ophiuroid fauna of the southeastern United States. Because this specimen is a partial arm, and comparative material is limited, it is unwise to attempt generic identification. However, as this partial arm is stout, tapering, and with short vestigial spines, characteristics of the family Ophiuridae as described by Spencer and Wright (1966), we assign this specimen to the family ? Ophiuridae. We report the occurrence with hope that other researchers may be encouraged to look for and identify more material from this understudied group.

Class	**ECHINOIDEA Leske, 1778**
ORDER	**ARBACIOIDA GREGORY, 1900**
Suborder	**ARBACIINA Gregory, 1900**
Family	**ARBACIIDAE Gray, 1855**
Genus	***Arbacia* Gray, 1835**
Species	***Arbacia* sp. cf. *A. sloani* (CLARK IN CLARK AND TWITCHELL, 1915)**

Material — Two specimens: 1 complete test (NCSM 9715: Figure 3.1 - 3.3), 1 broken test (but glued), portion of peristomal region missing (NCSM 9716: Figure 4.1 - 4.3)

Measurements —

Measurements in mm	NCSM 9715	NCSM 9716
Diameter	29.5	39.0
Height	15.8	22.5
Diameter of peristome	14.5	16.2
Greatest width of ambulacrum	6.1	9.0
Height of interambulacral plate at ambitus	2.9	3.2
Width of interambulacral plate at ambitus	5.7	7.0
Greatest width of apical system	9.1	10.7

Description — Test medium size, horizontal diameters 29.5 mm and 39.0 mm; moderately high, height 54-58 percent of diameter.

Apical system — Dicyclic, more visible in NCSM 9715 than NCSM 9716 due to cemented grains on test of NCSM 9716. Oculars small, exsert, with one imperforate tubercle: genital plates large, pentagonal, rugose with genital pore toward adoral point.

Periproct — Roughly diamond shaped, elongate from interambulacra 3 to 1.

Ambulacra — Narrow, regularly expanding to maximum width, approximately one-half width of interambulacra, at ambitus, maintaining nearly maximum width to peristome; poriferous zones relatively straight from apical system to near margin. Pore pairs below ambitus in oblique groups of three. Tubercles generally absent adapically, increasing in number and size to large tubercles in offset pairs adorally. Number of tubercles varies from 17 in 29.5 mm diameter specimen to 25 in 39.0 mm diameter specimen. One large pit in each ambulacrum near peristome.

Interambulacra — Plates low, wide, rugose; primary tubercles somewhat smaller from apical system to slightly above ambitus, no tubercles in median region, one tubercle on each plate near adradial suture. Tubercles largest at ambitus then reducing in size to peristome; usually two tubercles one each plate.

Peristome — Very large, approximately one-half as wide as horizontal diameter of test, round to sub-pentagonal; gill slits wide, continuing fair distance on test surface.

Tuberculation — Tubercles imperforate, smooth on finely rugose bosses, largest at ambitus. Where absent, plates rugose.

Occurrence — Spoil piles, Yorktown Formation (Lower Pliocene), PCS Phosphate (Lee Creek) Mine, near Aurora, Beaufort County, North Carolina.

Discussion — Cooke (1941; 1959) described several species of *Arbacia* from Late Miocene-Early Pliocene sediments of Virginia and South Carolina including: *A. waccamaw* Cooke, 1941, *A. rivuli* Cooke, 1941, *A. sloani*, and *A. improcera*. Kier (1963) described *A. crenulata* from the Tamiami Formation of Florida and reported *A. improcera* from the Yorktown Formation of Virginia and at the PCS Phosphate (Lee Creek) mine (Kier, 1972; 1983). Specimens described here differ from *A. waccamaw* by lacking a depressed test and by having much lower and wider interambulacral plates. *Arbacia rivuli* differs from specimens referred here in lacking conspicuous bare spaces on the interambulacra and having some insert ocular plates.

Specimens here are most similar in size, shape and ornamentation to *A. crenulata*, *A. improcera*, or *A. sloani*. However, Kier (1972) in his discussion of *A. improcera* suggested all three species may be synonymous. Kier (1972) cited that *A. improcera* might be conspecific with *A. crenulata*, because new material representative of *A. improcera* showed crenulations similar to those of *A. crenulata*. The two may still be separate species, however, as specimens described as *A. crenulata* by Kier (1963) lack tubercles on ocular plates, whereas tubercles are present on *A. improcera*. The specimens described here each possess tubercles on their ocular plates and thus are likely to be either *A. improcera* or *A. sloani*.

Cooke (1941) stated the most notable difference between *A. sloani* and *A. improcera* as greater height in *A. sloani*. Cooke (1959) noted that *A. improcera* is flatter with a more rugose sculpture than *A. sloani*. Kier (1972) observed that, with more specimens differences in heights between *A. improcera* and *A. sloani* were slight: 49 percent versus 53 percent; he

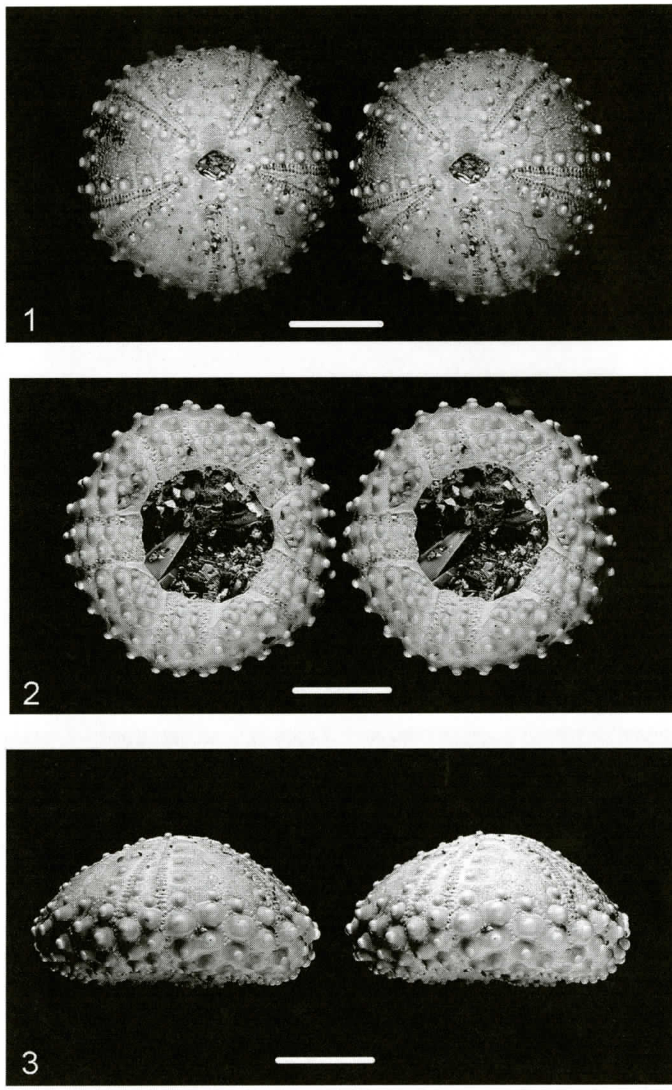

Figure 3: Stereo pairs of NCSM 9715, *Arbacia* sp. cf. *A. sloani* (Clark in Clark and Twitchell, 1915); 3.1 aboral view, 3.2 oral view, 3.3 lateral view. Scale bars = 10 mm.

could see no distinguishing features between the two species. Kier (1972) further suggested that *A. improcera* and *A. sloani* might be synonymous, but felt more specimens were required to resolve the issue. The two specimens we report herein have ornamentation similar to *A. improcera*, but have heights of 54 and 58 percent of their diameter, suggesting that they are most similar to *A. sloani*. Until the issue of synonymy between *A. improcera* and *A. sloani* can be resolved with more specimens these specimens are considered *Arbacia* sp. cf. *A. sloani*.

ORDER	**CLYPEASTEROIDA A. AGASSIZ, 1872**
Suborder	**SCUTELLINA Haeckel, 1896**
Family	**MELLITIDAE Stefanini, 1911**
Genus	***Encope* L. Agassiz, 1840**
Species	***Encope macrophora***

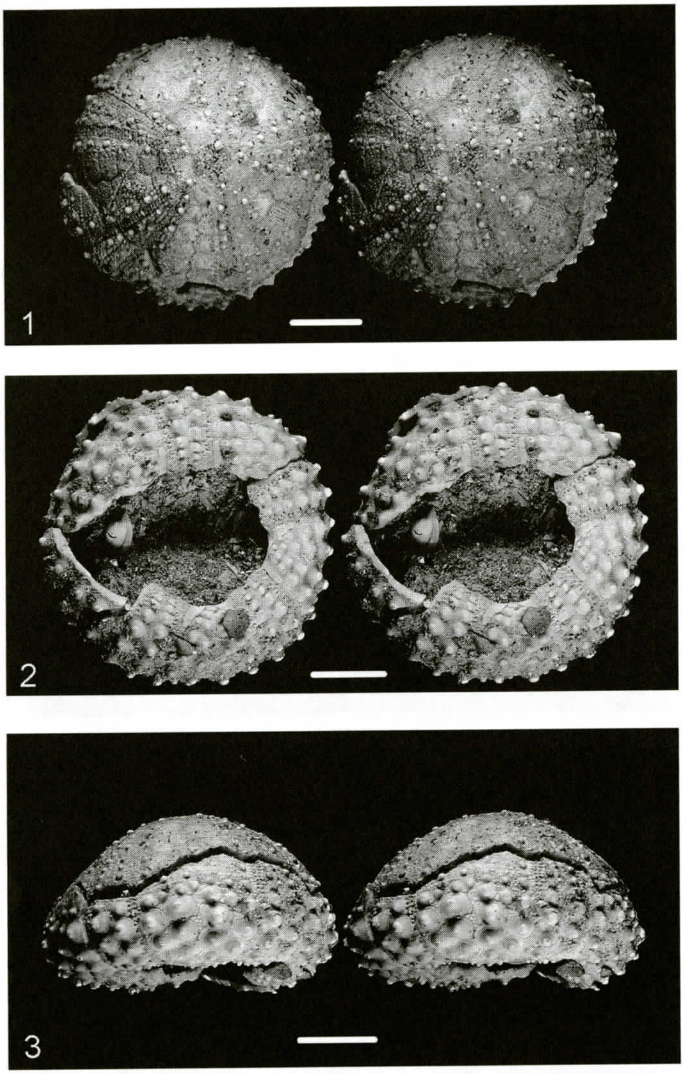

Figure 4: Stereo pairs of NCSM 9716, *Arbacia* sp. cf. *A. sloani*; 4.1 aboral view, 4.2 oral view, 4.3 lateral view. Scale bars = 10 mm.

(RAVENEL, 1842)

Material — One juvenile specimen (NCSM 9717: Figure 5.1 - 5.3)

Measurements — Length 29.1 mm, width 25.6 mm.

Description — Specimen heavily coated with calcite-cemented detritus, few surface details visible. Due to uniqueness and apparent fragility, no attempt beyond simple washing was made to clean the specimen.

Three anterior ambulacral notches not developed; two posterior notches form shallow indentations in edge of test. Interambulacral (anal) lunule large, well-developed. Anal pore (partially damaged) contained in wall of interambulacral lunule nearest peristome. Faint suggestions of feeding grooves. Interambulacral plates, barely visible, just to right of upper center on oral side. Unfortunately, a hole passes through center of test, completely obliterating peristome and aboral ambulacral center. Portions of five ambulacra visible. Fragments of several spines cemented to test, spine on adoral surface near interambulacral lunule resembles

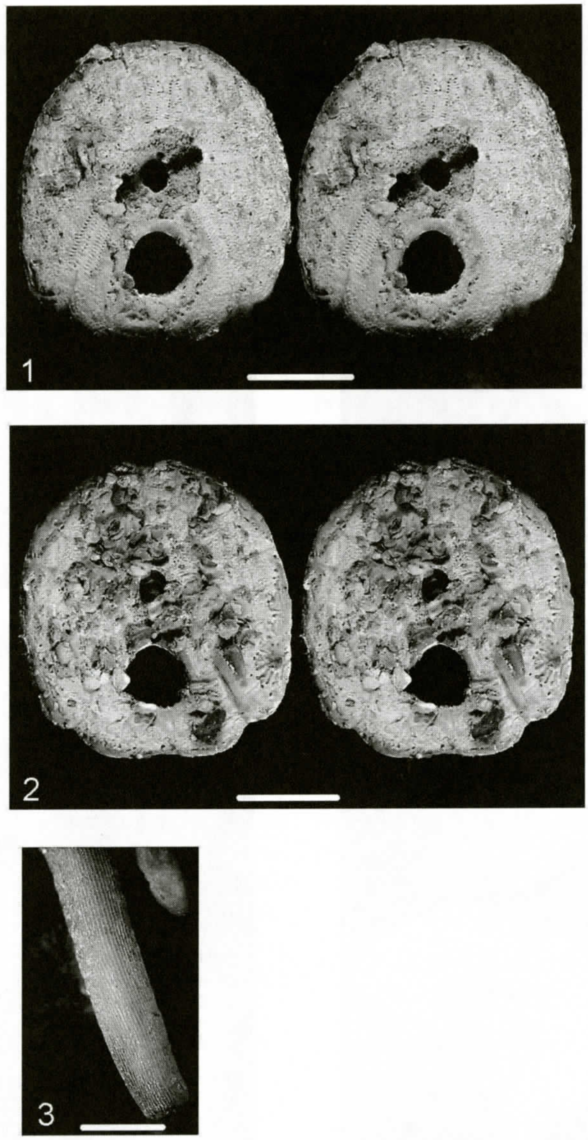

Figure 5: Stereo pairs of NCSM 9717, *Encope macrophora* (Ravenel, 1842); 5.1 aboral view, 5.2 oral view, 5.3 spine. Scale bars = 10 mm, except 1 mm for spine.

spines from extant *Encope* species.

Occurrence — Spoil piles, Yorktown Formation (Lower Pliocene), PCS Phosphate (Lee Creek) Mine, near Aurora, Beaufort County, North Carolina. Collector could not completely rule out that source was nearby James City sediments.

Discussion — Comparison with juvenile specimens of *Encope tamiamiensis* Mansfield, 1932 and *E. macrophora* of almost identical size indicates NCSM 9717 is *E. macrophora*. The primary difference between this specimen and *E. tamiamiensis* is the size of the interambulacral lunule and the complete absence of anterior notches. The height: width ratio of NCSM 9717 is 1.14 while that of the juvenile *E. tamiamiensis* was 1.07, much closer to Cooke's (1959) description of that species: "test as wide as long." The specimen described here also shows curving of the posterior ambulacra around the

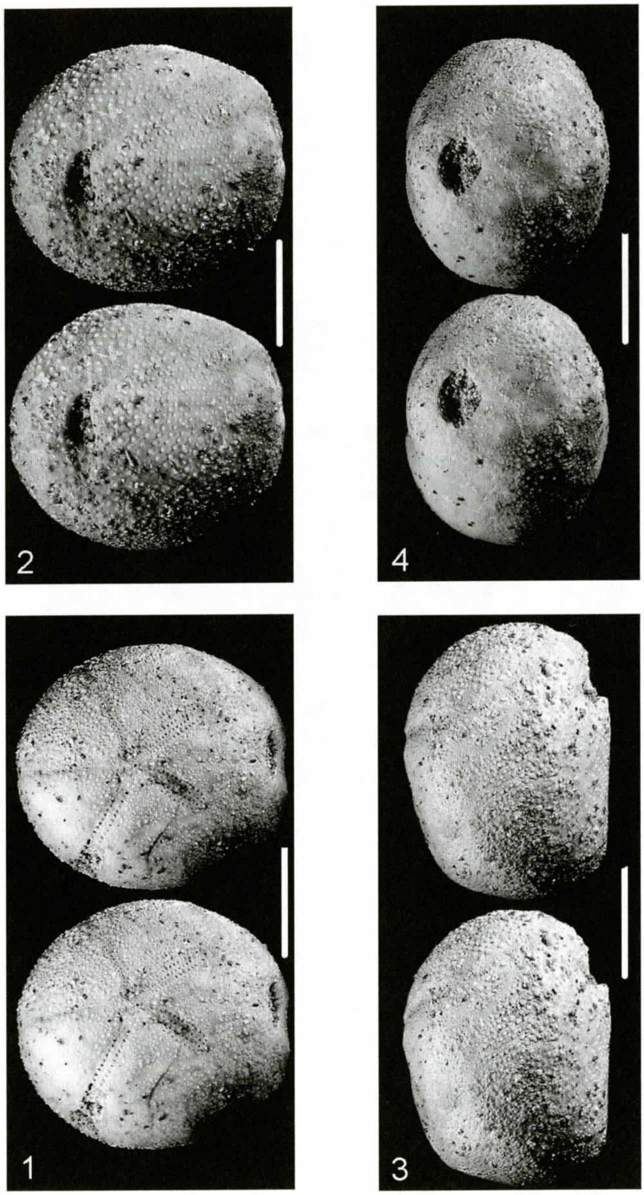

Figure 6: Stereo pairs of NCSM 9718, *Agassizia scrobiculata* Valenciennes, 1846; 6.1 aboral view, 6.2 oral view, 6.3 lateral view, 6.4 periproctal view. Scale bars = 10 mm.

lunule, which is consistent with *Encope macrophora*. In *Encope tamiamiensis* the posterior ambulacra are much more nearly straight.

Order **SPATANGOIDA Claus, 1876**

Suborder **HEMIASTERINA Fischer, 1966**

Family **SCHIZASTERIDAE Lambert, 1905**

Genus ***Agassizia* Agassiz and Desor, 1847**

Species ***Agassizia scrobiculata* Valenciennes, 1846**

Material — One complete test (NCSM 9718: Figure 6.1 - 6.4).

Measurements — Height 19.4 mm with greatest height posterior to center. Length 25.0 mm, width 22.7 mm. Width to length ratio is 0.93 with greatest width posterior to center. Height to length ratio is 1.07.

Description:

Test — Small, subglobular, slightly inflated anteriorly with posterior truncation not overhanging and dipping steeply. Horizontal outline ovate, widest in front, anterior portion convex, posterior portion concave below truncation.

Apical disk — Subcentral to slightly posterior (located 47 percent of test length from anterior margin). Detailed structure not visible.

Ambulacra — Ambulacra II and IV not petaloid, extend to margin, slightly sunken with single row of pore pairs. Ambulacra I and V petaloid, extend 60 percent of radius to margin, slightly sunken with a double row of pore pairs. Ambulacra III is subdued, not petaloid, extends to radius and too weathered to discern pores.

Interambulacral plates — Not visible.

Fascioles — Peripetalous fasciole anteriorly low, below margin, passing below petals II and IV, abruptly rising adapically and passing close, posteriorly below petals I and V. Lateroanal fasciole forms posterior to petals II, IV and extends to posterior just below periproct forming V-shaped sulcus. Four plates form sulcus, each with single large "bump."

Peristome — Large, with pronounced lip, crescent shaped, 3.9 mm long by 2.4 mm wide at maximum, located 21 percent of test length from anterior margin.

Periproct — High on posterior portion of test on truncation, transversely oval.

Plastron — Raised with large tubercles in biradial pattern.

Occurrence — Spoil piles, James City (Upper Pliocene/Pleistocene), PCS Phosphate (Lee Creek) Mine, near Aurora, Beaufort County, North Carolina.

Discussion — Cooke (1959) described *Agassizia scrobiculata* and *Agassizia porifera* (Ravenel, 1848) from Late Miocene and Kier (1963) reported *A. porifera* from the Upper Pliocene Caloosahatchee Formation in Florida. The specimen described here is clearly most similar to *A. scrobiculata* specimen described in Cooke (1959). This specimen does not have the rear truncation overhanging and is less inflated than *A. porifera*. The *A. porifera* specimens described by Kier (1963) ranged in length from approximately 50 mm to 79 mm and in width from 50 mm to 76 mm respectively. The specimen described here is significantly smaller and features in common with *A. scrobiculata* include overall test shape, petal pattern (including narrow and depressed ambulacra III), double pore spacing of ambulacra II and IV, lack of a deep sulcus and raised peristome with labrum.

ACKNOWLEDGMENTS

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EOLIAN DEVELOPMENT OF THE NAVAL LIVE OAKS DUNE ESCARPMENT, SANTA ROSA PENINSULA, FLORIDA (USA)

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ABSTRACT

A large sand dune ridge (herein referred to as the Naval Live Oaks Dune Ridge) occurs on the northwestern section of the Santa Rosa Peninsula. It is in the federally protected Naval Live Oaks Area, a part of the Gulf Islands National Seashore, located east of Gulf Breeze, Florida (USA). A large escarpment has formed on the north face of the dune ridge due to the landward encroachment of Pensacola Bay at Butcherpen Cove. The location, orientation, and unusual elevation of the Naval Live Oaks Dune escarpment suggest that eolian processes are responsible for its ongoing development. Winds predominately from the north carry sand from the shoreline and escarpment to the top of the sand dune. This continuing process has resulted in doubling the elevation of the existing ridge where the sand has accumulated. Vegetation on the lee side of the migrating dune is slowly being buried. The process has likely been ongoing since sea level conditions stabilized at or near present levels in the late Holocene. The escarpment will continue to supply sand to the slowly southward migrating dune ridge as long as the present shoreline and prevailing wind patterns remain relatively stable.

INTRODUCTION

The modern Florida coast has long held interest for coastal geologists and geomorphologists. Its development is typically defined by the action of water and related to changes in sea level position through the Quaternary (e.g., Belknap, 1985; Tanner, 1985; Davis, 1997). The shoreline extending along the panhandle of Florida has been the focus of numerous studies that have documented its sedimentological and geomorphic development (e.g., Kurz, 1942; Shepard, 1960; Kwon, 1969; Stapor, 1973; Stapor and Tanner, 1977; Otvos, 1985, 1992; Donoghue and Tanner, 1992; Balsillie and Clark, 2001).

While sea level change can dramatically impact shoreline geomorphology, its stasis generally allows for the maturity of the coastal environment. Features such as the 32-kilometer-long Santa Rosa Peninsula, suggest a relatively stable former sea level position coupled with energetic westward-directed longshore

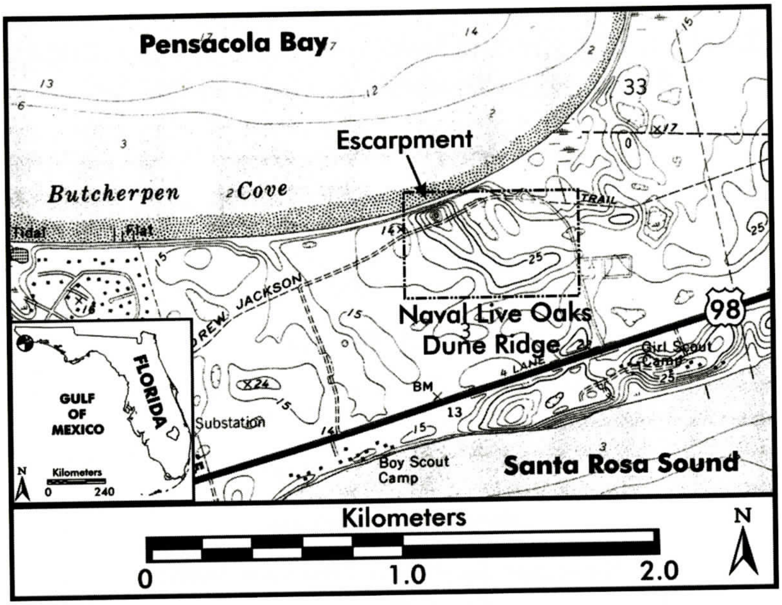

Figure 1. The Naval Live Oaks Dune Ridge is within the small box. Its northern exposure forms a large erosional escarpment along a portion of Pensacola Bay at Butcherpen Cove. Winds from the north transport loose quartz sand grains up the escarpment and deposit them along the top of the dune ridge. Map constructed from the U.S. Geological Survey 1992, Gulf Breeze, Fla., 7.5 minute quadrangle. Topographic contours are in five-foot intervals with elevations in feet.

drift. The large, isolated sand dunes and ridges along portions of the peninsula were likely initiated along the shoreline. With a drop in sea level, these features were further developed by eolian processes via dune migration, perhaps during a period of late Wisconsin aridity (Otvos, 2004). Eventually, the dunes were stabilized by various forms of vegetation (e.g., Hesp, 1983).

A prominent sand dune ridge and cliff scarp are located adjacent to Pensacola Bay at Butcherpen Cove on the northwestern section of the Santa Rosa Peninsula (Figure 1). The escarpment is part of a northwest-southeast trending curvilinear dune ridge. We examined the area to determine the cause of the large escarpment and sand dune to ascertain the geologic processes responsible for their historic and ongoing development.

SANTA ROSA PENINSULA GEOMORPHIC PROVINCE

Cooke (1945) subdivided the western Florida panhandle into the Western Highlands and Coastal Lowlands based on the 30-meter elevation contour. Further geomorphic refinement placed much of the western panhandle within the Southern Pine Hills District (Means and others, 2000). Recently, Scott (2005) proposed unifying the geomorphic divisions across northern Florida with those already in use across the Gulf Coastal Plain. He also changed the boundary separating the Western Highlands and Coastal Lowlands to the 15-meter elevation (Scott, 2005). As a result, the Santa Rosa Peninsula with its coastal setting and lower than 15-meter elevation occurs within the Gulf Coastal Lowlands.

GEOLOGIC HISTORY OF THE SANTA ROSA PENINSULA

The stratigraphy of the coastal portion of Santa Rosa County has been derived from several geological investigations (Heath and Clark, 1951; Musgrove and others, 1961; Barraclough and Marsh, 1962; Musgrove and others, 1965;

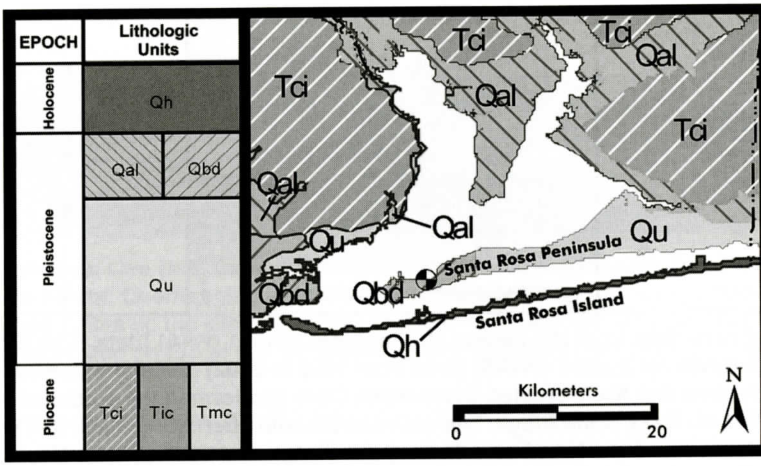

Figure 2. Geologic map of the coastal area shows the westward-directed Pleistocene age progression of the Santa Rosa Peninsula. The round circle with alternating black and white panels indicates the location of the Naval Live Oaks Dune Ridge and escarpment. Inset presents the geologic column of the stratigraphic layers, Qh-Holocene sediments, Qal-Quaternary Alluvium, Qbd-Quaternary Beach Ridge and Dune, Qu-Quaternary Undifferentiated, Tci-Tertiary Citronelle Formation (Pliocene). Map derived from Scott and others, 2001.

Marsh, 1966). The surface and shallow subsurface stratigraphy of the area likely reflect coastal processes consistent with changes in sea level position from the Pleistocene to the present (e.g., Balsillie and Donoghue, 2004). Initial geologic investigations of the area defined the surficial sediments as undifferentiated Pleistocene and recent deposits (e.g., Heath and Clark, 1951; Marsh, 1966). These units are paleontologically poor and as a result their age designation has remained unchanged even with recent geological mapping (Scott, 1991, 1993, 2001). The surficial sediments have likely been reworked on numerous occasions by water and wind in association with sea level changes. Features, such as the Santa Rosa Peninsula, are believed to have formed during a portion of the Pleistocene when sea level was considerably higher than at present (Marsh, 1966; Otvos, 1970; Tanner, 1985). Longshore drift carried siliciclastic sediments westward along the coastline forming the spit that eventually became subaerially exposed to form the modern peninsula (Figure 2).

Historically, along the Santa Rosa Peninsula, sea level has fluctuated sufficiently to allow the formation of low-lying nearshore beach ridges and dunes. Eolian processes later combined some of these features into larger coastal dune ridges with some reaching elevations approximating 7.6 meters. Storm washover, changes in the prevailing wind direction, and human activities have all contributed to the separation and flattening of the various linear dune ridges that at one time likely extended further along the peninsula. Today, several large isolated dune ridges still occur along portions of the Santa Rosa Peninsula. One such prominent feature is found in the federally protected Naval Live Oaks Area, a part of the Gulf Islands National Seashore, east of Gulf Breeze, Florida (USA).

HUMAN USE AND HISTORY OF THE SANTA ROSA PENINSULA

The Santa Rosa Peninsula has historical cultural significance. Archaeologists have identified numerous Indian middens situated along its shores, including sites near Butcherpen Cove and the Naval Live Oaks Dune (UWF Institute of Archaeology, 2005). During the European colonial periods, the British and Spanish used the peninsula as a source for wood, a careening ground, and a site for shipyards. Live oak timber was utilized for the construction of ships, and a few settlers herded cattle on the peninsu-

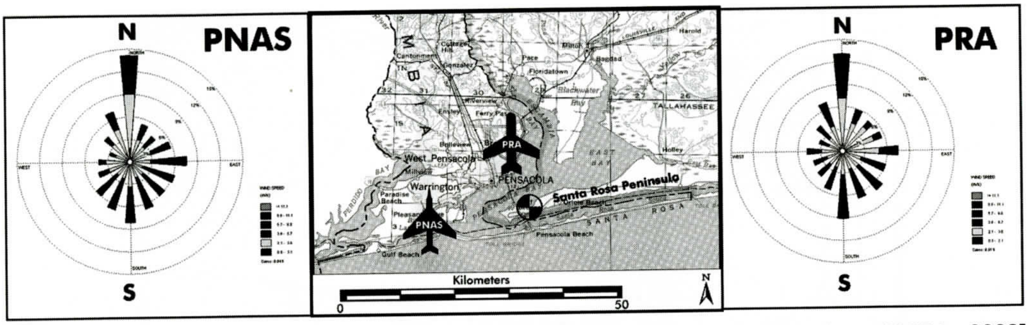

Figure 3. Wind rose data from the Pensacola Regional Airport (PRA) [data from 1995 to 2002] and Pensacola Naval Air Station (PNAS) [data from 1995 to 2002] reveal the dominant north-to-south wind direction that likely created Butcherpen Cove and formed the erosional escarpment along the Naval Live Oaks Dune Ridge. The round circle with alternating black and white panels indicates the location of the Naval Live Oaks Dune Ridge and escarpment. Map constructed from the U.S. Geological Survey 1967, State map of Florida, 1:500,000 scale. Topographic contours are in 50-foot intervals with elevations in feet.

la. (Rucker, 1990).

With the American acquisition of Florida in 1821, a viable transportation corridor was needed to link Pensacola with the new capital of Tallahassee and St. Augustine to the east. In 1824, the historic Pensacola-St. Augustine Road was constructed by U.S. soldiers stationed at Pensacola. From its terminus at the western end of the Santa Rosa Peninsula, the road proceeds eastward along the peninsula, passing over the Naval Live Oaks Dune Ridge only a short distance south of the escarpment. Today, this old military road remains in use as a hiking trail in the park (Boyd, 1935; Rucker, 1990).

In 1828, much of the western section of the Santa Rosa Peninsula was purchased by the U.S. government as the Naval Live Oaks Plantation. President John Quincy Adams, a guiding force in this project, was keen on preserving stands of southern live oak trees for the construction of U.S. naval warships. Local Judge Henry M. Brackenridge served as the manager of the nation's first experimental forestry station, nurturing and cultivating the valuable stands of live oak trees (Rucker, 1990; Snell, 1983).

After the Civil War, and with the advent of iron-clad warships, the forested area was no longer viewed as a strategic military resource. However, timbers from the live oak reservation were used in the 1920s for the restoration of the historic USS *Constitution* (Bowden, 1994). In the late 1800s, a cattle corral, pen, and slaughterhouse were constructed east of the Naval Live Oaks Dune escarpment providing Butcherpen Cove its name (UWF Institute of Archaeology, 2005). Various tracts of land within the Live Oak Plantation were utilized in the 1930s as sites for Boy Scout and Girl Scout camps and the large bluff and capping dune were noted by locals as early as 1934 (Bowden, 1994; Dawkins, 2005; Johnson, 1992). In 1972, the National Park Service incorporated the land associated with the Naval Live Oaks within the Gulf Islands National Seashore. Today, the Naval Live Oaks Reservation is the portion of that park located exclusively on the Santa Rosa Peninsula (Bowden, 1994).

PREVAILING WINDS ALONG THE WESTERN FLORIDA COASTLINE

Wind directional data from a number of coastal weather stations along this portion of the northern Gulf of Mexico were analyzed in an effort to determine prevailing surface level wind patterns. Results indicate that the wind direction is predominately north-to-south. The wind rose diagrams presented in Figure 3 document the surface level wind direction in the Pensacola area and are consistent with the wind patterns from other nearby coastal weather stations. Within the study area, the northerly winds reached top speeds ranging from 8.8 to 11.1 m/

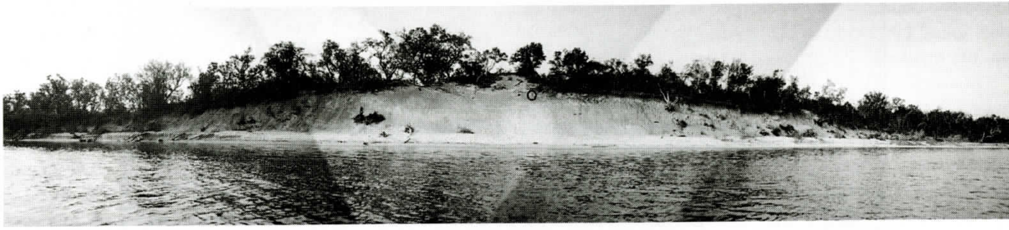

Figure 4. The Naval Live Oaks escarpment exposed at Butcherpen Cove. Person in black circle is 1.5 meters in height. Overhanging vegetation along the majority of the exposure tends to funnel the loose sand grains up the slightly lunate parabolic dune surface. As a result, the sand is transported and deposited on top of the dune escarpment elevating the existing dune ridge.

s, with the more predominant winds averaging between 2.1 to 8.8 m/s, based on hourly measurements obtained from 1995 to 2002. Some of these velocities exceed Bagnold's (1954) threshold value (i.e., 4.27 m/s) for sand grain movement. The actual measured wind velocities would transport all but the most coarse fraction (i.e., <0.6 mm median diameter) of dry quartz sand particles (US Army Corps of Engineers, 2002). While southerly winds reached the same velocities, they occurred less often than winds blowing from the north.

The local topography surrounding Pensacola Bay influences wind patterns in the area. Surface winds tend to follow north-south oriented topographic features such as river valleys or slopes. An examination of the topographic map shown in Figure 3 reveals terrain elevations of 15 meters to greater than 30 meters above mean sea level bordering Pensacola Bay and extending further north to Escambia Bay and its river basin. These terrain elevations are likely significant enough to affect the flow of surface level winds and channel them in a southerly direction across Pensacola Bay impacting the Santa Rosa Peninsula and resulting in the development of Butcherpen Cove.

GEOMORPHOLOGY OF THE NAVAL LIVE OAKS DUNE RIDGE

The Naval Live Oaks Dune Ridge is located on the north side of the western portion of the Santa Rosa Peninsula. Its northern exposure forms an erosional escarpment at Butcherpen Cove. The cliff scarp is a prominent feature that is unusual in both size and elevation and out of place in its present location (Figure 4). Typical coastal depositional models based on eustatic changes fail to address the morphology and location of this dune ridge escarpment. Rather, the size and shape of the large feature suggest an origin by eolian processes (see Goldsmith, 1982) that have been in operation since sea level position stabilized at or near its present position during the late Holocene.

The escarpment exhibits a slightly concave lunate parabolic type of dune surface. Its lee side is covered with scrub vegetation and a fully mature live oak (*Quercus virginiana*) forest. The encroachment of the escarpment into the vegetated portion of the dune and dune ridge has created an overhanging root zone which enhances the funneling of sand to the top of the dune. The highest point of the dune (approximately 18 meters) occurs only a few meters downwind (i.e., south) from the edge of the escarpment. Sand carried up the escarpment by eolian processes continues to accumulate on top of the dune. Vegetation on the lee side of the escarpment is slowly being buried by southward-directed dune migration. The movement of sand across the top of the dune ridge is doubling its original elevation, most notably in the area of sand accumulation. The addition of sand to the base of the escarpment adds to its further development (Figure 5). Eolian dune building has been reported from similar backshore settings in Australia (Carter and others, 1990) and Padre Island, Texas (Weise and White, 1980).

SUMMARY AND CONCLUSIONS

The modern Florida coast is interpreted to

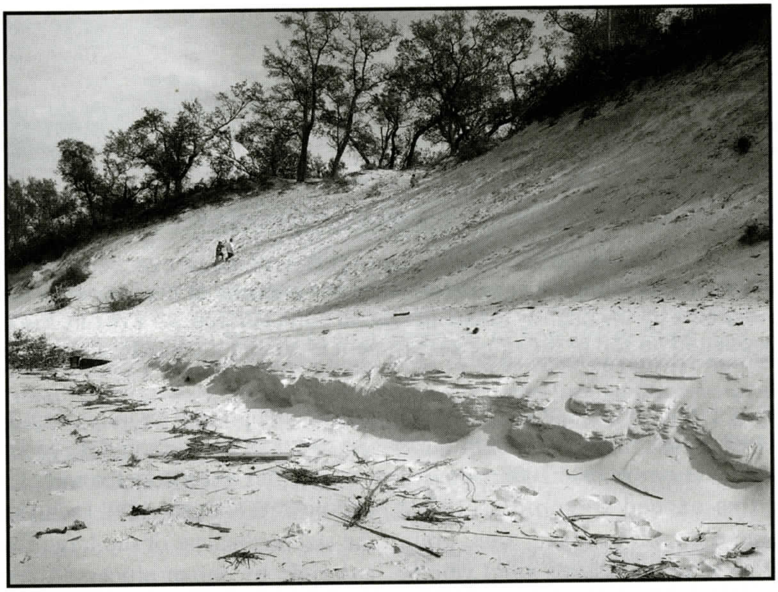

Figure 5. On September 16, 2004, a storm surge associated with Hurricane Ivan deposited this sand bench at the base of the Naval Live Oaks Dune Escarpment. This material is now available for transport (with sufficient wind velocity) to the top of the dune, further extending the elevated portion of the dune ridge.

have developed predominately by historic eustatic changes. Coastal areas along the western panhandle generally support this interpretation, but with certain exceptions. While features like the Santa Rosa Peninsula were likely formed due to an elevated Pleistocene sea level position coupled with westward-directed longshore drift, large isolated dunes and dune ridges identified on top of the peninsula were probably developed by eolian processes. One such large sand dune ridge occurs within the Naval Live Oaks Area adjacent to Butcherpen Cove.

The predominant north-to-south wind pattern in combination with the topographic channeling of surface wind in Escambia Bay and across Pensacola Bay has contributed to the formation of Butcherpen Cove and the Naval Live Oaks Dune Escarpment. The unique setting of the dune ridge both perpendicular to the Santa Rosa Peninsula and adjacent to Pensacola Bay has been essential to its ongoing eolian development. No other dunes occur in this manner along the back side of the Santa Rosa Peninsula. This large feature should continue to slowly migrate southward as long as sea level and prevailing wind patterns remain static.

ACKNOWLEDGMENTS

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SIMILAR INTENSITIES BUT DIFFERENT CAUSES IN BARRIER ISLAND EROSION: DISCUSSION OF A PAPER BY CARL R. FROEDE, JR.

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COMPARISONS: VARIABLE LEVELS OF HURRICANE IMPACTS ON DAUPHIN ISLAND AND ON MISSISSIPPI'S MAINLAND SHORE

My comments concern questionable conclusions in Froede's paper (2005), with the intention of adding pertinent geological context. The publication deals with effects of the 1998 hurricane, a Category 1 tropical cyclone at its impact on Dauphin Island. With the landfall site located 56-72 km west-northwest of the island, the magnitude of foredune destruction along most of Dauphin's Gulf beach and the extent of the overwash was comparable with that brought about both by much stronger and similar-intensity hurricanes. Effects of low-category ones, such as Elena (1985), Opal (1995), and Danny (1997), and of major cyclones (Frederic, 1979, Ivan, 2004, and Katrina, 2005) were remarkably similar in the highly vulnerable low, narrow central and western island sectors. The continuous foredune ridge behind the Gulf shoreline was repeatedly flattened by less powerful storms as well. Even with the relatively less destructive northwestern cyclone quadrants passing over Dauphin, foredune ridges that front the Gulf, dunes in the least resistant island sectors were washed away during Frederic's and Ivan's landfall just east of the island.

Several episodes of extensive foredune destruction along the low central-western island sectors in the past decade resulted from winter-spring storms, associated with Arctic and Pacific cold weather fronts that swoop down from the mainland. The degree of island erosion under Georges' destructive northeastern quadrant was commensurate with the storm's intensity. In contrast, the remarkable beach *aggradation* that took place on the Mississippi beaches at and west of Georges' landfall location, was a rare event even for low-category hurricanes (Otvos, 2004).

Froede correctly states that early in 2005 the constructive influence associated with the post-Georges storm-free fair weather phase was not detectable on Dauphin. The actual reason for this state, however, was the island's intensive re-erosion just a few months earlier. Ivan, a highly destructive, strong-Category 3 hurricane made landfall at nearby Gulf Shores, Alabama in late September, 2004.

RAPID SHORE AND DUNE RECESSION — ONLY BY HURRICANE EROSION?

Georges' erosion around the western shore pavilion (Froede, 2005; Figs. 6A and B), already subjected to dune retreat and undercutting, was but a minor sequel to episodes of remarkable, localized dune retreat in the Fishing Pier recreational area that started seven years earlier (Douglass, 1994, p. 312-314; Otvos, 1997, Fig. 50). The Fishing Pier area is located along the western fringe of the broad and high, forested eastern island sector. This sector is better protected against storms than the much more fragile central and western island segments are. Intensive erosion at the Fishing Pier resulted in the destruction or isolation of three large structures on a high shore dune ridge in 1993-1994. Storms during the spring-to-early summer season caused significant local erosion in 1991 and during the 1993-1994 winter-spring season, associated with the passage of cold fronts.

Shore erosion in 1991 was influenced by landward migration of a large sand bar; the narrowing and deepening of its landward flanking channel east of the Fishing Pier complex, in front of Park and Beach Board Beach. Shore retreat amounted to 13 m (Douglass, 1994). Wave current-related steady channel deepening and consequent intrusion of larger, erosive Gulf waves into Pelican Passage at the Fishing Pier area. This was caused by channel narrowing, driven by the steady northward progradation of Sand-Pelican Island. These changes triggered shore and associated dune recession (Figs. 2, 3, and 14, in: Douglass, 1994).

EXHUMED PALEOSOL OR OXIDIZED PLEISTOCENE BARRIER SURFACE SANDS IN THE ISLAND CORE AREA?

Eastern Dauphin Island formed as a separate island by beach and dune plain aggradation around a late Pleistocene barrier ridge during the mid-to-late Holocene sea-level rise. The Pleistocene island core was formed by the Sangamon Interglacial Gulfport Formation barrier ridge complex, comfortably underlain by the late Pleistocene Biloxi Formation (Otvos, 1985; Otvos and Giardino, 2004; Fig. 4). The Gulfport barrier sectors consist mostly of well to very well sorted medium eolian and intertidal-subtidal sand. These barriers form a discontinuous trend behind long stretches of the northern Gulf shoreline. The Gulfport Formation obviously is much too old for radiocarbon dating but has provided excellent OSL dates (Otvos, 2005).

In various other barrier sectors the Gulfport outcrops similarly display a thin, weakly oxidized yellowish-brown or tan top interval. This feature is often exposed in a low beach scarp that stretches eastward from the Fishing Pier. The yellowish-brown and tan sands, generally 0.5-1.0 m thick above ground level, are overlain by white Recent dune sand in the scarp. The top Gulfport sand interval was oxidized during its prolonged late Pleistocene-to-mid Holocene exposure in the land surface (Otvos, 1979, Fig. 7; 1997). English and Haywick (1996), did not supply any pedological structure and chemistry data to support their contention. Without proper analysis, they also rejected ample drillcore and stratigraphic data (Otvos, 1979; see also Otvos and Giardino, 2004) in stating that the oxidized interval represents the absurdly thick A and B horizons of an only 200 years old soil unit, formed when the Gulf beach was wider here. Unfortunately, Froede (p. 48-49, Fig. 5A), following them, has also accepted the darker-hued interval that caps the Gulfport, as (a contradiction in terms?) modern paleosol.

As in other Gulf coastal outcrops, the oxidized sands occasionally are moderately case-hardened but away from the outcrop are not indurated. Dead tree trunks that stand on dune slopes and in nearshore waters roots, and their thin gray buried forest soil (Froede, 2005, Figs. 7A, B) of subrecent age are remnants of a dense pine forest that still covers most of the eastern island. The overburden-sensitive pine roots were smothered by thick landward shifting eolian sand layers; the trees killed (Otvos, 1973, Fig. 26).

Penetrated locally by pine roots that were dated 235±80 yr B.P. from stumps in the surf zone (English and Haywick, 1996), the oxidized Pleistocene sand horizon should not be confused with the buried subrecent forest soil. The oxidized top Gulfport interval does not display a recognizable soil profile or an ancient buried soil horizon with prerequisite pedogenic features and chemical characteristics.

EROSIONAL STATE AND EASTERN SAND SOURCE OF DAUPHIN AND THE MISSISSIPPI BARRIER ISLANDS

Froede emphasizes what he regards as the island's precarious, fragile setting and erosional vulnerability; its allegedly tenuous natural sand supply. He asserts Dauphin's reliance on regularly continuing, intensive human intervention for its very existence. However, the large sand volumes transmitted by littoral drift from southeast Alabama along and across the huge ebb tidal delta and the tidal current transport from Mobile Bay suggest not a frail, human-dependent but rather robust natural sediment supply and transmission processes. Despite recurring storm- and fair weather-related sediment loss and because of this abundant source, islands in the Dauphin-Mississippi island chain downdrift were able to restore and maintain themselves rather adequately in the past. Dauphin and the other islands in the chain underwent recurring and very extensive segmentation and major area loss; mostly in the exposed updrift and low central sectors (Otvos and Giardino, 2004, p. 68).

Dauphin, being closest to the prime littoral sand source, was more capable of restoring and extending itself after major storm events and prolonged erosional phases than were other islands, such as Ship and Horn. Located in the distal part of the chain, these islands obtained diminished volumes from the westward transmitted littoral drift. Even if losing surface area in their very brief recorded history, the islands probably could not have maintained themselves without this dominant sediment source.

Along the Gulf shore of Dauphin Island net littoral drift downdrift from Pelican-Sand Island on the western flank of Mobile Bay ebb-tidal delta, was calculated as ca. 200,000 m^3/yr. This figure is about an order of magnitude greater than the westward-directed drift along the eastward adjacent low-energy eastern shore stretch that faces the ebb delta (Douglass, 1994). Dauphin only very rarely received repeated beach nourishment. The highly erosive eastern island tip and the adjacent 1909 groin field did receive nourishment but only on very few occasions. Another limited erosion hotspot, the Fishing Pier beach was nourished in June, 1991 and again after 1993 . A boulder riprap was placed on the foreshore (Froede, Figs. 5A, B) in 1994. These were relatively very minor and with the questionable exception of the recent, second eastern groin field-fill, rather ineffective attempts at beach conservation and enhancement. Such rare, limited and short-lived restoration efforts, of course, had zero influence on net littoral drift and long-term island maintenance.

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Reply to Ervin G. Otvos: "Similar Intensities But Different Causes in Barrier Island Erosion"

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INTRODUCTION

I am very appreciative of the help received from Ervin Otvos preceding the publication of my article regarding Hurricane Georges' impact to Dauphin Island. I am equally grateful for his careful reading and comment on the published paper. I have greatly benefitted from the published works by Otvos reflective of his many years of studying this portion of the Gulf of Mexico. While we have differences of opinion regarding certain ideas related to Dauphin Island, Alabama, I believe that we are predominately in agreement.

COMPARISONS: VARIABLE LEVELS OF HURRICANE IMPACTS ON DAUPHIN ISLAND AND ON MISSISSIPPI'S MAINLAND SHORE.

I agree with Otvos that strong storms adversely impact the morphology of Dauphin Island, especially the more low-lying western portion of that island. I also agree that storm erosion is not limited solely to hurricanes. However, the focus of my article was specific to Hurricane Georges' impact to Dauphin Island and the erosional changes that I photographically documented. This same hurricane impacted the Mississippi coast in a completely different manner (Otvos, 2004). Hurricane Georges changed the position and profile of the gulf-facing surf zone and beach (respectively). The gulf side of the island has not returned to its pre-Hurricane Georges' shape or position in the many years following this storm's passing. Subsequent hurricanes and storms have only served to further erode (at varying levels) the island.

RAPID SHORE AND DUNE RECESSION–ONLY BY HURRICANE EROSION?

I agree that erosion has occurred in the area adjacent to the fishing pier during the 1990s. However, my photographic documentation of this area during this time also reveals that some level of stasis related to beach position was occurring and that eolian sand was actually accumulating in and around the picnic area during the latter portion of the decade. The greatest beach erosion that I witnessed followed passing storms. I did not have the opportunity to investigate normal tidal cycle erosion. The dune beneath the picnic pavilion pictured in Figure 6A was lost *solely* as a result of the storm surge associated with Hurricane Georges. My focus in this article was to pictorially present the specific changes that occurred as a result of Hurricane Georges' impact to this portion of the island.

EXHUMED PALEOSOL OR OXIDIZED PLEISTOCENE BARRIER SURFACE SANDS IN THE ISLAND CORE AREA?

Otvos has conducted extensive investigations across Dauphin Island and I can only repeat his excellent work. The stratum in question is the Pleistocene age Gulfport Formation which he describes as:

> Medium and fine-grained, very well, and well-to-moderately well-sorted, white sand predominates...although lesser amounts of silty sands of poorer sorting are also present. Near the surface, the sands are often oxidized to a light orange-yellow color. Dark humate impregnations

> are common and extensive... Humate developed through the precipitation of soluble and colloidally dispersed humic substance. These substances were derived from plant material on the land surface and transported downward by percolating water. Brackish ground waters, due to their Mg, Al and Fe content, caused precipitation of these humic substances in pores of Gulfport sands (Otvos, 1985, p. 31).

Following the publication of my article, Otvos (2005, personal communication) shared with me his ideas regarding the possibility of Gulfport Formation pedogenesis:

> The oxidized top interval of the Gulfport Formation has probably been subjected to various pedological processes during and ever since the Sangamon Interglacial. Whether or not this actually represents a paleosol horizon appears to remain an open question until detailed and conclusive pedological studies can prove/disprove this.

I agree with Otvos. My work did not attempt to determine whether the soil classification described by English and Haywick (1996) for the stratified Gulfport Formation exposure is technically correct or not. Rather, all that I intended to convey was that the outcrop no longer existed as it was eroded away during the course of Hurricane Georges.

EROSIONAL STATE AND EASTERN SAND SOURCE OF DAUPHIN AND THE MISSISSIPPI BARRIER ISLANDS

Along with Otvos, I agree that Dauphin Island and the other adjacent Mississippi barrier islands have undergone extensive segmentation over the years since Europeans first came into the area. I also agree that the close proximity of Dauphin Island to the submerged Mobile Bay ebb-tidal delta has aided in the preservation of the eastern end of the island. This feature has likely contributed sand to the offshore bar system adjacent to the gulf-facing side of Dauphin Island predominately westward from the pier/picnic area. However, I respectfully disagree that the Mobile Bay ebb-tidal delta has contributed sand that has resulted in beach growth to the eastern portion of the island. I also disagree that this sand source has contributed to sustained beach development along the gulf side of the western and more low-lying portion of the island. Perhaps this growth would occur if passing hurricanes and large storms (generated from the north) did not impact the island so often.

Based on my multi-year photographic documentation from various gulf-facing sections of the island, the east-to-west longshore transport appears to contribute quartz sand particles to a shore-parallel submerged barrier bar. Sand from this feature moves shoreward in the spring and offshore during the winter. However, I have also witnessed exceptions to this sand cycle whereby the sand that was carried gulfward during a storm moved rapidly back to shore, resulting in the restoration of portions of the beach (Froede, 1998). The constant longshore drift of sand particles along the gulf side of the island ensures that sand is always available to maintain the offshore bar and contribute (to a lesser extent) to the surf zone profile. In my many years of visiting the island, I have not witnessed the permanent and continual gulfward extension of the beach (either in plan view or in profile) that might be expected from the drift of "200,000 m^3/yr" of sand from the Pelican-Sand Island/Mobile Bay ebb-tidal delta complex.

I must disagree with Otvos on his assessment regarding beach nourishment activities across Dauphin Island from the 1990s to the present. Otvos correctly stated that beach nourishment activities have occurred several times at the southeastern end of the island and within the fishing pier-picnic area. Considerable volumes of sand (on multiple occasions) have been added to the beach adjacent to the groin field on the southeastern end of the island. This added sand has repeatedly been eroded and transported westward along the gulf side of the island by longshore drift. This migrating sand temporarily extends a portion of the beach gulfward only later to recede as the sand moves further westward. Lesser amounts of sand have been added to the pier-picnic area. Not mentioned by Dr.

Figure 1. Facing west along the western low-lying developed portion of the island. The image was taken November 2000 following the construction of a berm designed to reestablish a beach and reduce flooding across this portion of the island. Approximately 229,366 cubic meters of sand was added to this portion of the beach (Henderson, 2005a). This was the largest beach renourishment project that I have witnessed in my many years of photographing various portions of the island. It was eroded away in only a few years.

Otvos is perhaps the largest beach nourishment project conducted on Dauphin Island. Following Hurricane Georges, a large artificial dune berm was constructed along a portion of the developed western low-lying portion of the island (Figure 1). This feature was subsequently eroded away by a combination of the daily tidal cycle, large storms, and hurricanes. Interestingly, another large manmade shore-parallel dune berm is currently being considered for this same portion of Dauphin Island (Henderson, 2005b). I am aware of these three sections of the gulf-facing side of Dauphin Island that have been renourished in the past. Beach renourishment activities will likely continue in the future. The added sand is intended to help the island retain its predefined shape and size–at least for the developed portion of the island. With the recent impact of Hurricane Katrina and its devastation to Dauphin Island, I remain convinced that much of this island is destined to undergo continual change even with renourishment activities due to its precarious position in the northern Gulf of Mexico.

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