GEOMORPHOLOGY AND FACES RELATIONSHIPS
OF QUATERNARY BARRIER ISLAND
COMPLEXES NEAR ST. MARYS,
GEORGIA

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Charlotte Fries: National Park Service
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28th Annual Field Trip
Georgia Geological Society

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Volume 13, Number 1  November, 1993
This guidebook is dedicated to the memory of our friend and colleague, Dr. Robert W. Frey.
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28th Annual Field Trip
Georgia Geological Society

Georgia Geological Society Guidebooks
Volume 13, Number 1  November, 1993
28th Annual Georgia Geological Society Field Trip Itinerary

DAY 1: November 20, 1993
0800  Assemble at National Park Service Headquarters, St. Marys, Georgia.
0830  Depart from National Park Service Dock.
0900  Arrive at Cumberland Island National Seashore.
0915  Stop 1: Overview of Cumberland Island Geology
      Contribution 1: A Regional Overview of the Geology of Barrier Complexes
                     Vernon J. Henry Jr., Kathleen M. Farrell, and Stephen V. Cofer-Shabica
      Stop 2: Maritime Forest (Contribution 2) Charlotte Fries
      Stop 3: The Oceanside of Cumberland Island.
      Contribution 3: Ocean Shoreline Changes  Stephen V. Cofer-Shabica
      Contribution 4: The Beach  James H. Darrell II, Nancy A. Brannen, and
                     Gale A. Bishop
      Contribution 5: Ecology and Paleocology of Georgia Ghost Shrimp
                     Gale A. Bishop and Nancy A. Brannen
      Contribution 6: Nesting Traces of the Loggerhead Sea Turtle
                     Nancy A. Brannen and Gale A. Bishop

1200 Lunch
      Stop 4: The Back Dune Ridge Complex (Contribution 7) Stephen V. Cofer-Shabica
      Stop 5: The Salt Marsh.
      Contribution 8: Beach Creek Salt Marsh  Stephen V. Cofer-Shabica
      Contribution 9: Backbarrier Geomorphology, Facies Sequences and
                     Processes  Kathleen M. Farrell, Vernon J. Henry Jr. and Stephen V. Cofer-Shabica

Stop 6: Cultural History of Cumberland Island (Contribution 10) Charlotte Fries
Stop 7: Backbarrier Processes.
       Contribution 11: River Trail Shoreline Recession  Stephen V. Cofer-Shabica
       Contribution 9: Backbarrier Geomorphology, Facies Sequences and
                     Processes  Kathleen M. Farrell, Vernon J. Henry Jr. and Stephen V. Cofer-Shabica

1630  Arrive at National Park Service Dock.

Day 2: November 21, 1993
0730  Assemble at National Park Service Headquarters, St. Marys, Georgia.
0800  Depart from National Park Service Dock.
0830  Stop 1: Bells/Roses Bluff: Pamlico Barrier Complex.
      Contribution 12: Yulee Heavy-Mineral Deposits
                     Fredric L. Pirkle, E.C. Pirkle and John G. Reynolds
      Contribution 14: Stratigraphy and Facies of Bells/Roses Bluff, Pamlico
      Note: Conditions may not permit on-shore examination of these outcrops.

1000 Stop 2: Reids Bluff: Pamlico Barrier Complex.
       Contribution 13: Palynology and Paleocology of Reids Bluff
                     Fredrick J. Rich and Fredric L. Pirkle
       Contribution 15: Stratigraphy and Facies of Reids Bluff, Pamlico
       Note: Participants will be ferried ashore from the Cumberland Princess. Please use
       caution in transferring to and from the smaller vessels. Lunch will be served on the
       outcrop.

1400  Depart from Bells Bluff.
1430  Arrive National Park Service Dock. End of Field Trip.
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THE 28TH ANNUAL FIELD TRIP OF THE
GEORGIA GEOLOGICAL SOCIETY

INTRODUCTION

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Georgia Southern University, Statesboro, Georgia 30460
Kathleen M. Farrell North Carolina Geological Survey
P. O. Box 27687, Raleigh, North Carolina 27611

OVERVIEW

The purpose of this field trip is to compare the
geomorphic features and sedimentary facies of
Cumberland Island, Georgia, and its backbarrier regions
with nearby outcrops of Pleistocene barrier island deposits.
This field trip includes discussions of: (1) modern barrier
island facies, geomorphology, and processes; (2) the
preservation and origin of barrier-associated facies in
Pleistocene deposits; (3) the significance and origin of
Ophiomorpha nodosa, a shoreline indicator; (4) the age of
the Pleistocene deposits; and (5) the origin of economically
valuable heavy mineral deposits.

The field trip originates in St. Marys, Georgia, and
includes localities on Cumberland Island, Georgia,
and along the Bells and St. Marys River, northeastern
Florida, that will be accessed by boat (Fig. 1 of Contribution
1). The first day consists of a walking tour of Cumberland
Island, a Pleistocene/Holocene-age barrier complex. The
second day is an upriver excursion along backbarrier tidal
channels to classic Pleistocene localities that are exposed
along bluffs being actively eroded by migrating tidal
channels.

The tour of Cumberland Island will demonstrate the
distribution of modern geomorphic features including
backbarrier salt marsh, tidal channel, tidal flat and point
bar environments as well as the dunes, beach ridges, and
beaches that are typically associated with sandy barrier
complexes. Discussions will focus on the dynamic
processes of shoreline change, as well as lithologic,
mineralogic, and biogenic attributes of beaches and dunes.
Facies relationships of a commonly neglected coastal
environment, the muddy backbarrier region, will be
examined in cores and the dynamic processes of
backbarrier evolution will be discussed. A review of the
taxonomy, occurrence, and life habits of the Carolinian
ghost shrimp (formerly Callianassa) will be provided.
For the first time (in print) the occurrence of nesting traces
of the loggerhead sea turtle (Caretta caretta (Linne)) is
presented. Archeological aspects and vegetation patterns
on Cumberland Island will also be explained.

On the second day of the field trip, we will visit Bells/
Roses Bluff and Reids Bluff, two classic Pleistocene
localities exposed inland along the Bells and St. Marys
Rivers. Exposures at both bluffs are oriented
approximately parallel and/or perpendicular to the long
axes of the ancient Pamlico barrier island complex. Both
localities contain fossiliferous strata that have been
sampled for age dates and palynology. Facies exposed in
these outcrops, such as parallel-laminated sand and
complexly laminated clay and sand, have been compared
to modern barrier island deposits. The vertical sequence
of facies has been used to determine sea level stands
during Pleistocene time. Of particular importance are the
facies containing the trace fossil, Ophiomorpha nodosa,
and fossil Cypress stumps, both indicators of sea level
conditions. The economically valuable Yulee heavy
mineral deposits are exposed at Bells/Roses Bluff and
Reids Bluff.

ORGANIZATION OF THE FIELD TRIP
AND GUIDEBOOK

A large number of authors have contributed to this
guidebook and to the discussions enumerated above. The
authors intend to build on the voluminous literature that
exists for modern and ancient environments of the Georgia
Coast. References are presented at the end of each
contribution. The fifteen contributions here, which consist
of both complete papers and short write-ups of individual
field stops, are arranged and numbered in this guidebook
according to the content and order of field stops. The field
trip map (Fig. 1 of Contribution 1) shows the location of
the field stops. The Itinerary of the field trip (page ii) lists
the field stops, identifies the stop presenter(s), and lists the
guidebook contributions relevant.
REGIONAL GEOMORPHOLOGY

Cumberland Island is an elongate, north-trending barrier island and the southern-most of Georgia’s Sea Islands (Fig. 1). It is separated from Amelia Island, Florida, by St. Mary’s Sound and from Jekyll Island by St Andrews’s Sound. The sounds function as major tidal inlets that connect the Atlantic Ocean with the St. Marys River, the Satilla River and the Cumberland River. The St. Mary’s River and the Satilla Rivers are east-west flowing, small-scale estuaries with meandering channels. The larger Cumberland River occurs landward of Cumberland Island and parallels it. These estuaries are connected with numerous tidal creeks that dissect the extensive salt marshes in the backbarrier regions. Tidal flats, point bars and creek banks are exposed at low water in and along these tidal channels.

The lower Coastal Plain of Georgia can be summarized as a series of successively older Holocene-Pleistocene barrier island complexes from east to west. Figure 2 is a map from Huddleston (1988) that shows the six Pleistocene shoreline complexes and the Holocene shoreline complex in southeast Georgia. Figure 3 is a schematic cross-section through the area of Figure 2 (from Hails and Hoyt, 1969) and shows the stairstep topography of the Georgia Coastal plain that formed as successively younger and lower barrier complexes were left stranded on the coastal plain during successive transgressive/regressive events. This figure also shows generalized geomorphic relationships between each sand-rich barrier and the intervening muddy, lagoonal-marsh deposits. Table 1 summarizes elevations above Mean Sea Level (MSL) for each barrier island complex. This field trip includes stops at deposits associated with the Holocene, Silver Bluff and Pamlico barrier systems.

Table 1. Summary of elevations above mean sea level (MSL) for each barrier island complex (modified after Hails and Hoyt, 1969)

<table>
<thead>
<tr>
<th>SHORELINE</th>
<th>EL. (+MSL)</th>
<th>AGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Holocene Barrier Complex</td>
<td>0</td>
<td>Holocene</td>
</tr>
<tr>
<td>Silver Bluff PaleobARRIER</td>
<td>1.4 m</td>
<td>Late Pleistocene</td>
</tr>
<tr>
<td>Princess Anne PaleobARRIER</td>
<td>4.0 m</td>
<td>Late Pleistocene</td>
</tr>
<tr>
<td>Pamlico PaleobARRIER</td>
<td>8.0 m</td>
<td>Pleistocene</td>
</tr>
<tr>
<td>Talbot PaleobARRIER</td>
<td>12-14 m</td>
<td>Pleistocene</td>
</tr>
<tr>
<td>Penholoway</td>
<td>21-23 m</td>
<td>Pleistocene</td>
</tr>
<tr>
<td>Wicomico PaleobARRIER</td>
<td>29-31 m</td>
<td>Early Pleistocene</td>
</tr>
</tbody>
</table>

STRATIGRAPHIC NOMENCLATURE

Huddleston (1988) recently reviewed the stratigraphic nomenclature for the Coastal Plain of Georgia. Because deposits of various ages are lithologically identical, he includes the Pamlico, Princess Anne and Silver Bluff Formations of Hails and Hoyt (1969) and Holocene deposits as part of the Satilla Formation of Veach and Stephenson (1911). The Satilla Formation is a heterogenous unit that consists of variably fossiliferous shelly sands, sands and clays of shelf, backbarrier and barrier island origin (Huddleston, 1988). Thus, Cumberland Island and its backbarrier salt marsh is underlain by the Satilla Formation. Exposures of the Pamlico paleobARRIER at Roses, Bells and Reids Bluffs are reference localities for the Satilla Formation (Huddleston, 1988).

HOLOCENE/SILVERBLUFF SHORELINE COMPLEX

The geomorphic relationships of Cumberland Island have been previously described by Martinez
Figure 1. Location map showing Cumberland Island, its backbarrier regions, and field trip stops. Inset shows position of Cumberland Island relative to the Georgia Sea Islands.
Figure 2. Map showing the relationships between six Pleistocene shoreline complexes and the Holocene shoreline complex in the outer Coastal Plain of southeast Georgia. (From McLemore et al., 1981 and Huddleston, 1988).

Figure 3. Cross-sectional schematic of Figure 2 showing the stairstep topography of the Georgia Coastal plain that formed as successively younger and lower barrier complexes were left stranded during successive trangressive/regressive events and generalized geomorphic relationships between each sand-rich barrier and the intervening muddy, lagoonal-marsh deposits (modified after Hails and Hoyt, 1969).
1 - Regional Overview

(1980) and McLemore et al. (1981). Cumberland Island is a complex barrier consisting of both Pleistocene and Holocene deposits (Fig. 2). The core of the island consists of the Pleistocene Silver Bluff paleo barrier and is associated with a sea level high stand of approximately 1.8 m above Mean Sea Level (MSL). Amino acid ratios on shell material from the Silver Bluff paleo barrier landward of Sea Island, Georgia, suggest an age of about 100,000 years for deposits associated with the Silver Bluff paleo barrier (John F. Wehmiller, personal communication, 1993). This Pleistocene core has been reactivated during the Holocene transgression so that it is now enlarged, heightened, and fringed by Holocene dunes and beach ridges, and an extensive, backbarrier salt marsh. The next landward and older Pleistocene paleo barrier is the Princess Anne shoreline complex, associated with a high stand in sea level of approximately 4.0 m above MSL (Fig. 2). 'Lagoonal-marsh' deposits of Holocene age extend between the compound Cumberland Island barrier and the Princess Anne paleo barrier. Facies attributes of Holocene depositional environments associated with barrier complexes are summarized in Frey (1973) and Howard et al. (1980). The 'lagoonal-marsh' deposits of Holocene age include salt marsh, estuarine and tidal creek deposits. These backbarrier deposits underlie a salt marsh with a surface elevation of approximately 1.5 m above MSL.

SUBSURFACE GEOLOGY AT CUMBERLAND ISLAND

During previous studies (Martinez, 1980, McLemore et al., 1981), the subsurface geology of Cumberland was determined from a series of cross-sections constructed from both well borings and and seismic lines (Fig. 4). This geology is summarized in Figure 5 which shows that the island is underlain by Pleistocene and Holocene deposits that overlie a 'basement' of middle Pliocene deposits. The following summary is taken from McLemore et al. (1981).

Deposits of Pleistocene age are predominantly composed of sand, clayey sand and sandy clay consisting of a barrier island facies as well as a backbarrier facies. The barrier island facies is characterized by fine clean sand whereas the backbarrier environment is represented by a more clayey facies. Holocene or Recent sediments include those deposited from 25,000 years ago to the present. The oceanward side of Cumberland Island began forming about 5000 years before present (B.P.) during the Holocene transgression which allowed the formation of a barrier island system somewhat seaward of its present location (Henry et al., 1973). As this Holocene barrier island system retreated landward, it became welded onto the pre-existing Pleistocene Silver Bluff barrier. Thus, Cumberland Island is a composite of Holocene and older Pleistocene barrier island complexes. A minor regression in sea level during the Holocene may have occurred between 4500 and 2400 years B.P. (Deprat and Howard, 1981).

PAMLICO PALEOBARRIER

The Pamlico paleo barrier is associated with high stand in sea level of approximately 8 m (Fig. 2). Reids, Roses and Bells Bluffs expose segments of the Pamlico paleo barrier in Florida, south of the St. Mary's River, and occur as cross-sectional profiles through this barrier complex (Fig. 1). Many previous workers (Pirkle et al., 1984; Kussel and Jones, 1986) believe that the Yulee Ridges formed as part of the Pamlico barrier complex and not some other paleo barrier for several reasons: (1) the toe of the elevation of surficial ridges is about 25 ft above present sea level and is consistant with this origin; (2) numerous occurcences of fossil burrows of the marine decapod Callianassa (Ophiomorpha nodosa) indicate a sea level of at least (8 m) above MSL; and (3) the highest elevation of highest burrows is about 25 ft (8 m) above MSL. Hails and Hoyt (1972) suggested a date of 110,000 years B.P. for Bells Bluff based on regional correlation. Recently obtained radiometric dates of wood and shells from Reids Bluff indicate a younger, late Pleistocene age, of about 25,000 to > 38,000 yrs BP, however (see Rich and Pirkle, this guidebook). Amino acid ratios on mollusks from Reids Bluff suggest that the fossiliferous unit sampled is at
Figure 4. Bore hole, seismic line, and cross-section locations, Cumberland Island, Georgia. Cross-sections are shown in Figure 5. (From McLemore et al., 1981).
Figure 5. Cross-sections through Cumberland Island show that the island is underlain by Pleistocene and Holocene deposits that overlie a 'basement' of middle Pliocene deposits. Location of sections and wells shown on Figure 4. (Modified after McLemore et al., 1981).
least 250,000 years in age (Wehmiller et al., 1988). The geology of these deposits will be discussed further in Contributions 12 through 15, this guidebook.

RECENT GEOMORPHIC HISTORY OF CUMBERLAND ISLAND

Griffin and Henry (1984) summarized the recent geomorphic history of Cumberland Island for the period 1957-1982. For the most part, the ocean side of Cumberland has experienced a net accretion in a seaward direction for period 1857/1868-1974 (Fig. 6). The north central strand as well as several small areas however have experienced net erosion for the same period. The southern tip of Cumberland as well as the northern extension of Amelia Island have accreted in a seaward direction on the ocean side for the period 1857-1982 probably in response to the construction of a pair of jetties in 1881 by the U.S. Army Corps of Engineers (Fig. 7). Jetty construction apparently reversed an earlier trend of shoreline retreat in the immediate area, but may have promoted backbarrier erosion. Shorelines in backbarrier regions landward of St. Mary's Entrance have retreated during the interval 1857-1982 causing the estuaries and tidal creeks to widen and/or migrate laterally. This topic is discussed further in Contribution 3 of this guidebook.

REFERENCES

Frey, R.W., editor, 1973, The Neogene of the Georgia Coast: Guidebook, 8th Annual Field Trip, Georgia Geological Society, Dept. Geology, University of Georgia, Athens, GA.
Frey, R.W., editor, 1973, The Neogene of the Georgia Coast: Guidebook, 8th Annual Field Trip, Georgia Geological Society, Dept. Geology, University of Georgia, Athens, GA.

1 - Regional Overview

Coast: Guidebook, 8th Annual Field Trip, Georgia Geological Society, Dept. Geology, University of Georgia, Athens, GA.


Figure 6. Little Cumberland/Cumberland Islands net Mean High Water (MHW) shoreline change 1857/68-1974 (from Griffin and Henry, 1984 and McLemore et al., 1981).
Figure 7. St. Marys Entrance MHW shoreline change, 1857-1982 (from Griffin and Henry, 1984).
Cumberland Island, the largest and southernmost of the Georgia barrier islands, supports a variety of wildlife and plant communities. Primary habitats include salt marsh, maritime forest, fresh water wetlands, and eighteen miles of pristine beach and dunes. Covering sixty percent of the island, the maritime forest is the dominant plant community of the uplands. Spanish moss draped live oaks create a dense canopy under which grows a thick under-story of palmetto and several species of shrubs (Fig. 1). Deer, squirrel, raccoons, wild turkey, armadillo, and many species of songbirds depend on the maritime forest for food and shelter.

At its eastern boundary the maritime forest abuts the dune system. In some places shifting dunes are encroaching on the forest. Live oaks growing in this transition zone are twisted and pruned by salt-laden winds blowing off the ocean. A waxy cellulose cover protects the leaves from salt spray, sun-scald, and even occasional freezes. Old leaves drop off in the spring, as new leaves emerge, so that the entire tree is never without greenery. Hence the name “live” oak. During the Colonial period and continuing into the early 20th century the live oaks of Cumberland Island's maritime forest were harvested for ship building timber. During the winter months, crews of laborers and ship-wrights from Massachusetts would take up residence in tent-camps on the island. The cut timbers were generally shipped north to naval yards for the construction and repair of America's wooden boat fleet.

Figure 1. Maritime forest with live oaks up to four feet in diameter.
Contribution 3

OCEAN SHORELINE CHANGES

Stephen V. Cofer-Shabica
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Cumberland Island National Seashore, Box 806, St. Marys, Georgia 31558

The information that follows (including text and figures) is taken from Volume I: Main Text and Appendix A of the Kings Bay Coastal and Estuarine Physical Monitoring and Evaluation Program: Coastal Studies prepared by the U.S. Army Corps of Engineers, Waterways Experiment Station, N. C. Kraus, L. T. Gorman, and J. Pope, editors, for the U.S. Department of the Navy, Office of the Chief of Naval Operations, 1993.

The magnitude and direction of shoreline position change were evaluated for Cumberland Island using six different surveys (1857/70, 1924, 1933, 1957, 1973/74, 1991). Although a number of significant changes in shoreline position are identified for local areas of the study region, the dominant trend for all time periods is relative stability to progradation for the Cumberland barrier island system (Figs. 1 and 2). The greatest amount of change occurred between 1857/70 and 1924, apparently in response to jetty construction.

Cumulative and incremental changes in position of the high-water shoreline were calculated to estimate historical trends for Cumberland Island. Table 1 illustrates the magnitude and direction of the rate of change for Cumberland Island, indicating shoreline progradation for all time periods relative to the 1857/70 survey. Net shoreline advance predominates at an average rate of 1.5 m/year for the period 1857/70 to 1991. In fact, all combinations within the change matrix for Cumberland Island (Table 1) show average coastal progradation for the 134-year time period. Only small sections of Cumberland Island exhibit retreat for any time interval, resulting in lower variability. However, because the number of points sampled was large for each shoreline (calculated at 50-m intervals), the 95 percent confidence limit was relatively small. This suggests that the computed sample mean provides a reasonable estimate of the population mean for the shoreline.

The overall trend in cumulative shoreline position change for Cumberland Island is similar throughout, although the magnitude of change varies. Minor deviations in trend are associated with change rates for Cumberland Island (except for a slight increase in shoreline progradation to 1957), although the rate of shoreline advance for the entire time period is lower than any other historical time interval. The fact that the rate of shoreline change has remained relatively constant and progradational indicates a surplus in sand supply to the subaerial beach. The southern half of Cumberland Island exhibits shoreline progradation between 1857/70 and 1924. Two areas of substantial change include the northern coast of the Cumberland Embayment and the southern prograding shoreline near the jetty. Figures 1 and 2 illustrate historical shoreline response for central Cumberland Island (the northern half of the Cumberland Embayment and the southern 4 km of the Stafford Shoal morphologic compartments) where accretion of sand on the subaerial beach caused shoreline advance for most of the area. Only the northern 2 km of coast in this area have experienced shore retreat since 1870, and this occurred after the 1957 survey. Just south of this zone of erosion is a 2-km reach of rapid shoreline progradation. Most of the change recorded for this area occurred between 1870 and 1924. However, this zone of accretion continued to prograde to the south and east between 1924 and 1991 at a slower rate. The shorelines adjacent to St. Marys Entrance show the greatest amount of change for the area. The rate of shoreline advance peaked in 1924 and 1933; however, it has decreased since this time.

Potential error estimates were calculated to gauge the significance of shoreline change measurements. In most cases, measurements of change exceeded potential errors; however, in some cases, areas showing small amounts of change over short time intervals were considered insignificant. Four primary results can be used to summarize the
findings of the historical shoreline change study. First, average long-term shoreline position change is net progradational for the Cumberland barrier island systems. In other words, the magnitude of shoreline advance exceeded retreat for the study. Second, two areas of substantial shoreline movement were identified for the area. The northern margin of the Cumberland Embayment and southern area. Between 1857/70 and 1991, Cumberland Island prograded at a net rate for 1.5 m/year. Cumberland Island at the jetty show large shoreline advances since 1857/70.

Figure 1. Change in historical shoreline position for central Cumberland Island (modified after Griffin and Henry, 1984).
Figure 2. Change in historical shoreline position near St. Marys Entrance (modified after Griffin and Henry, 1984).
Table 1
Average Shoreline Position Change for Cumberland Island

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¹ Average shoreline change rate (m/year).
² Sample standard deviation (± m/year).
³ 95 percent confidence limit (± m/year).
⁴ Length of analyzed shoreline (km).

Table 1. Average shoreline position change for Cumberland Island — 1924-1991.
INTRODUCTION

Beaches on Georgia’s Barrier Islands are dynamic systems dominated by tides with moderating effects of organisms, waves, and winds. Tidal energy is generated by an average tidal range of ≈2-3 m during semidiurnal tides (two highs and two lows each lunar day) (Thurman, 1993) with two sets of higher-range spring tides and two sets of lower-range neap tides per lunar month. The Georgia coast is characterized by low wave energy (normal summer wave heights = 0.25 m) regimes dissipated on barrier island beaches. Transfer of sand by wind is a significant process, particularly because of the wide beaches and extensive ebb deltas exposed during low tides. Major modifications of the shore are made during short but intense storms which may include occasional hurricanes, but typically include storms systems which generate intense northeasterly winds, called Nor’easters. Nor’easters periodically occur throughout the year and, particularly during the Fall and Winter. When combined with high spring tides, these storms can rapidly rework the sediment of the beach and erode a prominent scarp at the back of the beach. This may erase the physical and biologic sedimentary structures generated during intervals between storms (ie. “during normal sedimentation”).

These processes give rise to beaches (Fig. 1) which are typically 100-150 m wide with a 1°-2° seaward slope. The beach can be subdivided by the line of demarcation of the latest high tide line onto the backshore lying above high tide line, and the foreshore, lying below the high tide line. The area of the backshore can be termed the backbar and that of the foreshore, the forebar. The high tide line moves higher onto the beach with spring tides and lower with neap tides.

Lying behind the beach may be accretionary dune ridge systems, Pleistocene deposits vegetated by mixed maritime forest, or marsh meadows. Bluffs are commonly developed along high standing sections of beach, spits, or berms along low areas. Abundant washover fans are deposited in low areas by storms occurring during spring tides. Erosive areas which are forested develop prominent areas

Figure 1. Cross-section of a Georgia beach summarizing active sedimentary processes, physical sedimentary structures, and biogenic sedimentary structures.
of skeletal trees on the beach, which are called boneyards. Below low tide level the sediment is constantly being bioturbated by organisms characterized by bivalves, gastropods, crabs, and fish.

THE BACKBEACH

The backbeach typically consists of a complex of sediment deposited by the most recent storm event which leaves its signature by eroding into inactive sediment underlying the beach and forming a backbeach scarp on high areas and washover fans on low areas. Basic sedimentary structures of the storm event are horizontally laminated quartz sand interlaminated with heavy-mineral sands. A basal heavy-mineral sand 10 to 25 cm thick is often found at the back of the beach and in a 7-10 m wide band at the base of the active storm beach along the base of scarps or bluffs. Washover fans form in areas backed by marshes and often exhibit reverse density sorting resulting in a basal layer of quartz sand overlain by black heavy-mineral sand.

In the intervals between storms or high spring tides, the backbeach is modified by wind-dominated processes which result in the building of secondary dunes and wind ripples accentuated by sorting of quartz sand from the heavy minerals (Howard, Frey, and Reieck, 1972). These processes result in festoon type crossbedding, cross laminations on dune slip faces and in blowouts, and small-scale ripple cross laminations. Occasional heavy rains erode the backbeach and form a series of rills which may result in minor discontinuities (scour and fill structures) in beach laminations.

Biogenic sedimentary structures commonly seen on the backbeach are ubiquitous Ghost Crab (Ocyopode quadrata) burrows, occasional mole trails, and various trackways of foraging insects and vertebrates, including raccoons, pigs, alligators, and river otters as well as various birds and occasional deer. Seasonal visitors to the backbeach include nesting Loggerhead sea turtles and Diamondback terrapins. Loggerhead sea turtles produce crawlways, large-scale disturbed masses called body pits, small vertical urn-shaped egg chambers, and their hatchling produce small scale crawlways.

THE FOREBEACH

The Forebeach typically consists of a complex of sediment redeposited by the most recent storm event and daily sorted by wave swash on flood and ebb tides. The resulting sediments consist predominantly of quartz sand interlaminated with thin laminae of heavy minerals which all dip seaward at 1° - 2° and then, on the lower foreshore, become more level and rippled. Sand “ridges” often advance onto the foreshore from oceanward giving rise to prominent beach ridge and runnel systems. Ridge and runnel systems produce typical seaward-dipping laminated sands and shoreward-facing crossbed sets developed as the sand ridge advances onto the beach. Repeated migration of ridges up the shoreface can result in the accretion of backbeach terraces, occasional berms, and small-scale erosional scarps as the terraces subsequently are eroded by spring tide sets. Runnels, characterized by ripple marks, are usually connected to the ocean by shallow channels cutting across seaward ridges. Where beach obstructions are present, and often where they are not, complex, chaotic scour-and-fill systems may form. Margins of sounds separating barrier islands usually have steeper forebeaches and complex ripple-marked tidal flats formed at mid-tide level at corners of islands, and dominant megarippled structures on the sound side of ebb deltas.

Biogenic sedimentary structures commonly seen on the forebeach are ubiquitous Ghost Shrimp burrow openings and a plethora of trails of sand dollars and snails, and burrows of tube-secreting worms (some tubes are made of cemented sand), stomatopods, polychaete worms and a host of transient organisms which include crabs, flounder, stingrays, etc (Frey and Howard, 1972). Amphipods are important bioturbators and constantly rework beach sediment causing internal lamination to become diffuse (Howard and Elders, 1970).

Carolinian Ghost Shrimp (Callichirus major) burrows (see Bishop and Brannen, Contribution 5 of this guidebook) on open beaches are found from about mid-tide level to below low tide level,
normally reaching their most abundant density (ca. 40 burrows/square meter) at just below mid-tide level. Each burrow consists of an upper, small-diameter vertical aperture from 5 to 25 cm long which transects the active beach and connects with a larger diameter burrow shaft which is lined with fecal mud and penetrates the sediment beneath the active forebeach to depths of 3-5 m. These burrows are characteristically 1-2 cm in diameter and, being lined with fecal mud from within, develop a knobby exterior as clay is packed into the sand to stabilize it, resulting in traces which fossilize into *Ophiomorpha*.

REFERENCES
INTRODUCTION
Fossorial (burrowing) thalassinidean shrimp are among the most common macroinvertebrates in modern beach environments, but have seldom been studied because of their fossorial mode of life. In the fossil record, callianassid claws and burrows are among the most common and abundant remains of decapod crustaceans. Research on the Carolinian Ghost Shrimp (*Callichirus major* Say 1818) has been conducted by us on nearby St. Catherines Island (Fig. 1) since August 1987. This research was enlarged to include *Callianassa biformis* Biffar 1970, when it became obvious that the burrow distributions of these two species overlapped on beaches of the Georgia coast. Aspects of this research on ghost shrimp (Fig. 2) have been previously published (Bishop and Bishop, 1992). Carolinian Ghost Shrimp are particularly pertinent to this field excursion because we will be able to directly compare Recent burrows and organisms.

Figure 1. Index map of the southeastern United States, the Georgia coast, and St. Catherine’s Island, Georgia. (from Shadrue, 1990).
5 - Shrimp Ecology

Figure 2. General morphology of the Ghost Shrimp as depicted in dorsal and lateral views. (from Bishop and Bishop, 1992 after Biffar, 1971).

on Cumberland Island with Pleistocene burrows at Reid’s Bluff.

The Carolinian Ghost Shrimp (Fig. 3a) was described and named Callianassa major by Say in 1818. Stimpson (1870) established a new genus, Callichirus, with Callianassa major as its type species. Manning and Felder (1986) reviewed the usage of the systematics of this species indicating that it was considered by various authors to be a subgenus of Callianassa, a separate new genus, or a synonym of Callianassa. In a major systematic revision, Biffar (1971) reviewed its relationships to other callianassids of the southeastern United States. Manning and Felder (1986) reassigned it to Callichirus. At this time no ghost shrimp assignable to Callianassa is found on the Georgia coast. Aspects of the ecology of the Carolinian Ghost Shrimp have been described by Lutz (1937), Pearse, Humm, and Wharton (1942), Pohl (1946), Pryor (1975), Rodrigues (1976, 1983, 1984, 1986a, 1986b), Rodrigues et al. (1986), (Williams 1984), Frey, Howard, and Pryor (1978), and Bishop and Bishop (1992). The Carolinian Ghost Shrimp was reported to reach densities of 1,780 animals per acre on South Carolina beaches (Pohl, 1946:75). Calculations for South Beach, St. Catherines Island indicate an abundance of 400 Ghost Shrimp per linear meter of beach. Carolinian Ghost Shrimp occur from Beaufort Inlet, North Carolina to Santa Catarina, Brazil (Rodrigues, 1983).

The Georgian Ghost Shrimp (Fig. 3f), Callianassa biformis, described by Biffar (1970) from the south end of Sapelo Island, McIntosh County, Georgia, is known to range from South Carolina to Georgia but probably has a more extensive range. This diminutive ghost shrimp, usually less than a centimeter in length, is commonly found from mid-tide to low tide level in protected areas behind sand bars associated with sound channels where it burrows shallowly into ripple-marked, muddy, sand flats and contributes to the fluidity of the upper 30 cm of these tidal flats. Populations have been observed on the south end of Sapelo Island (type locality), on the south end of St. Simon’s Island, on the north end of St. Catherines Island (this study), on the northern part of the McQueen Inlet ebb delta on St. Catherine’s Island, and at the mouth of Big Bay Creek, Edisto Island,
Figure 3. Ghost Shrimp from St. Catherines Island, Georgia. A) Dorsal view of a male Carolinian Ghost Shrimp (*Callichirus major*) released in beach runnel; B) Carolinian Ghost Shrimp burrow mouth on North Beach showing a flood of sand being pumped onto the beach by a ghost shrimp; C) Four active burrow mouths on North Beach showing open burrow mouths, sand mounds, and a ring of fecal pellets of Carolinian Ghost Shrimp; D) Burrow mouth of Carolinian Ghost Shrimp burrow showing sand mound capped by a ring of rod-shaped fecal pellets; E) Collapsed burrow mouth of Carolinian Ghost Shrimp burrow indicating diameter of major shaft; F) Lateral view of Georgian Ghost Shrimp (*Callianassa biformis*) above a Carolinian Ghost Shrimp to illustrate the difference in size of adults of these two species; G) Active burrows of the Georgian Ghost Shrimp on Wilburg Creek Shoal showing great density, tiny size, and abundant fecal pellets (hoop diameter = 36 cm); and H) Active burrows of Georgian Ghost Shrimp (*Callianassa biformis*) in fossoria in an aquarium. Scales indicated in most photographs. (from Bishop and Bishop, 1992).
5 - Shrimp Ecology

NATURAL HISTORY OF GHOST SHRIMP

The Animals: The Carolinian Ghost Shrimp is a medium-sized (30-100 mm) decapod crustacean adapted to its burrowing mode of life (Fig. 2) by being elongate, dorsally-ventrally flattened, cylindrical, and lightly mineralized (Fig. 3a). The cuticle is thin and flexible over most of the shrimp's body, consisting of a short cephalothorax, appendages, and an elongate abdomen. The abdomen consists of two very soft and flexible segments with very reduced pleopods used for reproduction, followed by three more highly mineralized segments with well developed diaphram-like pleopods, in turn followed by the hardened 6th segment and telson with attached uropods forming a tail fan.

The Carolinian Ghost Shrimp spends most of its life burrowed into the sand of beaches and sand flats (Fig. 4) fronting the open ocean or sounds. The presence of Carolinian Ghost Shrimp on a beach is indicated by burrow mouths (Fig. 3c), often marked by water spouts or sand floods (Fig. 3b) which accumulate as volcano-like mounds (Fig. 3c) ringed by extruded rod-shaped black fecal pellets composed of clay (Fig. 3d). Few observations have been made of the animal outside its burrow (Pohl, 1946:78; G. Bishop in 1988 and 1993). The animal apparently feeds on detritus and particulates in mud-laden water pumped through its burrow. It periodically extrudes masses of rod-shaped, cylindrical, channeled fecal pellets consisting largely of mud containing 3-10% organic carbon (Pryor, 1975:1246). The production of fecal pellets is copious and often results in accumulation of layers of mud in environments otherwise characterized by sand-sized quartz (Howard and Reineck, 1972:92; Pryor, 1975).

Carolinian Ghost Shrimp reproduce sexually. The mechanism of sperm transfer is not known. Visible egg masses develop internally and are extruded throughout the summer. The eggs hatch and the larvae enter a planktonic existence as is

Figure 4. Photograph of Callichirus major burrow strand looking north at low tide along North Beach at Picnic Bluff, St. Catherine's Island. Ten cm scale in foreground. (From Bishop and Bishop, 1992).
indicated by their abundance in plankton samples. Larval settlement into the substrate is an unknown process.

The Georgian Ghost Shrimp (Callianassa biformis Bifar 1970) resembles the Carolinian Ghost Shrimp but is differentiated by morphologic details and by being an order of magnitude smaller (Fig. 3f). Ecologically the Georgian Ghost shrimp tends to inhabit lower intertidal muddy sand flats in sheltered areas associated with abundant acorn worms, mud snails, hermit crabs, and other invertebrates. Burrows of the Georgian Ghost Shrimp are smaller, shallower, and found in densities of nearly 500 burrows per square meter on St. Catherine's Island (Fig. 3g-h).

The distributions of Carolinian and Georgian Ghost Shrimp burrows overlap but appear to be of different depths, effectively partitioning them from one another.

**The Burrows** Carolinian Ghost Shrimp burrows on the beach (Fig. 5) consist of three major parts: (1) a vertical upper constricted burrow aperture, (2) a nearly vertical main shaft of the burrow, and (3) one or more horizontal burrow mazes. The constricted burrow aperture is vertical, about 5 mm in diameter, usually 15-20 cm in length, and opens into the main burrow shaft which is vertical to inclined, several meters in length, and about 1-2 cm in diameter (Fig. 6) with periodic enlarged turnarounds and numerous horizontal and/or inclined branches. The lower part of the burrow is thought to consist of anastomising, interconnected, horizontal tunnels. Rodrigues (1984) described correlation between the size of the burrow and the size of the animal inhabiting it. Felder (personal communication) has correlated the number of surficial burrow apertures with the number of burrow shafts to establish a closer estimate of the true population density. Ghost Shrimp probably spend most of their time in the lower part of the burrow, occasionally rising in the shafts to eliminate sand or waste from the burrow. In unconsolidated sand the Carolinian Ghost Shrimp must line its burrow with a mucal-mud binding agent, packing mucus-laden fecal pellets into the burrow walls, thereby building a substantial burrow wall and imprinting a knobby texture onto the exterior of the burrow. When lithified, these burrows are assigned to the ichnogenus Ophiomorpha. Burrow morphology varies dramatically in different sediment types as well as interspecifically (Frey, Howard, and Pryor, 1978). Many burrow morphologies are constructed by a single species and similar burrow morphologies are constructed by different species. Burrows occasionally collapse, forming chevron-like sedimentary structures in the surrounding sediment (Frey, Howard, and Pryor, 1978). When erosional conditions scour a beach containing Carolinian Ghost Shrimp burrows, the burrows may be eroded to a level where the beach surface intersects the main shaft of the burrows. When this occurs the main shaft is plugged and abandoned or a small, inclined shunt is constructed to the beach surface (Frey, Howard, and Pryor, 1978).

The burrows of different ghost shrimp species exhibit a wide range of morphologies ranging from solitary burrows to communal burrow systems (Frey et al., 1978), necessitating great caution when interpreting Holocene and ancient data on the ecology of ghost shrimp (e.g., the interpretation of the ecology of Callichirus major, if based on what is known about the ecology of Callianassa californiensis, would lead to largely erroneous conclusions.) Erickson and Sanders (1991) described Pleistocene burrows of the ghost shrimp, Callichirus major.

Carolinian Ghost Shrimp burrows in analogous Pleistocene ancient barrier island sediments (Howard and Scott, 1983:176) indicate the burrows extend 3 to 5 m into the beach. Ancient burrows of analogous, but not necessarily conspecific ghost shrimp from more ancient sediments (Fig. 7), preserve similar morphologies and are assignable to the ichnogenus Ophiomorpha but should be interpreted with extreme caution. The presence of Ophiomorpha usually indicates the presence of a thalassinoid burrowing shrimp but does not necessarily indicate a nearshore environment because many thalassinoids range to the edge of the continental shelf. To emphasize this point, one should consider the distribution of Callianassa
Figure 5. Hypothetical burrow morphology of Carolinian Ghost Shrimp seen in side view.
subterranea (Montagu) in the North Sea (Witbaard and Duineveld, 1989) indicating the maximum density of this thalassinooid increases toward the center of the North Sea and decreases to total absence along the shores of the North Sea. Bird (1982) pointed out that over half of the known ghost shrimp did not inhabit shallow waters.

**Burrow Distribution** The density and distribution of burrows of the Carolinian Ghost Shrimp have been described by Pohl (1946), Pryor (1975), Eric Bishop (1988), and Bishop and Bishop (1992). Pohl (1946:74) found the maximum burrow density to lie just above mid-tide level “...in the quarter of the beach between the half-way and three-fourths mark...” Pryor (1975) cited Ghost Shrimp burrow mouth densities of up to 450 burrows per square meter in tidal pools and protected shores of the Mississippi Sound (although these probably belonged to a smaller ghost shrimp such as Callianassa biformis). Eric Bishop (1988) surveyed and contoured the distribution of Carolinian Ghost Shrimp burrow mouths on beaches at Tybee Island, St. Catherines Island, and St. Simon’s Island, concluding that maximum burrow mouth development occurs on stable beaches just above mid-tide level. Burrow density increases rapidly from 0 bp/m² to a maximum just above the mid-tide level then decreases slowly to low densities along
the low water level. Bishop and Bishop (1992) found that highest density follows the beach trend just above mid-tide level, but changes with physical conditions such as revetments, beach slope, steep sound channel walls, or presence of relict marsh sediment under the beach surface. The density of burrows varies along the beach, building into local density maxima separated by minima, but burrows are always present except where excluded by sedimentological factors cited above. On disturbed beaches (renourished beaches and ebb deltas) burrow density generally decreases dramatically, the distribution pattern broadens and becomes more dispersed. On normal, open oceanic beaches the burrow distribution occupies a continuous narrow zone about 50 meters wide along the beach from below low water level to just above mid-tide level. When contoured and plotted on a true-scale map this pattern is string-like, hence it can be referred to as the “burrow strand”.

The subtidal occurrence of the Carolinian and Georgian Ghost Shrimp have not been investigated in this project because turbid water conditions prohibit subtidal sampling. Reports from the literature (Dörjes, 1972; Frey, Howard, and Pryor, 1978) indicate that the Carolinian Ghost Shrimp occurs subtidally to a depth of a few meters on the forebeach and on shallow shoals. Monitoring of burrow mouth density on St. Catherines and St. Simon’s Islands over a two year period has demonstrated the stability of the density distribution patterns.

Burrows of the Georgian Ghost Shrimp (Callichirus biformis) are much smaller, typically ringed by muddy sand mixed with fecal pellets, and have smaller fecal pellets (Fig. 3g). Their depth (based on success in extracting these animals from shallow depths with a Yabby Pump and observation in an aquarium (Fig. 3h)) appears to be much less than that of Callichirus major. Burrow density data from the north end of St. Catherine’s Island indicate the density increases from 0 bpm² just above mid-tide line to 483 bpm² just above the low tide line with a lateral distribution of abundant Georgian Ghost Shrimp along the shore for 700 meters in an elongate hill-like pattern. This pattern of distribution appears to be much more dependent upon the presence of mud in the substrate than upon any other physical parameter.

Burrow Morphology The morphology of the upper part of burrows of the Carolinian Ghost Shrimp has been examined by trenching at dead low tide (Fig. 8). A square meter has been laid off on the beach and excavated using archeological-like technique. Each burrow opening is mapped on the surface and its depth determined by probing. After excavation through the 15-20 cm of active beach, mud-lined main burrow shafts are encountered. Virtually every burrow examined has had an significant offset at the bottom of the active beach at the juncture of the burrow aperture and the main burrow shaft. Trenching demonstrates that the number of burrow mouth openings on the beach surface is the same number of burrow shafts present beneath the surface. This technique allows direct comparison of active Recent burrows with ancient burrows. To illustrate this point, Pleistocene burrows exposed at the Colleton Plantation, on the Colleton River, South Carolina (Erickson and Sanders, 1991) were exposed by scraping a horizontal surface to provide an analogous comparison between Recent and fossil burrows attributable to the Carolinian Ghost Shrimp. The excellent exposures at Bell’s and Reid’s Bluffs will provide analogous vertical cross sections of burrows (Fig. 9).

CONCLUSIONS

The distributions of the burrows of the Carolinian and Georgian Ghost Shrimp on St. Catherines island are characteristically parallel to the strand-line. Their distributions are related to the presence of available preferred substrate and marine exposure. The burrow strand occupies the area from the midbeach nearly to the low water line and is variable in burrow density along strike.

The following conclusions can be made regarding of ghost shrimp in coastal Georgia:

1. The Carolinian Ghost Shrimp (Callichirus major) is one of the most abundant and nearly ubiquitous macroorganism on sandy, oceanic Georgia beaches.

2. Burrows of the Carolinian Ghost Shrimp (C.
Figure 8. Upper portion of main shafts of Recent burrows of the Carolinian Ghost Shrimp exposed in a one meter square trench dug to approximately 25 cm depth on South Beach, St. Catherine's Island, Georgia. Scale = 10 cm.

Figure 9. Fossilized ghost shrimp burrows, assignable to the ichnogenus *Ophiomorpha*, exposed in Reids Bluff. Notice the vertical shaft exterior parallel to the left of scale (10 cm on left) and transverse cross section showing mud lining analogous to those of Figure 8.

*major* are distributed and reach their maximum density as a narrow band, termed the "burrow strand," parallel to the mid-tide line on open ocean sand beaches.

3. Protected muddy, rippled sand substrates at entrances to sounds are inhabited by abundant Georgian Ghost Shrimp (*Callianassa biformis*) whose presence is indicated by abundant, small, active burrows.

4. The geographic distribution of the Carolinian Ghost Shrimp (*C. major*) and the Georgian Ghost Shrimp (*C. biformis*) overlap spatially but are partitioned vertically.

5. Carolinian Ghost Shrimp burrows consist of of a small, unlined, vertical aperture, the mud-lined nearly vertical main shaft with turn-around chambers and possible side branches, and undescribed burrow mazes at depths of 3-5 m.

6. The Carolinian Ghost Shrimp burrow is constructed by impressing fecal pellets into
unconsolidated sand leading to a knobby exterior that would be preserved as a trace fossil known as *Ophiomorpha*.

7. Burrows of the Recent ghost shrimp *Callichirus major* and those found exposed along the St. Mary’s River at Reid’s and Bell’s Bluffs are similar in structure and association with sequences of sedimentary structures leading to the conclusion that they were both made by *C. major* in a nearshore environment.

8. Not all *Ophiomorpha* are indicators of nearshore environments.

ACKNOWLEDGEMENTS

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REFERENCES


Contribution 6
NESTING TRACES OF THE LOGGERHEAD SEA TURTLE
(Caretta caretta (Linne)), ST. CATHERINES ISLAND, GEORGIA:
IMPLICATIONS FOR THE FOSSIL RECORD
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ABSTRACT
Although sea turtles have an extensive geological record extending at least into the Early Cretaceous, traces of their terrestrial nesting activities have not been documented in the literature. Studies of nesting of loggerhead sea turtles on St. Catherine's Island, Georgia, have demonstrated that several types of traces are made by nesting recent sea turtles, and these are potentially represented by trace fossils in ancient near-shore marine sedimentary rocks. These traces include large crawlways produced by egress of mature female turtles from the sea crawling across the beach to nest, small scale crawlways made by hatchlings as they emerge from the nest and scamper to the sea, and distinctive disrupted sediments of the nest. Nests consist of superjacent disrupted layering having a broad, shallow depression overlying a small, vertical-walled nesting chamber. Nesting chambers of loggerhead turtles tend to be about 20 cm in diameter and about 25 cm deep. Erosional events may truncate or obliterate the sediments containing the nests, making them easy to overlook or misinterpret.

INTRODUCTION
Modern sea turtles (Bjorndal, 1979) spend virtually their entire life in the sea, except for the laying and development of eggs into hatchlings. Sea turtles characteristically lay their eggs in nests excavated into the backbeach or sand dunes behind sandy beaches (Caldwell, Carr, and Ogren, 1959). Nesting behavior is capable of leaving several sedimentological imprints on the nearshore sedimentary record including nesting crawlways, nest excavations which disrupt "normal" sedimentary structures, and hatchling crawlways made as the hatchlings make their way back to the sea.

Sea turtles are known to have existed for approximately the last 100 million years (Weems, 1988), from the Early Cretaceous to Holocene. Marine turtles are significantly modified to live an active nektonic life on the open ocean, including a significant increase in body size, reduction of carapace and plastron armoring, and modification of appendages into oar-like flippers. While graceful in the marine habitat, sea turtles are ill adapted for terrestrial activities such as egg laying because their very size and highly modified appendages are so inefficient for moving on land. The paleontological record of sea turtles is represented by body fossils, which are more or less fragmentary. Modern sea turtles utilize sandy beaches in the tropical and subtropical regions for their nesting. Evidence now indicates that individual sea turtles may not nest every year, but when they do nest, they commonly deposit multiple clutches of eggs throughout the nesting season. The specifics of nesting behavior vary from species to species, but are similar in all sea turtles. The annual nesting of sea turtles on sandy beaches provides ample opportunity for the preservation of nesting crawlways, turtle nests, and hatchling crawlways to be incorporated into the stratigraphic record. Evidence of these trace fossils made by sea turtles is lacking in the literature.

The purpose of this note is to briefly describe traces produced by sea turtles during nesting activities, hoping that this note will provide a model for description of sea turtle trace fossils from Mesozoic and Cenozoic sedimentary rocks.

SEA TURTLE NESTING TRACES
Nesting by sea turtles involves a sequence of events which include, but may not be confined to: (1) locating a nesting site, (2) construction of the

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nest by digging, (3) laying a clutch of eggs, (4) covering the nest, (5) returning to the sea, and (6) hatching and emergence of the hatchlings (Caldwell, Carr, and Ogren, 1959). Traces of these behaviors are thus of two general types: (1) trackways made as the female or hatchlings cross the beach and (2) disruption of back-beach or dune stratigraphy by digging and backfilling of the nest. In spite of the low preservation potential of this environment, the long geologic history of sea turtles and the intensity of annual nesting should result in preservable traces in the fossil record.

Typical nesting history involves selection of a nesting site, egress of the turtle from the sea onto the beach, selection of a suitable nest site located at the back beach, the digging of an egg chamber, covering of the nest, and re-entry of the ocean (Hailman and Elowson, 1992). The nest is constructed by digging a broad depression (the body pit) in the loose surficial sand which allows the turtle to then dig her nest into the underlying damp sand to a proper depth to insure incubation. The depth of the egg chamber is controlled by the length of the turtles rear paddles, therefore the nest depth is controlled by the length of rear paddles and body pit depth (Carr, 1967).

Frey and Pemberton (1987) briefly described and figured a sea turtle crawlway made during nesting on a Georgia beach (Frey and Pemberton, 1987: Fig. 17). Nesting sea turtles emerge primarily at night, apparently to escape desiccation and predation. They somehow select a nesting beach, sometimes exhibiting great fidelity, then crawl across the exposed beach to a nesting site which is typically on the extreme back beach or on the seaward slope of the first dune ridges. The crawlway is typically a linear path of medial drag marks made by the turtle’s plastron bordered by lateral tracks made by the flippers pushing backward against the beach surface. Nesting sea turtles often crawl onto the beach without completing a nest, producing a trace called a “false crawl,” a misnomer because the crawlway is real but the nesting is nonexistent. Crawling activity may indicate an aborted nesting attempt or could be an attempt to distract predators by laying lines of misinformation on the beach.

Nesting crawlways often indicate an extended search by the female for a nest site as she crawls about on the back part of the beach before depositing her clutch. A second crawlway is constructed as the turtle returns to the sea. Similar crawlways are made by emergent sea turtle hatchlings as they scamper en masse from the nest site toward the sea. The extent of these crawlways is dictated by the width of the beach which in turn is controlled by the height and stage of the tide at a given point. Nesting and hatchling crawlways are extremely ephemeral traces, being nearly obliterated by subsequent high tides. Hatchling crawlways are similar to nesting crawlways, differing from them by being much smaller and by being very abundant, often so abundant that they cannot be differentiated one from another.

OBSERVATIONS FROM ST. CATHERINES ISLAND, GEORGIA

Monitoring nesting behavior of the loggerhead sea turtles nesting on the beaches of St Catherines Island, Georgia, between 1990 and 1993 has allowed us to observe approximately 250 nests and 500 crawlways made by Loggerhead sea turtles involved in nesting activity. A description of these traces on St. Catherines Island presents a matrix of recent sea turtle nesting traces which might lead to the description of similar fossil traces. We would point out that the observations are still limited to a few nests laid by one species on one barrier island, and ought to be interpreted using great caution and not be overextended in application to the interpretation of nesting by all sea turtles, fossil or living.

Crawlways Crawlways of nesting females (Frey and Pemberton, 1987: 342, Fig. 17) consist of large, linear plastron drags paralleled by rows of paddle prints made by the sea turtle’s flippers as they push and pull the large animal across the beach toward back-beach nesting habitat (Fig. 1). These traces are the size of mature loggerhead turtles, approximately 60-110 cm in width, extending from waters edge at the time of ocean emergence onto the back-beach. Nesting crawlways are very ephemeral traces, especially those lying below the high tide line which are easily obliterated or nearly
Figure 1. Crawlway of female loggerhead sea turtle attempting to nest on St. Catherines Island. Flipper marks laterally and plastron drag medially. The direction of crawling is indicated by V’s made by front flipper claws (opening in direction of crawling, toward top of picture) and push marks made by rear flippers. (Scale = 10 cm).

obiterated by subsequent high tides.

Crawlways of hatchlings (Fig. 2) mirror the morphology of those of gravid females differing from them by being much smaller, by generally being unidirectional, by occurring in more or less large groups rather than as individual crawlways, and by originating from emergence craters. The crawlways are linear trails or drags made by the plastron paralleled by two pairs of “paddle prints” on either side of the plastron drag. The patterns made by the hatchlings in their scamper for the sea are largely parallel, but have overtones of a braided, anastomising pattern made as the hatchlings cross one another’s crawlways. When the number of hatchlings gets large (perhaps above 20 or 30) the pattern of individual crawls becomes obliterated by interweaving of their crawlways.

Nests  Loggerhead nests (Fig. 3) typically consist of the entrance and exit crawlways which lead to and from the nest itself. The nest consists of a a circular area of disturbed or bioturbated sand with a swirled appearance on the surface of the nest where the turtle has engaged in rotating during her covering activity after nesting. This area, underlain by the body pit, is excavated by the turtle to about 20 cm prior to digging the egg chamber and subsequently backfilled during covering. Its surface is typically swirled and of variable relief. The relief of the nest surface is rapidly modified by wind and rain until its presence is virtually undetectable on the beach surface within hours or days.

Loggerhead nests have been excavated by trenching on St. Catherines Island, Georgia. Using a flat sand shovel and trowel, we excavate a trench in front of nests, level the back face into a vertical wall and shave it back until we encounter the egg chamber of the nest. The wall is then shaved back to approximately the center of the nest from both sides and carried into the egg mass in an attempt to
6 - Sea Turtles

Figure 2. Emergence craters and crawlways of Loggerhead sea turtle hatchlings made as they exited from nest 93-44 and scampered to the sea (toward top of photograph). (Scale = 10 cm).

Figure 3. Nest and crawlway of Loggerhead sea turtle deposited on South Beach, St. Catherines Island. Nest 93-33 showing entrance crawlway to right (south) and longer exit crawlway to left (north) indicating the turtle spent considerable time on the beach. Body pit in foreground. (Scale 1.5m).

thick, basal heavy-mineral layer which is horizontally laminated having sparse layers of quartz sand and may be cut by quartz sand-filled Ghost Crab burrows. Overlying the basal heavy-mineral layer normally is 20-40 cm of quartz sand horizontally interlaminated with sparse layers of heavy minerals. The upper layer usually consists of approximately 10-30 cm of quartz sand interlaminated with festoon crossbeds marked by laminations of heavy minerals.

A loggerhead nest is approximately 50 cm deep, reaching only a few centimeters into the basal
heavy-mineral bed or into the inactive sediment underlying the beach. The upper half of the nest is a broad nesting depression having diffuse, disjunct, gently-dipping boundaries. The lower half of the nest is a bulbous, cylindrical egg chamber about 18-23 cm in maximum diameter, with near vertical to undercut, sharply-defined boundaries. The eggs are deposited in the urn-shaped egg chamber, usually restricted to the lower 20-30 centimeters of the egg chamber. Most of the egg chamber is occupied by eggs prior to hatching and subsequently collapses upon emergence of hatchlings.

Typical nest morphology is interpreted to consist of a nesting depression (the body pit) dug by the turtle in loose surficial sand (upper part having gentle, poorly defined sides) until she reaches wet sand into which she can dig an egg chamber (lower part having vertical walls and egg mass) with her rear flippers. The nest is subsequently backfilled by the female with homogenized surface and excavation sand, then further disrupted as the hatchlings worked their way upward through the

Figure 4. Vertical cross section of loggerhead sea turtle nest dug parallel to shoreline showing beach stratigraphy disrupted by nesting depression and egg chamber. (Scale 10 cm).

Figure 5. Horizontal cross section of egg chamber neck of Loggerhead sea turtle egg chamber showing disruption of interlaminated back beach heavy mineral layers by mottled backfilling of quartz-rich sand in egg chamber neck. (Scale = 10 cm).
sand plug during their emergence. Both processes lead to vertical cylindrical sedimentary structures approximately 20 cm in diameter and 20 - 30 cm high which dramatically cut across back beach horizontally laminated sedimentary structures and are characterized by being filled with mottled (bioturbated) sand that stands in stark contrast to the surrounding laminated sands (Fig. 5).

The sequencing of post-depositional erosional events are important in producing a suite of final sedimentary structures by a nesting event. A typical nest could be modified significantly by erosion and/or could be truncated by beach erosion. Deposits of interlaminated quartz and heavy-mineral, wave-deposited sand could be deposited across the body pit or the egg chamber or it could be covered by prograding dunes, giving rise to the possible preservation of more or less complete back beach sequences. These scenarios could lead to very small sedimentary structure which might be easily overlooked or misinterpreted in sedimentary rocks.

Nearly all nests deposited in a given year are located along the backshore either at the base of spring tide-storm scars, on the front face of primary dunes, or just over the top of backbeach berms or washovers, all within a few meters of the backshore boundary.

CONCLUSIONS

The nesting behavior of loggerhead sea turtles produces discrete traces which consist of adult crawlways, hatchling crawlways, and nests. These sedimentary structures ought to have the same stratigraphic range as body fossils of sea turtles. The absence of cited sea turtle nesting trace fossils may indicate they have a low preservation potential or, more likely, that they are currently unrecognized by geologists. The total spectrum of nesting structures remains undescribed and may not be typified by the structures described in this paper. Especially pertinent to study are beaches which have intensive nesting, perhaps with so many nests as to produce intersecting nests, which could produce a disrupted layer along the backbeach for several kilometers. The following conclusions can be drawn:

- Nesting sea turtles leave a suite of distinctive traces on the beaches used for nesting, consisting of ephemeral crawlways and more permanent nesting structures.
- Nesting females leave large crawlways on the beach as they unsuccessfully (false crawls) or successfully nest on the back beach.
- Sea turtle nests consist of a body pit which forms a broad depression in the drier sand and a narrow, vertical-walled, cylindrical egg chamber.
- Sea turtle nests form discordant sedimentary structures which are characterized by a homogenized, bioturbated texture cutting across and downward into back beach structures characteristic of near-shore dunes and back-beach sediments.
- Sea turtle nests may be preserved intact or attenuated by erosion into very small structures.
- Sea turtle hatchlings leave unidirectional, small scale, subparallel crawlways on the beach when they scamp for the sea.
- Crawlways would be difficult to recognize in vertical exposures while nesting cavities should be easily recognizable.

ACKNOWLEDGMENTS

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Landward of the beach are a series of low dunes that comprise the fore dune zone. Continuing eastward, one enters an inter-dune meadow consisting initially of a marsh-shrub complex and then a higher and dryer swale. This area faces Cumberland Island’s back dune ridge that extends, unbroken, from the south end of the island, northward for over 24 km. Except for one 3 km section in the Lake Whitney area where the active beach fronts the back dune ridge, a fore dune complex and an inter-dune meadow separate the back dune ridge from the beach. This landform is composed of crossbedded, fine-grained quartz sand with dark heavy minerals. The dunes are formed from the prevailing easterly winds and thus migrate in a westward direction (McLemore, et al., 1981).

The back dune ridge at the Dungeness Dune Crossing is the subject of this stop. Anecdotal evidence suggests that at one time, as late as the 1960’s, this ridge was over 60 feet high (Smith, 1983) in this area. During a research trip to Cumberland Island in August 1978, the elevation of the crest of the dune ridge was measured (Shabica and Kirkland, unpublished field notes). The highest point was selected by visual inspection from Beach Field west of the ridge. The stadia rod was placed at that point, approximately 75 m north of the road bed. Using a K & E transit, the elevation, above Beach Field ground level was measured as 7.91 m. On 22 September 1993 the crest of the dune ridge was again measured and found to be 3.86 m above Beach Field ground level (Cofe-Shabica and Zoodsma, unpublished field notes). The Beach Field dune toe of the Dungeness dune ridge was levelled on 8 October 1993 from a USGS/NPS well site (near the historic Dairy Manager’s quarters) and was found to be +4.86 m NGVD-29 (Cofe-Shabica and Fries, unpublished field notes). If one assumes that the elevation of the field west of the dune ridge has not changed greatly since 1978 that would put the high point of the dune ridge at +12.8 m (42.2 ft) NGVD-29 in 1978.

Between August 1978 and September 1993, the crest of the dune ridge decreased from 7.91 m to 3.86 m. Thus a total reduction of 4.05 m occurred over 15 yrs or 0.27 m of erosion per year. If one extends the data back to 1965 (the “1960’s” of Smith, 1983) it is possible that the crest of the dune was 16.3 m (53.8 ft) NGVD-29.

What has occurred in this back dune ridge complex that has resulted in a documented reduction in elevation of over four meters since 1978 and where has the sand gone? Simply stated, erosion will occur when the sand supply to an area is less that the sand being removed from the area. It is clear that the sand supply to the back dune complex at the Dungeness Dune Crossing has changed (there is a deficit) resulting in a decrease in the height of the dune crest as sand is reworked. The reduction in sand supply may be related to the maturation of the interdune meadow and the fore dune complexes eastward of the ridge. This may have resulted in more sand being trapped in the fore dune and interdune areas as sand is transported east to west from the beach. From the previous stop it is clear that there is sufficient sand supply to the beach (the beach is accreting) so system sand starvation is not a factor here. Or as we shall see below, perhaps the sand supply has not changed, only the dynamics of the area.

Figure 1 (from McLemore et al., 1981) was taken around 1980. In this photograph it appears that the palm is approximately 1 to 2 m from the leading edge of the dune. The photograph in Figure 2 was taken on 8 October 1993 from approximately the same horizontal location as that in Figure 1. The distance from the palm to the shrub complex (Fig. 3) measured 15.5 m, and 24.2 to the toe of the dune in Beach Field (Fig. 3). The influence of vegetation on dune migration is evident here. It is
Figure 1. Leading edge of back dune at Dungeness Crossing (ca. 1980). Note location of palm. (From McLemore et al., 1981).

Figure 2. Photograph from same location, 8 October 1993. Note Palm.
clear that "a lot" of sand is still in the system and that the dune is migrating into Beach Field at a rate of 1.2 to 1.9 m per year. The resulting landform that we see at the Dungeness Dune Crossing is a function of, among others, the sand supply, physical forces (wind, waves, and rain) and the vegetation over which a dune is migrating. The landform is not static and its morphology and rate of migration are ever changing.

REFERENCES
The Beach Creek salt marsh consists of both high and low marsh habitats. The low marshes, characterized by the marsh grass *Spartina alterniflora*, are topographically lower and commonly have a clay and silt substrate. As the name implies, high marshes are topographically higher, characterized by the marsh grass *Spartina patens* and the black rush *Juncus sp.*, and are composed of a fine- to medium-grained, argillaceous sand substrate (McLemore, et al., 1981). Like other Georgia coastal marshes, Beach Creek is built on sands and muds reworked from Pleistocene coastal outcrops, including those from the core of Cumberland Island (Howard and Frey, 1985), and Holocene beach and dune sands. These features are a rich source of sediments that have been largely retained within individual barrier/shoal complexes (BSSC), rather than exported offshore or downcoast (DePratter and Howard, 1977). The St. Marys and Satilla Rivers carry very little sediment to the coast, and localized zones of land-building and land loss occur throughout the Cumberland BSSC (Griffin, 1982; Griffin and Henry, 1983). The total areas of marsh and mudflat within this system have, at least until recently, remained relatively constant throughout historical time, despite eustatic sea level rise (Granant, 1990; Letzsch and Frey, 1980). Cumberland Island National Seashore encompasses 10,211 ha of which 3,440 ha or 33.2% are salt marsh (Cofer-Shabica, unpublished file report). The stability of this region reflects a balance between sediment supply and demand within the BSSC system that has been described by Brunn (1978) as one of “dynamic equilibrium”. An accreting marsh will generate underlying layers of material suitable for peat formation. Such peat deposits are visible along creek meanders in the Beach Creek area (Fig. 1). Studies conducted since 1989 suggest that Beach Creek marshes are accreting on the order of 1-4 mm per year (Woodward-Clyde Consultants, 1993).

Figure 2, a photograph taken from the back barrier dune ridge along the eastern margin of Beach Creek shows the marsh and the Raccoon Keys. This active dune is migrating over, and burying Beach Creek marsh, peat and oyster deposits (Figure 3). A photopoint marker was established at the base of this dune on 11 June 1983 (Rhyne, unpublished field notes). The distance between the toe of the dune and the reference marker was 27.8 m at that time. On 13 October

Figure 1. Peat layer below marsh surface. Note oyster reef.
Figure 2. Looking west over Beach Creek and Raccoon Keys from crest of back-barrier dune.

Figure 3. Westward migrating dune (same dune as Fig. 2).

1993, the distance was found to be 24.9 m (Cofer-Shabica, unpublished field notes). Thus, in a little over ten years, the dune has migrated 2.9 m (0.3 m/yr) into Beach Creek marsh.

REFERENCES
Contribution 9
GEOMORPHOLOGY, FACIES SEQUENCES AND PROCESSES
IN BACKBARRIER SALT MARSHES, CUMBERLAND ISLAND, GEORGIA
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OBJECTIVES
Backbarrier regions in coastal Georgia consist of large estuaries bordered by extensive salt marshes that are dissected by meandering tidal channels. This salt marsh-estuary system is physiographically analogous to the meandering channel, levee, splay and floodplain deposits of fluviatile systems (Edwards and Frey, 1977; Howard and Frey, 1985). Tidal channels in Georgia salt marshes are known to migrate laterally (Ragotzkie, 1959; Letsch and Frey, 1980) and it has been proposed that this process produces fining-upward sedimentary sequences comparable in thickness to channel depth (Land and Hoyt, 1966; Barwis, 1978; Howard and Frey, 1985). Ignoring the effects of sea level rise, Howard and Frey (1985) also proposed that the process of lateral channel migration is capable of rapidly producing broadly distributed sand sheets enclosed in impermeable facies landward of a sand barrier. This paper examines the geomorphology of these backbarrier regions, the facies attributes of the mud-rich sequences preserved beneath the Holocene salt marsh, and the processes that produce these sequences.

PREVIOUS WORK
The surficial distribution and characteristics of facies across intertidal, sandy point bars (Land and Hoyt, 1966; Pryor, 1967; Pierce, 1976; Howard, Elders and Heinbokel, 1975; Barwis, 1978; and Mercer, 1984), tidal flats and inlet shoals (Greer, 1975), and the salt marsh, creek banks and estuary channels (Frey and Howard, 1969; Edwards and Frey, 1977; Letsch and Frey, 1980, 1980a; Howard and Frey, 1985; Frey and Howard, 1986; Frey and Basan, 1978, 1985) are well known from box core studies in coastal Georgia and South Carolina. Backbarrier deposits in Coastal Georgia have been mapped in three-dimensions as a "lagoonal-marsh facies" (e.g. Hails and Hoyt, 1969; Martinez, 1980; McLemore et al., 1981) and many authors have pointed out the need for more detailed vertical sequence studies in lagoon-marsh accretion (Fischer, 1961; Kraft, 1971; Edwards and Frey, 1977; Howard and Frey, 1985). Few studies, with the exception of Duc and Tye (1987), have discussed facies attributes, facies successions, and the processes that produce specific morphologic features in the backbarrier environment. Tidal point bar deposits have been compared and contrasted with their fluvial counterparts (Land and Hoyt, 1966; Barwis, 1978; Mercer, 1984). Tidal flat deposits also exhibit a fining-upward sequence overlying a basal lag (Greer, 1975). Thus far, vertical sequences proposed for the lateral migration of tidal channels in a backbarrier environment are based on surficial box coring studies (Barwis, 1978; Howard and Frey, 1985). In this study, we vibracored salt marsh deposits to examine in detail, facies sequences preserved beneath the salt marsh.

REGIONAL SETTING
The physiography of Holocene salt marshes and estuaries of coastal Georgia is summarized by many previous authors (e.g. Basan and Frey, 1978; Basan and Frey, 1985; Howard and Frey, 1985; Frey and Howard, 1986: Duc and Tye, 1987). This sector of the eastern seaboard is classified as a mesotidal coast (Davies, 1964); semidiurnal, ebb-dominated tides have an average range of 2.4 m (7.9 ft) and a spring tide range of 3.4 m (Howard...
and Frey, 1985). Sea level is rising along the southeast coast of the U.S. at the net rate of 2 mm per year (Hicks et al., 1983).

The regional geomorphic setting for the north-trending Cumberland Island, its backbarrier region, and nearby paleobarriers is discussed in Henry et al. (Contribution 3, this volume). The salt marsh under investigation is landward of Cumberland Island, Georgia, an elongate, north-trending barrier island and the southern-most of Georgia’s Sea Islands (Fig. 1). The Cumberland River, an estuary, occurs landward of Cumberland Island and parallels it. The St. Mary’s River and the Satilla Rivers are east-west flowing, smaller-scale estuaries with meandering channels that join the Cumberland River near St. Marys and St. Andrews Sounds, respectively. These estuaries are connected with numerous tidal creeks that dissect extensive salt marshes. Several intergradational types of bars including point bars, tidal flats, and longitudinal bars occur within the estuaries and tidal creeks. Many of these bars are partially vegetated by salt marsh. Salt marshes are considered ‘vegetated tidal flats’; unvegetated tidal flats are rare on the Georgia coast.

The Holocene salt marsh under investigation occurs between Cumberland Island and the next landward Pleistocene paleobarrier, the Princess Anne paleobarrier (Fig. 1). Terminology for naming Quaternary barrier island complexes is adapted from Hails and Hoyt (1969) and Huddleston (1988). Cumberland Island has a core of late Pleistocene Silver Bluff paleobarrier deposits that is fringed by Holocene beach ridges and an extensive, backbarrier salt marsh. The Silver Bluff and Princess Anne paleoshorelines are associated with sea level high stands of approximately 1.8 m and 4.0 m above MSL (respectively). Each barrier complex consists of a sandy barrier island facies and a clay-rich backbarrier or ‘lagoonal-marsh’ facies. Huddleston (1988) includes Holocene, Silver Bluff and Princess Anne deposits as part of the Satilla Formation. The subsurface geology at Cumberland Island is summarized in Henry et al. (Contribution 3, this volume). This paper will further elaborate on the facies attributes and sequences in these backbarrier deposits.

Figure 1. Regional location map showing location of vibracore profiles presented in this paper and geomorphic features of the backbarrier region behind Cumberland Island.

**METHODOLOGY**

Salt marsh deposits were vibracored along two profiles (Figs. 2 and 3) that extend westward from the margin of the Pleistocene-Holocene beachridge accretion plain of Cumberland Island to the Pleistocene Princess Anne paleobarrier (Figs. 1 and 2). The two profiles are positioned most closely to Profiles A and E of the McLemore et al. (1981) report (see Henry et al., Contribution 3, this volume, Figures 4, 5). The deposits were vibracored with a 2.5 H.P. ultralight vibracorer designed by Smith (1987) for wetlands coring using previously
Figure 2. Maps showing areal distribution of surficial environments, profile lines, and vibracoring sites. A) Backbarrier salt marsh near the southern tip of Cumberland Island and Profile B. B) Backbarrier salt marsh near Profile A.
Figure 3. Cross-sections through the backbarrier salt marsh near Cumberland Island, Georgia. The map view of Profile A is shown in Figure 2B. The map view of Profile B is shown in Figure 2A.
9 - Salt Marshes

Eighteen vibracores were acquired in 9 cm diameter seamless irrigation tubes with a maximum penetration depth of about 9 m. After cutting into meter-long segments and splitting lengthwise with piano wire, the cores were logged and photographed. Details in stratification were enhanced by several episodes of scraping the cores flat, slowly drying the cores under plastic wrap to avoid cracking and desiccation, and jetting water onto the core with a spray bottle. This treatment enhanced relief and color differences between sand and clay lithologies, allowing details in sedimentary and biogenic fabrics to be revealed without X-ray radiography. Where observable as detrital grains, fecal material, pellets, and ripup clasts were logged as sand-sized and larger particles according to Folk's (1980) classification scheme. Graphic logs and cross-sections are corrected for compaction effects.

SITE SPECIFIC GEOMORPHOLOGY

The Holocene salt marsh is bordered on the east and west by sand barriers mantled by dunes and ridges with localized maximum surface elevations as high as 12 m (Huddleston, 1988). Figure 2 shows the distribution of surficial features, including salt marsh, estuary and tidal creek channels near Profiles A and B.

Near St. Marys Sound, the Cumberland River has a natural depth of about 12 m and a dredged depth greater than 12 m (Fig. 2A). It's thalweg (deepest part of channel) occurs close to the eastern margin of the estuary, but to the north it branches and flows around large, longitudinal (parallel to flow), subtidal to intertidal bars demarcated by the 5 m depth contour. Here, near St. Marys Sound, the main channel of the Cumberland River appears to be deeply entrenched and bordered by bench-like, tidal flats or bars (Fig. 2A).

Deep within the salt marsh, the Cumberland River becomes narrower and shallows to a depth of about 7 m (Fig. 2B). Here it has a meandering thalweg that is associated with smaller, subtidal, longitudinal bars. Where major tributaries such as the Crooked River (Fig. 2B) and the St. Marys River (Fig. 2A) join the Cumberland River, broad, elongate, partially vegetated, tidal flats extend more than half-way across the channels. Marginal tidal channels separate these bars or tidal flats from the salt marsh. Point bars which may or may not be partially vegetated occur where channel curvature permits (Fig. 2A and 2B).

Small-scale, meandering tidal creeks dissect the salt marsh and are locally eroding the Cumberland Island and Princess Anne barriers. These tidal creeks have general maximum depths of about 5 m but the wider creeks (the Brickhill River) attain greater depths (10 m) than the narrower creeks. Point bars, tidal flats and longitudinal bars are also present at a much smaller scale within these tidal creeks. Between Cumberland Island and the Cumberland River (Fig. 2A) isolated dunes separated from the beach-ridge accretion plain of Cumberland Island occur within the salt marsh near the Cumberland River estuary. We did not study the origin of these features, but they are probably wind-blown deposits that post-date underlying salt marsh-tidal creek-estuary deposits.

The salt marsh maintains a surface elevation close to 1.5 m above MSL, but is slightly higher along levees adjacent to channels and slightly lower in pond-like regions in the interior of the salt marsh. Trench-like channels or crevasses (10 m wide) cut-across levees at regular intervals and connect the main tidal channels with small-scale distributary channel networks in the interior of the salt marsh. The crevasses are about 1-3 m deep, occur at right angles to the levee, and distally narrow and shallow. In many areas, the levee along tidal creeks geomorphically appears to consist of overlapping crevasse splays. Many crevasses are associated with splay-shaped channel networks, but other crevasses distally meander across the salt marsh.

BACKBARRIER FACIES DISTRIBUTION

The backbarrier facies assemblage beneath the salt marsh consists of two principal facies, a mud-rich facies and a sand-rich facies, which exhibit a full range of intergradational lithologies. Shell beds form a minor component. In two dimensions, a sand 'sheet' extends laterally from the Cumberland
River in the subsurface of the salt marsh and mud encloses it. The sand sheet is exposed at the surface of the large ‘tidal flat’ that occurs at the confluence of the Cumberland and Crooked Rivers. Towards the west, the sand sheet thins and interfingers with mud near the Princess Anne paleobarrier. To the east, the sand sheet appears to extend from the Cumberland River to Cumberland Island. It maintains a constant thickness on Profile A, but becomes thinner and more discontinuous near Cumberland Island on Profile B (Fig. 3). Its maximum thickness near the Cumberland River is unknown except for the east bank (Profile A). Black horizontal lines within the sand unit represent lags of shell hash, oyster valves, ripup clasts, or any type of coarser sediment that sharply overlies finer sediment.

Across most of the salt marsh, the mud facies occurs as a thin, surficial unit. The thickest muds occur distal to the principal estuary, the Cumberland River, near the Princess Ann paleobarrier (Profiles A and B) and near the Cumberland Island barrier on Profile B. Mud also occurs beneath the areally extensive sand sheet.

In three-dimensions, the sand ‘sheet’ that is enclosed by mud is more accurately conceptualized as a branching sand ribbon that approximately parallels a zone within which an estuary and its tributary tidal creeks migrate in the backbarrier environment. Figure 4 shows an example of a sand ribbon enclosed by mud and its distal (relative to major estuary or channel) changes in dimensions. Figure 5 shows a thinner, sharp-based sand ribbon being generated by a laterally migrating tidal channel in a very muddy region of the salt marsh such as proximal to the Princess Anne paleobarrier.

**CHARACTERISTICS OF BACKBARRIER DEPOSITS**

Figures 6 and 7 show the variation in composition, grain size, and primary physical and biogenic structures observed in the cores from the salt marsh deposits. Compositionally these deposits are a mixture of detrital quartz (and other terrigenous
clastics), resedimented ripup clasts and fecal material, detrital plant debris and shell hash. The relative abundance of quartz sand generally reflects the local availability of quartz. For example, the further the salt marsh deposits are from the main, present-day estuaries or sound regions, the less detrital quartz they contain. Quartz-starved regions contain sediments composed mainly of fecal material and ripup clasts. Fecal material and ripup clasts (or intraclasts) composed of mud are generated, transported and redeposited as detrital (sand and gravel-sized) particles in backbarrier regions. Ripups and fecal material that is less than 2 mm diameter are referred to as pellets. Smearing of pellets and intraclasts creates a muddy looking lithology but proper treatment of cores allows individual grains to be observed.

Grain sizes in salt marsh deposits range from boulder-sized slumps blocks of massive sandy clay-silt (non-resolvable in cores) to the massive, in place, sandy clay-silts of the surficial marsh (Fig 7A). The coarsest deposits resolvable in the cores consist of pebble to cobble-sized intraclasts and layers (< 30 cm thick) of ripup clasts (Fig. 6A), shell hash (Fig. 6B and 6C), and thin concentrations of quartz granules. Figure 6C shows an imbricated oyster valve bed. Elsewhere, oyster valves are randomly oriented and occur in pavement-like layers or as small biostromes enclosed in mud (Fig. 7A). Ripup clasts commonly were 1 to 5 cm in diameter, flat, and imbricated (Fig. 6A). Larger, equant mudball intraclasts were observed locally.

Fine-medium grained quartz sand (Fig. 6D,E,F) may be cross-cut by large (2-3 cm diameter) mud-lined burrows (Fig. 6D). Commonly it is crossbedded or flaser bedded with pellet layers as flasers and drapes (Fig. 6E). Bioturbation of sand, as shown in Figure 6E, results in a micro-mottled pelletal texture (Fig. 6F) superimposed on current-bedded sand.

Tidal bedding is common with interlayers of quartz, pellets and intraclasts (Fig. 6A) arranged as wavy, flaser and lenticular bedding (Reineck and Wunderlich, 1968). For example, Figure 6E shows quartz sand with pellet flasers. Figure 7B shows wavy bedded deposits with equal thickness interlayers of quartz sand and pellet sand. The pellet-rich layers are ripple cross laminated. Figure 7C shows lenticularly bedded deposits with pellet layers predominating over quartz layers. Other primary features include cut and fill structures, large-scale cross-bedding (interpreted from inclined layers), small-scale, bidirectional cross-bedding, and minor ripple bedding.

These interlayered deposits are commonly bioturbated to varying degrees. In Figures 6F and 7D for example, the quartz-rich layers have a micro-mottled appearance but remnant tidal bedding is still visible. Other fabrics are more extensively bioturbated with only minor traces of primary fabric preserved (Fig. 7E, 7F, 7G). Surficial clay-silt of the salt marsh generally is very fine-grained and completely bioturbated (Fig. 7A and 7F).

Distinct, identifiable, Ophiomorpha nodosa were rare but occurred as burrows cross-cutting fine-grained, flaser-bedded quartz sand. Large (several cm diameter), distinct, mud-lined burrows, lacking the knobby wall structure of Ophiomorpha, were also observed in quartz sands (Fig. 6D). These are interpreted as Thalassinoideas-type burrows. In Figure 7G, distinct (cm-sized) clay-lined burrows form a reticulate network (cross-sectional view) in loose quartz sand. Figure 7C shows numerous small escape burrows disrupting primary layering. Fine (mm) reticulate networks of mud-lined(?) burrows also occur in quartz sand (Fig. 6C). Locally, pellets appear to be matrix supported and/or clast-supported within a quartz sand matrix (Figs. 7E, 7F). These fabrics probably originated as multiple generations of overlapping backfill or pelleted wall structures (Fig. 7G). We recognize that this area of research needs further literature review and study.

**BACKBARRIER SEQUENCES**

Several figures have been prepared that show core photographs, graphic logs and facies characteristics (Figs. 8, 9, 10). The sand 'sheet' beneath the salt marsh overlies fine-grained mud that is locally lenticularly bedded (Fig. 8), rooted, interlayered (Fig. 9) or massive. These muds are
Figure 6. Photographs of sediment cores from backbarrier salt marsh deposits. Cores are about 10 cm wide. (A) BC-1 - Channel lag deposit with imbricated layers of mud intraclasts (dark gray). (B) BC-17 - Channel lag deposit consisting of mud intraclasts, shell hash, quartz granules and fine sand. Lag deposit overlies tidally bedded sand with faint reticulated bioturbate texture. (C) BC-1 - Imbricated oyster valves sharply overlie bioturbated sand with reticulated network of tiny mudlined burrows. The oyster lag is probably a creek bank or lateral channel deposit adjacent to a salt marsh bank. It is overlain by the extensively bioturbated clay-silt typical of some creek banks and the surficial salt marsh. (D) BC-17 - White quartz sand with cross-section through Thalassinoides type burrow (lower left). Faint mottling indicates bioturbate texture. Minor flaser bedding. (E) PO-24 - Flaser bedded quartz sand with pellets forming flat laminations, flaser laminations and ripple cross-laminations. An example of sub-tidal deposition in a channel; physical sedimentary structures predominate over biogenic structures. (F) Flaser bedded quartz sand with micro-pelletal (bioturbate) texture superimposed on primary stratification. Thicker clay layer in center of photograph may have been a hardground because it contained Pholad boring clams. Hardgrounds in the subsurface may provide support for a Holocene lowstand in sea level about 2400 years B.P. (Deprattar and Howard, 1981).
Figure 7 (A-D). Photographs of sediment cores from backbarrier salt marsh deposits. Cores are about 10 cm wide. (A) Surficial salt marsh deposits consists of a rooted mud less than 30 cm thick that overlies a bed of oyster valves, oyster hash, and mud. The oyster shell zone overlies more massive rooted mud typical of the salt marsh. These deposits probably represent creek bank or deposits in a small tidal channel or crevasse that dissect the marsh. (B) PP-3 - Wavy bedded quartz sand and pellet sand with minor superimposed bioturbation. Ripple laminations present. (C) Predominantly flaser to parallel laminated deposits consisting of interlayers of quartz sand and fecal pellet sand. Locally laminations are distorted or disrupted by burrowers. Traces of cross-stratification are present. (D) BC-1 - Bioturbated sand and mud with traces of primary stratification still visible. Sediment is micro-mottled and pellets appear to be dispersed in quartz sand. Fabric may actually consist of overlapping burrow backfill.
Figure 7 (E-F). Photographs of sediment cores from backbarrier salt marsh deposits. Cores are about 10 cm wide. (E) Extensively bioturbated sand and pellets. Only traces of primary stratification are preserved. Two large tabular shaped mud intraclasts occur near the bottom of the photo. This fabric probably represents intertidal bar or tidal flat deposits. (F) Micromottled sand with pelleted texture fines upwrd into massive clay-silts of the surfical salt marsh. Pelleted texture here looks like multiple generations of overlapping backfill. (G) BC-17 - Bioturbated sand changes upsection into sand with numerous distinct mud-lined burrows that form a reticulate network near top of photo. Note 'holes' lined with mud and filled with loose sand.
interpreted as preexisting salt marsh or possibly lagoonal (?) deposits. Above these basal muds are fining-upwards sequences of facies that are interpreted as thalweg, channel, intertidal bar, channel margin, creek bank and salt marsh deposits. Tidal flat deposits are considered similar to intertidal bar deposits.

Figures 8 and 9 show complete fining-upwards sequences. Each fining sequence has a basal lag (thalweg deposit) that is overlain in succession by (1) well-stratified, tidally bedded sediment, (2) poorly stratified, extensively bioturbated mud or sand, and (3) rooted mud of the surficial salt marsh. Sequences such as this are interpreted as channel migration or bar accretion deposits. Figure 8 shows a mud-rich fining-upward sequence characterized by wavy- to flaser-bedded quartz-pellet sands and the overlying bioturbated sands. Figure 9 shows a similar sequence that contains more quartz sand. Here the basal lag is overlain by flaser bedded sand that is replaced upward by bioturbated sand and bioturbated mud with oyster valves. Well stratified deposits represent high energy channel and channel margin deposits. Bioturbated deposits are interpreted as intertidal bar, tidal flat, or bank margin deposits. Shell lags, lenticular bedding, sand layers and slump blocks are associated with channel margins and creek banks. Some salt marsh sequences consist of thick intervals of bioturbated sand and mud (Fig. 10). Rooted mud of the salt marsh occurs at the top of all three vertical sections. Locally sequences are amalgamated or stacked due to several events of lateral migration.

**FACIES INTERPRETATIONS**

**Thalweg and Channel Floor Deposits** Sequence bases are marked by abrupt facies changes from a finer-grained facies to a coarser grained facies (Figs. 8, 9). Where quartz sand is scarce, contacts may be subtle. A lag of oyster or shell hash, slump blocks, intraclasts or coarser sediment commonly occurs at the contact. Such lag deposits typically mark the passage of the deepest part of a tidal channel, or thalweg, past the site during the lateral migration process. A thalweg typically is coincident with the highest velocity currents and longest duration of flow over the tidal cycle. Barwis (1978) observed large (>1m) undeformed, rooted slump blocks, shell beds, ripup clast beds, and scattered isolated mudball ripups (>20 cm) on channel floors and in the thalwegs of tidal creeks.

**Channel and Tidal Bar/Tidal Flat Deposits** Ripples, megaripples and sandwaves have been observed on intertidal point bars in tidal creeks (Barwis, 1978; Howard and Frey, 1985). Internally these bars exhibit alternating layers, lenses or stringers of sand and mud (Barwis, 1978; Howard and Frey, 1985). Stratification within individual beds includes large-scale trough cross-stratification, ripple bedding, wavy, flaser and lenticular bedding, and flat laminations (Howard et al., 1973: Howard et al. 1975). Mud is present as fecal pellets, as ripup clasts, and as finely disseminated suspension mud (Barwis, 1978). However, detrital fecal pellets are volumetrically more important than suspension muds (Howard and Frey, 1985). Wavy, flaser, and lenticularly bedded intraclasts, quartz, and pellet sands were the most common bedding types observed in this study of salt marsh stratigraphy. These complexly interlayered deposits were deposited on intertidal to subtidal bars in tidal channels.

In small tidal creeks, large-scale bedforms are completely confined to subtidal bar flanks, usually consisting of ebb-oriented sand waves and megaripples (Barwis, 1978). Bioturbated muddy sand (massive quartz-pellet sand, this report) occurs on the inner bar adjacent to the marsh and biogenic structures are common on bar crests (Barwis, 1978). Although bioturbation is much more significant along channel margins and on point bar crests, physical structures are common in channels (Howard and Frey, 1973; Barwis, 1978). As a consequence, point bars typically consist of 2 principal facies: the high-energy, channel side exhibits distinct layers, laminae or flasers, whereas the low-energy marsh side consists of bioturbated muddy sand with few or no traces of clean sand (Howard and Frey, 1985). Bioturbation is least in the most seaward point bars along an estuary (Howard et al., 1975). Well stratified channel-margin mud is primarily derived from reworked, eroded, and ripup clasts arranged as wavy or flaser bedding.
Figure 8. A muddy, fining-upward sequence generated by tidal channel migration. Core photograph, graphic log, facies descriptions and interpretation for Core PO-XX. The location of PO-XX is shown in map view on Figures 1 and 2 and in cross-section view on Figure 3. A shell lag has yielded a radiocarbon date of 989 ± 73 B.P.
Figure 9. A sandy, fining-upward sequence generated by tidal channel migration. Core photograph, graphic log, facies descriptions and interpretation for Core BC-17. The location of BC-17 is shown in map view on Figures 1 and 2 and in cross-section view on Figure 3. Two radiocarbon dates on shell material yield dates of $1780 \pm 73$ and $1584 \pm 83$ years B.P.
Figure 10. Core photograph, graphic log, facies descriptions and interpretation for Core PO-12. The location of PO-12 is shown in map view on Figures 1 and 2 and in cross-section view on Figure 3. Closely spaced samples yield disparate radiocarbon dates of $2990 \pm 78$ and $1615 \pm 53$ B.P.
(Howard and Frey, 1985).

The fining-upward sequences discussed here consist of a lower section with well preserved primary sedimentary structures and an upper section dominated by bioturbated deposits. We interpret these facies as subtidal to intertidal bar deposits with the well stratified lower section deposited on the high energy, channel side of the bar and the upper section of bioturbated sand deposited on the low-energy marsh side of the intertidal bar. Physical processes dominate over biogenic processes subtidally in the main channel. The more massive-appearing, bioturbated fabrics are interpreted as bar-crest and channel-margin deposits, where biogenic processes predominate over physical processes. Ripup clasts, slump-blocks, shell hash and other other coarse debris aggregate as pavements in small tidal channels between the salt marsh and the intertidal bar during periods of submergence and high-velocity flow. At low water, burrowers actively rework the strata. Localities deep within the marsh that lack a source of quartz sand may have intertidal bars composed of mudball ripups clasts, slump blocks and thick sequences of massive pellet-quartz sand.

**Creek-Bank Deposits** Creek-bank deposits are characterized by chaotic bedding, especially where thick sequences of mud are exposed in creek banks. Chaotic bedding as originally used by Howard *et al.* (1973) refers to the characteristic internal structure caused by the reincorporation of large slump blocks (generally composed of sandy clay-silt) into the sediment on the lower portion of creek banks. Chaotic bedding was further described as a collective term referring to particularly distinctive features unique to the creek-bank environment and including contorted and loadcasted laminae, slump and fault structures, and local discontinuity surfaces (Edwards and Frey, 1977). Barwis (1978) also noted chaotic bedding in the lower banks of tidal creeks.

Chaotic bedding is very difficult to recognize in vibracores because of the width of the core (<10 cm). Mottled, patchy fabrics, rotated slump blocks and irregular bedding all indicate chaotic bedding. Slump blocks would not be as prominent along sandy salt marsh banks near barriers and sound areas because slump blocks composed of sand are easily disaggregated by tidal currents. Here, creek-bank and bank-margin deposits would probably consist of extensively bioturbated sand and mud with or without chaotic bedding.

Chaotic bedding is attributed to several factors (Frey and Howard, 1985; Edwards and Frey, 1977): (1) intense bioerosion by decapods (Edwards and Frey, 1977; Frey and Basan, 1978; Letsch and Frey, 1980); (2) dessication of channel banks leading to the formation of mudcracks (Edwards and Frey, 1977); (3) tidal current plucking of mud and sod clasts from the banks and adjacent marshes (Edwards and Frey, 1977); and (4) slumping of creek banks which produces slump blocks, sod clasts and ripup clasts. The process of creek bank accretion occurs as slump blocks and grass growth emanating from blocks trap and bind sediment, eventually coalescing with other blocks or the marsh edge (Letsch and Frey, 1980).

Creek-bank deposits also have other characteristics besides chaotic bedding. Stratification along creek bank and estuary margins also includes wavy and lenticular bedding with occasional flasers (cf. pl. 2, Fig. 5; Reineck and Wunderlich, 1968; Howard and Frey, 1975) and intercalated laminae or cross laminae of clay, silt and sand (Edwards and Frey, 1977). Creek banks may consist of thixotropic muds deep within the marsh (Barwis, 1978).

**Salt Marsh** The massive mud that mantles the salt marsh was deposited on channel margin levees and as overbank or “through-the-bank deposits” within the marsh. Marsh deposits in Georgia are almost totally bioturbated (Frey and Basan, 1978; Howard and Frey, 1985) with intense sediment reworking attributed to the deposit-feeding activities of abundant decapods and plant roots (Edwards and Frey, 1977; Basan and Frey, 1977; Howard and Frey, 1985). Mud layers and lenses for the most part originate as accumulations of fecal pellets and pseudofeces rather than as discrete or flocculated particles of clay. Compaction amalgamates these detrital particles into a cohesive, structureless deposit (Howard and Frey, 1985). Thus, deposition
is primarily by vertical accretion of mud derived from a number of sources but reseedimented pellets contribute the greatest volume of mud (Frey and Basan, 1985; Smith and Frey, 1986). Vertical accretion rates for Georgia salt marshes are very low, less than 1 cm per year (see Smith and Frey, 1986; Frey and Basan, 1985; Letsch and Frey, 1980, 1980a; Pethick, 1981). Mussel biodeposition alone constitutes a large proportion of the annual marsh sediment budget (Smith and Frey, 1985). Rooted mud occurs only surficially or beneath channel migration sequences. Rooted massive mud that occurs at depth beneath the thinning-upward sand sheet may represent slumped blocks.

BACKBARRIER PROCESSES

Lateral Migration of Tidal Channels Because lateral migration of tidal creeks reworks or removes a thickness of sediment equivalent to the thalweg depth, it is considered the most active sedimentologic process presently operating in the marsh (Letsch and Frey, 1980). Tidal channels may migrate via several related processes: (1) bank erosion (Letsch and Frey, 1980); (2) bank accretion (Letsch and Frey, 1980); or (3) intertidal bar-tidal flat accretion. These three processes act together to generate large-scale (m thick) lateral accretion sets in a backbarrier environment.

Creek banks commonly slump along a series of fault planes that parallel the channel. Coherent slump blocks are generated along the channel margin (Fig. 11A). If the disarticulated slump blocks and sod clasts are removed by currents, the bank retreats and the channel widens (Fig. 11B). If bar accretion or bank accretion occurs along the opposite bank, the tidal channel migrates laterally. The curvature of the creek channel (whether straight or meandering) has little direct relation to the amounts of slumping and erosion observed (Letsch and Frey, 1980).

The process of creek bank accretion (Fig. 11C and 11D) occurs as slump blocks and grass growth emanating from blocks trap and bind sediment, eventually coalescing with other blocks or the marsh edge (Letsch and Frey, 1980). In this manner, the channel bank, although slumped along fault planes, accretes laterally towards the channel axis. The channel narrows if bank erosion does not occur along the opposite bank. Bank accretion generates chaotic bedding and is an important process along tidal creeks with thick muddy banks. Slump blocks remain coherent if the banks are composed of clay-silt rather than sand-rich. Bank accretion generates lateral accretion sets (Fig. 11D).

Bar accretion occurs when successive beds of sand are emplaced along the channel side of point bars, tidal flats, or other tidal bars that effectively cause lateral accretion of the bar towards the channel (Fig. 12). The channel may simply narrow and deepen. If bank erosion occurs concurrently along the opposite bank, the channel migrates laterally. Bar accretion towards a channel generates lateral-accretion bedding. Bar accretion is an important process where a large supply of sand is available to construct barforms, usually near coastal barriers or sound regions.

Although known rates of channel-margin retreat are as high as 7.2 m/yr with a mean of 1.9 m/yr, rates of creek bank accretion have not been documented (Letsch and Frey, 1980). Freely meandering channels probably accrete laterally at the same rate. Rates of bank accretion and bar accretion can be determined where historic maps exist. Historically, however, very little shoreline data has been collected for tidal creeks and estuaries in backbarrier regions.

Beach Creek is a small tidal creek that meanders across the salt marsh between the Cumberland River and the beach ridge accretion plain of Cumberland Island near Profile B. Figure 13 shows Beach Creek meandering patterns for two different time periods, 1857 and 1981. Over a period of 124 years, Beach Creek behaved in a manner analogous to fluvial channels; thalwegs migrated laterally, tightly curving meander loops were cut-off, meander loops were abandoned and infilled, creek banks were eroded, and point bars laterally accreted deep within the muddy salt marsh. Beach Creek has a maximum channel depth of about 5 m below MSL. The salt marsh surface occurs at about 1.5 m above MSL. Therfore, the lateral migration of Beach Creek via bank erosion
9 - Salt Marshes

has reworked a 6.5 m thickness of salt marsh deposits. At the same time, a thickness of about 6.5 m of salt marsh was generated via bank and/or bar accretion. Thicker sequences were generated in abandoned channels. In areas where Beach Creek migrated laterally between 1857 and 1981, a tentative rate of lateral erosion/accretion derived from the two maps is on the order of 1 m/year for that time period.

Thus, the Holocene salt marsh landward of Cumberland Island is being reworked rapidly to depths comparable to tidal channel depths (5-10 m). At the same time, salt marsh deposits are generated rapidly at vertical thicknesses comparable to channel depth (5-10 m). Because laterally migrating tidal creeks are eroding into Pleistocene ‘headlands’, sharp facies contacts are probably generated between barrier island sand facies and the ‘lagoonal-marsh’ backbarrier facies. Where the Cumberland River is shallow (7m) it probably freely meanders across the salt marsh between the Princess Ann paleobarrier and the Cumberland Island. As the Cumberland River approaches St.

Figure 11. The processes of creek bank erosion and bank accretion as proposed by Letsch and Frey (1980). A. Creek banks commonly slump along a series of fault planes that parallel the channel and coherent slump blocks are generated along the channel margin. B. If the disarticulated slump blocks and sod clasts are removed by currents, the bank retreats and the channel widens. If bar accretion or bank accretion occurs along the opposite bank, the tidal channel migrates laterally. C. Slump blocks may be generated along fault planes at the channel margin, but instead of being transported away, the slump blocks become vegetated by salt marsh grass (Spartina alterniflora). D. Creek bank accretion occurs as slump blocks and grass growth emanating from blocks, trap and bind sediment, eventually coalescing with other blocks or the marsh edge. In this manner, the channel bank accretes laterally towards the channel axis and lateral accretion bedding may be produced as the channel migrates to viewers right.

Figure 12. Channelward migration of tidal bar, point bar or tidal flat deposits may generate an areally extensive sandbody in the subsurface of the salt marsh. Channel flow is perpendicular to the plane of the figure. The tidal channel is migrating to the right as successive beds of sediment are emplaced along the channel margin. A basal lag is produced during bar accretion and the overlying sequence consists of well-stratified subtidal bar deposits, bioturbated intertidal bar or tidal flat deposits, and a surficial salt marsh deposit. In this manner lateral accretion bedding is produced via tidal channel migration.
9 - Salt Marshes

Figure 13. The Beach Creek salt marsh near the southern tip of Cumberland Island. The eastern section of Profile B extends across the Beach Creek salt marsh. Figures 1 and 2 show the location of Profile B. Beach Creeks channel pattern changed considerably between 1857 and 1981. During this period, Beach Creek behaved in a manner analogous to fluvial channels: thalwegs migrated laterally, tightly curving meander loops were cut-off, meander loops were abandoned and infilled, creek banks were eroded, and point bars laterally accreted deep within the muddy salt marsh. The lateral migration of Beach Creek reworked and regenerated a 6.5 m thickness of salt marsh deposits.

Marys Sound, it is entrenched to depths of at least 12 m below MSL and probably migrates laterally only in a restricted sense. Thus, the thickest sand bodies are generated in a narrow zone marginal to the thalweg of the Cumberland River.

The salt marsh landward of Cumberland Island probably evolved as vegetated, bar-accretion deposits—especially near St. Marys Sound. The salt marsh appears to be locally expanding. Many tidal channels contain bars that originated (based on their morphology) as mobile intertidal bars that are now stabilized by salt marsh vegetation. Historical photographs near Sea Island show that mobile intertidal bars near inlets have enlarged and heightened overtime. Once low-marsh vegetation (Spartina alterniflora) becomes established, it traps sediment, and rapidly expands in area. Eventually the new salt marsh coalesces with the former channel bank. An example landward of Cumberland Island is the large tidal flat or tidal bar near the confluence of the Cumberland River and the St. Marys River. It is now partially vegetated but probably underwent an initial period of progradation in both a seaward (downcurrent) direction (ebb-dominated currents) and in a channelward direction as sediment filled in and narrowed a once broader (and shallower) Cumberland River. Eventually this bar will probably completely convert to salt marsh. This common backbarrier process locally has caused the Cumberland River estuary channel thalweg to shift eastward and to narrow and deepen.

**Vertical Accretion** Vertical accretion rates for Georgia salt marshes are very low—less than 1 cm per year (see Smith and Frey, 1986; Frey and Basan, 1985; Letsch and Frey, 1980a, 1980b; Pethick, 1981) Mussel biodeposition alone constitutes a large proportion of the annual marsh sediment budget (Smith and Frey, 1985). Long term vertical accretion rates in these salt marshes due to the Holocene transgression are probably only about 2 mm/year (the rate of sea level rise). Lateral accretion rates in salt marshes are thus several orders of magnitude greater than vertical accretion rates.

Salt water enters the marsh through crevasses connected to tidal channels during flood tide. At highest tides, levee overtopping can occur. Suspended fecal material is deposited at slack water as a drape over the marsh surface and levee, but is rapidly reworked via bioturbation. Because crevasses are connected to distributary channel networks deep within the salt marsh, it is likely that resedimented pellets are deposited as splays associated with the crevasses. Amalgamated fecal material, however, looks like massive mud; and from observations at Sea Island splay deposits are
not differentiable lithologically from other massive
muds. At high-water slack, the entire marsh surface,
including the levees, may be covered with water
and resembling a lake. During ebb tide, sea water
leaves the marsh through the distributary channel
networks which are connected to crevasses and
tidal channels. As the water level drops and becomes
more channelized, the network draining the marsh
comes eroded and entrenched. Thus both vertical
accretion and erosion are taking place on the present
salt marsh surface.

THE AGE OF THE DEPOSITS

Several radiocarbon dates were acquired from
intervals of predominantly reworked shell material.
For example, at BC-17 (Fig. 9) two samples were
dated. Shell material from the basal lag of a fining-
upward sequence yielded a date of about 1780 years
B.P. A radiocarbon date of an oyster lag deposited
along a bank margin yielded a date of approximately
1584 B.P. Elsewhere (Fig. 10), two dates were
attained on oyster valves that were probably associated
with a small biostrome that developed on a bar crest.
However, the valves were neither articulated nor in
place. These shells yielded ages of about 2990 and
1615 years B.P respectively. A shell lag from a
subtidal bar was also dated (Fig. 8) yielding a date of
about 989 BP. All of the dates were less than 3000
years B.P.

Radiocarbon dates could potentially help us
determine whether a Holocene low-stand in sea level
(-3 to -4 m below present MSL) occurred about 2400
years ago along the Georgia coast (Depratt and
Howard, 1981). However, because of the dynamic
nature of tidal channels in the backbarrier region, and
its ability to rework and redeposit shell material, the
radiocarbon dates on disarticulated shells and shell
hash tell us very little—except that the deposits
sampled are less than 3000 years old. In addition, the
dates can not be used to calculate either vertical or
lateral rates of accretion for the salt marsh because the
salt marsh is continually being reworked over vertical
distances comparable to channel depth.

CONCLUSIONS

• Backbarrier salt marshes are underlain by two
principal facies, a clay-rich facies and a sand-rich
facies which exhibit a full range of intergradational
lithologies. Shell beds form a minor component.
• Beneath the salt marsh, sandbodies probably
occur as ribbons bordering estuaries and tidal creeks.
These sand ribbons are irregular in thickness, highly
variable lithologically, and constructed of coalescing
lenses that represent individual bar deposits. The
width of the sand ribbon represents the zone in which
a tidal creek or estuary migrated.
• Deposits in salt marshes are composed of
resedimented fecal material, ripup clasts, shell hash,
and slump blocks as well as quartz and other detrital
mineral grains. Surficial salt marsh deposits consist
of amalgamated fecal pellet muds; bar deposits are
tidally bedded (wavy, flaser, and lenticular bedded)
and composed of layers of quartz, pellets, and ripup
clasts. Bioturbation is superimposed on tidally bedded
deposits to form massive appearing quartz-pellet sands.
• The relative abundance of quartz sand within salt
marsh deposits generally reflects the local availability
of quartz. For example, deposits deep within the
marsh, although often depleted in quartz, are enriched
in fecal material, intraclasts and slump blocks and are
more extensively bioturbated than deposits adjacent
to estuary channels or sound areas.
• Salt marsh deposits consist of a fining-upward
sequence of facies that overlies pre-existing probable
salt marsh deposits. The fining-upwards sequence is
interpreted as tidal channel migration deposits that
include thalweg, channel, intertidal bar, creek bank
and salt marsh deposits.
• Tidal channels migrate via the processes of bank
erosion, bank accretion and bar accretion. The channel
migration process reworks or removes a thickness of
sediment equivalent to the thalweg depth of the
migrating channel.
• Lateral migration of tidal channels probably
produces large-scale (m thick) lateral accretion sets
that are internally tidally bedded.
• Lateral accretion rates in salt marshes are several
orders of magnitude greater than vertical accretion
rates.

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ABORIGINAL OCCUPATION (2000 B.C.-1562 A.D.)
Timucuans, part of the Muskogean linguistic group, were probably the first settlers of Cumberland Island. Archaeological excavations show that pottery-making aboriginal people inhabited the island as early as the second millennium B.C. The Timucuans on Cumberland Island were the Tacatacori tribe. Their main village was located in the vicinity of Dungeness dock.

EUROPEAN ARRIVALS (1562-1733)
French explorers discovered Cumberland Island in 1562 and planned to return to build a fort. However, the Spanish arrived in 1564 and claimed it for King Philip II of Spain. The Tacatacori had been friendly with the French, but were distinctly unfriendly toward the missionary efforts of Jesuits led by Father Pedro Martinez. They killed him, three other Jesuits and eleven Spanish soldiers on Cumberland Island in September, 1566. Jesuits were replaced by Franciscans in 1574. Franciscan influence reached its height in 1650. Two reasons for its subsequent decline were preeminent: coastal Indians sickened and died by the thousands from the "pestilence" (smallpox), and Yamassee Indians, displaced by English settlers to the north, pushed the less aggressive and weakened Timucuans further south and inland. By 1700 there were no Timucuans or Spanish on the island.

ENGLISH OCCUPATION (1733-1776)
While Frederica and its fort (St. Simons) were nearing completion, James Edward Oglethorpe, founder of Georgia, set off to see "where His Majesty's Dominions and the Spaniards join." He was accompanied by 40 Yamacraw Indians, including the chief, Tomochichi, and his nephew, Toonahowic. In 1736, they landed on an island that

Figure 1. Ruins of Dungeness.
the indians knew as Missoe or Wissoe (sassafras or beautiful, resp.), and that the Spanish knew as San Pedro. It was Toonahowie that asked that the island be named Cumberland in honor of his friend William Augustus, Duke of Cumberland. Oglethorpe ordered the construction of two forts—Fort St. Andrews on the north end and Fort Prince William on the south end. In 1742, these forts played an important role in helping the English defeat the Spanish at Frederica during the War of Jenkins Ear.

With the cession of Florida by Spain to England in 1763, the English Board of Trade began to pass on applications for land grants in Georgia and Florida. Nearly all of Cumberland was granted to some nine individuals. Due to the threat of war, pirates, and hostile indians, the wealthy land owners did not live on the island. During the American Revolution, 4000 British soldiers camped on Cumberland, and used it as a staging and supply area for their attack on Savannah.

**PLANTATION ERA (1786-1880)**

General Nathaniel Greene received a plantation near Savannah from the state of Georgia for his services during the American Revolution. He then purchased land on Cumberland Island for the purpose of harvesting the live oak trees for ship building. He died from sunstroke in Savannah in 1786. Ten years later, his widow, Caty, married Phineas Miller and moved to Cumberland where she built a four story tabby mansion named “Dungeness”. “Liveoaking” and growing sea island cotton were the main commercial ventures. Other crops included oranges, lemons, pomegranates, figs, olives, corn, and sweet potatoes.

**ESTABLISHMENT OF PRIVATE ESTATES (1880-1916)**

When the Civil War ended, Cumberland Island's plantation economy was in a shambles, and the fields and mansions in ruins. Thomas Carnegie (brother and partner of Andrew Carnegie)
acquired Dungeness and presented it to his wife, Lucy, in 1881. They built a new Dungeness which was to become a winter home. After Thomas’s death in 1886, it became the primary residence of Lucy and her nine children. Before her death in 1916, Mrs. Carnegie established a trust to maintain her 16,000 acres on Cumberland Island as a home for her children.

PRIVATE AND PUBLIC OWNERSHIP
(1916-PRESENT)

The synergistic effects of the Great Depression, the emigration of the labor force to industries developing in Florida, and a cumbersome trust arrangement induced severe problems in the management of Cumberland by the Carnegie heirs. They began to seek a product which could make the island financially self-sufficient. T. Morrison Carnegie experimented with growing tung trees. Tung nuts would produce tung oil for waterproof paints, varnishes and cloth compounds for insulating motor dynamos and cables, and serve as other substitutes for rubber. It was a commercial failure.

Between 1955 and 1957 various mining companies became interested in strip and hydraulic mining for titanium and ilmenite, found in the sands of the sea islands. The extraction of this natural resource would have ensured retention of the land by the Carnegie heirs but would have changed radically the topography, floral and faunal habitats, and future suitability of Cumberland as a private retreat. The high bidder was Glidden Company of Cleveland, who planned to employ 100 persons, build a 150-acre village, and erect 70 residences. Expected annual production was 100,000 tons of ilmenite and leucoxidene concentrates, 10,000 tons of rutile, and 15,000 tons of zirconium.

The offer to lease was refused, as the result of litigation brought by one Carnegie heir. The Superior Court determined that acceptance of the 20-year lease was in violation of the Trust established by Mrs. Carnegie that Cumberland was to be maintained as a home for her children until the last one had died. The trust came to an end in 1962 when the “last surviving child” died. The estate was divided among the Carnegie heirs, some of whom decided to sell their land. One developer, Charles Frazier, bought 2,000 acres and planned to build a resort comparable to his Sea Pines development at Hilton Head. Other Carnegie heirs began to consider sale of their land to the National Park Foundation. To start the ball rolling, the heirs that owned Plum Orchard mansion donated the mansion, 13 acres, and $50,000 to the Foundation. The Foundation acquired approximately 90 percent of the island. In 1972, Cumberland Island National Seashore became an official part of the National Parks system of the U. S. Department of the Interior.
The western shore of Cumberland Island National Seashore is characterized by extensive salt marsh habitat. Like, the ocean shore, the estuarine shore has remained relatively constant with respect to elevation above sea level throughout historical time, despite eustatic sea level rise (Granant, 1990; Letzsch and Frey, 1980). In several areas experiencing recession, the marsh has eroded back to the island core. The presence of two archeological sites along the estuarine shoreline of Cumberland Island in the vicinity of Sea Camp provide an opportunity to evaluate shoreline change over a six year period. Between Sea Camp and Dungeness Docks is an eroding shoreline were Holocene sand deposits and forest habitat are eroding and exposing underlying peat and mudflat deposits (Figs. 1,2, and 3). In October 1987, in an effort to determine how threatened the archeological sites were along the river, permanent reference markers were installed by a team from the Southeast Archeological Center of the National Park Service (Brewer, pers. comm.). Profile surveys of the bluff and beach were carried out at that time. In June of 1993, with the assistance of an EarthWatch research team, profiles with a K & E transit were run at both sites. The results of these two surveys are presented in Figure 4. At Station 3, the shoreline (measured at the toe of the escarpment) retreated 9.2 m over the period of record for an average loss of 1.6 m per year. At Station 1, located near Dungeness Dock, 7.9 m of erosion occurred (-1.4 m/yr). A combination of factors are related to the observed trend here. This shoreline is exposed to severe weather from the northwest and southwest, over relatively long fetches. As a result, storm waves attack the bluff directly during periods of high tide. Even at low tide, run-up and erosion of the bluff and underlying peat outcrops occurs. Acting in a similar manner to a seawall, the peat probably offers some degree of protection from scour and direct impact because of its consolidated nature. Boat wakes are also a factor; especially large displacement hulls and again at high tide.

Figure 1. Wide view of western shore between Sea Camp and Dungeness Docks, at low tide.
11 - Shoreline Recession

Figure 2. Shoreline near CUIS 6, Station #1 (see Fig. 4).

Figure 3. Indian midden along western shore.

REFERENCES
INTRODUCTION

Heavy minerals have been mined from surface sands and near surface sands of the Atlantic Coastal Plain in the southeastern United States for decades (Table 1). Important heavy-mineral deposits in northern Florida and southern Georgia (Fig. 1) are present in beach ridges composed of quartz sand. Some of these ridges formed in creastal areas of major marine transgressions. Others developed during times of temporary halts or at times of slight transgressions of seas during periods of general marine regression.

The Yulee heavy-mineral deposits in northeastern peninsular Florida are in low, north-south trending beach ridges that formed in the creastal area of the Pamlico marine transgression in late Pleistocene time. The general surface elevations of these sand ridges range from about 25 feet to slightly more than 60 feet above mean sea level (Fig. 2). A plains region with surface elevations generally from about 30 to 35 feet above mean sea level extends westward from the Yulee ridges to the eastern edge of the Duval Upland (Fig. 1). Surface elevations east of the Yulee ridges are lower than those west of the ridges. These lower surfaces pass into salt marshes which extend to Amelia Island along the Atlantic Coast (Fig. 3).

Even though the individual Yulee ridges are not wide, their combined width is slightly over 4.5 miles. The northwestern most ridge, the Crandall Ridge, is clearly delineated (Fig. 2). The ridge lying just east and southeast of the Crandall Ridge has summit elevations similar to those of the Crandall Ridge, but this more easterly ridge is not as clearly delineated. Still farther east the summit elevations of ridges are lower, and the ridges are less pronounced. In some cases the ridge positions are best reflected by drainage patterns.

YULEE DEPOSITS

Location The Yulee heavy-mineral deposits are located on approximately 5,000 acres of land

Table 1. Chronology of Heavy-Mineral Mining — Atlantic Coastal Plain of the Southeastern U.S. (from Pirkle et al., 1991; modified from Garnar 1978; 1980).

<table>
<thead>
<tr>
<th>Dates of Operation</th>
<th>Operator (Deposit)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1895-1916</td>
<td>North Carolina Monazite Mining</td>
</tr>
<tr>
<td>1916-1929</td>
<td>Buckman-Pritchard, Inc. (Mineral City, FL)</td>
</tr>
<tr>
<td>1940-1955</td>
<td>Riz Mineral Company/Florida Ore Processing (Palm Bay near Melbourne, FL)</td>
</tr>
<tr>
<td>1943-1964</td>
<td>Humphries Gold Corp. (Arlington, FL)³</td>
</tr>
<tr>
<td>1956-1963</td>
<td>Hobart Brothers (Vero Beach, FL)</td>
</tr>
<tr>
<td>1949-Present</td>
<td>Humphries Gold Corp.*/E. I. Du pont de Nemours &amp; Co., Inc. (Trail Ridge, FL)</td>
</tr>
<tr>
<td>Late 1950’s - early 1960’s</td>
<td>Marine Minerals (Horse Creek, Aiken, SC)</td>
</tr>
<tr>
<td>1964-1974</td>
<td>Humphries Gold Corp. (Folkston, GA)</td>
</tr>
<tr>
<td>1972-Present</td>
<td>Titanium Enterprises/Associated Minerals USA (Green Cove Springs, FL)</td>
</tr>
<tr>
<td>1974-1980</td>
<td>Humphries Mining Company (Bolougne, FL)</td>
</tr>
</tbody>
</table>

³Humphries Gold Corp. later became Humphries Mining Company.
Figure 1. Location map showing physiographic features and the sites of heavy mineral concentrations. The Highland ore body is also known as the Maxville ore body. Modified from Pirkle et al., 1991.
Figure 2. Yulee heavy-mineral sand deposits. These deposits, outlined by dark lines, have a minimum thickness of 6 feet and a minimum TiO₂ content of 1 percent. A few contour lines (in feet) are shown in the areas of the ridges. No contour lines above 40 feet were shown and no contour lines were drawn on the plains region to the west or to the east of the ridges. (The ore outline and values were furnished through the courtesy of ITT Rayonier, Inc., and the Pennsylvania Glass Sand Corporation). Modified from Pirkle et al., 1984.
Figure 3. General setting of the Yulee heavy-mineral deposits (shaded areas). Amelia Island, a present-day barrier island, occurs along the Atlantic Coast in the same general latitudes of the Yulee heavy-mineral concentrations. Vast salt marshes separate Amelia Island and the Yulee ridges. Modified from Pirkle, et al., 1984.
north, east, and southeast of the town of Yulee in northeastern peninsular Florida (Figs. 1, 2, and 3). These deposits have been described by Pirkle et al., 1984, and the water table within the sand ridges containing the heavy-mineral deposits has been characterized by Spangler et al., 1989. These heavy-mineral deposits are bordered on the north by the St. Marys and Bells rivers, and on the south by the Nassau River. Ore grade heavy-mineral concentrations occur from the Crandall Ridge on the west to ridge areas just seaward of Chester Road on the east (Fig. 2).

Correlation with Pamlico Shoreline Hoyt and Hails (1974) consider the regions in which the Yulee ridges are present to be parts of a Pamlico barrier island complex of Pleistocene age. White (1970, page 34) points out areas in eastern Florida where the Pamlico shoreline, along which these Yulee ridges occur, truncates older relic beach ridges. From such studies of the Pamlico shoreline and from elevations of fossil burrows of a marine decapod, Ophiomorpha, the ridges containing the Yulee heavy-mineral concentrations can be correlated with the crestal area of the Pamlico transgression.

Size of Deposit and Heavy Mineral Suite The Yulee heavy-mineral concentrations contain an estimated 1.2 million short tons of available TiO$_2$ product (ilmenite, leucoxene, and rutile) at approximately 66 percent TiO$_2$. These concentrations begin at the land surface and extend downward to an average depth of about 9 feet. The heavy-mineral content of the sands varies from place to place but averages from 3 to 4 percent.

The most abundant heavy minerals in the Yulee deposits are ilmenite, zircon, epidote, rutile, sillimanite, stauroite, leucoxene, tourmaline, kyanite, and hornblende. Small amounts of garnet and monazite are present. Of these minerals ilmenite, leucoxene, rutile, and zircon are the most important ore minerals. The titanium minerals (ilmenite, leucoxene, and rutile) make up about 60 percent of the heavy minerals, and zircon constitutes approximately 14 percent of the heavy minerals (volume percent corrected for size).

Grade of the Ilmenite The ilmenite in the Yulee deposits has been upgraded through weathering, but not as much as has the ilmenite in some of the other heavy-mineral concentrations of the Coastal Plain. The Yulee ilmenite contains an average of about 59 percent TiO$_2$. In contrast, the average content of the TiO$_2$ in the ilmenite of the heavy-mineral deposits on Trail Ridge (the Trail Ridge and Highland ombodies, Fig. 1) is about 64 percent.

Ridges Containing Yulee Heavy-Mineral Deposits The Yulee ridges, parts of a Pamlico barrier island complex, have a counterpart in Amelia Island, a more modern barrier island in the same general latitudes along the present Atlantic Coast. Amelia Island is located just south and somewhat east of the mouth of the St. Marys River (Fig. 3). Where large rivers enter oceans, coastal current patterns may be locally interrupted and river sediments added to the ocean waters. At such localities, heavy minerals may be concentrated in favorable environments in a downdrift direction from the river mouthes. Amelia Island, in a downdrift direction from the mouth of the St. Marys River, contains substantial heavy-mineral deposits of ore grade. Estimates of approximately 320,000 short tons of TiO$_2$ in sands containing an average of 4 percent heavy minerals have been determined for these deposits.

The Yulee heavy-mineral concentrations in beach ridges of Pamlico barrier islands, like the concentrations of Amelia Island, accumulated in favorable environments in a downdrift direction from a point where the St. Marys River entered the Atlantic Ocean. Just as Amelia Island occurs somewhat to the east of the present mouth of the St. Marys River, the Pamlico barrier islands containing the Yulee deposits were formed a little to the east of the mouth of the ancestral St. Marys River. As the ridges of the Pamlico barrier islands were forming, heavy minerals were concentrated by current, wave and wind action.

CONCLUSIONS

The Yulee heavy-mineral concentrations are located in northeastern peninsular Florida on an ancient barrier island complex, downdrift from a
point where the ancestral St. Marys River entered the Atlantic Ocean in Pamlico time. The correlation of the Yulee heavy-mineral ridges with the Pleistocene Pamlico shoreline can be drawn from considerations of topography and elevations of fossil burrows of a marine decapod, *Ophiomorpha*.

**REFERENCES**


CONTRIBUTION 13
PALYNOLOGY AND PALEOECOLOGY OF REIDS BLUFF
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INTRODUCTION
The economic and stratigraphic significance of Quaternary deposits of the Atlantic Coastal Plain of Georgia and Florida has been recognized for many years as attested by numerous investigations of the sediments. Only recently, however, has a combination of radiometric analyses, paleontological investigations, and stratigraphic/sedimentological evaluations been employed to construct a synergistic paleoecological framework for this part of the Coastal Plain.

The present study presents the results of a variety of different analyses undertaken to provide information to help understand the geological history of sediments near the mouth of the St. Marys River. Results from studies of Reids Bluff and Bells Bluff, in particular have important implications for an understanding of the depositional histories of more inland localities and of coastal sites such as Cumberland Island. The data and interpretations presented here are the result of several years of investigations, and constitute just one part of a very extensive effort to understand the paleoecology of parts of the Atlantic Coastal Plain.

LOCATION OF THE STUDY AREA
Reids Bluff, along the St. Marys River, lies in Nassau County, Florida, approximately 6 linear km upstream from the waterfront at St. Marys, Georgia. The bluff is quite prominent as seen from the river, and appears clearly on the St. Marys, Ga.- Fla. 7.5-Minute Series Orthophotomap, where it lies near the boundary of Township 4 North and Township 3 North. The bluff has been eroded by a southward-advancing meander of the St. Marys River. Continued erosion, slumpage, and removal of the bank materials keep the exposures relatively fresh, and result in a nearly vertical scarp which stands a little more than 19 m above river level at its highest point. Thicknesses of sedimentary units as seen along the cliff face may vary from one time to another because of the continual slumpage and the lens-like nature of the units. A striking panoramic view of the meander and coastal marshes which lie at the foot of Reids Bluff appears in the fold-out on pages 38 and 39 of the April, 1992, issue of National Geographic magazine (see Conniff, 1992). About 2.5 km east of Reids Bluff, and lying along the west bank of Bells River lies another bluff, Bells Bluff. This bluff also rises almost vertically from river level. A detailed description of the Bells Bluff sediments and their associated biogenic structures is presented in Howard and Scott (1983).

SEDIMENTS OF REIDS BLUFF
The geology of Reids Bluff and Bells Bluff has been the subject of several recent investigations, including those of Kussel and Jones (1986) and Pirkle et al. (1991). The latter authors provide detailed descriptions of the deposits comprising Reids Bluff, as well as photographs that show an ancient tree stump and the sequence of sediments that compose the bluff.

The tree stumps at Reids Bluff have been identified as those of cypress (Taxodium distichum). At the level of the river and for about 2 m above it there is a layer of loose quartz sand. This sand contains some of the fossil tree trunks and root systems. The presence of upright trunks with their spreading, knee-bearing root systems intact provides conclusive evidence that the trees are in place. The stumps are, furthermore, associated with an organic-rich horizon which may be soil of the ancient swamp forest. Some ghost-shrimp burrows (Ophiomorpha) are visible in the sands which lie in and around the trees.

Blue-gray clayey sediments about 4 m in thickness lie just above the quartz sands containing the fossil trees; these clayey sediments contain lenses of oyster shells (Crassostrea sp.). At one site two lenses of oyster shells, one above the other, were noted. The lower lens, from 0.6 to 1.8 m in thickness, occurs directly over cypress stumps still
in growth position (see Fig. 7, Pirkle et al., 1991). The upper lens at this site is approximately 0.3 m in thickness. In places the oyster lenses contain large, articulated shells of the clam *Mercenaria* sp. The clayey sediments containing the oyster lenses are overlain by about 13 m of quartz sand. Pirkle et al. (1991) note that at some sites the upper parts of these loose sands are wind-blown and were deposited as sand was carried up over the present cliff face to accumulate as subtle ridges bordering the St. Marys River.

**AGE AND STRATIGRAPHIC LOCATION OF REIDS BLUFF AND BELLS BLUFF**

The age of the deposits at Reids and Bells Bluffs has been a matter of conjecture for a number of years. Huddleston (1988) provides a relatively recent analysis wherein he reintroduces, redefines, and revises the concept of the Satilla Formation which he considers to be of late Pleistocene and Holocene age. In Huddleston’s classification scheme, the “Satilla Formation... includes the Pamlico Formation of Cooke (1943); the Pamlico Formation; Princess Anne, and Silver Bluff formations of Hails and Hoyt (1969); and the Pamlico, Princess Anne, Silver Bluff, and Holocene shoreline complexes of Mann (Georgia Geological Survey, 1976).” In other words, the Satilla Formation includes the lower four beach ridge and swale complexes which are familiar to those people who work on the Outer Coastal Plain of Georgia. These have been mapped by a number of individuals, but perhaps are most clearly shown on the map entitled *Georgia’s Coast - Wetlands and Geologic Resources*, printed by the Georgia Department of Natural Resources (1976).

According to Huddleston, “The best and most instructive exposures of the Satilla Formation in the type area are at Roses and Bells Bluffs along the Bells River, a tidal distributary of the St. Marys River, and at Reids Bluff on the lower St. Marys River... These bluffs are all in Nassau County, Florida, across the St. Marys River from St. Marys, Georgia. Roses and Bells Bluffs, which form one continuous bluff, and Reids Bluff are here designated reference localities of the Satilla Formation.” Markewich et al. (1992) take some issue with the “single formation” concept proposed by Huddleston (interestingly, Huddleston is one of the authors on that paper). The actual stratigraphic assignment of the lower four beach ridge and swale complexes on the Georgia coast remains to be determined to everyone’s satisfaction. In any case, Huddleston, based upon his correlation of the strata at Reids Bluff with the Satilla Formation, considers the Reids Bluff sediments to be of late Pleistocene and Holocene age. Seven samples of organic material have been collected and radiometrically dated. The dates corroborate the contention that Reids Bluff is of late Pleistocene age (Table 1).

The range of ages for all samples from Reids and Bells Bluffs leads to a number of conclusions.

1) The dates indicate that the deposits are of late Pleistocene age. Different mollusc genera and wood samples from different locations on the bluffs represent a reasonably broad sampling of materials. The ages obtained for these samples offer a strong argument for a late Pleistocene age for the sediments exposed at the bluffs.

2) Another significant interpretation drawn from the range of ages of the trees is that the “cypress forest” is actually “cypress forests.” The ages for the trees range from >38,130 to 25,830 ± 373 years, so for at least twelve thousand years trees grew at the site. At first glance one might be able to see only one horizon of trees at Reids Bluff, but upon closer investigation it appears that the stumps actually occupy at least two horizons (each with its own soil). This interpretation is consistent with the range of ages determined for the trees by radiocarbon analysis. Evidently a sequence of forests grew at this site during the Late Pleistocene.

3) The *Anadara* shells from Bells Bluff are approximately the same age as the *Mercenaria* shells from Reids Bluff (37,395 ± 2155 and 36,030 ± 610, respectively). The *Anadara* site lies seaward of Reids Bluff, and contains a normal marine mollusc assemblage (including articulated Anatina, disarticulated but well preserved *Dinocardium*, and *Polinices*). The *Mercenaria* site, by contrast, looks more like a brackish lagoon/salt marsh deposit, where large, articulated clams lie mingled with large, articulated oysters. The blue clayey matrix
Table 1 — C¹⁴ Age Dates

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Location</th>
<th>Lab</th>
<th>Sample Material</th>
<th>C¹⁴-Date YBP</th>
<th>Date Collected</th>
</tr>
</thead>
<tbody>
<tr>
<td>R-1</td>
<td>Reids Bluff</td>
<td>Univ. of GA</td>
<td>Charcoal and charwood from 3-3.6 meters below land surface.</td>
<td>405±55</td>
<td>Oct. 1987</td>
</tr>
<tr>
<td>X-1</td>
<td>Reids Bluff</td>
<td>Univ. of GA</td>
<td>Wood, pulpy wood, extracted from the base of Reids Bluff.</td>
<td>34,290±155</td>
<td>Sept. 1987</td>
</tr>
<tr>
<td>Mercenaria Shell</td>
<td>Reids Bluff</td>
<td>Beta Analytic</td>
<td>Mercenaria shell from shell layer above trees. Taken from about 2 meters above the river.</td>
<td>36,030±610</td>
<td>Feb. 1988</td>
</tr>
<tr>
<td>Cypress</td>
<td>Reids Bluff</td>
<td>Beta Analytic</td>
<td>Wood sample taken from beneath oyster lens. Stump at river level.</td>
<td>&gt;38,130</td>
<td>Feb. 1988</td>
</tr>
<tr>
<td>Reids Bluff Newsome</td>
<td>Reids Bluff</td>
<td>Univ. of GA</td>
<td>Wood sample extracted from near base of bluff (about 1 meter above the base of bluff). Oyster lens was not present.</td>
<td>29,222±597</td>
<td>July 1992</td>
</tr>
<tr>
<td>Reids Bluff 2nd Upper Tree</td>
<td>Reids Bluff</td>
<td>Univ. of GA</td>
<td>Wood sample extracted from about 3.1 meters above the base of the bluff.</td>
<td>25,830±373</td>
<td>July 1992</td>
</tr>
<tr>
<td>Anadara Shell</td>
<td>Bells Bluff</td>
<td>Univ. of GA</td>
<td>Sample collected from fossil lens on Bells Bluff just above river level.</td>
<td>37,395±2155</td>
<td>Aug. 1989</td>
</tr>
</tbody>
</table>

which encloses the shells is, furthermore, the type of sediment one might expect to accumulate in such an environment. It appears reasonable to assume that the Bells and Reids Bluff assemblages may represent different mollusc-bearing facies which lay on a shoreline which had advanced landward and over the tops of the oldest trees at Reids Bluff. The Anadara assemblage would represent the shoreface community, and the Mercenaria assemblage would represent the back-barrier, or at least back-beach community.

PALYNOCLOGICAL COMPOSITION OF REIDS AND BELLS BLUFF SAMPLES

Ten samples of various kinds have been collected from Reids and Bells Bluffs and have been palynologically analyzed. The samples are as follows:

1. Reids Bluff, lower cypress forest soil - orange sand and clay with abundant, relatively large (1 x 2 cm in one case) twigs or roots; collected from near the base of one of the lower stumps at Reids Bluff in July, 1992.

2. Reids Bluff, upper cypress forest soil - several small (4 x 4 cm), thin (1 cm) blocks of sand which are held together by brown rootlet-like structures; collected from near the base of one of the upper stumps at Reids Bluff in July, 1992.

3. X-1 - a large piece of moist, humified wood with clayey sand adhering to its surface; collected September, 1987.

4. Reids Bluff, sediment from around a tree - clayey sand collected from sediments immediately surrounding a freshly exposed stump low on the bluff; collected February, 1988.

5. Mercenaria shell-infilling - sediment washed from a Mercenaria shell collected in the upper prominent shell horizon at Reids Bluff; the shell was subsequently sent for radiocarbon dating; collected February, 1988.

6. Dinocarodium in-filling - sediment found within an inverted Dinocarodium valve preserved in the shell-rich layer at Bells Bluff; collected August, 1989.

7. Dinocarodium in-filling - sediment found within a Dinocarodium shell collected at Bells Bluff, April, 1991.

8. Bells Bluff shrimp burrow - a clayey infilling taken from a ghost shrimp burrow found within the shell bed at Bells Bluff; collected April, 1991.


SAMPLE PROCESSING TECHNIQUES

A small amount of each sample was put in a beaker and dispersed with water. Shell-bearing samples, such as those washed from within the mollusc shells were put in a sieve and the fine sediment washed from the shells with distilled water. In every case, the sediment was further processed as indicated:

1. It was covered with 10% HCl until all reaction ceased; this removed carbonates.
2. The residues were washed with distilled water, centrifuged, and covered with 52% HF for several days; this removed silicates.
3. The samples were then washed free of HF and boiled in 10% KOH for 10 minutes; this removed soluble humic substances.
4. The remaining sediment was washed free of soluble organic substances and mixed with a 50:50 mixture of distilled water and glycerin jelly.

Several slides were prepared from the glycerin jelly suspension of insoluble organic matter; enough residue was placed on each slide to cover an area of 22 x 22 mm. The residues were of variable character and were composed of microscopic particles of many probable origins. The slides were observed at 400X magnification using a Jenaval research photomicroscope. Slides were systematically observed using a mechanical stage with X-Y movement. Pollen grains and dinoflagellate cysts were observed and counted until, whenever possible, at least 200 identifiable pollen grains had been seen. Two samples had so few grains that 200 could not be identified. The results were tabulated, and simple percentages of pollen taxa were computed.

RESULTS OF ANALYSES

Among the ten samples listed above, only two had so few pollen grains on a slide (the *Polinices* and *Mercenaria* samples) that two slides had to be observed. Actually, the *Polinices* and *Mercenaria* samples had insufficient pollen on two slides to allow the identification of 200 grains.

Figures 1 and 2 illustrate the relative abundance, by percent, of 16 taxa found in amounts equal to, or greater than, 1% in at least one of the ten samples.

Table 2 — Taxa which did not constitute at least 1% of pollen and spores in any sample.

<table>
<thead>
<tr>
<th>Taxon</th>
<th>Taxon</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alnus</td>
<td>Betula</td>
</tr>
<tr>
<td>Lycopodium</td>
<td>Osmunda</td>
</tr>
<tr>
<td>Sabatia</td>
<td>Salix</td>
</tr>
<tr>
<td>Castanea</td>
<td>Cephalanthus</td>
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<tr>
<td>Cyperaceae</td>
<td>Cyrrilla</td>
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<td>Sagittaria</td>
<td>Sphagnum</td>
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<tr>
<td>Fagus</td>
<td>Fraxinus</td>
</tr>
<tr>
<td>Tilia</td>
<td>Vitis</td>
</tr>
<tr>
<td>Gordonia</td>
<td>Juniperus</td>
</tr>
<tr>
<td>Woodwardia</td>
<td></td>
</tr>
</tbody>
</table>

Among all of the 35 taxa encountered in these samples, one can recognize a fairly typical Southeastern flora, with one notable exception, *Tsuga* (hemlock) whose importance is discussed below. The other plants indicate a community of freshwater trees, shrubs, and herbs which could be found almost anywhere along the current east coast of the United States south of Chesapeake Bay. Some taxa, notably *Cyrrilla* (ti-ti), *Gordonia* (loblolly bay), *Myrica* (wax myrtle), and *Taxodium* (cypress) suggest a warm, humid climate, probably very similar to that which currently exists in southeastern Georgia. This interpretation stands in rather interesting contrast to what the presence of *Tsuga* indicates. *Tsuga*, the common hemlock of the Blue Ridge is present in the Reids and Bells Bluff pollen flora, and its appearance really comes as a surprise. Both paleosol samples, the shrimp burrow in-filling sample, and the blue-gray clay sample from Bells Bluff contained measurable amounts of hemlock. It is a minor constituent, to be sure, but it is still more abundant than the pollen of plants that one might expect to find there, based upon the general paleoecological reconstruction that has been drawn (e.g. *Cephalanthus* (button bush), *Cyrrilla*, *Betula* (birch), or *Alnus* (alder)). As more coastal pollen samples have been observed from the Georgia, South Carolina, and North Carolina coasts it has become clear that *Tsuga* left
a distinct, though subtle record in the pollen floras. Whether or not the hemlock trees grew near the sites of deposition is impossible to tell at present. The pollen grains are small, buoyant, and are easily transported by air and water and so could have been carried to the sites of deposition from distant forests. If we could find *Tsuga* needles or its very distinctive small cones preserved in the sediments we would have the best evidence for the trees’ actual presence on the southern coastal plain. For now, it seems plausible that hemlock could have extended its range down across the Piedmont, perhaps as far as the Outer Coastal Plain, at a time when general climatic cooling was significant. At 30-34,000 years ago, in Georgia, that may certainly have been the case. During the course of analyzing the Bells Bluff and Reids Bluff samples for their pollen content, and then relating the results to the presence of pyrite framboids, dinoflagellate cysts, and microfossil tests, another interesting discovery was made concerning four otherwise unremarkable pollen types. Relative to scores of samples which come from dozens of sample sites in the Southeast, there are four taxa which appear together in the Reids and Bells Bluff samples and which are most abundant as a group in marine shoreline samples. Those pollen are *Pinus* (pine), *Quercus* (oak), *Liquidambar* (sweet gum), and *Carya* (hickory/pecan). Figures 1 and 2 illustrate the relative abundances of these four genera in the ten samples from Reids and Bells Bluffs.

Bearing these histograms in mind, consider the following distributional characteristics of the genera, as published by Rich (1985) (his Figures 9 and 10) in his detailed analysis of samples from organic-rich sediments and an ancient brown coal buried beneath Trail Ridge:

1) *Pinus* is typically less than 20% in Trail
Figure 2. Relative abundances of pollen in the 20-90% range, Reids Bluff and Bells Bluff
Ridge samples in general, is always less than 20% in the TR2 core and is always less than 10% in core TR1.

2) *Quercus* is always less than 10% in Trail Ridge samples in general, and in the TR1 and TR2 core samples is always less than 10%.

3) *Carya* and *Liquidambar* are always less than 1% in Trail Ridge samples in general; Rich (1985) states that *Carya* and *Liquidambar* are always less than 1% in the TR1 and TR2 core samples.

The unusually high percentages of these four pollen taxa in the Reids and Bells Bluff samples is very likely to be more than mere coincidence. While their true value as coastal/maritime forest indicators remains to be evaluated, there is strong reason to suppose they might be useful in that regard. Rich and Pirkle (in press) have discussed that possibility. The following information is taken in modified form from their text.

Brush (1988) presents the results of her work in establishing pollen abundance gradients in the Potomac estuary and Chesapeake Bay. She particularly mentions the fact that the gradients of plants of *Pinus taeda* (loblolly pine), *Liquidambar*, and *Carya* adjacent to the estuarine tributaries “...are similar to their pollen gradients in surface sediments. Pollen distributions suggest that pollen, and hence fine sediment, are not transported far in the estuary before being deposited.”

Still further evidence that *Carya, Liquidambar*, and *Quercus*, may be reliable indicators of a marine coastal environment has been produced by Darrell (1989). He, and Pratt *et al.* (1989) have jointly undertaken a study of pollen and vertebrate fossil remains from the Early Miocene Mark’s Head Formation in southeastern Georgia. Pratt *et al.* (1989) have identified a small but diverse assemblage of vertebrates which “... is consistent with a subtropical nearshore marine depositional environment...” for the Mark’s Head Formation. Darrell observed pollen/spores from precisely the same sediments and found that *Quercus, Ulmus*, Chenopodiaceae, Gramineae, and Cyperaceae dominate the assemblage, with lesser quantities of *Carya, Liquidambar*, and fern spores. The palynological composition of the sample led Darrell to postulate “... a very nearshore marine environment...” for his Mark’s Head samples. Darrell’s data, coupled with those of Brush and the independently drawn conclusions which are based on the study of the Bells Bluff and Reids Bluff samples all point to the same conclusion: in southeastern North America high relative abundances of *Quercus, Pinus, Carya*, and *Liquidambar* may characterize deposits of shallow marine or brackish water sediments.

CONCLUSIONS

Prominent bluffs occur along the St. Marys River near its mouth in northeastern Florida. The bluffs are composed of stratified sediments including cypress-stump-bearing sands, oyster and clam shell beds, blue-gray clays, and eolian sands. Radiocarbon dates of wood and shells reveal Late Pleistocene ages (~25,000 to > 38,000 yrs. BP) for all the organic remains. Stratigraphic relationships further illustrate a classic transgressive/regressive sequence wherein a freshwater forest is overlain by brackish/marine strata which are in turn overlain by the Holocene sand ridge and forest.

Ten samples of sediment from Reids and Bells Bluffs were prepared for palynological analysis. Most of the samples contained abundant pollen/spores of various types of plants which are common to the southeastern coastal plain of the United States. While the general aspect of the pollen diagrams suggests a flora and, consequently, a climate which is very similar to that which prevails in the area now, the presence of hemlock pollen (*Tsuga*) suggests that these more typically cool climate trees may have grown on the coastal plain 30,000 years ago. The ten samples also contain a common association of oak, pine, hickory/pecan, and sweet gum pollen. This, in addition to the presence of pyrite framboinds, dinoflagellate cysts, foraminifera, and the close stratigraphic proximity of shell beds all suggest that the oak-pine-hickory/pecan-sweet gum assemblage is diagnostic of nearshore, marine deposits.

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Contribution 14
Stratigraphy and Facies Sequences at Bells/Roses Bluff, Pamlico
PaleobARRIER, NortheasTERn Florida
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PURPOSE
The purpose of this contribution is to provide a
stratigraphic framework for reviewing facies
interpretations at Roses and Bells Bluffs. Bells
Bluff forms an east-west profile or depositional dip
section through the Pamlico paleobARRIER (Fig. 1).
Roses Bluff extends parallel to the Pamlico
paleobARRIER and represents a depositional strike
section. These Bluffs have been previously
described (Scott, 1976; Howard and Scott, 1983;
Kussel, 1984; Pirkle et al, 1984; Kussel and Jones,
Bluff as a prograding beach to offshore sequence
and provided detailed facies interpretation on
individual units. At this point in time, we will not
reinterpret the sections, but will provide some new
observations and descriptions to stimulate on-site
discussions.

Based on regional correlations, Hoyt and Hails
(1974) considered Bells Bluff to be at least 110,000
years old and interpreted it as a prograding barrier
sequence that formed when sea level stood 8 m
higher than present. This conclusion was based on
the highest occurrence of Ophiomorpha nodosa, a

Figure 1. Map showing locations of Bells, Roses, and Reids Bluffs profiles and localities.
trace fossil considered to be the fossil analog of the burrow of the shrimp *Callianassa* (Frey *et al.*, 1978). Recently obtained radiometric dates of wood and shells from nearby Reids Bluff indicate a younger late Pleistocene age of about 25,000 to > 38,000 years B.P. (see Rich and Pirkle, this guidebook). Amino acid ratios on mollusks from Reids Bluff suggest that the fossiliferous unit sampled is at least 250,000 years in age (Wehmiller *et al.*, 1988).

Howard and Scott (1983) believe that a complete vertical section of offshore through probable beach dunes, shoreface, upper and lower foreshore, and back shore facies is preserved here. According to their model, the lower 4.5 m of bluff is equivalent to the upper offshore facies and is composed of muddy sand with occasional mud laminae and discontinuous coarse sand lenses. The sediments of this lower section are bioturbated and few identifiable burrows are present. The degree of bioturbation decreases upward. Abundant identifiable burrows and discontinuous primary physical structures characterize the shoreface facies from 4.5 to 6.5 m. A density break in the abundance of *Ophiomorpha* occurs at 6.5 m (approximately the elevation of Pamlico mean low water) and primary physical structures become dominant above 6.5 m. Few *Ophiomorpha* are found above Pamlico mean sea level (7.5 m).

**BELLS BLUFF: EAST-WEST (DIP) SECTION THROUGH THE PALEOBARRIER**

Bells Bluff is very difficult to describe because it is poorly exposed. The basal two meters of the outcrop is covered with sandstone boulders that have fallen from the tops of the bluff. The sandstones in the upper part of the bluff are cemented with dark brown humate which obscures physical and biogenic sedimentary structures. Major bedding surfaces were traced from a photomosaic at the east end of Bells bluff and two localities were described (Fig. 2).

At locality 1, the lowest exposed unit consists of green, micaceous, fine silty sand which is extensively bioturbated (Fig. 3). It grades in color upward into tan bioturbated sand which is overlain by an unconformity. The unconformity is marked by a line of pea pebbles and quartz granules and a reworked zone about 20 cm thick. A similar contact occurs at about the same elevation at the top

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Figure 2. Tracing from photomosaic showing bedding and stratigraphy at east and west ends of Bells Bluff (Profile A). Locations of localities 1 and 2 are shown on Figure 1.
of a fossiliferous unit along segments of Roses Bluff.

Above the unconformity are multiple beds of sand containing *Ophiomorpha nodosa* burrows (Fig. 3). Each bed (10 to 60 cm thick) becomes slightly muddier and more bioturbated near its upper boundary. Many of the burrows have slightly muddy linings (Fig. 4). A distinct, dark-brown, humate-cemented sandstone overlies these burrowed sands (Fig. 4). The humate-cemented rocks also contain *Ophiomorpha nodosa* burrows and traces of parallel laminations and large-scale cross-stratification. Humate-cemented sandstones form resistant traceable units along Bells Bluff.

Locality 2 occurs at the west end of Bells Bluff, where only the upper section is well exposed. Here, humate-cemented sandstones are absent and we have an opportunity to view uncemented analogous deposits. Unconsolidated quartz sands with distinct *Ophiomorpha nodosa* burrows are overlain by parallel-laminated quartz sand with shell hash layers (Fig. 5). The shells, however, which have a *Mulinia*-like or *Donax*-like size and form have been dissolved away and occur only as secondary bivalve moldic porosity (Fig. 6). The parallel-laminated sands also contain *Ophiomorpha nodosa* burrows.

*Donax variabilis* is ubiquitous along modern coastlines and the following information was obtained from Kussel and Jones (1986). *Donax*
**Variabilis** is a rapid burrower in shifting beach sands (Frey, 1970). *Donax* has been identified in the foreshore region of the Georgia and Florida coasts (Frey, 1970) and has been found in close association with *Callianassa islagrande* in the bar trough system of the foreshore to offshore region (Hill and Hunter, 1976).

**ROSES BLUFF: NORTH-SOUTH (STRIKE PARALLEL) THROUGH THE PALEOBARRIER.**

Figure 7 shows tracings of stratification from a photomosaic constructed for a limited segment of the strike-parallel section along Roses Bluff (Fig. 7). Several sections were measured and described in detail (Fig. 8); many of the contacts were mapped in the field. The cross-sectional profile for Roses Bluff will be referred to as Profile B. The location of profile B is shown in Figure 1.

An overview of Roses Bluff (see Fig. 7) shows that: (1) a fossiliferous unit occurs at the base of the bluff; (2) an unconformity occurs at the top of the fossiliferous unit; (3) clay-rich layers form bounding surfaces that can be traced for great distances along the outcrop; (4) lens-shaped sand bodies pinchout laterally; (5) sandbodies internally exhibit lateral accretion sets; and (6) horizontally bedded sands can be traced for great distances along the upper section of the outcrop.

At the south end of Roses Bluff, between Localities 3 and 4, three units are traceable along the lower third of the outcrop. A fossiliferous unit...
Figure 7: Tracing of photomosaic showing bedding and stratigraphy for a limited segment of the strike-parallel section along Roses Bluff. This profile is referred to as Profile B and its location is shown on Figure 1.
Figure 8. Graphic logs for vertical sections at Localities 3, 4, and 6, Roses Bluff.
is exposed at the base of the outcrop. At Locality 3 it consists of olive-gray silty sand and its top is cemented with calcite. The cemented zone is only 5 cm thick and dissolution has removed primary shell material from it, leaving secondary molluscan moldic porosity.

Kussel and Jones (1984) and Kussel (1984) collected two bivalves (Divaricella dentata and an unidentified mactrid) and two gastropods (Terebra dislocata and an unidentified naticid) from this unit. Their interpretations follow. These are small delicate shallow marine forms, moderately well preserved and largely unbroken, implying minor transport. Crassostrea virginica is also present but the valves are typically transported implying substantial transport, perhaps from an adjacent estuary.

Abundant quartz granules occur in conjunction with the cemented zone at the top of the fossiliferous unit (Locality 3). The “marker bed” of granular quartz was mapped northward to locality 4. It is interpreted as an unconformity. The unconformity is overlain by extensively bioturbated, slightly silty fine sand with clay-lined burrows. Because of slumping, it is not possible to map the fossiliferous unit, the unconformity, or the bioturbated sand north of Locality 4.

Dark-gray, clay-rich layers form bounding surfaces that can be traced for great distances along Roses Bluff (Fig. 7). At the south end of Roses Bluff, two of these clay-rich layers occur. Both have maximum thicknesses of about 0.5 m and are internally tidally bedded. Internal stratification consists of wavy, flaser, and lenticular bedding. The unit includes inclined layers locally that indicate lateral-accretion bedding. Figure 9 shows an example of the upper “clay” at Locality 3. Here, extensively burrowed sand is overlain by lenticularly bedded clay that weathers into a fissile unit. The fissile bed is sandwiched between two layers of plastic clay. Figure 7 shows that this upper “clay unit” drops in elevation to the north and probably joins the lower clay unit just north of Locality 4. The single resultant clay body is mappable along the entire length of Profile B. Figure 10 shows a macropore about 0.5 m in diameter that developed in sand above the “upper” clay bed. Macropores are preferential flowpaths for groundwater flow; in this case, a conduit-like feature along an impermeable boundary.

The dark-gray, tidally bedded clays are overlain by sandbodies which pinchout laterally. At the south end of the Roses Bluff between localities 3 and 4, a distinct sand body occurs between the lower clay and the upper clay. This sand body has a sharp base with overlying large-scale cross-stratification. Mudball ripup clasts occur along inclined foresets or lateral accretion sets. At Locality 4 the cross-stratified unit is overlain by the upper clay. At Locality 3, the large-scale, cross-bedding is replaced upsection by parallel-laminated to cross-stratified sand and then Ophiomorpha-burrowed sand.

Near the north end of Profile B, the tidally
bedded layer is overlain by a second, well defined sand body that pinches out to the south near Locality 3. Figure 11 shows that this sand body (near localities 5 and 6) sharply overlies dark-gray plastic clay. The sand body contains inclined layers demarcated by mudball ripup clasts that dip southward. These inclined layers are interpreted as lateral-accretion bedding, indicating channel migration towards the south. The sand body locally thickens and contains distinct *Ophiomorpha* burrows. It is overlain by extensively burrowed sand.

Near the north end of Profile B (Fig. 7, Localities 5 and 6), the upper section consists of laterally-continuous horizontal beds. These beds consist of alternating layers of burrowed/bioturbated sand and parallel-laminated/cross-stratified sand. Hardpan layers form in association with bedding planes at contacts between unbioturbated and bioturbated units. Figure 12 shows a hardpan that has developed at the contact between a parallel-laminated sand and an *Ophiomorpha*-burrowed sand. The hardpan forms a ledge at the top of the
parallel-laminated sand.

At the top of the sequence (Fig. 7, near localities 5 and 6) are black sands that occur as inclined layers that dip to the north. These beds are associated with the Yulee heavy-mineral sands. Figure 13 shows a sketch of these layers from outcrop. The blackest layers are slightly bioturbated and form a hardpan at the top of an inclined set of parallel laminated sand. They are probably cemented with some kind of black humate. Sets of parallel-laminated sand (Fig. 14) alternate with beds of extensively bioturbated sand with swirling laminations (Fig. 15). The sketch (Figure 13) shows the bedding arrangement on these features and their progradation northward. Parallel laminated beds occur as inclined "foreset beds"; the bioturbated beds occur as "bottomset beds". According to Howard and Scott, (1983) these are backshore deposits.

Further north along Roses Bluff (Fig. 7, locality 7), the deposits on Profile B are replaced by thickly bedded cemented sandstones (Fig. 16). We were not able to determine facies relationships between these sandstones and the horizontally bedded deposits at the north end of profile B because of the lack of exposure. These sandstones are several meters thick and are internally cross-stratified (Fig. 17). These deposits represent a very dynamic, 

![Figure 12. Photograph showing horizontal beds near north end of Profile B (Fig. 7, Localities 5 and 6). Two units are present; a lower parallel-laminated sand and an upper Ophiomorpha-burrowed sand. A hardpan has developed at the top of the parallel-laminated bed and forms a resistant ledge.](image)

![Figure 13. A sketch of black sands at the top of Roses Bluff (Fig. 7, near Localities 5 and 6). The blackest layers are bioturbated and form a hardpan. They are probably cemented with some kind of black humate. These deposits were emplaced via progradation from the south towards the north. Sets of parallel-laminated sand are emplaced as "foreset beds" over pre-existing bioturbated "bottomset beds". The tops of the parallel-laminated beds are then bioturbated.](image)
Figure 14. Photograph of black sand deposits (Roses Bluff, localities 5 and 6). Black sands sharply overlie a burrowed sand. Small (2 mm diameter) burrows extend downward from this contact. A black, bioturbated layer overlies this contact. Above this are alternating zones of bioturbated black sand and parallel-laminated black sand. The hard pans that form in the blackest layers probably are cemented by some type of black humate cement.

Figure 15. Photograph of black sand deposits (Roses Bluff, Localities 5 and 6). Cleaned outcrop shows bioturbated black sand with swirling laminations and bioturbated fabrics. Some of swirling laminations appear to be distorted primary bedding. Elsewhere it resembles burrow backfill deposits.

Figure 16. Photograph of thickly bedded, cemented sandstones that occur further to the north along Roses Bluff (Locality 7). Individual beds are several meters thick and exhibit large-scale and small-scale cross-stratification. Burrows were not observed. The facies relationship between these cemented sandstones and the deposits on Profile B are unknown because of poor exposure.
high-energy environment such as shoals or channels associated with a sound or inlet.

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Figure 17. Photograph of thickly bedded, cemented sandstones showing large-scale and small-scale cross-stratification and absence of biogenic structures. North end of Roses Bluff near Locality 7.
Contribution 15
Stratigraphy and Facies Sequences at Reids Bluff, Pamlico Paleobarrier, Northeastern Florida
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PURPOSE
Reids Bluff (St. Marys Florida-Georgia 7.5-Minute Quadrangle) exposes sections of the Pamlico paleobarrier complex. Hails and Hoyt (1969) associated the Pamlico paleobarrier with an 8 m high stand in sea level and considered the deposits to be greater than 110,000 years in age. Recently obtained radiometric dates of wood and shells from Reids Bluff indicate a younger, late Pleistocene, age of about 25,000 to > 38,000 years B.P. (see Rich and Pirkle, 1993, this guidebook). Amino acid ratios on mollusks from this locality suggest that the fossiliferous unit sampled is at least 250,000 years in age (Wehmiller et al., 1988).

The purpose of this contribution is to provide a stratigraphic framework for reviewing facies interpretations at Reids Bluff. Reids Bluff is well exposed along most of its length and provides profile views of a strike-parallel section and a section parallel to depositional dip through the paleobarrier deposit (Fig. 1). Many workers have studied Reids Bluff (e.g. see Belknap, 1979 referenced in Wehmiller et al., 1988; Kussel, 1984; Kussel and Jones, 1986; Huddleston, 1988; Wehmiller et al., 1988; Pirkle et al., 1989). Huddleston (1988) used Reids Bluff as a reference locality for the Satilla Formation. We believe that further work will resolve the discrepancies in the

Figure 1. Map showing locations of profiles and localities for Bells, Roses, and Reids Bluffs.

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age dates, clarify facies relationships, and establish sequence boundaries.

In order to clarify facies relationships and establish sequences, we prepared a photomosaic for about 80 percent of the length of the well-exposed section at Reids Bluff. We traced bedding planes and stratigraphic layers from the strike-parallel profile (Fig. 2) and the profile parallel to depositional dip (Fig. 3) to develop cross-sections. Using the cross-sections in conjunction with field observations, several mappable units were identified. We consider our results to be preliminary and hope to benefit from discussions on the outcrop.

**REIDS BLUFF**

Well-stratified white sands lacking abundant biogenic features predominate along the strike-parallel profile (Fig. 2). Cross-stratification is the most prominent bedding type here, especially at the north end of the profile (Fig. 2). *Ophiomorpha*-burrowed sand and parallel-laminated sand are also important facies. Burrowed sand is present but not conspicuous. A white sand that locally contains Cypress stumps and *Ophiomorpha* burrows occurs at the base of the outcrop. The upper boundary of this sand is traceable in the photomosaic from Locality 9 (Fig. 3) to the north end of our strike-parallel section (Fig. 2). Along the strike-parallel profile, these basal sands exhibit large-scale cross-stratification and Figure 4 shows an example of this facies. Higher up in the section near Locality 5, a prominent *Ophiomorpha* sand occurs in which densely distributed *Ophiomorpha* burrows are preserved in relief (Fig. 2). This bed pinches out to the southwest (Figs. 2 and 3). Aeolian cross-stratification occurs in the upper section along most of the length of the outcrop (Fig. 2 and 3).

Dark bluish gray clay and sand, interpreted as a backbarrier facies assemblage, occurs along the dip parallel section (Fig. 3). It is locally fossiliferous and contains *Crassostrea* and *Mercenaria*. The unit pinches out near the east end of Figure 3 and and thickens towards the west. The backbarrier facies assemblage is tidally bedded to extensively bioturbated (Fig. 5). Lateral-accretion sets which internally exhibit wavy, flaser and lenticular bedding are observable—especially at Locality 7 (Fig. 3). Lateral-accretion sets occur as steeply inclined, clay-rich layers that dip toward the west at Locality 7. Locally, beds are slump ed and convoluted. Near Locality 8, shell hash occurs as a series of inclined beds that dip toward the west (Fig. 6). These are also interpreted as lateral-accretion deposits associated with the westward movement of the thalweg and bank of a channel. The backbarrier deposits are overlain by white, well-stratified sands, most of which are probably aeolian dune deposits. The prominent *Ophiomorpha* bed preserved near Locality 5 pinches out over the top of the backbarrier deposits near locality 6 (Fig. 3). Many of the bedding surfaces or “reflectors” appear to dip toward the east along the profile parallel to depositional dip.

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Figure 2. Tracing from photomosaic showing bedding and stratigraphy along strike-parallel section at Reids Bluff (Profile C). Location of profile is shown on Figure 1.
Figure 3. Tracing from photomosaic showing bedding and stratigraphy along section parallel to depositional dip at Reids Bluff (Profile D). Location of profile is shown on Figure 1.
Figure 4. Photograph from vicinity of Locality 4 (Fig. 2) that shows large-scale cross-stratified sands in basal sand unit. This basal sand locally contains Cypress stumps.

Figure 5. Photograph showing complex tidal bedding in backbarrier facies assemblage near Locality 10 (Fig. 3).
Figure 6. Photograph showing inclined shell hash layers near Locality 8 (Fig. 3).
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