Geologic Field Studies of the Coastal Plain in Alabama, Georgia, and Florida

Articles and field trips in conjunction with the Annual Meetings of
Southeastern Section, Geological Society of America
Southeastern Section of the SEPM
Southeastern Section of the Paleontological Society
and
Southeastern Section of the National Association of Geology Teachers

Meeting at
Tallahassee, Florida
April 1993

Editor: Stephen A. Kish

Southeastern Geological Society
Tallahassee, Florida
Guidebook 33
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Guidebook 33

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Tom Scott  
Florida Geological Survey  
903 West Tennessee Street  
Tallahassee, Florida  
32304-7700

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Mark G. Steltenpohl  
Department of Geology  
Auburn University  
Auburn, Alabama  
36849

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William Tanner  
Department of Geology B-160  
Florida State University  
Tallahassee, Florida  
32306-3026

Suggested bibliographic citation format for articles in this publication:

Author(s), 1993, Title of article: in Kish, S.A., editor, Geologic field studies of the Coastal Plain in Alabama, Georgian, and Florida: Southeastern Geological Society, Guidebook 33, p. ___.

Guidebook cover design - Gabrielle Lee, Department of Geology, Florida State University.

Photographic collage of Silver Lake, 6 miles southwest of Tallahassee, Florida. The lake is karst-related, and is situated near the Woodville Karst Plain (see article by Rupert in this volume)
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AUTHORS

Jon Bryan
Antarctic Research Facility B-160
Florida State University
Tallahassee, FL 32306-3026

Valerie Croup
Department of Geology B-160
Florida State University
Tallahassee, FL 32306-3026

Katherine Kelly Ellins
Department of Geology; 1112 Turlington Hall
University of Florida
Gainesville, FL 32611-2231

Richard H. Fluegeman, Jr. (contributing author)
Department of Geology
Ball State University
Muncie, IN 47306

David Furbish
Department of Geology B-160
Florida State University
Tallahassee, FL 32306-3026

James A. Heller (contributing author)
Department of Geological Sciences
University of Tennessee
Knoxville, TN 37996-1410

Richard A. Hisert
Department of Geology; 1112 Turlington Hall
University of Florida
Gainesville, FL 32611-2231

Todd Kincaid
Department of Geology; 1112 Turlington Hall
University of Florida
Gainesville, FL 32611-2231

Editor:
Stephen A. Kish
Department of Geology B-160
Florida State University
Tallahassee, FL 32306

Field Trip Coordinator:
William C. Parker
Department of Geology B-160
Florida State University
Tallahassee, FL 32306

Ernest Mancini
Geological Survey of Alabama
420 Hackberry Lane
Tuscaloosa, AL 35486

Gary S. Morgan (contributing author)
Florida Museum of Natural History
University of Florida
Gainesville, FL 32611

William Parker
Department of Geology B-160
Florida State University
Tallahassee, FL 32306-3026

Roger W. Portell (contributing author)
Florida Museum of Natural History
University of Florida
Gainesville, FL 32611

Frank Rupert
Florida Geological Survey
903 West Tennessee Street
Tallahassee, FL 32304-7700

Nick Tew
Geological Survey of Alabama
420 Hackberry Lane
Tuscaloosa, AL 35486

Stephen Thorne
Department of Geology B-160
Florida State University
Tallahassee, FL 32306-3026
INTRODUCTION

Surface exposures of the Cretaceous-Tertiary (Maestrichtian-Danian) contact are located across east-central Mississippi and southwestern and south-central Alabama (Fig. 1). In particular, the section exposed in a roadcut along Highway 263, southeast of Braggs, Lowndes County, Alabama, has been studied extensively (Worsley, 1974; Baum and others, 1984; Donovan and Vail, 1986; Jones and others, 1987; Donovan and others, 1988; Habib and others, 1992). The studies of this locality concluded that the exposure represents a nearly continuous section across the Cretaceous-Tertiary boundary (Fig. 2, section 5). In addition, Baum and others (1984), Donovan and Vail (1986), and Donovan and others (1988) described the interval that spans the Cretaceous-Tertiary contact exposed near Braggs as a stratigraphically condensed section associated with maximum landward transgression of the coastline during relative sea-level rise. Although the latest Maastrichtian and earliest Danian planktonic foraminiferal zones have not been recognized in the Braggs section (Cepek and others, 1968; Gibson and others, 1982), the latest Cretaceous and earliest Tertiary calcareous nannoplankton zones have been identified at this locality (Cepek and others, 1968; Worsley, 1974; Thierstein, 1981; Zemo, 1982; Habib and others, 1992). The earliest Danian planktonic foraminiferal zones have been recognized in core samples from Millers Ferry damsite, Alabama, 68 km from the Braggs locality (Liu and Olsson, 1992).

LITHOSTRATIGRAPHY

The Prairie Bluff Chalk (Fig. 3) is the youngest exposed Cretaceous formation in east-central Mississippi and southwestern and south-central Alabama and attains thicknesses of 21-38 m in parts of the study area (Monroe and Hunt, 1958; Copeland, 1968). At the type locality (Fig. 2, section 4) along the Alabama River in Wilcox County, south-central Alabama, the Prairie Bluff Chalk is 3.6 m thick (LaMoreaux and Toulin, 1959). Due to post-depositional erosion, the unit is absent in parts of Wilcox, Marengo, and Dallas Counties, Alabama (LaMoreaux and Toulin, 1959; Newton and others, 1961; Copeland, 1968). Typically, the Prairie Bluff Chalk is composed of blue-gray, fossiliferous, dense chalk and fossiliferous, glauconitic, sandy marl (Monroe, 1941; Russell and Keady, 1983). The basal Prairie Bluff Chalk usually contains coarse quartz grains, glauconite, phosphatic pebbles and fossil molds and casts and unconformably overlies the Ripley Formation (Monroe, 1941). The upper Prairie Bluff Chalk consists of blue-gray, massive-bedded, dense, silty to sandy, fossiliferous, phatic, micaceous chalk, and the lower Prairie Bluff Chalk consists of blue-gray, silty to sandy, very fossiliferous, glauconitic chalky marl.

A 15-30 cm bed consisting of quartz grains, phosphate pebbles, fossil molds and casts, shark teeth and Cretaceous macrofossils occurs 0.6-0.9 m below the top of the formation at Moscow Landing (section 2). This phosphatic macrofossil bed is also present 1.2-1.5 m below the top of the formation along Shell Creek (section 3) and 2.1-2.4 m below the top of the formation at Prairie Bluff Landing (section 4). This bed generally separates the upper Prairie Bluff Beds from the lower beds. A layer of quartz grains, phosphate pebbles, and shark teeth, which is interpreted as part of the basal unit of the overlying Paleocene Clayton Formation, is present at Moscow Landing (section 2), in the roadcut southeast of Braggs (section 5), and along Mussel Creek (section 6). This transgressive lag bed should not be confused with the Prairie Bluff phosphatic macrofossil bed observed in Sumter and Wilcox Counties.

Lenticular, discontinuous, irregularly-bedded, quartzose, fine- to coarse-grained sands which weather yellow-orange overlie the Prairie Bluff Chalk at most localities (Fig. 2). These sands were observed in the roadcut south of Lynn Creek, at Moscow Landing, along Shell Creek, at Prairie Bluff Landing, and in the streamcut along Mussel Creek. Historically, these sands have been assigned to the Paleocene Clayton Formation (Monroe and Hunt, 1958; LaMoreaux and Toulin, 1959) and have been reported by LaMoreaux and Toulin (1959) to contain Tertiary macrofossils. Reworked Cretaceous microfossils, macrofossils, shark teeth, phosphate pebbles, and chalk clasts up to 0.9 m in diameter occur in these sands. The sands fill depressions on the eroded surface of the underlying chalk and represent lowstand fill of incised valleys in the Sumter County area and of scour channels and depressions in the Wilcox County area. Paleocene microfossils (planktonic foraminifera) and/or macrofossils (e.g., Ostrea pulex beans) were recovered from the sands at Moscow Landing (section 2b), along Shell Creek (section 3), at Prairie Bluff Landing (section 4), and along Mussel Creek (section 6) indicating a Paleocene age for this unit. Assignment of the sands to the Paleocene differs from the interpretation of Donovan and others (1988), who assigned them to the Cretaceous, but is consistent with the findings of LaMoreaux and Toulin (1959) and Liu and Olsson (1992). At Moscow Landing, calcareous silts containing Paleocene microfossils conformably overlie these sands. Where the sands are absent at Moscow Landing, the calcareous silt beds conformably overlie the Prairie Bluff Chalk. The Clayton basal sands or silts are generally overlain along a sharp contact by a Clayton sandy marl.

limestone bed containing quartz pebbles, phosphate pebbles, and reworked Cretaceous fossils. In the absence of the sands and silts, this bed, which is interpreted as a transgressive lag deposit, directly overlies the Prairie Bluff Chalk. In these cases, the upper surface of the Prairie Bluff Chalk can be highly buried, and the burrows are filled from above with Clayton sediments containing Paleocene microfossils.

In the Sumter County area, southwestern Alabama, the Clayton Formation includes less than 6 m of marl, limestone, and silt, but the formation thickens to almost 61 m in Wilcox County, south-central Alabama, where it is divided into two members. The lower member, the Pine Barren Member, consists of about 46 m of glauconitic sand, calcareous silt and clay, and sandy limestone at the type locality along Pine Barren Creek, Wilcox County, south-central Alabama (LaMoreaux and Toulmin, 1959). A distinctive, very fossiliferous, glauconitic, sandy limestone unit referred to as the "Turritella rock" occurs at the top of the Pine Barren Member in the Wilcox County area. The upper member, the McBryde Limestone Member, includes about 15 m of light gray to white, sandy, argillaceous limestone in the type area near McBryde Station, Wilcox County, south-central Alabama (LaMoreaux and Toulmin, 1959).

The Clayton Formation is present throughout southwestern and south-central Alabama. In the roadcut south of Lynn Creek (section 1) and at Moscow Landing (section 2), the Clayton Formation consists of medium gray, calcareous, fossiliferous, glauconitic, sandy, argillaceous marl containing reworked Cretaceous fossils. The Pine Barren Member of the Clayton Formation along Shell Creek (section 3), at Prairie Bluff Landing (section 4), and at the Lowndes County localities (sections 5 and 6) is comprised chiefly of blue-gray, calcareous, micaceous silt.

The Porters Creek Formation conformably overlies the Clayton Formation in Alabama and Mississippi. In the Sumter County area, the Porters Creek consists primarily of 107-137 m of massive, black clay (Turner and Newton, 1971). The upper 0.9-1.8 m of the formation consists of gray to green, fossiliferous, glauconitic sand, which is recognized as the Matthews Landing Marl Member. To the east, the Porters Creek Formation decreases to 46-61 m in thickness and includes clays, marls, limestones and sands in the eastern Wilcox County area. The Matthews Landing Marl Member, however, increases in thickness eastward and is 6 m thick at its type locality on the Alabama River in Wilcox County (LaMoreaux and Toulmin, 1959).

The lower beds of the Porters Creek Formation are present in the roadcut south of Lynn Creek (section 1) and at Moscow Landing (section 2). At these localities, the Porters Creek Formation generally consists of massive, gray, blocky, calcareous, silty claystone, and marl.

PLANKTONIC FORAMINIFERAL BIOSTRATIGRAPHY

The Upper Cretaceous planktonic foraminiferal zonation used in this study was established by Pessagno (1967) and modified by Smith and Pessagno (1973). The Paleogene planktonic foraminiferal zonation employed was published by Bolli (1957, 1966) and Luterbacher and Premoli Silva (1964) and modified by Stainforth and others (1975) and Smit (1982). These zonations have been accepted
as biostratigraphic standards for warm-water areas of the world, including the Gulf Coastal Plain region.

The lower boundary of the Maestrichtian Globotruncana aegypitaca Zone, Globotruncana gansseri Subzone, Globotruncana contusa-stuartiformis Assemblage Zone (Pessagno, 1967; Smith and Pessagno, 1973) is recognized by the lowest occurrence of Guembelitria cretacea Cushman, Heterohelix glabrans (Cushman), Planoglobulina caseyae (Plummer), Pseudoguembelina excolata (Cushman), P. kempensis Esker, P. palpebra Bronnimann and Brown, Pseudotextularia deformis (Kikoine), Racemiguembelina powelli Smith and Pessagno, Globotruncana aegypitaca Nakkady, G. duwi Nakkady, G. navarroensis Smith and Pessagno, G. patelliformis Gandolfi, G. trinitadiensis Gandolfi, Rugoglobigerina hexacamerata Bronnimann, R. macrocephala Bronnimann and R. milamensis Smith and Pessagno (Smith and Pessagno, 1973).

Lower Prairie Bluff beds—that is, strata below the phosphatic macrofossil bed—and the phosphatic macrofossil bed as exposed at Moscow Landing (section 2), along Shell Creek (section 3), and at Prairie Bluff Landing (section 4), generally contain Guembelitria cretacea, Planoglobulina caseyae, Pseudotextularia deformis, Globotruncana aegypitaca, G. duwi, G. gansseri Bolli, G. stuartiformis Dalbiez, and Rugoglobigerina hexacamerata. Species diagnostic of the underlying early Maestrichtian planktonic foraminifer zonule, Rugotruncana subpenyri Zone, Rugotruncana subcircumnodifer Subzone, Globotruncana fornicata-stuartiformis Assemblage Zone of Pessagno (1967) and Smith and Pessagno (1973), are absent. Therefore, the lower Prairie Bluff beds are assigned to the Globotruncana aegypitaca Zone.

The lower boundary of the Maestrichtian Racemiguembelina fructicosa Zone, Globotruncana gansseri Subzone, Globotruncana contusa-stuartiformis Assemblage Zone (Pessagno, 1967; Smith and Pessagno, 1973) is recognized by the lowest occurrence of Planoglobulina brazoenensis Martin, Pseudoguembelina cornuta Seiglie, Racemiguembelina fructicosa (Egger), Ventilabrella multicamerata de Klasz, Globigerinoides rosebudensis Smith and Pessagno, Globotruncana conica White, G. contusa (Cushman), Rugoglobigerina reicheli Bronnimann, R. rotundata Bronnimann and Globotruncanella monmouthensis (Olsson) (Smith and Pessagno, 1973). The upper boundary of this zone is marked by the first occurrence of Abathomphalus mayaroensis (Bolli), A. intermedius (Bolli), and Pseudotextularia intermedia de Klasz, and the extinction of several species, such as Heterohelix glabrans, Globigerinoides prairiehillensis Pessagno, and Globotruncana stephensoni Pessagno (Pessagno, 1967; Smith and Pessagno, 1973).

Upper Prairie Bluff beds—that is, strata above the phosphatic macrofossil bed at Moscow Landing (section 2), along Shell Creek (section 3), and at Prairie Bluff Landing (section 4)—and the Prairie Bluff beds exposed in the roadcut south of Lynn Creek (section 1) and along Mussel Creek (section 6) generally contain Planoglobulina brazoenensis,
<table>
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<tr>
<th>Period</th>
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<th>Distribution of Key Planktonic Foraminifera</th>
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<td>Subbotina intricadensis Interval Zone</td>
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Figure 3. Upper Cretaceous and lower Paleocene lithostratigraphy and biostratigraphy in east-central Mississippi and southwestern and south-central Alabama (modified from Mancini and others, 1989).

**Racemiguembelinita fruticosa**, R. powelli, Ventilabrella multicamerata, Globigerinelloides praireihiillensis, Globotruncanca eugubina, G. conica, G. gansseri, G. stephensoni, G. stuartiformis, and Rugoglobigerina reicheli and, therefore, are assigned to the **Racemiguembelinita fruticosa Zone**.

The Prairie Bluff Chalk exposed southeast of Braggs (section 5) contains **Globotruncanca eugubina** and G. gansseri. Species diagnostic of the **Racemiguembelinita fruticosa Zone** were not recovered from this section, probably because of a combination of preservational and paleoenvironmental reasons. Therefore, the Prairie Bluff beds exposed at Braggs are assigned to the **Globotruncanca gansseri Subzone**.

No specimens of the planktonic foraminifera **Abathomphalus intermedius**, A. mayaroensis or **Pseudotextularia intermedia**, which are diagnostic of the latest Maestrichtian **Abathomphalus mayaroensis Subzone**, **Globotruncanca contusa-stuartiformis Assemblage Zone** (Pessagno, 1967; Smith and Pessagno, 1973), were recovered from Prairie Bluff beds in this study. Additionally, Prairie Bluff faunas throughout the study area lack morphologically advanced forms of **Planoglobulina brazoensis**, Pseudotextularia deformis, P. elegans, Racemiguembelinita fruticosa, Ventilabrella multicamerata, and **Globotruncanca conica**. At the Braggs section, these forms are absent. Species such as **Heterohelix glabrana**, Globigerinelloides praireihiillensis, and **Globotruncanca stephensoni**, which are not known from strata assignable to the **Abathomphalus mayaroensis Subzone**, occur in the uppermost Prairie Bluff beds throughout the study area. Either sediments representing the late Maestrichtian were never deposited in this area or sediments that might have accumulated during this time period have been subsequently removed.

The lower boundary of the Danian **Guembelitria cretacea Interval Zone** of Smit (1982) is defined by a reduction in the number of other Cretaceous species and the upper boundary is marked by the lowest appearance of **Parvularugoglobigerina eugubina** (Luterbacher and Premoli Silva). This zone is characterized by an abundance of the species **Guembelitria cretacea**.

**Basal Clayton (Pine Barren) sands** in core samples from Millers Ferry damsite contain a zone of abundance of **Guembelitria cretacea** (Liu and Olsson, 1992). The abundance of **Guembelitria cretacea** and the absence of many other Cretaceous species and the later Danian species **Parvularugoglobigerina eugubina** in the basal sands place these beds in the **Guembelitria cretacea Interval Zone**.

The lower boundary of the Danian **Parvularugoglobigerina eugubina Range Zone** of Luterbacher and Premoli Silva (1964) is defined by the lowest appearance of **Parvularugoglobigerina eugubina**, and the upper boundary is marked by the last occurrence of this species.

**Basal Clayton (Pine Barren) sands** in core samples from Millers Ferry damsite contain **Parvularugoglobigerina eugubina** (Liu and Olsson, 1992). The presence of this
species in these sands place these beds in the Parvularugoglobigerina eugubina Range Zone.

The lower boundary of the Danian Subbotina pseudobulboideis Interval Zone of Bolli (1966) is defined by the lowest occurrence of Subbotina pseudobulboideis (Plummer), and the upper boundary is delineated by the lowest occurrence of Subbotina trinidadensis (Bolli). Subbotina pseudobulboideis, S. triloculinoides (Plummer), and Globoconusa daubjergensis (Bronnimann) are characteristic of the zone, while Planorotalites compressa (Plummer) occurs in the upper part of the zone (Stainforth and others, 1975).

Basal Clayton (Pine Barren) sands exposed along Shell Creek (section 3), at Prairie Bluff Landing (section 4), and along Mussel Creek (section 6) generally contain Subbotina pseudobulboideis, S. triloculinoides, and Globoconusa daubjergensis. Also, the Clayton lower silts, limestones, and marls exposed at Moscow Landing (section 2), Pine Barren lower silts exposed at Prairie Bluff Landing (section 4), Pine Barren basal limestone bed and lower silts at Bragg (section 5), and Pine Barren lower silts along Mussel Creek (section 6) generally contain Subbotina pseudobulboideis, S. triloculinoides, and Globoconusa daubjergensis. The presence of S. pseudobulboideis and the absence of S. trinidadensis place these sand, silt, limestone, and marl beds in the Subbotina pseudobulboideis Interval Zone.

The lower boundary of the Subbotina trinidadensis Interval Zone of Bolli (1957) is delineated by the lowest occurrence of Subbotina trinidadensis and the upper boundary is marked by the lowest appearance of Morozovella uncinata (Bolli). Subbotina trinidadensis, S. pseudobulboideis, S. triloculinoides, Globoconusa daubjergensis, and Planorotalites compressa typify the zone (Stainforth and others, 1975).

Middle and upper Clayton beds exposed in the roadcut south of Lynn Creek (section 1), upper Clayton and Porters Creek beds exposed at Moscow Landing (section 2), and upper Pine Barren beds exposed at Bragg (section 5) generally contain Subbotina trinidadensis, S. pseudobulboideis, S. triloculinoides, S. trivalis (Subbotina), S. inconstantis (Subbotina), Globoconusa daubjergensis, and Planorotalites compressa. The occurrence of S. trinidadensis and the absence of Morozovella uncinata place these beds in the Subbotina trinidadensis Interval Zone. The lower Pine Barren beds along the Alabama River near Shell Creek (section 3) and Prairie Bluff Landing (section 4) were assigned by Olsson (1970) to rest within the Subbotina trinidadensis Interval Zone. The middle and upper clays of the Porters Creek Formation in southwestern Alabama and the clays and marls of the Porters Creek Formation in south-central Alabama have been placed in the Morozovella uncinata Interval Zone by Berggren (1965) and Mancini (1984). The Matthews Landing Marl Member of the Porters Creek Formation has been assigned to the Morozovella angulata Interval Zone (Berggren, 1965; Mancini, 1984).

Although the Cretaceous-Tertiary contact outcrops in east-central Mississippi and southwestern and south-central Alabama appear to represent continuous sections on the basis of the distribution of calcareous nanoplankton in these strata, evidence from planktonic foraminifers indicate that these sections are incomplete. Deposits representing the time interval defined by the Maestrichtian Abathomphalus mayaroensis Subzone are not present, indicating a hiatus of approximately 1 million years resulting from either nondeposition or erosional removal. We concur with Montgomery and others (1992) that the absence of Abathomphalus mayaroensis is not a result of paleoenvironmental exclusion. These authors have observed this species in shallow neritic Maestrichtian strata in south Texas and report that A. mayaroensis has been described from neritic Maestrichtian strata from northwestern Australia, the Sinai Peninsula, and Scandinavia. Further, no species of planktonic foraminifera restricted to the A. mayaroensis Subzone have been recovered from the Prairie Bluff Chalk in Alabama and several of the species, such as Globoburcana gansseri, Globoburcana stuartiformis, and Globigerinellinae prairiehillensis, occurring in the uppermost Prairie Bluff beds in Alabama have been reported as not surviving into the latest Maestrichtian (Smith and Pessagno, 1973; Montgomery and others, 1992).

Therefore, we believe, as did Smith (1975), that the Maestrichtian calcareous nanoplankton Nephrolithus frequens Zone of Cepek and Hay (1969) is assignable not only to the planktonic foraminifer Alabathomphalus mayaroensis Subzone but also to the upper part of the Globoburcana gansseri Subzone (Fig. 3). This biostratigraphic relationship, in which the base of the Nephrolithus frequens Zone lies within the upper part of the Globoburcana gansseri Subzone, has been documented previously in outcrop localities in western Europe (Van Hinte, 1976) and in cores drilled during the Deep Sea Drilling Project (Blow, 1971; Douglas, 1971; Hay, 1971; Smith and Poore, 1984; Percival, 1984). Recently, the presence of Micula prinsii Perch-Nielsen, a morphotype of Micula marus (Martin) that occurs in the Nephrolithus frequens Zone, has been used by some workers as an indicator of a late Maestrichtian age for strata in Alabama (Habib and others, 1992; Liu and Olsson, 1992). M. prinsii has been reported from the Bragg section (section 5) (Habib and others, 1992) and from core samples from Millers Ferry damsite, just downstream of the Prairie Bluff Landing section (section 4) (Liu and Olsson, 1992). Although calcareous nanoplankton workers have maintained that the Nephrolithus frequens Zone is present in the Prairie Bluff Chalk, they believed that beds representative of the latest Maestrichtian are absent at the Bragg section (section 5) (Worsley, 1974; Thierstein, 1981). Until the stratigraphic distribution of Micula prinsii relative to the distribution of key Maestrichtian planktonic foraminifer species is documented, the presence of the beds deposited during the latest Maestrichtian at the Brags section and in the vicinity of the Prairie Bluff Landing section will be equivocal.

SEQUENCE STRATIGRAPHY

The Cretaceous-Tertiary boundary at the Bragg section in south-central Alabama (section 5) has been described as a stratigraphically condensed section that occurs within the Clayton Formation (Baum and others, 1984; Donovan and Vail, 1986; Donovan and others, 1988). The boundary has been reported to be within a type 1 depositional sequence which is bounded by a basal type 1 unconformity in the Maestrichtian (67 Ma) and an upper type 2 unconformity in the Danian (62.5 Ma). The sequence includes Clayton (Pine Barren) incised-valley-fill, transgressive, condensed-section and highstand-regressive deposits (Baum, 1986; Donovan and others, 1988). The Prairie Bluff-Clayton formational
contact represents the base of the sequence, and the contact is described as disconformable, with Clayton sediment-filled channels scoured into the underlying Prairie Bluff Chalk. The contact of the "Turritella rock" beds of the Pine Barren with the underlying part of the Pine Barren is also disconformable and represents the top of the sequence (Fig. 4). The interpretation of the Cretaceous-Tertiary boundary as a stratigraphically condensed section is based, in part, upon the reported presence of Late Cretaceous microfossils in the basal beds of the Pine Barren.

The planktonic foraminifera and macrofossils recovered from the Prairie Bluff Chalk, basal Clayton (basal Pine Barren) sands and lower Clayton (lower Pine Barren) limestones, marls, and silts at Moscow Landing (section 2), along Shell Creek (section 3), at Prairie Bluff Landing (section 4), near Bragggs (section 5), and along Mussel Creek (section 6) indicate that the basal and lower Clayton (Pine Barren) sand, silt, marl, and limestone beds are of Danian rather than Maestrichtian age. The sequence interpretation of previous studies is an accurate representation of the depositional dynamics of the stratigraphy; however, the condensed section in this depositional sequence, at least in Alabama surface exposures that represent shelf deposition, is of Danian age rather than straddling the Maestrichtian-Danian boundary. This sequence, as observed in this study, would include Danian Clayton lowstand-shelf deposits, Clayton transgressive deposits, Clayton or Porters Creek condensed-section deposits and Clayton or Porters Creek highstand-regressive deposits. The Prairie Bluff-Clayton formational contact represents the lower boundary, and the base of the "Turritella rock" in south-central Alabama or the base of the glauconitic sand bed containing quartz and phosphate pebbles in the lower unnamed member of the Porters Creek Formation in southwestern Alabama represents the upper boundary of the sequence (Fig. 4). The Cretaceous-Tertiary contact at all localities is disconformable, and Danian lowstand or transgressive deposits disconformably overlie Maestrichtian highstand-regressive deposits.

The absence of the lowstand-shelf sands at the Bragggs section (section 5) is atypical; however, it can be explained based on the discontinuity of the basal Clayton sands at Moscow Landing (section 2) and at Prairie Bluff Landing (section 4). The presence of the sands in a streamcut along Mussel Creek, Lowndes County, about 3.2 km from the Bragggs section, is evidence that these deposits are present in the Lowndes County area. Danian planktonic foraminifera were recovered from these sands at Mussel Creek during this study. Further, the Pine Barren bed that disconformably overlies the lowstand-shelf sands at Mussel Creek contains phosphate pebbles, quartz grains, and reworked Cretaceous fossils. This unit is overlain by beds that contain Danian macrofossils (Reinhardt and others, 1986) and Danian planktonic foraminifera. A basal Pine Barren limestone bed containing phosphate pebbles, quartz grains, shark teeth and reworked Cretaceous macrofossils disconformably overlies the Prairie Bluff at the Bragggs section and is probably equivalent to the Pine Barren pebble bed at Mussel Creek. This limestone bed and the lower Pine Barren silts above this limestone at Bragggs contain Danian planktonic foraminifera.

There is no bed at the Bragggs section equivalent to the basal Clayton (Pine Barren) sands present at the other exposures observed in this study. At these other sections, the sands are sharply overlain by Clayton silts, marls, or limestones that commonly contain phosphate pebbles and reworked fossils. In the absence of the basal sands, these lithologies disconformably overlie the Prairie Bluff Chalk, as at the Bragggs section. The contact of the Clayton limestones, marls, or silts with the basal Clayton sands or the Prairie Bluff is interpreted as a transgressive surface associated with a sea-level rise in the early Danian.

This interpretation implies that relative topographic highs on the Prairie Bluff depositional surface were either: (1) subaerially exposed and therefore not a site of deposition during the accumulation of the Clayton lowstand deposits in topographic lows or depressions; or (2) preferentially subjected to transgressive reworking that destroyed any lowstand deposits that might have accumulated. Further, in the absence of the lowstand-shelf sands, the transgressive surface coincides with the type 1 sequence boundary that defines the base of the depositional sequence. This interpretation agrees with Donovan and others (1988) in that the basal Pine Barren beds at the Bragggs section represent deposition on an "interfluvial" area occurring between sediment-filled incised valleys and are underlain by a transgressive surface that has merged with the type 1 sequence boundary. Based on paleontologic and sedimentologic data and stratigraphic relationships, we believe that this type 1 sequence boundary corresponds to the base of the Danian in southwestern and south-central Alabama. In the Wilcox and Lowndes Counties area (sections 3, 4, 5, and 6), the early Danian depositional sequence includes lowstand-shelf sands (basal Clayton (Pine Barren) marine-shelf glauconitic sands), transgressive deposits (Pine Barren marine-shelf limestone and silts), condensed-section deposits (Pine Barren marine-shelf silts and limestones), and highstand-regressive deposits (Pine Barren marine and marginal marine silts and clays). A type 2 unconformity caps the sequence and is overlain by the "Turritella rock" marine-shelf limestone beds of the Pine Barren, which represent shelf-margin deposits of an overlying Danian depositional sequence. The transgressive and condensed-section deposits of this sequence are represented by the marine-shelf marls and limestones of the McBryde Limestone Member of the Clayton Formation. The marine-shelf marls of the McBryde and marine-shelf marls and clays of the lower unnamed member of the Porters Creek Formation represent the highstand-regressive deposits of the sequence.

At Moscow Landing along the Tombigbee River in Sumter County, southwestern Alabama (section 2), the lower Danian sequence includes lowstand-shelf, incised-valley-fill deposits (basal Clayton marine-shelf glauconitic sands) and lowstand-shelf "interfluve" deposits (Clayton marine-shelf calcareous silts), transgressive deposits (Clayton marine-shelf silts, limestones, and marls), condensed-section deposits (Clayton/Porters Creek marine-shelf marls and limestones), and highstand-regressive deposits (Porters Creek marls and calcareous clays). The upper Danian depositional sequence consists of a basal type 2 unconformity, transgressive deposits (marine-shelf, glauconitic sands and sandy marls of the lower unnamed member of the Porters Creek Formation), condensed-section deposits (marine-shelf calcareous clays and limestones of the Porters Creek), and


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Figure 4. Sequence stratigraphy of the Upper Cretaceous and lower Paleocene strata in southern Alabama (modified from Baum, 1986).

highstand-regressive deposits (marine and marginal marine clays of the Porters Creek). The lowstand-shelf deposits of the lower Danian sequence at Moscow Landing overlie upper Prairie Bluff beds assigned to the middle Maestrichtian *Racemiguelbelina fructicosa* Zone or the phosphatic macrofossil bed underlying these beds, which is assigned to the early middle Maestrichtian *Globotruncanca aegyptiaca* Zone. This suggests either local relief on the shelf at the time of deposition or differential incision of the shelf after deposition of the Prairie Bluff Chalk.

An unconformity-bounded, type 1 depositional sequence (Fig. 4) is evident in the middle Maestrichtian strata in the study area. This entire sequence is exposed along Shell Creek (section 3) and at Prairie Bluff Landing (section 4). Along Shell Creek and at Prairie Bluff Landing, the Prairie Bluff Chalk unconformably overlies glauconitic, fossiliferous, calcareous sands of the Ripley Formation. The Prairie Bluff depositional sequence consists of glauconitic, fossiliferous, sandy, marine-shelf chalky marl (transgressive deposits), burrowed, glauconitic, marine-shelf chalk containing phosphate pebbles, shark teeth and mollusk molds (condensed section), and marine-shelf dense chalk (highstand-regressive deposits). The burrowed surface in the middle chalk bed represents the surface of maximum sediment starvation. This bed is best exposed at Moscow Landing (section 2). The accumulation of phosphate pebbles and reworked, clionid (sponge) bored mollusk shells that are covered with serpulid (worm) tubes and are partially replaced by iron sulfides, and, in some cases, phosphatized, in this middle chalk bed, indicates long exposure on the sea floor and, therefore, a very low sedimentation rate. The high concentrations of planktonic foraminiferal tests in the dense chalk of the highstand-regressive deposits overlying the condensed-section deposits are interpreted to be related to the continuation of slow sedimentation rather than to an increase in water depth.

The depositional history of the Maestrichtian and Danian in southwestern and south-central Alabama reflects a fall in sea level at or after the end of the middle Maestrichtian and a relative sea-level rise in the early Danian (Fig. 5). The foraminiferal fauna of the Prairie Bluff Chalk is indicative of open-marine shelf deposition, whereas the fauna of the Clayton is indicative of more nearshore deposition than the upper Prairie Bluff fauna. The foraminiferal fauna of the Clayton reflects an overall increase in shelf water depths from the Prairie Bluff-Clayton contact upward to the condensed section that occurs in the Clayton/Porters Creek transition beds in the Sumter County area or the lower part of the Pine Barren Member of the Clayton in the Lowndes County area. The Prairie Bluff and Clayton foraminiferal faunas in the Sumter County area (section 2) represent more offshore marine deposits than the Prairie Bluff and Clayton foraminiferal faunas in the Lowndes County area (sections 5 and 6).

Physical and biologic events associated with the Cretaceous-Tertiary chronostratigraphic boundary are the topics of considerable current debate and research. Recent studies have hypothesized that a bolide impact on the Yucatan Peninsula of Mexico at the boundary resulted in the mass extinctions of numerous taxa that occurred at this time (e.g., Hildebrand and others, 1991). Evidence cited for this hypothesis include the presence of seemingly anomalous,
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1 Section 2-- Moscow Landing (ML)  2 Section 4-- Prairie Bluff Landing (PBL)

Figure 5. Upper Cretaceous and lower Paleocene lithostratigraphy, vertical changes in foraminiferal assemblages, and sequence stratigraphy in southern Alabama. See Figure 1 for location of sections and Figure 2 for legend of symbols (modified from Mancini and others, 1989).

crude-grained beds at the Cretaceous-Tertiary transition at many localities in the circum-Gulf of Mexico area; supporters of the Yucatan bolide scenario attribute these beds, which exhibit features indicating deposition in a high-energy hydrologic regime, to a tsunami wave resulting from the bolide impact (Smit and others, 1992a, 1992b).

As stated earlier, the lowermost Tertiary unit in the field trip area is the Paleocene (Danian) Clayton Formation, which unconformably overlies the Upper Cretaceous (Maestrichtian) Prairie Bluff Chalk. The basal Clayton sands comprise a sequence of discontinuous sand bodies that, in general, are confined to topographic depressions on the upper Prairie Bluff surface. The basal Clayton sands consist of lenticular beds of cross-bedded, fine- to coarse-grained, locally calcareous, quartz sand that contains shark teeth, phosphate pebbles, and reworked Cretaceous microfossils, macrofossils, and clasts, as well as Danian microfossils (planktonic foraminifera) and macrofossils; these beds range in thickness up to ~1.5 m. In southwest Alabama, the basal Clayton sands are interpreted to have been deposited in a shallow marine shelf setting, whereas these strata represent an estuarine environment in south-central Alabama. In both areas, the underlying Prairie Bluff Chalk and carbonate and mixed carbonate-clastic strata of the overlying lower part of the Clayton Formation were deposited in relatively deeper water marine shelf settings as compared to the basal Clayton sands.

While not discounting the possibility of a bolide impact, an extraterrestrial causal mechanism is not necessary to explain the origin of the basal Clayton sands and, further, is
not entirely consistent with outcrop observations. We contend that the basal Clayon sands were deposited as the lowstand systems tract of a third-order, unconformity-bounded depositional sequence, as indicated above.
FIELD TRIP ROAD LOG

00.0  Figure 6 is the route map for this field trip. We will begin in the Windwood Inn parking lot, Demopolis, Alabama. Turn left on U.S. Highway 80 (westbound).

00.2  Intersection with U.S. Highway 43 on right at traffic light. Continue on U.S. 80.


10.2  Intersection with Alabama Highway 28 on left. Continue on U.S. 80.

11.3  Tombigbee River.

14.7  Intersection with Alabama 28 on right and Sumter County Road 25 on left. Turn left on County Road 25.

17.0  Dead end of County Road 25. Park at end of Road and we will walk to Stop 1-Moscow Landing on the west bank of the Tombigbee River.

STOP 1. Exposure along west bank of Tombigbee River at Moscow Landing, Sumter County, Alabama, Coatopa 7.5-minute quadrangle, sec. 24, T. 17 N., R. 1 W.

The geologic section exposed at Moscow Landing includes the Upper Cretaceous (Maestrichtian) Prairie Bluff Chalk of the Selma Group, the Paleocene (Danian) Clayton Formation, and the lower beds of the Paleocene (Danian) lower unnamed member of the Porters Creek Formation of the Midway Group (Fig. 7). The Prairie Bluff consists of more than 9.8 m of light-gray burrowed, silty chalk and marl. A 15- to 30.5-cm thick bed consisting of quartz grains, phosphate pebbles, fossil molds and casts, and shark teeth generally occurs 0.6 to 0.9 m below the top of the formation. At the north end of the bluff, this phosphatic macrofossil bed is present 3.1 to 3.4 m below the top of the formation. A discontinuous bed containing *Exogyra costata* Say occurs about 0.3 m below the phosphatic macrofossil bed. The lenticular, discontinuous, irregularly bedded, calcareous, glauconitic, fossiliferous, phosphatic, quartzose, fine- to coarse-grained basal sands of the Clayton Formation overlie the Prairie Bluff Chalk. These sands, which attain a maximum thickness of 1.5 m at this locality, weather yellowish-orange and contain chalk clasts reaching 0.9 m in diameter, reworked Cretaceous macrofossils, such as *Exogyra costata*, and Paleocene macrofossils, including *Ostrea pulaskensis* Harris. The basal Clayton sands rest above an eroded surface on the underlying chalk and represent lowstand fill of incised valleys. The basal Clayton sands are overlain along a sharp contact by Clayton sandy marls containing quartz pebbles, phosphate pebbles, reworked Cretaceous fossils, and abundant *Ostrea pulaskensis.* In the absence of the sands, this bed, which is interpreted as a transgressive lag deposit, overlies Clayton dark-gray calcareous silt containing *Ostrea pulaskensis* or directly overlies the Prairie Bluff Chalk. In the latter case, the upper surface of the Prairie Bluff is generally highly burrowed, and the burrows are generally filled with Clayton sediments from above. The Clayton calcareous silt beds attain a maximum thickness of 1.5 m and are more prominent at the south end of the bluff. Olive-gray calcareous, fossiliferous, glauconitic, sandy marls of the Clayton are interbedded with two 15- to 30-cm thick sandy limestones. Together, the interbedded Clayton marls and limestones are about 1.2 m thick. Gray, fossiliferous, calcareous, silty clays of the lower member of the Porters Creek Formation that are 4.6 m thick conformably overlie the Clayton. A sandy, glauconitic marl bed, which is about 46 cm thick and contains phosphate and quartz pebbles, discontinuously overlies these Porters Creek calcareous clays. A fossiliferous limestone bed which is approximately 0.3 m thick occurs 0.6 m above the sandy marl bed. The Porters Creek clays immediately above the sandy marl bed are calcareous and fossiliferous; however, about 1.8 m above this bed and upward through the section, the beds become less calcareous and fossiliferous. At about 6.4 m above the marl bed, the clays are black, massive, carbonaceous, and sparsely fossiliferous. Blocky-weathering, black clays are typical of the Porters Creek in this area. The Upper Cretaceous and Paleocene strata are faulted at this locality, and the normal faults complicate the geologic interpretation of these units. The oldest faults, which displace only the Prairie Bluff Chalk, may have had an effect on Paleocene sediment deposition and distribution. Younger faults, characterized by slickensides and calcite mineralization, displace all of the units exposed at this locality.

The lower Prairie Bluff beds—that is, the phosphatic macrofossil bed and strata below this bed—contain *Guembelia cretacea* Cushman, *Planoglobulina caseyea* (Plummer), *Pseudotextularia deformis* (Kikoina), *Globotruncanina aegypitica* Nakkady, *G. duwi Nakkady, G. gansseri Bolli, G. stuartiformis Dalbiez, and Rugoglobigerina hexacamerata* Bronniman. The lower Prairie Bluff beds are assigned to the middle Maestrichtian *Globotruncanina aegypitica* Zone, *Globotruncanina gansseri* Subzone, *Globotruncanina contusa-stuartiformis* Assemblage Zone of Pessagno (1967) and Smith and Pessagno (1973) based on the occurrence of these species. The upper Prairie Bluff beds—that is, strata above the phosphatic macrofossil bed—contain *Planoglobulina brouensis* Martin, *Racemiguelbellina fruticosa* (Egger), *R. powelli* Smith and Pessagno, *Ventillabrella multicamerata* de Klasz, *Globigerinelloides praetextilis* Pessagno, *Globotruncanina aegypitica*, *G. conica* White, *G. gansseri, G. stephensonii* Pessagno, *G. stuartiformis,* and *Rugoglobigerina reicheli* Bronniman and, therefore, are assigned to the middle Maestrichtian *Racemiguelbellina fruticosa* Zone, *Globotruncanina gansseri* Subzone, *Globotruncanina contusa-stuartiformis* Assemblage Zone of Pessagno (1967) and Smith and Pessagno (1973). The basal Clayton silts and the lower part of the sandy marl and sandy limestone unit contain *Subbotina edita* (Subbotina), *Subbotina pseudobulloidea*
Figure 6. Index map of field trip area locating stops.

(Plummer), Subbotina triloculinaeoides (Plummer), and Globoconusa daubjergensis (Bronnimann). The presence of S. pseudobulbuloides and the absence of S. trinidadensis (Boll) place these beds in the Danian Subbotina pseudobulbuloides Interval Zone of Stainforth and others (1975). The remainder of the Clayton Formation and the lower Porters Creek calcareous clays are assigned to the Danian Subbotina trinidadensis Interval Zone of Stainforth and others (1975) on the basis of the occurrence of S. trinidadensis in the absence of Morozovella uncinita (Boll). Subbotina edita, S. inconstans (Subbotina), S. pseudobulbuloides, S. triloculinaeoides, S. trivialis (Subbotina), Globoconusa daubjergensis, and Planorotalites compressa (Plummer) also occur in these beds.

The Prairie Bluff beds have been reported to rest within the Lithraphidites quadratus Zone of Cepek and Hay (1969) on the basis of the occurrence of Lithraphidites quadratus Bramlette and Martini and Micula murra (Martini) (Smith and Mancini, 1983). The Clayton has been reported to be contained in the NP1 Zone of Martini (1971) based on the abundance of Thoracophera spp. (Siesser, 1983) and the NP2 Zone of Martini (1971) on the basis of the occurrence of Cruciplacolithus tenuis (Stradner) and the absence of Chiasmolithus consuetus (Bramlette and Sullivan) (Gibson and others, 1982; Siesser, 1983). The Porters Creek has been assigned to the NP3-4 Zone of Martini (1971) by Siesser (1983) based on the presence of C. consuetus and/or Helioritus concinnus (Martini) and the absence of Fasciculithus spp. and to the NP3 Zone of Martini (1971) by Gibson and others (1982).

The Cretaceous-Tertiary contact is unconformable, and as exposed in the field trip area, the strata at this boundary do not represent a continuous record of geologic time. The boundary occurs at the top of the Prairie Bluff Chalk. No specimens of the planktonic foraminifera Abathomphalus intermedius (Boll), A. mayaroensis (Boll), or Pseudotextularia intermedia de Klaz, which are diagnostic of the latest Maestrichtian Abathomphalus mayaroensis Subzone, Globotruncana contusa-stuariiformis Assemblage Zone of Pessagno (1967) and Smith and Pessagno (1973), were recovered from Prairie Bluff beds by Mancini and others (1989) in a study of the Cretaceous-Tertiary contact in Mississippi and Alabama. Additionally, Prairie Bluff faunas throughout the study area lack morphologically advanced forms of Planoglobulina brazeoensis, Pseudotextularia deformis, P. elegans (Rzevik), Racemiguembelina fructiosa, Ventilabrella multicromerata, and Globotruncana conica. Species such as Heteropelis glabrans (Cushman), Globigerinelloides prairiehilensis, and Globotruncana stephensoni, which are not known from strata assignable to the Abathomphalus mayaroensis Subzone, occur in the uppermost Prairie Bluff beds throughout the study area. Either sediments representing the late Maestrichtian were never deposited in this area or sediments that might have accumulated during this time period have been subsequently removed. The presence of reworked Cretaceous microfossils in
Figure 7. Measured section at Moscow Landing, Tombigbee River, Sumter County, Alabama.
the basal Clayton sands, the Clayton silts and marls, and even into the lower beds of the Porters Creek Formation indicates that, at least locally, erosion and reworking of fossils continued throughout much of the early Danian in the field trip area.

The Prairie Bluff Chalk represents a type 1 unconformity-bounded depositional sequence. The lower sequence-bounding unconformity is not exposed at this locality, but can be observed in the field trip area, particularly to the east along the Alabama River near the type locality of the Prairie Bluff Chalk at Prairie Bluff Landing, Wilcox County, Alabama and in the vicinity of Bragg's, Lowndes County, Alabama. The depositional sequence consists of sandy, fossiliferous chalks of the lower Prairie Bluff (transgressive systems tract deposits), burrowed, glauconitic chalks containing phosphate pebbles, shark teeth and mollusk shells (condensed section), and dense chalks of the upper Prairie Bluff (highstand regressive systems tract deposits). The burrowed surface in the middle chalk bed is the surface of maximum starvation (point of maximum transgression) within this depositional cycle. The accumulation of phosphate pebbles and reworked, clamidial (sponge) bored mollusk shells that are encrusted with serpulid (worm) tubes and partially replaced with iron sulfides, and, in some cases, phosphatized, indicates long exposure on the sea floor and, therefore, a very low sedimentation rate. At Moscovy Landing, the Clayton Formation and the lower calcareous clays of the Porters Creek Formation represent a type 1 sequence deposited during the early Paleocene. The basal Clayton sands are interpreted as lowstand fill of incised topography developed on the lower sequence-bounding unconformity of the sequence, which is also the Cretaceous-Tertiary contact. This outcrop illustrates the discontinuous nature of the basal Clayton sands, which is characteristic of these strata throughout the outcrop area in Alabama. Where present along the outcrop, the basal Clayton sands are sharply overlain by interbedded marls and limestones of the Clayton; this well-defined contact represents the transgressive surface of the depositional sequence. Where the basal sands are absent, the transgressive surface and sequence boundary have merged and the contact of the Clayton Formation with the underlying Prairie Bluff Chalk is highly burrowed. The interbedded Clayton marls and limestones are the transgressive systems tract strata and the Clayton/Porters Creek transitional marls and limestones are the condensed section deposits of the depositional sequence in southwest Alabama. The lowermost calcareous clays of the lower unnamed member of the Porters Creek are the highstand regressive systems tract strata of this sequence. The glauconitic marl bed in the lower unnamed member of the Porters Creek disconformably overlies the lowermost calcareous clays of the unit and represents the transgressive systems tract of the superjacent depositional sequence, which is a type 2 sequence. The calcareous clays and limestone overlying the marl are the condensed section deposits, and the black, massive, carbonaceous clays of the Porters Creek are the highstand regressive systems tract strata.

Return to U.S. Highway 80 along Sumter County Road 25.

19.2 Turn right (eastbound) on U.S. 80.
22.6 Tombigbee River.
23.7 Intersection with Alabama Highway 28. Turn right on Alabama 28.
31.0 Town of Jefferson.
38.6 City of Linden.
40.0 Intersection with U.S. Highway 43/Alabama Highway 69. Turn right on Highway 43/69/28.
40.3 Intersection with Alabama Highway 28 on left. Turn left on Alabama 28.
51.1 City of Thomaston.
56.3 Y-junction with County Road 66. Bear to right and continue on Alabama 28.
63.4 Intersection with Alabama Highway 5. Continue on Alabama 28.
69.2 Intersection with Alabama Highway 162. Continue on Alabama 28.
69.6 Entrance to Shell Creek public boat launch on left. Turn left into the entrance and drive down to the boat ramp. Here we will board the waiting boats and proceed upriver approximately 1 mile to Stop 2-Prairie Bluff Landing on the west bank of the Alabama River.

STOP 2. Exposure along the west bank of the Alabama River at Prairie Bluff Landing, Wilcox County, Alabama, Catherine 7.5-minute quadrangle, sec. 6, T. 13 N., R. 7 E. and sec. 32, T. 14 N., R. 7 E. This is the type section of the Prairie Bluff Chalk.

The geologic section exposed at Prairie Bluff Landing includes the Upper Cretaceous (Maestrichtian) Ripley Formation and Prairie Bluff Chalk of the Selma Group and the Paleocene (Danian) Pine Barren Member of the Clayton Formation of the
Midway Group (Fig. 8). The Prairie Bluff Chalk disconformably overlies glauconitic sandy marls of the Ripley Formation. A continuous bed comprised of reworked macrofossils, fossil molds and casts, phosphatic clasts, and coarse quartz grains occurs at the base of the Prairie Bluff Chalk and this bed separates the Prairie Bluff from the Ripley Formation at this locality. The lower Prairie Bluff beds consist of 0.6 m of blue fossiliferous, glauconitic, sandy marl, and the upper Prairie Bluff includes 3.0 m of blue fossiliferous dense chalk. A bed consisting of phosphatic pebbles, quartz grains, fossil molds and casts, shark teeth and Cretaceous fossils divides the lower Prairie Bluff marls from the upper chalk beds. Lenticular discontinuous, irregularly bedded, quartzose, fossiliferous sand of up to 1.2 m of the Paleocene (Danian) Pine Barren Member of the Clayton Formation disconformably overlies the Maestrichtian Prairie Bluff beds. Where the basal Clayton (Pine Barren) sands are absent, brown sandy marl of the Pine Barren Member disconformably overlies the Prairie Bluff.


At Millers Ferry, 3.7 km from Prairie Bluff Landing, core samples from the basal Clayton (Pine Barren) sands have been reported by Liu and Olsson (1992) to contain *Guemmeliitria cretacea* and *Parvularugoglobigerina eugubina* (Luterbacher and Premoli Silva). Liu and Olsson (1992) assigned the lowermost beds of the basal Pine Barren sands to the *Guemmeliitria cretacea* Interval Zone of Smit (1982) because of an abundance of this species in the absence of most other Cretaceous planktonic foraminifer species and an absence of any Paleocene forms. Those beds containing *Parvularugoglobigerina eugubina* were assigned by Liu and Olsson (1992) to the *Parvularugoglobigerina eugubina* Range Zone of Luterbacher and Premoli Silva (1964). Liu and Olsson (1992) also reported the Maestrichtian calcareous nanofossil, *Micula prinsii* Perch-Nielson from core samples of the Prairie Bluff at Millers Ferry.

The Prairie Bluff Chalk at Prairie Bluff Landing constitutes a third-order depositional sequence. The lower marl beds represent deposits of the transgressive systems tract, and the upper chalk beds are the deposits of the highstand systems tract. The phosphatic macrofossil bed separating the lower and upper Prairie Bluff deposits is the surface of maximum sediment starvation associated with the condensed section of this Maestrichtian third-order depositional sequence. The unconformity at the base of the Prairie Bluff is interpreted as a type 1 regional unconformity. The basal Clayton (Pine Barren) sands that disconformably overlie the Prairie Bluff are interpreted as shelf deposits of a Danian lowstand systems tract that overlies the Maestrichtian Prairie Bluff sequence. The Pine Barren marls are the transgressive deposits of this Danian sequence.

Return to Alabama Highway 28 from boat launch.

69.6 Turn left on Alabama 28.
70.3 Alabama River. Millers Ferry lock and dam to the right (downstream).
76.1 Intersection with Alabama Highway 21. Continue on Alabama 28.
81.1 Intersection with Alabama Highway 10. Turn left on Alabama 28/10 (truck route).
82.1 Stop sign. Turn left and continue on Alabama 28/10.
85.7 Y-junction. Bear to left and continue on Alabama 28.
94.1 Intersection with Alabama Highway 21. Alabama 28 ends at this point; bear to left and proceed on Alabama 21.
95.8 Intersection with Alabama Highway 89. Continue on Alabama 21.
99.2 Town of Snow Hill.
101.6 Town of Furman.
108.3 Lowndes County line.
112.5 Town of Braggs.
Figure 8. Measured section at Prairie Bluff Landing, Alabama River, Wilcox County, Alabama.
Stop 3. Road cut exposures along Highway 263, southeast of Braggs, Lowndes County, Alabama, Braggs and Fort Dale 7.5 minute quadrangles.

The geologic section exposed in these roadcuts includes the Upper Cretaceous (Maestrichtian) Prairie Bluff Chalk, the lower Paleocene (Danian) Pine Barren Member of the Clayton Formation, including the "Turritella rock" limestone in the upper part, the McBryde Limestone Member of the Clayton Formation, and the Paleocene Ports Creek Formation, including the Matthews Landing Marl Member. Figure 9 is a measured section at the road cut that exposes the Cretaceous-Tertiary contact. The Prairie Bluff Chalk consists of about 4.3 m of gray, micaceous, clayey silt containing *Exogyra costata* at this locality. The Prairie Bluff-Pine Barren contact is disconformable and is marked by quartz grains and phosphate pebbles at the base of the lowermost Pine Barren limestone. The Pine Barren Member includes approximately 33.1 m of olive gray, micaceous, glauconitic, calcareous, fine-grained sand; olive-gray, argillaceous, glauconitic limestone; greenish-gray, micaceous, calcareous, glauconitic, silty clay and silt; and gray silt and clay. *Ostreopsis pulaskensis* occurs in the Pine Barren. The "Turritella rock" is a sandy, fossiliferous limestone that is particularly abundant in species of *Turritella*. The McBryde Limestone Member is comprised of 10.3 m of gray, glauconitic marl and pale orange, argillaceous limestone. The Ports Creek, which is approximately 30.4 m in thickness, grades from gray, glauconitic marls and nodular limestone at the base to greennish-gray, calcareous clays near the top. White, micaceous, glauconitic, fine- to medium-grained sand approximately 29.7 m thick is present at the top of the lower member of the Ports Creek Formation. The Matthews Landing Marl Member is comprised of 2.7 m of gray, glauconitic, calcareous, very fine-grained sand and marl.

*Globotruncana aegyptiaca* and *G. gansseri* were recovered from the Prairie Bluff at this locality indicating that these beds are placed in the *Globotruncana gansseri* Subzone of Pessagno (1967) and Smith and Pessagno (1973). The lower Pine Barren beds are assigned to the Danian *Subbotina pseudobulloides* Interval Zone of Stainforth and others (1975) based on the occurrences of *Subbotina pseudobulloides*, *S. trilobulinaeoides*, and *Globocoma daubergensis* and the absence of *Subbotina trinidadiensis*. The remainder of the Pine Barren, the McBryde, and the lower Ports Creek rest within the Danian *Subbotina trinidadiensis* Interval Zone of Stainforth and others (1975). *Subbotina inconstans*, *S. pseudobulloides*, *S. triloculinaeoides*, *S. trinidadiensis*, *S. trivialis*, *Globocoma daubergensis* and *Planorotalites compressa* are found in these units. The presence of *S. trinidadiensis* in the absence of *Morozovella uncinata* indicates this assignment. The upper Ports Creek marls and clays contain *M. uncinata* in the absence of *M. angulata* (White) and therefore are assigned to the Danian *Morozovella uncinata* Interval Zone of Stainforth and others (1975). *Subbotina inconstans*, *S. pseudobulloides*, *S. triloculinaeoides*, *S. trivialis*, *Planorotalites compressa* and *?Pararotalia perclara* (Loeblich and Tappan) also occur in these marls and clays. The Matthews Landing is characterized by the presence of *Morozovella angulata* in the absence of *Planorotalites pusilla pusilla* (Bolli) and, therefore, is placed in the Selandian *Morozovella angulata* Interval Zone of Stainforth and others (1975). *Subbotina pseudobulloides*, *S. triloculinaeoides*, *S. trivialis*, *Planorotalites compressa*, *?Pararotalia perclara*, and *Morozovella uncinata* also occur in the Matthews Landing. Neither planktonic foraminifera characteristic of the latest Maestrichtian *Abathomphalus mayaroensis* Subzone of Pessagno (1967) and Smith and Pessagno (1973) nor of the earliest Danian *Globigerina eugubina* Range Zone of Stainforth and others (1975) were recovered from the Braggs section.

The Prairie Bluff beds at the Braggs section have been assigned to the *Nephrolithus frequens* Zone of Cepek and Hay (1969) based on the presence of *Nephrolithus frequens* Gorka (Cepek and others, 1968; Worsley, 1974; Thierstein, 1981; Zemo, 1982). *Micula prinssii* was reported from this section by Habib and others (1992). Although calcareous nanoplankton workers have maintained that the *Nephrolithus frequens* Zone is present in the Prairie Bluff Chalk at the Braggs locality, they believed that the latest Maestrichtian beds are absent at this section (Worsley, 1974; Thierstein, 1981). The Pine Barren has been reported to rest within the NP1 Zone of Martini (1971) based on the abundance of *Thoracosphaera* spp. (Siesser, 1983) and the NP2 Zone of Martini (1971) based on the occurrence of *Cruciplacolithus tenuis* and the absence of *Chiasmolithus consuetus* (Gibson and others, 1982; Siesser, 1983). The McBryde and Ports Creek have been assigned to the NP3/4 Zone of Martini (1971) by Siesser (1983) on the basis of the occurrence of *C. consuetus* and/or *Heloirthus concinnus* and the absence of *Fasciculolithus* spp. and to the NP3 Zone of Martini (1971) by Gibson and others (1982). The Matthews Landing Member was assigned to the NP4 Zone of Martini (1971) based on the presence of *Toweius craticulus* (Gibson and others, 1982).

The Pine Barren Member is interpreted to represent an unconformity-bounded, type 1 depositional sequence. The Pine Barren marine shelf glauconitic silts, sands, and limestones comprise the transgressive and condensed section deposits of the transgressive systems tract, and the Pine Barren marine to marginal marine silts and clays comprise the highstand regressive deposits of the highstand systems tract. The "Turritella rock," the McBryde, and the interbedded clays and limestones of the lower member of the Ports Creek represent an unconformity-bounded, type 2 depositional sequence (Mancini and Tew, 1991). The "Turritella rock" marine shelf sandy limestones are the deposits of the shelf margin systems tract. The transgressive and condensed section deposits consist of McBryde marine shelf marls and limestones, and the marine shelf interbedded calcareous clays and limestones of the lower part of the lower member of the Ports Creek are the highstand regressive deposits of the sequence. The marine shelf marls and clays of the middle part of the lower member of the Ports Creek represent a type 2 depositional sequence. The marine shelf, cross-bedded, glauconitic sands in the upper part of the lower member of the Ports Creek are shelf margin deposits of an overlying depositional sequence. Within this type 2 sequence, the Matthews Landing marine shelf glauconitic marls and sands represent the transgressive and condensed section deposits and the marginal marine clay and silts of the Oak Hill Member of the Naheola Formation represent the highstand regressive deposits.
Figure 9. Measured section along Highway 263, south of Braggs, Lowndes County, Alabama.
Continue along Alabama 263.

121.5 Butler County line.

121.6 Turn left onto dirt road from Alabama 263.

122.4 Park before crossing wooden bridge over Mussel Creek. We will walk down the path beside the bridge to Stop 4- Mussel Creek section.

STOP 4. Exposure along the west bank of Mussel Creek, Lowndes County, Alabama, Fort Dale 7.5-minute quadrangle, sec. 31, T. 12 N., R. 14 E.

The geologic section exposed at Mussel Creek includes the Upper Cretaceous (Maestrichtian) Prairie Bluff Chalk of the Selma Group and the Paleocene (Danian) Pine Barren Member of the Clayton Formation the Midway Group, including the basal Clayton (Pine Barren) sands (Fig. 10). The Prairie Bluff consists of 2.4+ m of gray, micaceous, fossiliferous, argillaceous, sandy marl. A lenticular, discontinuous bed of micaceous, carbonate, lignitic, quartzose, fine- to medium-grained, cross-stratified sand of the basal Clayton (Pine Barren) sands disconformably overlies the Prairie Bluff Chalk. This sand of Danian age attains a maximum thickness of 0.5 m at the southern (upstream) end of the exposure and pinches out to the north. The basal Clayton (Pine Barren) sands at this locality are interpreted to have been deposited in an estuarine setting; a minor interval of fossiliferous, glauconitic marine sand occurs near the top of the unit along the southern part of the exposure. An indurated brown, fossiliferous, glauconitic, phosphatic limestone bed of the Pine Barren Member of the Clayton Formation which ranges in thickness from 18 to 31 cm, disconformably overlies the basal Clayton (Pine Barren) sands. Where the sand is absent, this Danian bed disconformably overlies marl of the Maestrichtian Prairie Bluff Chalk. The limestone bed contains clasts of the underlying Prairie Bluff, reworked Cretaceous macrofossils, and pebble- to small cobble-size phosphate clasts. Brown, glauconitic, micaceous, fossiliferous, argillaceous, sandy marl interbedded with indurated, brown, glauconitic, fossiliferous limestone of the Pine Barren Member conformably overlies the phosphatic limestone bed. This unit is 1.8+ m thick.

The Prairie Bluff beds at this locality are assigned to the Maestrichtian Racemiguembelina fructicosa Zonule, Globotruncanina gansseri Subzone, Globotruncanina contusa-quadiformis Assemblage Zone of Pessagno (1967) and Smith and Pessagno (1973). These beds contain Guembelliria cretacea, Heterohelix glabrans, Planoglobulina brazoensis, Globigerinelloides prairiehillensis, Globotruncanina aegyptiaca, G. gansseri, G. stephsoni, and Rugoglobigerina reicheli. The basal Clayton (Pine Barren) sands are placed in the Danian Subbotina pseudobulboides Interval Zone of Stainforth and others (1975) based on the presence of Subbotina pseudobulboides. The Pine Barren marl beds are also assigned to this zone. These beds contain Subbotina pseudobulboides, S. triloculinaeoides, and Globococcus daubergensis.

The basal Clayton (Pine Barren) sands at the Mussel Creek section are interpreted as incised valley-fill estuarine and marine deposits of a lowstand systems tract. These Danian sands disconformably overlie Maestrichtian Prairie Bluff marls of the highstand systems tract of the underlying depositional sequence. The Danian Pine Barren interbedded marls and limestones accumulated in the transgressive systems tract of the Pine Barren depositional sequence. The Pine Barren phosphatic lag bed in the lower limestone unit represents the transgressive surface. Although the Danian incised valley-fill sands have not been identified at the Bragg section, the transgressive phosphatic lag bed is present. At the Bragg locality, the transgressive surface and the lower sequence boundary of the Pine Barren depositional sequence have merged.

This stop concludes the field trip.

REFERENCES


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LATE EOCENE AND EARLY OLIGOCENE CARBONATE FACIES AND PALEOENVIRONMENTS OF THE EASTERN GULF COASTAL PLAIN

Jonathan R. Bryan

Department of Geology, Florida State University, Tallahassee, FL 32306

With Contributions by: Richard H. Fluegeman, Jr. (Ball State University) James A. Heller (University of Tennessee, Knoxville) Gary S. Morgan (Florida Museum of Natural History) Roger W. Portell (Florida Museum of Natural History)

INTRODUCTION
Strata of the Gulf Coastal Plain of the southeastern United States preserve an exceptional record of Tertiary sediments. Some sections have been championed as among the most fossiliferous and stratigraphically complete shallow marine sequences in the world. Much of the western Gulf Coast, from Texas to Mississippi, consists of thick siliciclastic sequences, reflecting the long depositional history of the Mississippi Embayment. In contrast, the eastern Gulf Coast, from Alabama to Florida, is largely composed of a massive complex of Paleogene carbonate sequences. Not until the Miocene did prograding marine and fluvial clastics begin to progressively restrict most shallow marine carbonate production to its modern limit in south Florida.

During this fieldtrip we will examine and collect a variety of Paleogene fossiliferous carbonate facies, representing many paleoenvironments, including: open shelf, large foraminiferan banks (Ocala Limestone, Late Eocene); backreef, miliolid-peloidal limestone (Suwannee Limestone, Oligocene); coral reef (Bridgeboro Limestone, Oligocene); and deep shelf-to-forereef mudstones and algal banks (Marianna and Floral Limestones, Oligocene). The Eocene-Oligocene boundary will be examined (in core), and the nature of this boundary in Florida will be discussed in the context of the well-known Mississippi-Alabama section.

MID-LATE EOCENE (JACKSONIAN) OCALA GROUP: OVERVIEW AND PROSPECTS
There are numerous summaries of Upper Eocene carbonates of Florida and Georgia (Huddleston and Herrick, 1986; Toulmin, 1977; Hunter, 1976). Jones (1982) has summarized the history of stratigraphic nomenclature of what was originally called the Ocala Limestone by Dall and Harris (1892) and Cooke (1915). While Puri (1957) gave group status to the Ocala (i.e., Ocala Group), and assigned to it the Inglis, Williston, and Crystal River Formations (Fig. 1), it has been argued by Randazzo (1976) that these three formations cannot be adequately differentiated lithologically and might properly be combined into one litostratigraphic unit, the Ocala Limestone, with further subdivision into members or biozones as needed.

Although upper and lower subdivisions of the Ocala can generally be recognized lithologically, Puri's three formations are essentially biostratigraphic in nature (although he defined them before publication of either the International or the North American stratigraphic codes). For this reason, the Florida Geological Survey has officially adopted the original litostratigraphic term, Ocala Limestone (Scott et al., 1991). Nevertheless, Puri's terminology has been widely used in the literature (e.g., Murray, 1961; Cheetham, 1963; Toulmin, 1977) and should be understood. It is therefore used in Figure 1. The use of Ocala Group in Georgia, consisting of the distinctive Tivola Limestone and Ocmulgee Formation (Huddleston and Herrick, 1986), is within the guidelines of both the North American Stratigraphic Code and the International Stratigraphic Guide, and is advisable as a unifying concept for closely related (laterally and vertically) but distinctive carbonates lithologies of Jacksonian age in Georgia (see Siesser, 1984, p. 444). Any revisions on the litostratigraphic status of the upper and lower portions of the Ocala, and further biostratigraphic subdivision, will require additional study.

The Eocene age of the Ocala Limestone was recognized fairly early from molluscan and foraminiferal faunas (Cooke, 1915; Richards & Palmer, 1953; Gravel and Hanna, 1938). Using calcareous nanoplankton from localities in Alabama, Siesser (1983) assigned Puri's Crystal River Formation to NP 19/20 (late Eocene). Siesser assigned the Moodys Branch Formation (an "Inglis-Williston" equivalent in Alabama and Mississippi) to NP 17 (late Middle Eocene).
The Ocala Group carbonates contain a wealth of untapped paleontologic, sedimentologic, and stratigraphic information and have the potential for much interesting research.

Paleontologically, although several interesting papers have appeared recently on various aspects of the Ocala fauna (e.g., McKinney and Zachos, 1986; Carter et al., 1989; Jones and Nicol, 1989; Ivany et al., 1990), work is still needed on almost all elements of the fauna. There remains a diverse and largely undescribed molluscan fauna. Much of it is moldic, but with the use of good silicone peels, accurate descriptions are possible. Although Cheetham (1963) and Canu and Bassler (1920) did much work with the Ocala bryozoan faunas, there remains much to describe. The bryozoan-rich Tivola Limestone, for example, was not even considered in Cheetham’s much-cited GSA Memoir 91. The abundant larger foraminifera need reconsideration in the context of recent taxonomic developments and large forum paleoecology, and there is a diverse arthropod fauna in need of description. When adequately described, these exceptional faunas will be the subject of much interesting paleobiological, paleoecologic, and paleobiogeographic research.

Sedimentologically, with the increasing interest in the distinction between tropical and temperate water carbonate assemblages in the rock record (Nelson, 1988a,b), the Ocala Group poses some interesting research problems. These carbonates span some six degrees of latitude (at least 25°-31° N), from south Florida to southcentral Georgia. Despite the occurrence of "tropical" carbonate assemblages in the overlying Oligocene Suwannee Limestone (with green algae and coral reefs; Bryan, 1991), and underlying middle Eocene limestones (with green algae; Hunter, 1976; Applin and Applin, 1944), most of the Ocala Group contains little green algae or corals (i.e., the chlorozoan assemblage indicative of tropical carbonates). Bryozoans, however, may be exceptionally abundant. The Tivola Limestone of Georgia, for example, is largely a bryozoan hash, and should be classified as a bryomol limestone, a strong indicator of temperate water deposition (Nelson, 1988b). Coralline red algae, however, another good "temperate" water indicator, appear to be rare in the Ocala Group.
The nature of the Florida shelf during the Eocene, whether it was a ramp or rimmed platform, has received some consideration in the literature. Cheetham (1963) reconstructs the northern margin of the Florida shelf as a flat-topped platform, progressively shoaling throughout Jacksonian time along the Suwannee Channel, but notes the absence of true reef development. Chen (1965) suggested, but did not demonstrate, that the Florida Platform may have been rimmed on its northern and eastern flanks by reefs. Randazzo and Saroop (1976) describe very shallow marine bank conditions for the Avon Park and Inglis Formations, but also note the seemingly anomalous absence of reef corals. Carter (1989) and Manker and Carter (1989) document progressively muddier substrates through Ocala time in Florida and Georgia, with high energy conditions around the Suwannee Channel. Like Cheetham, they suggest a rimmed bank rather than ramp morphology for the northern Florida Platform, with protected, muddy, lagoonal, perhaps seagrass community environments behind the coarse-grained "rim", a situation not unlike that of the modern western Bahama bank margin. The influence of the Suwannee Channel on carbonate facies development in the Ocala Group needs additional study (see McKinney, 1984).

The question of the extent of reef development (if any) on the margins of the Florida Platform during the Late Eocene has not been satisfactorily answered. While reef corals suffered severely during the Late Cretaceous extinctions, and were slow to recover (Bryan, 1992), they did form reefs in the Late Eocene in the Caribbean (Budd et al., 1992). Monospecific scleractinian patch reefs (consisting of Actinastrea) are present locally in the Crystal River Formation (Jones, 1982), but widespread bank margin reefs are not known. However, the margins of the Eocene Florida Platform are buried, and scleractinian framework (composed of aragonite) is rarely preserved in pure carbonate lithologies. Furthermore, repeated bioerosion and recementation can quickly turn coral framestone into mudstone (Hubbard, 1992; James and Ginsburg, 1979), further complicating the recognition of reef framework. These are persistent problems in the study of Tertiary reefs (James, 1983).

Stratigraphically, the Ocala Group is of much interest in that it has been described as a deepening-upward sequence (see Manker and Carter, 1989), an atypical condition for shallow marine carbonates (see James, 1984). There is certainly a general deepening of environments from the lower to upper Ocala. This raises the question of whether Florida may have become a drowned platform during the terminal Eocene sea level highstand. The response of the Florida Platform during this sea level rise is especially important in the context of sequence stratigraphy, which now recognizes variable responses of carbonate systems during sea level changes (Schlager, 1992). Was the Florida Platform drowned during the Late Eocene? If so, why could it not "keep-up" with sea level? If it could not keep-up, was it because of the temperate water character of its fauna, making the system inherently prone to drowning (Simone and Carannante, 1988; Smith, 1988)? Why was there apparently so little coral reef development in the climatically equable ("polytaxic") Late Eocene of Florida, whereas it was common during the cooler ("oligotaxic") Early Oligocene (Bryan, 1991)? Careful microfacies and paleoecologic analyses should help resolve these questions.

THE EOCENE-OLIGOCENE BOUNDARY IN FLORIDA

There is extensive literature on the Eocene-Oligocene (E-O) boundary in the Gulf Coastal Plain (e.g., Mancini and Tew, 1991; Pasley and Hazel, 1990; Mancini, 1986; Baum and Vail, 1988; Deboo, 1965). This interval is considered by many Gulf Coast geologists to be one of the most stratigraphically complete in the world. In addition, the well-preserved and diverse marine faunas have attracted a number of important paleontological studies (e.g., Dockery, 1986; Hansen, 1987).

Recently, at least two debates have emerged over the E-O boundary in the Gulf Coast: (1) The first concerns the exact placement of the E-O boundary (see Keller, 1985)—a good problem as it indicates the stratigraphic completeness of the section involved; (2) there has also been a controversy over what type of sequence boundary the E-O contact represents. Sequence stratigraphers working in Alabama maintain that in this area, the boundary is contained within a condensed section (Tew and Mancini, 1992; Mancini and Tew, 1991), reflecting seal level highstand. Some Mississippi geologists, however, maintain that in Mississippi the boundary is clearly a Type 1 sequence boundary (i.e., a classical erosional unconformity; Dockery, 1992). Oddly enough, both may be right. The Alabama section, however, appears to be more complete.

In the shallow marine carbonate section of Florida, the location of the E-O boundary has also been controversial. The two, related concerns are the extent of the Bumpnose Limestone on the Florida Platform; and the faunal affinities, age, and correlation of certain beds between definitively Eocene (Ocala Limestone) and Oligocene (Suwannee Limestone) rocks.

The nature and extent of the Bumpnose Limestone deserves some consideration. The typical Bumpnose is a glauconitic and micritic limestone with abundant larger foraminifera, bryozoa, and molluscs, and is considered a calcareous facies of the Red Bluff Formation (Fig. 1). It ranges from 3 to 4.5 m in thickness, and is predominantly found in southern Alabama and portions of northern Florida. Characteristic fauna include the larger
foraminiferan *Lepidocyclina chaperi*, bivalves *Spondylus dumosus*, *Chlamys anatipes*, *Lopha vickshurgensis*, and the echinoid *Clypeaster rogersi*. *Lepidocyclina chaperi* and *Spondylus dumosus* are index fossils for the formation (Fig. 1). Although lithologically similar to, and conformable with, the underlying Ocala Limestone (and equivalents) of Late Eocene age, the Early Oligocene age of the Bumpnose has been confirmed by calcareous nannoplankton (*Siessier, 1983*), planktonic foraminifera (*Mancini, 1986*), and macrofauna (*Cheetham, 1957, 1963*). The Bumpnose is considered part of a condensed section, coinciding with seal level highstand (TE 3.3 depositional sequence; *Loutit et al., 1988*).

Other than the type area in Jackson County (*Moore, 1955*), the extent of the Bumpnose Limestone in Florida is uncertain. The Bumpnose has been reported from various localities in central Florida (viz., the subsurface of Polk County--*Cheetham, 1957, 1963*; and quarries in Hernando and Citrus Counties--*Hunter, 1972*), but the name has sometimes been applied loosely to lithologies of questionable Bumpnose affinities. Distinct lithostratigraphic units are difficult to separate in the shallow marine carbonate section of Florida, and rigorous biostratigraphic correlation is hindered by the general exclusion of planktonic microfossils (*Bryan and Huddleston, 1991*) and obvious facies changes. Without additional detailed petrographic and biostratigraphic studies, the use of the lithostratigraphic term "Bumpnose Limestone" in peninsular Florida should probably be discontinued.

A more difficult problem is the question of the stratigraphic affinities of certain strata between the Upper Eocene Ocala Limestone and Lower Oligocene Suwannee Limestone (and equivalents). This interval includes the *Turritella martinensis* Zone of older literature and the *Rotularia vernoni* Zone of peninsular Florida.

The *Turritella martinensis* Zone was first reported by *MacNeill (1944)* from the banks of the Suwannee River near Ellaville, Florida, and was considered Early Oligocene in age. This now classic Ellaville section is critical to understanding the nature of the E-O boundary in Florida. *MacNeill's T. martinensis* Zone was named for beds 1 and 2 of *Cooke and Mossom (1929, p. 72)* and *Cooke (1945, p. 85-86)* along the east bank of the Suwannee River. While the Ellaville section has generally been considered the type area of the Suwannee Limestone, it has been subdivided by *Huddleston (in press)* into the Ellaville Limestone (beds 1 & 2 of *Cooke, 1945*) and the Suwanacoochee Dolomite (beds 3 & 4 of *Cooke, 1945*). The age of the Ellaville is Early Oligocene as determined by *Huddleston (in press)* on the basis of the larger foraminiferan *Lepidocyclina mantelli*, gastropod *Turritella mississippiensis* (*T. martinensis* is also present), and echinoids *Clypeaster rogersi* and *Rhyncholampas gouldii*.

*Hunter (1972, 1976, 1981)* reports the *T. martinensis* Zone from a number of localities in peninsular Florida and considers this zone to be equivalent to the *Rotularia vernoni* Zone. The *Rotularia vernoni* Zone (*R. vernoni* is an annelid worm tube) contains a fauna with both Eocene and Oligocene affinities. The zone is currently considered to be Eocene in age by *Nicol and associates (Nicol et al., 1989, 1984, 1976; Jones and Nicol, 1989)*, but *Hunter (1976, 1981)* has argued for an Oligocene age, and has previously referred this interval to the basal Oligocene Bumpnose Limestone (*Hunter, 1972*), along with the *T. martinensis* Zone (*Hunter, 1976*). *Puri (1957)* included the *R. vernoni* jointly with the larger foraminiferan *Asterocyclina* into his *Asterocyclina-R. vernoni* Zone (implying an Eocene age), but *Hunter (1981)* reports (from personal communication with *Puri*) that this zone was created on the basis of the relative occurrence of the index taxa in their respective sections. The two index species have reportedly never been found together.

According to *Huddleston (in press)*, the *Rotularia vernoni* Zone lies approximately 3 feet below the Ellaville Limestone at the type area along the Suwannee River, although this interval is not well-exposed. These stratigraphic complexities are illustrated in Figure 1 and will be observed in the Ellaville 1 Core (W-10657) of the Florida Geological Survey prior to the departure of the fieldtrip.

If we accept *Huddleston's assignment of the T. martinensis* Zone (= beds 1 & 2 of *Cooke, 1945*) to the Ellaville Limestone, and accept an Oligocene age for the Ellaville, the unresolved questions would seem to be: (1) Is the *R. vernoni* Zone correlative with the *T. martinensis* Zone at Ellaville?; and (2) Is the *R. vernoni* Zone and/or the *T. martinensis* Zone equivalent to the Bumpnose Limestone (i.e., earliest Oligocene in age)? The first question can probably be resolved with additional work along the Suwannee River and with cores from the area (housed at the Florida Geological Survey). The question of the precise age of the *R. vernoni* and *T. martinensis* Zones will not be resolved without a more thorough faunal analysis. Both the *R. vernoni* Zone and the *T. martinensis* Zone have extremely diverse invertebrate faunas that have not been described. Only when these questions are resolved will the placement of the E-O boundary in Florida be possible.

It must be remembered, however, that these stratigraphic uncertainties suggest continuous deposition, transitional facies, and a relatively complete section across the Eocene-Oligocene boundary in peninsular Florida. Although an unconformity has been reported between the Ocala and Suwannee Limestones in some places in Florida (*Randazzo, 1982*), many sections show every indication of continuous deposition from the Eocene into the Oligocene (regardless
of the placement of the E-O boundary), with no evidence of unconformity.

The Florida section may be a critical one for understanding both faunal and environmental changes during the E-O transition. For example, the extensive corallal reef tract of the Bridgeboro Limestone that developed in north Florida and south Georgia in the Early Oligocene may have begun immediately after the Eocene.

Mabey and Applin (1968, p.22,23) document the occurrence of a limestone "composed mainly of coral and coralline algae" from approximately 1000 to 1300 feet in the Magnolia Petroleum Company No. 1-A Well, Franklin County, Florida. The sequence contains the larger foraminifera Lepidocyclina chaperi and Helicostegina polygyralis, the former species being indicative of the Bumpnose Limestone. Additional study of deep wells containing the Bumpnose Limestone in north Florida is needed.

The nature of the Eocene-Oligocene faunal transition in shallow marine carbonate sequences can be addressed in the Florida section. Although the E-O boundary interval is represented by a shallowing-upward sequence in peninsular Florida (as will be seen in the Ellaville 1 Core), with different environments on each side of the boundary, sections in the panhandle (as seen in various cores from Walton and Okaloosa Counties) may contain a continuous sequence across the boundary within similar, densely fossiliferous facies.

This again raises the question of the response of Florida Platform carbonates during sealevel change and the relation of the Florida E-O section to that of Mississippi and Alabama. If the E-O boundary does indeed occur during a highstand (Mancini and Tew, 1991), does the shallowing-upward E-O interval in peninsular Florida represent a "catch-up" in growth of the carbonate platform? Or is it reflecting the earliest Oligocene sea level fall? Unlike most of the Cenozoic sequence of the Gulf Coastal Plain, the Florida Platform carbonates have not been evaluated in the context of sequence stratigraphy. The high stratigraphic resolution of the Alabama-Mississippi section will provide a good comparative standard for the Florida sequence.

LOWER OLIGOCENE CARBONATE PLATFORM EVOLUTION AND REEF DEVELOPMENT IN THE EASTERN GULF COASTAL PLAIN

Throughout the Early Oligocene (Vicksburgian Provincial Stage; Fig. 2), the eastern Gulf Coastal Plain was characterized by three paleogeographic-carbonate facies provinces (Huddleston, in press; Bryan, 1991; Figs. 3, 4):

1) Shelf Province (Alabama/NW Florida), with foramol/bryomol, glauconitic limestones and local red algal pavements (Marianna, Glendon, Florala Limestones);
2) Gulf Trough/Apalachicola Embayment Province (N Florida/SW Georgia), a deep, current-swept structure with shallow, flanking corallal reefs (Bridgeboro Limestone); and
3) Florida Platform Province (peninsular Florida/SE Georgia), a millolid, peloidal chlorozoan limestone with local patch reefs and coral thickets (Suwannee Limestone).

During the mid-Oligocene highstand (TO1.2 depositional sequence), the Shelf Province became a drowned ramp with a shelf margin condensed section (Glendon Limestone). This carbonate system was unable to keep-up with sea level rise because of the temperate-water character of its fauna (30°N paleolatitude). Around the Gulf Trough, however, corallal reefs (Bridgeboro Limestone) kept pace with sealevel rise and formed a rimmed platform (Fig. 5). Despite its comparatively high paleolatitude (29°-32°N), the tropical fauna of these carbonates thrived because of the influence of warm Gulf Trough waters originating in the Caribbean (Fig. 6). The Florida Platform also kept pace with sea level rise and was partially emergent ("Orange Island" of Vaughan, 1910). This mosaic of adjacent tropical and temperate carbonates, developed on rimmed platform and ramp settings (Fig. 7), is comparable to the modern shelf margin from south Florida (Florida Reef Tract) westward to the west Florida carbonate ramp slope (Fig. 8). The Gulf Trough- Florida Platform complex likewise has a compelling modern analog in the Florida Straits Bahama Bank system (Fig. 9).

During the 30 Ma eustatic sea level fall, corallal reefs moved from the flanks of the Gulf Trough (Bridgeboro Limestone) into the Trough (Okapiello Member, Suwannee Limestone). This time-transgressive shift in reef development continued to step down the bathymetric gradient (tracking sea level fall) until by the Late Oligocene, reefs were growing along the northern Gulf shelf margin in southern Mississippi and Alabama (Heterostegina Zone) as a rimmed shelf (Fig. 10; Krutak and Beron, 1990; Forman and Schlager, 1957). Isolated, time-equivalent reefs grew on the shoals of emerging salt domes in Texas and Louisiana (Bryan, 1991).

Acknowledgements

This fieldtrip would not have been possible without the assistance of many people. Thanks to Tom Scott and Frank Rupert of the Florida Geological Survey for permission to examine the Ellaville 1, Bass 1, and Alum Bluff 1 cores. Thanks also go to several land owners for their kind permission to visit their property: Mr. Ottis Murray (Mayo Quarry), Ms. R.V. Singletary (Dowling Park Quarry), Mrs. Laura E. Leary (Bridgeboro Quarry), Ms. Elaine Howard of the U.S. Army Corps of Engineers (Franklin Landing Boat Ramp), White Construction Company (Duncan Church Quarry), and Mr. Danny Stovall (Stovall Quarry).
Figure 2. Regional stratigraphy of Vicksburgian strata of the eastern Gulf Coastal Plain (from Bryan, 1991).

Figure 3. Lower Oligocene paleogeographic map of the eastern Gulf Coast showing stratigraphic associations, facies provinces, and transect lines for cross sections of Figure 4. 
EGSA = Eastern Gulf Shelf Association, SSCP = Shelf Carbonate Facies Province, GT-AEA = Gulf Trough-Apalachicola Embayment Association, GT-AEAFP = Gulf Trough-Apalachicola Embayment Facies Province, FPA = Florida Platform Association, FPPFP = Florida Platform Facies Province, OI = Orange Island, a-j = localities of Figure 4 (from Bryan, 1991).
Figure 4. Cross sections of Vicksburgian strata across the eastern Gulf Coast. A, transect a-f of Fig. 3. a/HB = Haynes Bluff, Mississippi; b/SL = St. Stephens Bluff, Alabama; c/Sto = Stovall Quarry, near Florala, Alabama; d/W-8104 = Florida Geological Survey core W-8104 (Brown #1), Walton County, Florida; e/DC1 = Florida Geological Survey core Duncan Church #1 (W-11487), Washington County, Florida; f/AB1 = Florida Geological Survey core Alum Bluff #1 (W-6901), Liberty County, Florida. Vertical scale: 1 cm = 15 m. B, transect g-j of Fig. 3. g/BBQ = Bridgeboro Quarry (type section), Mitchell County, Georgia; h/MPF = Georgia Geological Survey core Mobley Plant Farm (GGS-3535), Colquitt County, Georgia; i/GGS-3208 = Georgia Geological Survey core GGS-3208, Brooks County, Georgia; j/B1 = Florida Geological Survey core Bass #1 (W-10480), Madison County, Florida. SL = mean sea level (modern). Vertical scale: 1 cm = 24 m (from Bryan, 1991).
Figure 5. Block diagram reconstruction of Vicksburgian carbonates across the entire eastern Gulf Coastal Plain from Mississippi to Florida, as an extension of Coleman's (1983) model (from Bryan, 1991).

Figure 6. Modern Loop Current of the Gulf of Mexico. The Loop Current enters the Gulf via the Yucatan Strait, loops clockwise, turns into the Florida Current in the Straits of Florida, and ultimately joins the Antilles Current to form the Gulf Stream. During higher sea levels in the Oligocene, the Loop Current would have flowed naturally through the Gulf Trough area (broken line) (from Bryan, 1991).
Figure 7. Block diagram facies and paleogeographic model for Eastern Gulf Coast Carbonates. SL = sealevel (from Bryan, 1991).

Figure 8. (Paleo)geographic map showing the Lower Oligocene paleoshelf margin and condensed section in the northern Gulf (i.e., the Glendon Limestone-GLS), and corresponding Bridgeboro corralgal reef (BBR) along the Gulf Trough; and the modern shelf margin and condensed section on the West Florida Shelf Ramp (FLR), and corresponding Florida Reef Tract (FRT) along the Florida Straits (from Bryan, 1991).

Figure 10. Time-transgressive movement of Oligocene reef facies in the northern Gulf Coastal Plain. Areas marked 1 show location of the Bridgeboro coralgal buildup of early Vicksburgian age. Area 2 shows the location of reef facies within the Gulf Trough (Okapilco Member of the Suwannee Limestone) which developed during late Vicksburgian time, just prior to the maximum sealevel lowstand at 30 Ma. Areas marked 3 show the location of late Oligocene (Chickasawhayan) to early Miocene reef facies along the Late Oligocene shelf margin of the northern Gulf Coast. Reef development tracked sea level fall throughout the Oligocene from Area 1 (early Oligocene), to Area 2 (late early Oligocene), to Area 3 (late Oligocene to early Miocene) (from Bryan, 1991).
Acknowledgment is made to the donors of The Petroleum Research Fund, administered by the American Chemical Society, for support of all phases of fieldwork and research on this project. Thanks to Douglas S. Jones and Thomas M. Scott for reviewing an early draft of this fieldguide. Finally, I especially thank Paul Huddlestun for spending many hours with me in the office, over the phone, and in the field, tutoring me in Gulf Coast stratigraphy, and helping develop my particular "worldview" of the Gulf Coast Lower Oligocene.

ORIENTATION

Friday Evening, April 2.

Before we depart for the fieldtrip, we will examine three cores which show various facies and stratigraphic relationships not conveniently seen in outcrop. The Ellaville 1 core (W-10657, Florida Geological Survey) contains the Eocene-Oligocene boundary. The Bass 1 core (W-10480, Florida Geological Survey) is the hypostratotype of the Suwannee Limestone. The Alum Bluff 1 core (W-6901, Florida Geological Survey) extends through the Gulf Trough.

Ellaville 1 Core (W-10657, Florida Geological Survey).

**Location:** Suwannee County, Florida. T1S, R11E, S24, 1/4 cb (Johnson, 1986, p.374).

**Surface Elevation:** 55 feet.

**Comments:** This core is critical for understanding the Eocene-Oligocene boundary in peninsular Florida. The section in this core is partially exposed on the banks of the Suwannee River from Ellaville to Dowling Park. But the most critical interval--the transition or contact between the *Rotularia vernoni* Zone and the *Turritella martinesis* Zone--is not well exposed in outcrop. It can be observed in this core. Several interesting sedimentary fabrics can also be observed.

<table>
<thead>
<tr>
<th>Depth</th>
<th>Formation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-18'</td>
<td>Post-Oligocene</td>
<td>Siliciclastics; fine sandstone to mudstone with large rip-up limestone clasts from underlying unit; poor recovery.</td>
</tr>
<tr>
<td>18-23'6&quot;</td>
<td>Suwanneacochee</td>
<td>Dolomitic mudstone, very porous, with non-dolomitic white limestone clasts scattered throughout; good intraformational conglomerate from approx. 20'6&quot;-22', and fenestral fabric from 22-23'; dark brown dolomite at 23'.</td>
</tr>
<tr>
<td>23'6&quot;-25'</td>
<td>Dolomite</td>
<td>No recovery.</td>
</tr>
<tr>
<td>25-59'</td>
<td>Ellaville</td>
<td>Fine, sandy, very porous calcarenite; very moldic (molluscan), with abundant <em>Turritella</em> and some <em>Chione, Lumulites</em> and other bryozoa, lepidocyclinids, and rare solitary coral. (Huddlestun (in press) places the lower boundary for the Ellaville at 39', but it could be extended to the first moldic molluscan zone, which begins at about 57', and I would extend the lithology to 59'). Small <em>Lopha</em> from 50-52'; abundant <em>Lepidocyclina</em> from 43-50'; large <em>Nummulites</em> at 43-44'(same as seen at Dowling Park section); some algae at 59'. The Ellaville appears to alternate between a muddy- to fine-sandy, non-moldic limestone and moldic molluscan limestone.</td>
</tr>
<tr>
<td>59-72'</td>
<td>Ocala</td>
<td>Biocalcarenite; many well-preserved pectinids from 59-61'; <em>Oligopygus</em> at approx. 71'.</td>
</tr>
</tbody>
</table>
Bass 1 Core (W-10480, Florida Geological Survey).

Location: Madison County, Florida (see Johnson, 1986, p.259).
Surface Elevation: 83 feet.
Comments: Huddleston (in press) has designated this core as a hypostratotype for the Suwannee Limestone, and his unpublished core log is incorporated into this description. The Suwannee is thick in this core and very typical of the formation. The description begins at 45 feet.

<table>
<thead>
<tr>
<th>Depth</th>
<th>Formation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>45-51'</td>
<td>Suwannee Limestone</td>
<td>Clayey limestone. Huddleston identifies the Suwannee from 0-175'.</td>
</tr>
<tr>
<td>51-58'</td>
<td></td>
<td>Very porous, miliolid calcarenite to calcilitute. Variably macromoldic and cherty.</td>
</tr>
<tr>
<td>58-79'</td>
<td></td>
<td>Porous foraminiferal biomicrite with rare articulated coralline algae.</td>
</tr>
<tr>
<td>79-95'</td>
<td></td>
<td>Variably fossiliferous, macromoldic calcarenite, partially cemented. <em>Dictyoconus</em> present.</td>
</tr>
<tr>
<td>95-122'</td>
<td></td>
<td>Very fine-grained (&quot;mealy textured&quot;) foraminiferal biosparite to dismicrite. <em>Dictyoconus</em> present.</td>
</tr>
<tr>
<td>122-142'</td>
<td></td>
<td>Coarse, fossiliferous calcarenite with common <em>Dictyoconus</em> (locally forms a coquina). <em>Rhynchozamps gouldii</em> present. Rounded, fine-grained lithoclasts present (up to 2 cm in diameter), with irregular shape.</td>
</tr>
<tr>
<td>142-165'</td>
<td></td>
<td>Coarse, miliolid-<em>Dictyoconus</em> calcarenite. Small, muddy lithoclasts present.</td>
</tr>
<tr>
<td>165-175'</td>
<td></td>
<td>Muddy, fossil-rich calcarenite. Fossils poorly preserved.</td>
</tr>
<tr>
<td>175-204'</td>
<td>Suwannee; Dolomite/ Ellaville Limestone</td>
<td>Variable lithology, mostly dark brown, sucrosic dolomite with variable amounts of clay and lime sand. Local fenestral fabric with thinly-laminated dolomitic mud on coarser-grained dolomite (possibly mud-cracks). Some molluscan molds. This variable, dolomitic lithology is commonly associated with the Eocene-Oligocene boundary in the eastern Gulf Coast. Huddleston (unpublished log of this core) considers the interval from 175-192’as Suwannee Limestone Dolomite, and from 192-204’ as Ellaville Limestone.</td>
</tr>
<tr>
<td>&gt;204'</td>
<td>Ocala Limestone</td>
<td>Bioclastic limestone. Echinoids and solitary coral present.</td>
</tr>
</tbody>
</table>

Alum Bluff 1 Core (W-6901, Florida Geological Survey).

Surface Elevation: 82 feet.
Comments: This core occurs just off the central axis of the Gulf Trough. The stratigraphic affinities of some of the lithologies are uncertain (Huddleston, pers. comm.), and the core is in need of detailed microstratigraphic study. Unpublished core log of Huddleston was used in this description. Huddleston’s terminology is used in Fig. 4. Description begins at 300 feet.
<table>
<thead>
<tr>
<th>Depth</th>
<th>Formation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>300-376'</td>
<td>Suwannee Limestone</td>
<td>Porous, gray calcarenite, dolomitic limestone, and dolomite. <em>Kaphus</em> at 384', and is common at 350' and above. Coral penetrated at 350' and 360'. (Huddleston considers the interval from 287-357' as the Okapilco Member of the Suwannee, and from 357-371.5' as Wolf Pit Dolomite).</td>
</tr>
<tr>
<td>473-527'</td>
<td>Marianna Limestone</td>
<td>Pale orange to gray dolomitic limestone, with extremely large (up to 5 cm diameter), thin <em>Lepidocyclina</em>, and some smaller forms, as well as <em>Nummulites</em>. Some very large, robust <em>Lepidocyclina</em> at 521' in a calcarenitic matrix. Contact with overlying Bridgeboro is gradational and difficult to place.</td>
</tr>
</tbody>
</table>

**ROAD LOG**

**Saturday, April 3**

**Mileage**

**Description**

00.0

Depart from FSU Geology Department at 8:00AM. Turn right onto Woodward Street.

00.4

Turn left onto Jefferson Street.

01.4

Turn left onto Duval Street.
01.5 Turn right onto College Avenue.
01.6 Turn right onto Monroe Street.
01.7 Turn left onto Apalachee Parkway, in front of Florida State Capitol.
21.9 Waukeenhah.
54.3 Turn left onto US27 in Perry.
78.4 Turn right onto dirt road (next to old car junkyard). Drive into quarry. (approx. 1 1/2 hours from Tallahassee)

STOP 1. Mayo Quarry (Upper Eocene Crystal River Formation).

**Location:** Dell Limerock Mine, 4.5 miles northwest of Mayo, Lafayette County, Florida, on US 27, 0.4 mile on dirt road (west side of US 27). Sec 32, NW 1/4, SE 1/4, T4S, R11E, Mayo Quadrangle.

**Comments:** This locality is perhaps one of the best-exposed and most fossiliferous outcrops of the Ocala Limestone. It is one of the few localities in the Ocala in which the remains of archaeocete whales have been recovered, as well as other marine vertebrates. Echinoids, pectinid bivalves, and larger foraminifer are especially abundant. A complete species list is included below.

<table>
<thead>
<tr>
<th>Formation</th>
<th>Thickness(m)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Suwannee Limestone</td>
<td>residual boulders</td>
<td>Silicified, moldic limestone with common <em>Rhyncholampas gouldii.</em></td>
</tr>
<tr>
<td>Ocala Limestone</td>
<td>4.0</td>
<td>Fine-grained, sparsely fossiliferous limestone (wackestone); <em>Lepidocyclina</em> concentrations locally; at 2.5 m the lower bed is a 20 cm thick concentration of pectinids, traceable along the outcrop; other fossils include <em>Ocalina floridana</em> and <em>Weisbordella cubae</em>. Distinct, but gradational contact.</td>
</tr>
<tr>
<td></td>
<td>0.38</td>
<td>Oyster bed in fine-grained matrix; dense concentration of oysters, most valves disarticulated (some intact) and with random orientation; some valves cemented together; common epibionts on shells (esp. bryozoans), including interior of valves, as well as borings; large foram coquina below oyster bed; pectinids common. Distinct, but gradational contact.</td>
</tr>
<tr>
<td></td>
<td>4.0</td>
<td>Bioclastic grainstone; fossils include larger foraminifera (coquina), solitary coral, molluscan internal molds, <em>Oligopygus wetherbyi</em>. Sharp contact.</td>
</tr>
<tr>
<td></td>
<td>0.7</td>
<td>Densely fossiliferous pectinid bed with abundant <em>Amusium ocalanum</em>; large, pycnodonte oysters common (up to 13 cm long); gastropods; common trace fossils; irregular lower contact.</td>
</tr>
<tr>
<td></td>
<td>2.2</td>
<td>Fine-grained, bioclastic limestone (wackestone); extremely fossiliferous.</td>
</tr>
</tbody>
</table>
The following is a preliminary species list from the Mayo Quarry prepared by Roger Portell and Gary Morgan of the Florida Museum of Natural History, University of Florida, Gainesville.

**Cnidaria**
- cf. *Trochycathus* sp.

**Bryozoa**
- Chelostome, gen. and sp. undet.¹
- *Amulosisia* sp.

**Mollusca**
- Gastropoda
  - *Akeria* sp.
  - *Bellatara* sp.
  - *Calyptrea* sp.
  - *Campanile* sp.
  - *Caricella* sp.
  - *Cassidae, gen. and sp. indet.*²
  - *Cerithidae, gen. and sp. indet.*
  - *Conus* sp. A
  - *Conus* sp. B
  - *Cypraeida fenestralis* Conrad
  - cf. *Hipponix* sp.
  - *Laevelea* sp.
  - *Lithophysema grande* (Aldrich)
  - *Naticidae, gen. and sp. indet.*
  - *Pseudocrombium* sp.
  - *Rimella* sp.
  - *Scaphander* sp.
  - *Terebellum* sp.
  - *Turritella* sp.
  - *Volutidae, gen. and sp. indet.*
  - *Xenophoridae, gen. and sp. indet.*

**Bivalvia**
- *Amusium ocalanum* (Dall)
- *Barbatia* sp.
- *Barbatia cuculloides* (Conrad)
- *Barbatia palmerae* Richards
- *Callista* sp.
- *Chlamys incerta*e Tucker-Rowland
- *Chlamys spilimani clinchfieldensis* Harris
- *Crassatella* sp.
- *Cubitostrea* sp.
- *Divaricella* sp.
- *Glycymeris* sp.
- *Hyotissa podagrina* (Dall)
- *Lithophaga* sp.
- *Lucina* sp. A
- *Lucina* sp. B
- *Lucinidae, gen. and sp. indet.*
- *Miliha ocalana* Dall
- *Mytilidae, gen. and sp. indet.*
- *Nayadina ocalensis* (MacNeil)
- *Nomocardium* sp.
- *Ostreidae, gen. and sp. indet.*
- *Pectinidae, gen. and sp. indet.*
- *Pinna quadra*ta Dall
- *Plicatula* sp.
- *Pteria* sp.
- *Spisula* sp.
- *Spondylus hollisteri* Harris
- *Tawera* sp.
- *Trachycardium* sp.
- *Venericardia* sp.
- *Venericardia withlacoochensis* Richards
- *Venericardia, gen. and sp. undet.*

**Scaphopoda**
- *Dentalium* sp.

**Annelida**
- Polychaete worm tubes

**Arthropoda**
- Malacostraca
  - *Calappa robertsi* Ross et al.
  - *Calianassa* sp.
  - *Lophoranina georgiana* (Rathbun)
  - *Ocalina floridana* Rathbun

**Echinodermata**
- Echinoidae
  - *Agassizia clevei* Cotteau
  - *Amblypygus americanus* Desor
  - *Ditremaster beckleri* (Cooke)
  - *Eupatagus ocalanus* Cooke
  - *Fibularia vaughani* ( Twitchell)
  - *Neolaganum durhami* Cooke
  - *Oligopygus haldemani* (Conrad)
  - *Oligopygus wetherbyi* deLoriol
  - *Phyllocanthus mortoni* (Conrad)
  - *Schizaster ocalanus* Cooke
  - *Weisbordella cubae* (Weisbord)

**Asteroidae**
- cf. *Gonioidiscaster* sp.

**trace fossils**
- *Ophiomorpha* sp.
- *Lithoplaesion ocalae* Diblin et al.

**Chondrichthyes**
- *Carcharhinus* sp.
- *Isurus praecursor*
- *Myliobatis* sp. A
- *Myliobatis* sp. B

**osteichthyes**
- *Cylindracanthus rectus*
- *Diodon* sp.
- *Ostracionidae, gen. and sp. indet.*
- *Sparidae, gen. and sp. indet.*
- *Sphyraena* sp.
REPTILIA
  Cheloniidae, gen. and sp. indet.
  Syllodus sp.
MAMMALIA
  Pontogeneus brachyspondylus
  Zygophiza kochii

1 material has not been sufficiently studied for more complete identification.
2 material is insufficient for more complete identification.

Return to US27, turn left (north).

83.1 Turn right onto State Road 53 (BP gas station across intersection).
87.3 Town of Day.
91.3 Turn right onto State Road 250.
93.2 Cross Suwannee River (singing of Florida’s state song optional).
93.4 Turn left into Dowling Park.
(approx. 15-20 minutes from Stop 1)

STOP 2. Dowling Park (Eocene/Oligocene? Rotularia vernoni Zone).

Location: Dowling Park boat ramp, Suwannee County, Florida, at water level along the east bank of the Suwannee River. Comments: The Dowling Park exposure is a continuous and easily collectable outcrop of the Rotularia vernoni Zone and contains abundant larger foraminifera, brachiozoa, molluscs, and echinoids. This relatively thin (perhaps 1-2 m) interval has generated much controversy. Named for a common index fossil (an annelid worm tube), this interval lies between definitively Eocene (Ocala Limestone) and Oligocene (Suwannee Limestone) rocks, and has faunal affinities with both Eocene and Oligocene assemblages. It is currently considered to be Eocene in age by Nicol and associates (Nicol and Jones, 1984, 1989; Nicol et al., 1976; Jones and Nicol, 1989), but Hunter (1976) has suggested an Oligocene age. Further study of this biozone is clearly warranted. The stratigraphic uncertainties may be indicative of a complete and transitional Eocene-Oligocene boundary interval in peninsular Florida.

Return to State Road 250, turn right, cross back over Suwannee River

94.6 Turn right onto paved road.
95.2 Turn right onto dirt road into quarry.

STOP 3. Dowling Park Quarry (Eocene/Oligocene? Rotularia vernoni Zone; Upper Eocene Crystal River Formation).

Location: Abandoned quarry on west side of Suwannee River, across river from Dowling Park section, Lafayette, County, Florida. NW 1/4, SE 1/4, sec 6, T3S, R11E, Madison SE quadrangle.
**Comments:** At this abandoned quarry near Dowling Park, 2 meters of Ocala Limestone is overlain by about 1.5 m of the *Rotularia vernoni* Zone.

<table>
<thead>
<tr>
<th>Formation</th>
<th>Thickness(m)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Rotularia vernoni Zone</em></td>
<td>1.5</td>
<td>Lower 0.3 meter: large foram and macrofossil coquina with reworked(?) pectinids from lower bed. Above this, sandy calcarenite, generally poorly fossiliferous except for top 0.3 meter, which has abundant molluscan molds (<em>Chione</em>, etc.) and echinoids. <em>Rotularia vernoni</em> is present. Sharp contact.</td>
</tr>
<tr>
<td>Ocala Limestone</td>
<td>1.0</td>
<td>Bottom is a poorly fossiliferous bioclastic calcarenite (just as in upper part of underlying bed). This grades upward into a macrofossiliferous limestone with abundant pectinids, coquinite at top with abundant larger forams. Trace fossils present. Sharp, but unbroken and conformable contact.</td>
</tr>
<tr>
<td></td>
<td>1.0</td>
<td>Tan, richly microfossiliferous, sandy calcarenite; very few macrofossils. Lower third is a large foram/pectinid grainstone which grades upward into the sandy calcarenite.</td>
</tr>
</tbody>
</table>

**(LUNCH AT DOWLING PARK)**

95.3 Return to SR 53 on SR 250. Turn right onto State Road 53.

96.5 Veer left at fork (i.e., stay on SR 53).

107.2 Turn left onto Interstate 10 West (Exit 37).

140.3 Turn right onto US19 north (Florida-Georgia Parkway) (Exit 33).

145.5 Monticello (town square). Continue north on US19.

154.0 Georgia State Line.

167.2 Thomasville (city limit).

199.4 Turn right onto State Road 112 in Camilla.

211.3 Park on right side of road. (approx. 1 hour 50 minutes from Stops 2 and 3)

**STOP 4. Bridgeboro Quarry (Lower Oligocene Bridgeboro Limestone, Type Section).**

**Location:** On SR 112, south of Bridgeboro, 1.6 miles south of intersection of SR 112 with highway 93, Mitchell County, Georgia.

**Comments:** This is the type locality of the Bridgeboro Limestone. The Bridgeboro is a coralgal, bioclastic grainstone that developed on the northern, and to a lesser extent southern, flanks of the Gulf Trough and Apalachicola Embayment. It outcrops in a northeast-southwest trend for nearly 280 km, and includes lithologies once mapped as Flint River Formation, Suwannee Limestone, and Duncan Church beds (previously considered a facies of the Suwannee Limestone in the Florida panhandle; Puri and Vernon, 1964, p. 106). The total thickness of the Bridgeboro is unknown, but there are some 20 meters exposed at the type section. The Bridgeboro is densely fossiliferous, with common larger foraminifera *Lepidocyclina undosa* and *Lepidocyclina yurnagunensis*, rhodoliths in rock-forming abundance, bivalve *Chlamys anatipes*, and echinoid *Clypeaster cotteau*. Massive reef corals are present locally, and the coral reef described by Vaughan (1900, 1919) from Bainbridge,
Georgia, occurs within the outcropping stike belt of the Bridgeboro. The Duncan Church beds, as seen in Washington County, Florida (Stop 6), lie adjacent to the wide Apalachicola Embayment and contain abundant branching coralline algae ("maerl") and fewer, smaller rhodoliths than the typical Bridgeboro of Georgia. The Duncan Church beds are therefore considered a distinct facies of the Bridgeboro. A biohermal/reefal origin for the Bridgeboro is clearly indicated. The Bridgeboro is locally overlain by the Suwannee Limestone. References: Manker and Carter (1987, 1989), Bryan and Huddleston (1991), Huddleston (in press). The following description is modified from Manker and Carter (1989):

<table>
<thead>
<tr>
<th>Formation</th>
<th>Thickness (m)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bridgeboro Limestone</td>
<td>2.8-3.7</td>
<td>Partly silicified limestone with large solution vugs and scalloped surfaces. Rhodoliths sparse near the base but more densely packed upward, up to 6cm in diameter. Large oysters. Unit 12.</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>Rhodolith-poor bioclastic limestone as in Unit 8. Clay clasts. Lepidocyclina. Unit 11.</td>
</tr>
<tr>
<td></td>
<td>2.9</td>
<td>Packed rhodolith limestone as in Unit 1. Rhodoliths more often discoid than below, randomly oriented. Molluscs, Clupeaster cotteau. Unit 10.</td>
</tr>
<tr>
<td></td>
<td>1.7</td>
<td>Bioclastic limestone as below, argillaceous (?). Few small (&lt;4cm) rhodoliths. Clay clasts and stringers. Abundant molluscs (Chlamys duncanensis, C. anatipes, Spondylus, and others), Clupeaster cotteau, Lepidocyclina. Unit 9.</td>
</tr>
<tr>
<td></td>
<td>0.6</td>
<td>Bioclastic limestone as below, but harder. Internal lamination. Unit 7.</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>Bioclastic limestone as in Unit 2, but coarser grained. C. duncanensis and other bivalves Unit 6.</td>
</tr>
<tr>
<td></td>
<td>2.45</td>
<td>Rhodolith limestone as in Unit 1. Inclined clay-rich bed (?lining channels). C. duncanensis, Lepidocyclina, gastropods, miliolids, cidaroid spines. Unit 5.</td>
</tr>
<tr>
<td></td>
<td>0.05</td>
<td>Thin, discontinuous green clay bed, similar to clasts below. Unit 4.</td>
</tr>
<tr>
<td></td>
<td>0.6</td>
<td>As Unit 2, but rhodoliths smaller (&lt;1cm) and less abundant. Large (up to 0.1m by 0.5m) clasts of green clay and fine unconsolidated quartz sand slumped into bedding depressions. Unit 3.</td>
</tr>
<tr>
<td></td>
<td>3.4</td>
<td>Finely granular bioclastic (algal) limestone, few small rhodoliths (&lt;2cm) becoming more common and larger (&lt;8cm) toward top. Irregular clay clasts. Unit 2.</td>
</tr>
<tr>
<td></td>
<td>4.9</td>
<td>Massive to thick-bedded limestone. Mostly densely packed rhodoliths up to 10 cm. Thin, discontinuous beds/lenses and clasts of waxy green clay to 10 cm. Thin, lensoid, discontinuous beds of bioclastic argillaceous (?) limestone, extending downward in to crevices and cavities. C. duncanensis, Lepidocyclina, Ampulina, oysters, miliolids, turritellid (?). Unit 1.</td>
</tr>
</tbody>
</table>

Continue north on State Road 112.

214.4 Turn left onto State Road 93 north (cross RR tracks immediately after turn).

40
224.5  
Turn left onto State Road 133.

229.4  
Veer left on fork onto SR 234.

230.1  
Turn right onto College Drive/Radium Springs Road (Dairy Queen on right).

231.7  
Turn left onto Ogelthorpe Blvd.

232.5  
Albany Days Inn.

**Sunday, April 4**

<table>
<thead>
<tr>
<th>Mileage</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>00.0</td>
<td>From Days Inn, proceed west on Ogelthorpe Blvd.</td>
</tr>
<tr>
<td>2.1</td>
<td>Turn left (south) on Slappey Blvd. (State Road 62).</td>
</tr>
<tr>
<td>5.6</td>
<td>Intersection of SR 62 and SR 91, turn right on SR 62.</td>
</tr>
<tr>
<td>18.3</td>
<td>Turn right onto State Road 37 in Leary.</td>
</tr>
<tr>
<td>34.9</td>
<td>Town of Fort Gaines (city limit).</td>
</tr>
<tr>
<td>36.0</td>
<td>Chattahoochee River (Alabama/Georgia state line).</td>
</tr>
<tr>
<td>36.4</td>
<td>Turn right into Walter F. George Dam and Powerhouse.</td>
</tr>
<tr>
<td>36.5</td>
<td>Turn right into boat ramp area parking lot. (approx. 1 hour from Albany)</td>
</tr>
</tbody>
</table>

**STOP 5. Franklin Landing (Paleocene Clayton Limestone).**

**Location:** Franklin Landing boat ramp on Chattahoochee River, Henry County, Alabama. Sec 6, T7N, R30E, Fort Gaines Quad.

**Comments:** The Chattahoochee River section has been the subject of many investigations (Toulmin and Winters, 1954; Toulmin and LaMoreaux, 1963; Swann and Poort, 1979; Vail et al., 1987; Gibson, 1989; Reinhardt and Gibson, 1981). Of interest to this fieldtrip is the Upper Clayton Limestone of Paleocene age. The top of the Clayton in this region is a deeply eroded paleokarst surface overlain by transgressive, channel-fill sands. Besides solution features, what appear to be tree root structures are common on the surface of the Clayton (cp. Adams and Horbury, 1989). Overlying the Clayton are the Gravel Creek Member and "Ostrea thirsae beds" (now Odontogryphaea thirsae) of the Paleocene Nanafalia Formation.

While Toulmin and LaMoreaux (1963) recognized the presence of calcareous algae within the Clayton, it is in fact a rhodolith limestone very much like the Oligocene Bridgeboro Limestone, but with fewer rhodoliths. The presence of these rhodoliths may be indicative of the location of the northern margin of the Suwannee Channel during the Paleocene. According to Toulmin (1955), the upper Clayton in southeastern Alabama directly underlies the Salt Mountain Limestone and has a smaller foraminiferal fauna similar to that of the type Salt Mountain (Toulmin, 1941). The upper Clayton at this locality also has an undescribed brachiopod fauna very much like that described by Toulmin (1940) from the Salt Mountain Limestone. The brachiopods are typically found attached to the undersides of the rhodoliths. Unfortunately, NO COLLECTING ALLOWED!
The following statement on the foraminifera of this section was provided by Richard H. Fluegeman (Ball State University):

"NOTES ON PALEOCENE FORAMINIFERA FROM THE FRANKLIN LANDING SECTION, HENRY COUNTY, ALABAMA

"Foraminifera are present in all samples of the Clayton collected from Franklin Landing. From the massive limestone in the lower part of the Clayton no planktonic foraminifera were collected and the benthic foraminifera obtained were recrystallized. The assemblage is assignable to the Cibicidoides-Anomalinooides assemblage and represents water depths of approximately 18 meters.

"The uppermost Clayton is a soft, fine-grained limestone which produced a well preserved and diverse foraminiferal fauna. A few planktonic foraminifera were collected from this unit. They were all identified as Globocoma daubjergensis, the range of which extends from the Subbotina pseudobulboides Zone (P1b) through the Morozovella trinidadensis Zone (P1c) and possibly into the Morozovella uncinata Zone (P2). While not a restricted index fossil, its presence does rule out correlating these limestones with the Salt Mountain Limestone in western Alabama (Zone P4 in age). Also, given the identification of the "Turritella" rock by Toulmin and LaMoreaux (1963) down section from Franklin Landing, it can be argued that the Clayton Limestone here is likely P1c or P2 in age. This would make it equivalent to the McBryde Limestone and lower Porters Creek Formation of western Alabama. Benthic foraminifera collected from this unit indicate an assemblage assignable to the Pulsiphonina-Anomalinooides assemblage. This assemblage indicates water depths of approximately 18-50 meters and reflects rising sea level compared to the underlying unit.

"No planktonic foraminifera were collected from the Nanafalia Formation at Franklin Landing. The presence of the stratigraphically important benthic foraminiferan Discorbidis washburni correlates the Nanafalia here with the "Ostrea thursoae" beds of the Nanafalia in western Alabama. The benthic foraminifera collected from most of the Nanafalia at Franklin Landing can be assigned to the Lenticulina-Anomalinooides assemblage. This is a nearshore assemblage with estimated water depths no greater than 5 meters. The uppermost marine beds in this exposure do contain an assemblage assignable to the Cibicidoides-Anomalinooides assemblage, indicating increasing water depth."

Fluegeman (pers. comm.) further suggests that the uppermost Clayton at Franklin Landing (Bed 11 of Toulmin and LaMoreaux) may be equivalent to the Matthews Landing Marl (NP4 in age), and the bed below this (Bed 10 of Toulmin and LaMoreaux) an equivalent of the lower to middle Porters Creek sequence. Given these ages, Fluegeman suggests that the upper Clayton (Beds 10 and 11 of Toulmin and LaMoreaux) might appropriately be assigned to the Cedar Keys Limestone of the subsurface of peninsular Florida.

Return to highway, turn right onto State Road 10.

49.6 Turn left onto State Road 27 in Abbeville.

50.4 Veer right on fork onto SR 27 (also 431 Business).

52.5 Turn left onto US 431 south (big intersection).

74.7 Turn left on US 231 south (Dothan by-pass).

81.5 Turn left US 231 south (end of Dothan by-pass).

94.9 Florida state line.

98.4 Turn right onto State Road 273 in Campbellton.

113.2 Turn left (south) onto State Road 77 in Chipley.
116.4 Go under Interstate 10.
119.9 Turn right onto County Road 276.
121.7 Turn right onto County Road 276.
122.0 Turn right on dirt road by Piney Grove Baptist Church.
122.2 Turn left and park at quarry entrance. (approx. 1 hour 50 minutes from Stop 5).

STOP 6. Duncan Church Quarry (Lower Oligocene Marianna Limestone and Bridgeboro Limestone).

**Location:** Small, active quarry, about 0.1 mile behind Piney Grove Church, Wausau, Washington County, Florida. SE 1/4 sec 36, T4N, R14W.

**Comments:** Excellent exposure of the Duncan Church facies of the Bridgeboro Limestone. This is the "type" area for the Duncan Church beds (formerly of the Suwannee Limestone) of Puri and Vernon (1964). Underlying the Bridgeboro is the Marianna Limestone, which retains its "chimney rock" texture, but starts to resemble the Bridgeboro Limestone, especially in its larger foram fauna (identical to the type section of the Bridgeboro); elsewhere, the Marianna generally has a chalky appearance with a monospecific assemblage of *Lepidocyclina mantelli*.

<table>
<thead>
<tr>
<th>Formation</th>
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</tr>
</thead>
<tbody>
<tr>
<td>Bridgeboro Limestone (Duncan Church Beds)</td>
<td>5.0</td>
<td>Fossiliferous grainstone with abundant rhodoliths (generally smaller in diameter and not as well-developed as those of the type Bridgeboro); local lenses of <em>Lepidocyclina</em> grainstone/coquina; abundant <em>Cypleaster coteau</em>, <em>Chlamys anatipes</em>, and <em>Chlamys duncanensis</em>; gradational contact marked by the appearance of algae and bryozoan. Gradational contact.</td>
</tr>
<tr>
<td>Marianna Limestone</td>
<td>20.0</td>
<td>Fossiliferous wackestone; abundant larger foraminifera (<em>Lepidocyclina undosa</em>, <em>L. yurnagunensis</em>, <em>L. mantelli</em>); common <em>Cypleaster coteau</em>.</td>
</tr>
</tbody>
</table>

5.7 Return to Interstate 10 on CR 276 and SR 77. Continue west on I-10 (Exit 18).

**LUNCH**

35.3 Exit I-10 onto US 331 north (Exit 14). DeFuniaq Springs.
2.3 Turn left on US 331-US 90 at DeFuniaq Springs.
1.9 Turn right on US 331.
19.2 Paxton City Limit.
2.7 Alabama state line.
0.4 Turn right on State Road 54.
6.2 Turn left into Stovall Lime and Cattle Quarry. (approx. 1 hour 30 minutes from Stop 6).
STOP 7. Stovall Quarry (Lower Oligocene Florala Limestone).

**Location:** On property of Stovall Lime and Cattle, Inc., about 7 miles (11 km) east of Florala on SR 54, Covington County, Alabama. NW 1/4, SE 1/4, sec 22, T1N, R20W. Hacoda Quadrangle.

**Comments:** This is the type section of the Florala Limestone. The Florala is an extremely fossiliferous wackestone and packstone, composed largely of platy, coralline red algae and larger foraminifera in rock-forming abundance. Twelve meters are exposed at the type locality. The full geographic extent of the Florala is not currently known, but it is thought to grade laterally into the Glendon and Marianna Limestones to the west, and the Bridgeboro Limestone to the east. The fauna consists almost exclusively of calcitic organisms, primarily the larger foraminifera *Lepidocyclina undosa* and *Nummulites panamensis*; coralline algae; bivalves *Chlamys anatipes* and *Lopha vickburgensis*; and at least three species of brachiopods (including *Argyrotheca* and *Lacazella*). James Heller (University of Tennessee, Knoxville) reports four species of echinoids from the Florala (*Lytechinus floridanus, Clypeaster cotteau, Brissus bridgeboroensis, Macropneustes mortoni*) and approximately 60 species of bryozoa. Heller (work in progress) has recognized four distinctive biofacies at this section. The Florala contains elements of both the Glendon and Bridgeboro faunas and is interpreted as an intermediate and transitional facies between the Glendon and Bridgeboro Limestones. References: Heller and Bryan (1992), Huddleston (in press), Giawse (1969, p. 88), MacNeil (1944, p. 1351-52; 1946, p. 44,47).

<table>
<thead>
<tr>
<th>Formation</th>
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<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bucatunna/Byram Marl</td>
<td>6.0</td>
<td>Dark brown to white marl. Upper portion may be Bucatunna Formation, lower portion appears to be more calcareous and carries a diverse molluscan fauna with Byram affinities (e.g., <em>Pecten byramensis</em>, <em>Nemocardium diversum</em>, <em>Panopea oblongata</em>, <em>Cassis</em> sp., and others). Irregular contact with underlying limestone.</td>
</tr>
<tr>
<td>Florala Limestone</td>
<td>12.0</td>
<td>Densely fossiliferous, large foraminiferal-algal limestone. Lithologies range from wackestone to packstone, with local micritic, algal bindstone (forming rubby, algal beds). Includes abundant <em>Lepidocyclina</em>, <em>Nummulites</em>, and coralline red algae, with common pectinids (<em>Chlamys anatipes</em>), <em>Lopha vickburgensis</em>, bryozoans, brachiopods, and echinoids.</td>
</tr>
</tbody>
</table>

7.0 Return to US 331 south.

Return to Interstate 10, east to Tallahassee.

Exit I-10 at Exit 28 south in Tallahassee.
(approx. 2 hours 45 minutes to Tallahassee from Stop 7).
REFERENCES


Hunter, M.E., 1972, Biostratigraphy and paleontology, in Oligocene stratigraphy—A study of the Lansing Quarry near Brooksville, Hernando County, Florida. Bay Area Geological
Nicol, D., and Jones, D.S., 1984, Chione (Chione) craspedon Dall in the Crystal River Formation (Eocene) in Peninsular Florida. Tulane Studies in Geology and Paleontology, v. 18, p. 73-75.
Guidebook 18, p. 49-62.
Toulin, L.D., 1941, Eocene smaller foraminifera from the Salt Mountain Limestone of Alabama. Journal of Paleontology,
Karst features of northern Florida

Frank Rupert

Florida Geological Survey, 903 West Tennessee St., Tallahassee, FL 32304

ABSTRACT

The shallow Tertiary carbonates of northern Florida have provided an ideal platform for the development of classic karst features, as well as some of the longest dissolution cave systems in the world. Outstanding examples of karst springs, sinks, dissolution depressions, natural bridges, and subaqueous conduits are situated south of Tallahassee, Florida in the Woodville Karst Plain. Foremost among these features are the Leon Sinks Geological Area, in southern Leon County, and Wakulla Springs, a first magnitude spring system in northern Wakulla County which forms the source of the Wakulla River. Both areas offer unique insight into the complex hydrogeologic processes at work in regional karst drainage systems.

REGIONAL GEOMORPHOLOGY

Northern Florida includes two broad geomorphic zones, the Northern Highlands and the Gulf Coastal Lowlands (White, 1970). The Northern Highlands are part of a nearly continuous series of siliciclastic uplands that span northern Florida from western Duval County westward to the Florida-Alabama state line. These highlands are thought to be dissected remnants of a once much larger and continuous highlands that extended from southern Alabama and Georgia southward into the Florida panhandle (Puri and Vernon, 1964). The Northern Highlands geomorphic zone extends from the Apalachicola River eastward to Duval County (White, 1970). They extend northward into Georgia, and are bounded to the south by the Cody Scarp, a relict marine erosional escarpment.

Locally, the highlands are named the Tallahassee Hills, a term applied by Cooke (1939) to the portion of the Northern Highlands zone lying between the Apalachicola River on the west, and the Withlacoochee River on the east. This subzone comprises most of central and northern Leon County, and includes the city of Tallahassee. The sediments comprising the Tallahassee Hills are shallow marine to deltaic in origin, with the hill tops composed largely of resistant clayey sands, silts, and clays. The modern hilly topography is the result of post-depositional dissection and erosion by running water. Land surface elevations vary from about 100 feet above mean sea level (MSL) at the southern limit of the Tallahassee Hills to nearly 300 feet above MSL near the Georgia state line.

The clayey nature of the sediments comprising and underlying the Tallahassee Hills has, for the most part, retarded downward percolation of acidic ground waters.

This has resulted in reduced dissolution of the underlying carbonates, and karst features are, with a few exceptions, rare in the highlands region. Some exceptions include the large sinkhole drained lakes in the northern part of Leon County, and smaller sinks located adjacent to the Cody Scarp in south-central Leon County.

Figure 1. Location map showing geomorphic features.

At the southern edge of the Tallahassee Hills, the east-west trending Cody Scarp forms the boundary between the Tallahassee Hills and the adjoining Gulf Coastal Lowlands to the south (Fig. 1). The name Cody Scarp was applied by Puri and Vernon (1959) to a distinct relict marine escarpment traceable across northern Florida and the panhandle. It trends slightly northwest-southeast across south-central Leon County, passing through southern Tallahassee near the Leon County Fairgrounds on south Monroe Street. Here the topographic break has been obscured by erosion, and the elevation drops gently down the former slope through a series of coalesced sand hills and reentrants. Tram Road, which extends southeast and then east from south Monroe Street near the Fairgrounds, runs approximately along the toe of the scarp. East of Tallahassee and into adjacent Jefferson County, the scarp becomes more prominent. It is particularly well developed near the small community of Cody in Jefferson County, for which
the feature is named.

The toe of the Cody Scarp commonly ranges between 75 and 100 feet above MSL, and the crest ranges from 125 to 170 feet (Puri and Vernon, 1964; Hendry and Sproul, 1966). However, dissolution of the underlying limestone has lowered the elevation in many areas along the scarp. Based on elevation, Hendry and Sproul (1966) reported that the Cody Scarp in Leon County appears to be related to the Pleistocene Okefenokee sea.

The Gulf Coastal Lowlands zone (Puri and Vernon, 1964) is an elevational lowlands spanning the coastal areas of the panhandle and the western peninsula of Florida. In northern Florida and the panhandle, these lowlands extend from the toe of the Cody Scarp southward to the modern Gulf of Mexico shoreline.

The Gulf Coastal Lowlands generally represent the area inundated by highstanding Pleistocene seas. In the eastern Florida panhandle the lowlands are underlain by a generally flat-lying, gently sea-ward sloping carbonate plain, comprised of Oligocene and Miocene limestones. This karstified surface is overlain by a variably thick veneer of unconsolidated Pleistocene sand and clayey sand. Relict marine sand bars, dunes, beach ridge systems, and terraces are common throughout the region.

In the Leon-Wakulla County area, two subdivisions of the Gulf Coastal Lowlands are recognized: the Apalachicola Coastal Lowlands and the Woodville Karst Plain. Hendry and Sproul (1966) proposed the name Apalachicola Coastal Lowlands for the flat, sandy region south of the Cody Scarp underlain by thick siliciclastic deposits in western Leon County. Rupert and Spencer (1988) extended this zone into western Wakulla County. It encompasses an area approximately bounded on the north by Highway 20 in Leon County, on the east by Highway 319 in Leon and Wakulla Counties, to the west and southwest by the Ochlockonee River lowlands, and on the south by the Coastal Marshes. The carbonates underlaying the Apalachicola Coastal Lowlands are overlain by poorly permeable sediments, and the area has not been lowered appreciably by dissolution. The terrain is characteristically flat and sandy, with numerous shallow swamp-like densely wooded "bays" and sluggish, poorly defined creeks. Occasional small sinks and dissolution depressions are present, but are much less common than in the karst plain to the east. Much of the area lies within the Apalachicola National Forest.

The Woodville Karst Plain includes the area south of the Cody Scarp and east of the Apalachicola Coastal Lowlands. This zone extends eastward into Jefferson County to the Aucilla River, and slopes southward to the Gulf of Mexico shoreline. The limestone plain continues offshore as a broad shallow continental shelf. Unlike the Apalachicola Coastal Lowlands, the carbonates underlying this zone are covered only by a thin veneer of largely unconsolidated sands and clayey sands. Rainfall percolates rapidly downward through these sediments, producing numerous doline features throughout the area.

Water-filled sinks, numerous dry dissolution depressions, natural bridges, extensive subaqueous cavern systems and disappearing streams are all found here.

Two major streams comprise a fluvio karstic drainage system in the Woodville Karst Plain. The St. Marks River originates in eastern Leon County and flows southwestward, flowing underground a short distance at Natural Bridge, east of Woodville, and ultimately empties into the Gulf of Mexico south of St. Marks. To the west, the Wakulla River originates at Wakulla Spring. It flows southeastward, joining the St. Marks River near the town of St. Marks. Both streams flow in limestone bedrock, and small springs contribute to the flow along their courses.

Throughout the karst plain, a surface veneer of generally less than 30 feet of quartz sand lies on karstic St. Marks Formation and Suwannee Limestone. The result is a surface topography of low sand dunes and other relict marine features. Two relict Pleistocene dune fields are situated in the northern portion of the karst plain. The Wakulla Sand Hills extend from Woodville in Leon County southeastward about two miles into Wakulla County, terminating at the Pamlico terrace shoreline. The dunes are of the barchan type, and suggest a paleowind from the northeast (Hendry and Sproul, 1966).

At the northwestern corner of the Woodville Karst Plain, the Lake Munson Hills extend from the base of the Cody Scarp near Tallahassee southwestward about a mile into Wakulla County. Hendry and Sproul (1966) believed these hills were once an offshore bar associated with the Wicomico sea level stand. The hills overlie an area of interfingering between resistant silts and clays on the west and the porous sands of the Woodville Karst Plain to the east. The resistant sediments under the hills may have retarded downward percolation of ground water and the subsequent dissolution of the underlying limestone. Hendry and Sproul (1966) theorized that the Lake Munson Hills may approximate the pre-solution elevation of the Woodville Karst Plain. Highway 61 (South Adams Street) skirts the northeastern edge of the Munson Hills from approximately the point where it diverges from Highway 363 to a point about one mile southwest of the Capital Circle intersection. Remnant low rolling hills are present along the route, and in places where the road cuts through them, the sugar-white to cream colored Pleistocene quartz sands forming the hills may be observed.

**GEOLOGY**

The eastern Florida panhandle is underlain by thousands of feet of interbedded marine carbonates and siliciclastic sediments, which rest on a complex basement of Paleozoic and Mesozoic igneous, metamorphic, and sedimentary rocks. The oldest rocks influencing the karst geology of the region are Eocene and younger marine carbonates, generally lying at depths of less than 500 feet. These units dip to the west-southwest into a broad
A sedimentary basin named the Apalachicola Embayment centered under Gulf County to the west. Miocene and Oligocene carbonates are brought close to the surface in the Leon-Wakulla-Jefferson Counties area as the units lap up on the flank of the Ocala Platform, a structural high situated to the east of the area.

**North**

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</table>

![Diagram of geological cross section](image)

**Figure 2.** North-south geological cross section in the Woodville Karst Plain.

**Ocala Limestone**

The Late Eocene Ocala Limestone (Dall and Harris, 1892) underlies most of the Florida Platform, and functions as part of the Floridan aquifer system. It is characteristically a white to pale orange to gray bioclastic-calcarenitic limestone. The unit was probably deposited on a broad, temperate, carbonate bank between 45 and 40 Ma. Near Tallahassee, the unit lies at depths of 400 to 500 feet below land surface (unpublished FGS well data). Well logs reveal it is often dolomitized and cavernous in the upper portion of the unit. Although the Ocala Limestone is not penetrated by any known surface-connected conduits, some of the deeper tunnels found at Wakulla Spring may extend down to within 50 to 100 feet of the top.

The Ocala Limestone shallows to the east as it laps up on the flank of the Ocala Platform, a structural high centered under Levy and Citrus Counties. It is exposed at the surface over the crest of this platform, forming a karst plain under the coastal Big Bend area similar to the Woodville Karst Plain.

**Suwannee Limestone**

The Suwannee Limestone (Cooke and Mansfield, 1936) unconformably overlies the Ocala Limestone in the Woodville Karst Plain. It is an Oligocene (33 to 30 Ma) calcarenitic marine limestone or dolomite, comprised largely of foraminifera tests and small mollusks. In the Woodville Karst Plain, the top of the unit varies from approximately 150 feet below land surface at the western edge of the plain to surface exposure in the eastern part of the plain (unpublished FGS well data). Like the underlying Ocala Limestone, the Suwannee shallows as it laps onto the flank of the Ocala Platform, and forms the uppermost carbonate in much of southern Jefferson and Taylor Counties.

Most of the larger mapped subsurface conduits in the Woodville Karst Plain are formed in the Suwannee Limestone. The known portions of the conduits feeding Wakulla Springs extend from a maximum depth of about 310 feet below the spring pool surface to about 140 feet below spring pool surface at the vent. Samples taken from the walls of these conduits show the Suwannee to be comprised of porous, loosely-cemented foraminifera tests, with rare mollusks, bryozoans, and coral molds (Rupert and Spencer, 1988). While the zig-zag course patterns of the conduit segments strongly suggest regional fracture control of tunnel orientation, the horizontal position of the conduits may have been determined by the location of highly permeable and easily-exploitable beds within the Suwannee Limestone. The sizes of some of the rooms developed in the Wakulla Springs conduits are truly spectacular. Divers have reported caverns on the order of 40 feet high and 100 feet wide, pitch black and full of crystal clear ground water.

**St. Marks/Chattahoochee Formations**

Two time-equivalent Lower Miocene carbonate formations, the Chattahoochee Formation (Langdon, 1899) and the St. Marks Formation (Finch, 1823), unconformably overlay the Suwannee Limestone in the eastern Florida panhandle. These units comprise what was once termed the Tampa Formation (Johnson, 1888), a unit which has undergone a number of divisions and redefinitions over the years (Vernon, 1942; Puri, 1953; Puri and Vernon, 1964; Scott, 1988).

The Chattahoochee Formation includes the updip, generally silty, clayey, dolomitic facies occurring from southwestern Leon County west into the panhandle. Chattahoochee Formation carbonates are identified as far south as the Leon Sinks area in southern Leon County (Johnson, 1989).

The St. Marks Formation underlies nearly all of Wakulla County and portions of southern and eastern Leon County. The unit extends eastward as far as Madison County in the northern peninsula. It crops out over much of the southern Woodville Karst Plain, and many area sinks expose St. Marks Formation carbonates in their walls. St. Marks Formation also forms the ledge overhanging the vent at Wakulla Spring, rims the spring pool, and forms the bed of the Wakulla River.

Water filled sinks perforate the St. Marks, and extend...
downward into the underlying Suwannee Limestones. Shallower sinks and cover-collapse depressions may have formed in cavities in the St. Marks Formation; explorations by cave divers have revealed a number of smaller conduits developed in the St. Marks, many of which link up with deeper tunnels in the Suwannee Limestone.

South of the Cody Scarp and east of U.S. 319, the St. Marks Formation is overlain by shallow relict Pleistocene undifferentiated sands and clayey sands. Erosion associated with high-standing Pleistocene seas has removed the younger Miocene and Pliocene units. In the Northern Highlands and in portions of the Apalachicola Coastal Lowlands, the younger units remain, forming the core of the Tallahassee Hills, and shielding the underlying carbonates from extensive dissolution. These younger units include the Miocene Torreya Formation and the Pliocene Jackson Bluff and Miccosukee Formations.

Hawthorn Group - Torreya Formation

The Miocene Torreya Formation of the Hawthorn Group includes all deposits in the eastern Florida panhandle previously referred to as Hawthorn Formation (Huddleston and Hunter, 1982; Scott, 1988). It is largely a siliciclastic unit, consisting of very fine to medium grained clayey sands to sandy, silty clays, and frequently containing variable amounts of limestone, dolomite, and phosphate pellets. Carbonate content generally increases in the basal portion of the unit.

The Torreya Formation occurs under northern and central Leon County, and extends into western Wakulla County under the Apalachicola Coastal Lowlands (Rupert and Spencer, 1988). It is present in the wall of Big Dismal Sink, but in most of the Woodville Karst Plain it has been removed by erosion.

Jackson Bluff Formation

The Upper Pliocene Jackson Bluff Formation (Puri and Vernon, 1964) is comprised of light gray to greenish-gray, mollusk-rich fossiliferous sands, clayey sands, and sandy clays. It occurs in western Leon County, and extends southward into western Wakulla County, underlining the Apalachicola Coastal Lowlands at a depth of approximately 20 feet below land surface. Due to its clay content, it retards downward percolation of ground water. Many of the bays and other shallow ponds in the Apalachicola Coastal Lowlands are probably perched on this unit.

Miccosukee Formation

The Late Pliocene Miccosukee Formation (Hendry and Yon, 1967) is a siliciclastic unit capping the Tallahassee Hills. It is characteristically grayish-orange to grayish-red, fine to medium grained clayey sands, containing quartz gravel and thin, discontinuous clay beds and laminae. Excellent exposures of the unit may be seen along roadcuts in and north of Tallahassee, especially along some of the famous "canopy roads" such as Centerville Road and Miccosukee Road. Due to its clayey nature, it holds a steep slope, with little or no slumping.

The Miccosukee Formation is restricted to the Northern Highlands, and does not occur south of the Cody Scarp. In western Leon County, it overlies the Jackson Bluff Formation. From Tallahassee westward and southward to the Cody Scarp, it unconformably overlies Torreya Formation sediments.

Precipitation runs off the Miccosukee surface rapidly. A series of creeks and streams drain southward from the Tallahassee Hills, and most are captured by sinks after crossing the Cody Scarp. Cave divers report common occurrences of a crusty "goethite or limonite-like" coating on the walls of conduit systems in the Woodville Karst Plain. These deposits may represent a precipitate from the iron-rich surface water drained from Miccosukee sediments to the north.

Undifferentiated sands and clays.

Pleistocene and Holocene undifferentiated quartz sands, silts, and clays comprise the undifferentiated surficial sediments in the Woodville Karst Plain (Schmidt, 1979; Rupert and Spencer, 1988). Most Pleistocene sediments are marine terrace deposits, lying unconformably on older formations.

In eastern Wakulla County, the St. Marks Formation and Suwannee Limestone are overlain by a veneer of fine, unconsolidated sands and clays, generally less than 20-feet thick. West of Crawfordville, these sediments thicken to as much as 100 feet and lie directly on the Jackson Bluff or Torreya Formations. Holocene alluvial and aeolian deposits are predominantly fine quartz sand, and are difficult to differentiate from Pleistocene sediments.

HYDROGEOLOGY

The Woodville Karst Plain comprises a unique hydrogeologic setting influenced by both regional ground water inflow and local surface water recharge. The Tertiary carbonates comprising the bedrock plain are part of the Floridan aquifer system, which carries high quality drinking water along a generally southeasterly gradient. In much of the northern portion of the Woodville Karst Plain this aquifer system is an unconfined water table aquifer, rising to within a few feet of the surface in some areas in the southern part of the karst plain. The porous nature of the overlying sediments as well as the presence of numerous wet sinks directly penetrating the aquifer system allow rapid local recharge by precipitation.

Along the northwestern edge of the Woodville Karst Plain, several tannic surface streams flowing are captured by siphoning sinks, contributing a significant surface water component to the local ground water. In addition, precipitation flushes tannin laden runoff into the small sinks throughout the karst plain, and various degrees of subterranean mixing occur between the clear regional
ground water and local tannic surface waters. This mixing often manifests itself in distinct color and clarity changes in resurgence points such as Wakulla Spring, especially after periods of high precipitation.

Greatly influencing the degree and speed of mixing is a vast, interconnected series of subaqueous dissolution conduits underlying the karst plain. The Woodville Karst Plain contains some of the longest conduit systems in the world. Some of the larger of these dissolution conduits have been explored and mapped by cave divers. Maps produced by the divers confirm that the primary conduit directions closely parallel regional bedrock-related lineaments. Linear surface features are quite common in Florida. A cursory examination of a map of Florida reveals many of the State's prominent geomorphic features, including segments of both coasts, many river courses, and major sinkhole trends assume distinct northwest-southeast or northeast-southwest bearings. These primary lineament directions may represent regional fractures.

The same northwest-southeast and northeast-southwest fracture patterns are evident in a local lineament map derived from airphoto examination (Figure 3). These lineaments are most obvious in the clayey sediments of the Northern Highlands. They are largely masked to the south by the unconsolidated sediments covering the Gulf Coastal Lowlands. However, the alignment of small sinkholes, as well as many of the larger karst features closely follow the primary lineament directions.

![Figure 3](image_url)

**Figure 3.** Regional surface lineament patterns derived from airphotos (from Windham and Sproul, unpublished FGS in-house study).

Several creeks and streams draining the Northern Highlands and the Apalachicola Coastal Lowlands are captured by sinks along the northwestern edge of the Woodville Karst Plain. On such example is Fisher Creek, a small meandering stream situated southwest of Tallahassee (Fig. 4). After flowing southeastward across the Cody Scarp, the stream siphons into a sink in the Leon Sinks Geological Area. Fisher Creek Sink (Figure 4) is the one of the northernmost siphoning sinks in a line of northwest-southeast trending sinks which line up very closely with Wakulla Spring at the southern end. This trend was observed early in the century, and some early workers presumed that Fisher Creek was the northern extension of the Wakulla River. Interestingly, the first soil survey map of Leon County (Wild et al., 1906) showed Fisher Creek labeled as the Wakulla River. Later, Sellards (1917) provided a rough map showing the nearly linear alignment of Fisher Creek and Wakulla Spring; he proposed that the entire system was one long subterranean conduit, and the surface flows of Fisher Creek and the Wakulla River were portions of conduit where the roof had collapsed.

Although perhaps oversimplified, Sellards ideas are being confirmed in part today by the on-going conduit exploration and survey work by cave divers. Many of these divers, working entirely on their own time and money, have generously shared their discoveries and personal observations with the staff of the Florida Geological Survey.

Figure 4 shows the conduit mapped to date. The main conduit trend begins north of Sullivan Sink, which is situated at the western edge of the Leon Sinks Geological Area. From this sink, divers mapped an upstream tunnel for a distance of 7,980 feet northwest of the sink. A downstream tunnel of similar distance was also explored. Next, they entered at Cheryl Sink to the southeast, exploring the conduit segments north and south of this sink as well. During an exploration push upstream from Cheryl Sink, the divers discovered the southern end of a dive line they had laid earlier while exploring Sullivan Sink to the north. This confirmed a Sullivan-Cheryl sink connection existed. Later, a four man team traversed the entire distance from Sullivan in Leon County and exit at Cheryl Sink in Wakulla County. This feat broke several world cave diving records, including the record at that time for longest underwater cave traverse.

Subsequent work by divers with the National Speleological Society's Woodville Karst Plain Project has resulted in the discovery of more continuous conduit system to the southeast. The general trend of the system appears to head for Indian Springs, and possibly ultimately to Wakulla Spring. It is a complex system, generally carrying clear regional ground water in a southeasterly direction. Along its course, small unmapped tributary tunnels contribute water of various colors and clarity, likely from nearby surface sources (Parker Turner, personal communication, 1990).

Recent mapping by divers centers on the area south of Val Halla (or Whiskey Still Sink). The major tunnel continues south of Valhalla Sink, through and southeast of Innisfree Sink to Turner Sink, an explored distance of about 6,000 feet. Such distances begin to pose severe endurance problems for the divers, especially at the depths in excess of 190 feet which these tunnels
Figure 4. Map showing extent of mapped karst conduit in the Woodville Karst Plain (modified from Irving, S., 1992).
commonly assume. Before the exploration can be pushed further south, new accessible sinks which penetrate the cave system must be located. Often, the smaller sinks are plugged with collapse breakdown or with tree logs, making them inaccessible. This has been a continuing problem for divers, and could preclude further downstream explorations.

In an effort to establish a connection without direct human exploration, dye tracing was attempted at Innisfree Sink, south of Valhalla. Twenty five pounds of Rhodamine WT dye were released 100 feet downstream of Clark Sink. Dye traps were placed in an unnamed sink 3000 feet southeast of Clark Sink, in Indian Spring’s upstream tunnel, and in all five tunnels leading into Wakulla Spring. No trace of dye was found in any of the traps, and it was later decided that too little dye had been used in the test.

In recent years, the water clarity at Wakulla Spring has deteriorated significantly, primarily from varying degrees of tannic coloration. This coloration appears to be becoming more frequent in occurrence in recent years, and concerns as to whether land use changes in the recharge area north of Wakulla may be affecting it. Plans are currently being formulated for a long term study, using water chemistry comparisons, to attempt to determine relationships between captured tannic surface water at the northern edge of the Woodville Karst Plain and the water quality at Wakulla Spring. Such a study, in conjunction with continued explorations by cave divers, should help shed light on this interesting hydrogeologic area.

**ROAD LOG**

<table>
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<tr>
<th>Mileage</th>
<th>Description</th>
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</thead>
<tbody>
<tr>
<td>00.0</td>
<td>Florida Geological Survey, corner of Woodward and Tennessee Streets, Tallahassee, Florida. From Antarctic Circle at the rear of building, turn right onto Woodward Street and proceed south.</td>
</tr>
<tr>
<td>00.6</td>
<td>Gaines Street intersection. Turn left, and proceed east on Gaines Street.</td>
</tr>
<tr>
<td>01.5</td>
<td>Bronough Street. Turn right and proceed south.</td>
</tr>
<tr>
<td>02.0</td>
<td>Bronough merges into Adams Street (SR 363) at light. Continue straight (south).</td>
</tr>
<tr>
<td>02.5</td>
<td>The crest of the hill marks the approximate southern extent of the Tallahassee Hills. From this point, with an elevation of approximately 150 feet above mean sea level, the land surface drops down the face of the Cody Scarp and onto the Woodville Karst Plain. The Cody Scarp is poorly defined in this area, having been modified by erosion and karst dissolution in the underlying carbonates. South of the Scarp, the road continues over gently-rolling sand hills.</td>
</tr>
<tr>
<td>04.0</td>
<td>Four Points intersection. Continue on south Adams Street, which curves to the right and becomes SR 61.</td>
</tr>
<tr>
<td>05.0</td>
<td>Church Sink, a 100 feet diameter water-filled sinkhole is situated out of sight in the woods on the left, just before the Baptist church. St. Marks Formation limestone is exposed in several pinnacles around the rim of the sink. The sink perforates the Florida aquifer system, and has gained notoriety because the Florida Department of Transportation funnels storm water runoff into the sink through a corrugated pipe on the west edge of the sink. The surrounding terrain is typical Woodville Karst Plain - flat to gently-rolling, sandy pine and scrub oak woods. Karst terrain becomes apparent from this point south, as small water-filled sink and cover-collapse depressions become more common.</td>
</tr>
<tr>
<td>05.5</td>
<td>Whittaker Road intersection on right. In 1990, a 50-foot diameter sink opened about 100 yards west of Adams, swallowing oak trees and a portion of Whitacre Road. The hole was filled, and the road repaved.</td>
</tr>
<tr>
<td>05.9</td>
<td>Intersection of Adams and Capital Circle SW (SR 263). Proceed south through the light. Adams street becomes Crawfordville Highway (SR 61 / US 319). The road traverses low, rolling, relict Pleistocene sand hills of the Lake Munson Hills geomorphic zone. These deposits may represent offshore bars associated with the Wicomico sea level highstand.</td>
</tr>
<tr>
<td>07.6</td>
<td>SR 61 turns off to the left. Continue straight on US 319.</td>
</tr>
<tr>
<td>11.8</td>
<td>Entrance to Leon Sinks Geological Area, Apalachicola National Forest, on left. Turn in and proceed to parking area. Meet at trail head, west end of parking lot. Take blue-marked trail east past Gopher Hole Sink.</td>
</tr>
</tbody>
</table>
STOP 1: The Leon Sinks Geological Area

The Leon Sinks Geological Area is a unique three-quarter square mile special interest area within the Apalachicola National Forest. A 3.1 mile marked sinkhole loop trail, which visits a variety of karst features, starts at the edge of the parking area. A crossover trail, marked with white blazes, heads west to the Fisher Creek Sink/Natural Bridge area. From this point north, the sink trail is marked with blue blazes on adjacent trees. Figure 5 is a map of the Leon Sinks Geological Area, showing the locations of named sinks and the routes of the marked trails.

Proceed west from the parking lot on the crossover trail. Gopher sink is on the right side of the trail. A large rock ledge may be observed in the south wall of this normally dry sink. Undifferentiated Pleistocene sands are visible in the sloping sides of the sink. Beyond Gopher Sink, the trail crosses a swampy lowlands, populated with cypress, slash pine, and palmetto. Wire grass, long leaf pine, and magnolia are present on the higher, sandy portions of the route.

The crossover trail ends at the natural bridge. Fisher Creek, which originates in eastern Gadsden County, flows southward through west-central Leon County, and siphons into Fisher Creek Sink, on the south side of the natural bridge. It flows underground a short distance, and reemerges at Fisher Creek Rise, on the north side of the natural bridge. The creek then flows northwest approximately 500 feet, and disappears underground into Lost Stream Sink. Fisher Creek is normally a dark tea color due to high tannic acid content. Its flow is largely dependent on local rainfall amounts.

North of Lost Stream Sink, the trail climbs northeastward across a sandy uplands to Black Sink. Black Sink is a large, deep sink which remains wet. From Black Sink, the trail makes a broad loop to the northwest, passing dry Magnolia and Big Eight sinks, then returns southeastward to Big Dismal Sink.

Big Dismal is the largest sink in the area, with a diameter of approximately 250 feet, and exposed walls reaching over 50 feet high, depending on the water level in the sink. Local divers report that the sink extends to a depth of about 90 feet below the water level, and has approximately 12,000 feet of mapped conduits associated with it (Mike Wisenbaker, personal communication, 1993). Big Dismal formed at the crest of a sand hill associated with the Lake Munson Sand Hills, which resulted in its large funnel and impressive depth. At average water levels, Miocene carbonates of the St. Marks and Chattahoochee Formations are exposed in the lower third of the sink wall, immediately above the water line. These units are predominantly light brown to very light orange, quartz sandy, dolomites and limestones, containing fossil mollusk molds. Approximately 16 feet of Miocene age Torreya Formation unconformably overlies the older Miocene units. The Torreya is comprised of light brownish gray to light grayish brown, quartz sandy, fossiliferous limestone, containing fossil oysters and pecs. The upper 25 feet of the sink wall is covered by undifferentiated Pleistocene sands and clayey sands. Figure 6 shows a generalized stratigraphic section in Big Dismal Sink.

Proceeding east from Big Dismal Sink, the trail gradually descends through a sandy pine uplands, past Tiny Sink, and curves around to Hammock Sink. Hammock Sink, or Little Dismal Sink as it has been known in the past, is connected to an extensive cave system. A profile map of this cave is shown in the park brochure, available at the trail head. The sink basin is about 30 feet deep, formed in Miocene carbonates. A shallow siphon tunnel extends about 360 feet northwest at a depth of 60 feet below the sink pool surface. A deeper tunnel, accessible through a narrow restriction, extends northeastward from the basin and feeds into a northwest-southeast trending flowing conduit system, lying at a depth of about 180 feet below the water surface. Several immense rooms or caverns occur along this route, some attaining dimensions of 150 feet long and 60 feet high. This deeper conduit carries regional
ground water, and may be linked to the major northwest to southeast trending conduit system which resurges at Wakulla Spring (Parker Turner, personal communication, 1989).

From Hammock Sink, the trail turns northeastward, then jogs south for the last leg. Along the last section, the trail passes between Back and Palmetto Sinks, both normally dry sinks. Turner Sink, a short distance further south, is the last wet sink on the trail. After passing Dry and Cone Sinks, the trail rejoins the entrance trail segment; turn left and proceed back to the parking area.

12.4 Entrance to Leon Sinks. Turn right onto Highway 319 and proceed southwest.

15.6 Intersection of Highway 319 and SR 267. Turn left at flashing light.

17.2 Quarry Springs, the site of a closed shallow limestone pit, from which the St. Marks Formation carbonates were mined for use as roadbed material.

18.4 Entrance to Indian Springs on right, a large, water filled sink which serves as a recreation area for the local YMCA. The conduits feeding the spring are currently being explored by cave divers in an attempt to determine their relationships to Wakulla Springs and the regional conduit network.

19.1 Intersection of SR 267 and SR 61. Proceed straight through flashing light.

19.2 Entrance to Edward Ball Wakulla Springs State Park. Turn right into park, and proceed through entrance station to the parking area in front of the Wakulla Springs Lodge.

STOP 2: Wakulla Springs

Wakulla Springs are comprised of the main spring, situated at the south end of a large spring pool which forms the headwaters of the Wakulla River, and nearby Sally Ward Spring. Smaller Sally Ward Spring is located about 0.7 mile northwest of the main spring, and discharges into the spring pool through a low, swampy spring run. Both springs may have originated as Pleistocene sinks formed from the collapse of caverns in the underlying limestone. The two springs are the largest spring system in the Woodville Karst Plain, with an average combined flow of 390 cubic feet per second for the 67 years prior to 1974 (Rosenau et al., 1977). The main spring is fed by a series of five conduits, carrying crystal clear regional ground water and local tannic surface water in varying mixtures (Rupert and Spencer, 1988). It shows considerable variation in discharge. A minimum recorded flow of 25.2 cubic feet per second was recorded on June 18, 1931, and a maximum flow of 1,910 cfs occurred on April 11, 1973 (Rosenau et al., 1977). Flow is measured downstream of the pool at the Highway 365 bridge.

The springs have been a popular swimming spot for much of the last century. They gained fame in 1930 when a nearly complete skeleton of American Mastodon (Mammut americanum) was recovered from the spring pool. This skeleton is now mounted in the Museum of Florida History in downtown Tallahassee. Until the advent of SCUBA, the karst conduit feeding the spring was largely unknown. In 1955, the first organized underwater exploration of the main spring was conducted by local divers. They explored a distance
of approximately 1100 feet into the main spring cave, where the cavern reaches a depth of over 200 feet below spring pool surface. An interesting narrative of the divers' exploration is given by Olsen (1988). Working in conjunction with the divers, Florida Geological Survey paleontologist Stanley Olsen prepared the cross sectional map of the main cave shown in Figure 7. Numerous Pleistocene fossils and aboriginal relics were discovered during the exploration. The divers recovered bones of Pleistocene mastodons, sloths, and deer, as well as over 600 bone spear points manufactured by indians residing near the spring.

The most extensive exploration of the Wakulla Springs cave system, and that which provided much insight into the hydrogeology of the spring, occurred during a two month period in 1987. The United States Deep Caving Team conducted a multifaceted study inside the conduits, which included filming the system, mapping, and the collection of lithologic samples from the cave walls (Stone, 1989). Over two miles of conduit system were mapped during this project (Figure 8). The primary spring conduit trends southwest from the spring entrance, and at a distance of 900 feet in from the vent, splits into five separate conduits. Tunnels "A" and "C" head southward, while Tunnel "B" heads east, and Tunnel "D" trends north to within 1200 feet of a similar conduit discovered by divers in Sally Ward Spring. The tunnels vary in size, and are generally elliptical in cross section. The sinuous nature and angular sweeps of the conduit segments may be a function of regional fracture directions coupled with high water flow, a common condition in flat-lying limestones (White, 1988). While portions of the tunnels may have been dry during low sea level periods of the Pleistocene, no dry cave speleothems are present. The cavern walls typically show scalloping, a feature produced by constant turbulent flow (White, 1988).

Large cavern "rooms" are common in the tunnels. Diver Wes Skiles (personal communication, 1987) reported one average size room in the main cavern with dimensions of 47 feet high and 117 feet wide. Time and diver endurance limited the extent exploration into the conduits. Remarkably, Tunnel "B" was explored a total of 4500 feet in from the spring vent, representing the greatest distance covered by the divers. The deepest depth reached during the project, 360 feet below spring pool surface, was recorded in tunnel "B" near the 4500 feet limit.

Water flow in all the tunnels converges near the Grand Junction Depot Room, from which it northeastern out the main cave to the spring vent. Many of the conduit segments assume directions approximately parallel to the region lineament directions, and they may be fracture-controlled. Water quality within the conduits showed somewhat differing characteristics. Tunnels "B", "C", and "D" carried "air clear" water, while Tunnel "A" carried tannic (tea-colored) water. This tannic water is believed to be derived from surface sinks, possibly as far north as the Leon Sinks Area in Leon County. The degree of tannic coloration increases with the amount of local precipitation. After periods of heavy rain, the normally crystal clear spring pool may be too turbid to see the spring vent.

Ever since Sellards (1917) proposed that Wakulla Springs represented the resurgence point of a karst conduit system originating at the northern edge of the Woodville Karst Plain, research has focused on establishing this link. For years prior to Sellard's work, Fisher Creek in western Leon County was named the Wakulla River (Wilder et al., 1906). This was probably due to an assumption at the time that it was the northernmost surface segment of a largely underground system extending southwestward, emerging for a short distance at river sink, and continuing to its final resurgence at Wakulla Spring. Recently, cave divers have in part confirmed Sellards idea. A four mile long subaqueous conduit trend has been mapped, extending from the Leon Sinks Geological Area in southern Leon County southeastward to within 3 miles of Wakulla Springs.

Water clarity permitting, a glass-bottom boat tour takes visitors out over the spring vent. The cavernous opening lies at a depth of about 140 feet below the pool surface, under a jutting limestone ledge.

22.5 Entrance to Wakulla Springs State Park. Turn left and proceed west on SR 267.

22.6 Intersection of SR 267 and SR 61. Turn right into SR 61 and proceed north.

30.5 Stop sign at SR 61 and U.S. 319. Blue Sink, a shallow sinkhole lake used as a National Forest recreation area, is on the right. Turn right towards Tallahassee.

32.2 Intersection of U.S. 319 and Capital Circle. Proceed straight through light. Road becomes South Adams Street.

34.2 Four points intersection. Stay on south Adams as it curves left.
Figure 8. Block diagram showing conduits feeding Wakulla Springs (from Rupert and Spencer, 1988).
35.3 Begin ascent up face of Cody Scarp, starting about 60 feet above mean sea level, and rising to 150 feet above mean sea level at the crest of hill, at 35.6 miles.

36.2 Veer left at light onto Bronough Street, which angles off to the left from Adams Street.

36.7 Intersection of Bronough and Gaines Streets. Turn left.

37.6 Intersection of Gaines and Woodward Street. Turn Right onto Woodward.

38.2 Pedestrian crosswalk with light. Proceed through light and immediately take the first right turn onto Florida State University campus. Proceed to stop sign, and turn left onto Palmetto Drive. Pass through tunnel under Woodward Street. Take the first left onto Antarctic Circle.

38.5 Florida Geological Survey building.

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—, and Vernon, R.O., 1959, Summary of the geology of Florida and a guidebook to the classic exposures: Florida geological Survey Special Publication 5, 225 p.


COASTAL PROCESSES OF THE FLORIDA PANHANDLE - AN INTRODUCTION FOR UNDERGRADUATES AND SCIENCE TEACHERS

William C. Parker

Department of Geology, Florida State University, Tallahassee, FL, 32306

ABSTRACT

The coastal environment is an ideal place to observe the dynamic interaction of a variety of geologic processes and forces that shape sedimentary deposits and affect humanity. A variety of simple techniques are introduced that allow teachers and students (without access to a large funding base) to document the dynamics of this environment. Aspects of the coastal environment discussed include beach profile, beach face sedimentary structures, beach drift, barrier island formation and migration, and sea level change. Examples from Florida Panhandle coastal environments are discussed. Interaction between human development on coastal lands and natural processes, and the State’s attempts to regulate development are discussed.

INTRODUCTION

For geoscience students in the southeast, and especially those in Florida, the geology of coastal areas is a fascinating and rewarding topic of study. The coastal zone is a dynamic realm of interaction between forces and processes of the marine and terrestrial environments. As such, it represents a useful natural laboratory for observing and measuring sedimentary processes, and for formulating the models necessary to interpret both the history of the coastline being observed as well as coastal sediments preserved in the rock record.

The purpose of this field trip is to acquaint the participant with easily observable coastal processes, their results, and simple techniques for measuring/recording/examining them.

While the number of published studies on coastal geology is large, this should not deter the beginning researcher from attempting to make a contribution. Coastal dynamics can be very complex and variable from location-to-location, and many fundamental relationships have yet to be well explained and/or quantified.

The relevance of coastal studies goes far beyond understanding coastal processes to embrace many of the "hot" topics of contemporary scientific concern. For example, the current interest in "global warming" has emphasized the need to unravel the immediate past history of sea level and project possible impacts of sea level rise. Rapid development of coastal areas, coupled with the public awareness of the destructive power of hurricanes (e.g. the destruction caused by Hugo and Andrew), have emphasized the need to understand how development of the coastal zone affects the dynamics and to what extent the future of a coastal area can be predicted.

From a practical point of view, many aspects of coastal geology are easy, inexpensive, and often enjoyable to study. All students of geology can benefit from first-hand observation of the processes operating there. Teachers may also want to consider class laboratory exercises on coastal processes to illustrate geoscience research.

FIELD AREA

The coast to the southwest of Tallahassee (see Fig. 1), along the Florida Panhandle, represents an ideal area to observe coastal processes. Here the coastal plain is drained by the Apalachicola River (largest river in Florida and 21st in the contiguous U.S.) which, along with its predecessors, has supplied fluvial-deltaic sediments to this region for most of the Cenozoic. Deposition of these sediments in the river estuary has built the Apalachicola Delta. Study of the most recent accumulations suggest that the delta is prograding rapidly (> 2m/yr, Donoghue, 1989). Offshore, the coast is protected by a line of barrier islands. From east-to-west the islands are: Dog Island, St. George Island, Little St. George Island (Cape St. George), St. Vincent Island, Cape San Blas and St. Joseph Peninsula.

The islands provide a protective barrier for the bay or sound waters (between the islands and the coast). These protected waters provide a "holding basin" for fine-grained sediments coming in from the Apalachicola and decrease mixing with open Gulf of Mexico waters. As a result the bay waters vary considerably in salinity (related to river discharge) and support considerable organic productivity in both the bay waters and surrounding tidal marshes. The primary mechanism of exchange for bay waters is through tidal activity in the passes between the islands. Circulation in the bay is primarily in from the Gulf through the eastern passes and out through the west.

The barrier islands are extremely dynamic entities, noticeably eroding, prograding, and migrating over the past century and a half. Most movement is attributable to sea level change and reworking by waves from the Gulf of Mexico. The dominant direction of regional wave approach is from the east-southeast (Walton, 1973), for both fair-weather waves and storm swell. Thus coasts which face east-southeast receive the brunt of wave energy. Most of the incident waves (> 70%) have heights under 1.2 m and periods under 5.5 seconds.

Due to the close proximity of this area to Tallahassee, considerable investigatory work has been done by personnel from the Florida Geologic Survey and the Geology Department of Florida State University. A good introductory guide to the geology of the area (Campbell, 1984) as well as additional information about this and other Florida coastal
Figure 1. Map of Gulf Coast southwest of Tallahassee.
areas is available through the Florida Survey office in Tallahassee. The Symposia Proceedings on Coastal Sedimentology, published by the Geology Department at FSU, contains numerous articles on coastal areas, particularly those in the northern portion of the state. In addition, numerous unpublished Master's and PhD theses/dissertations (available at the Departmental office or the FSU Dirac Science Library) also deal with various aspects of coastal geology.

LIFE AS A BEACH

Sandy beaches are characterized by some of the most intensive and constant physical reworking of all sedimentary environments. However, not all portions of the beach experience the same processes or are affected at the same rate.

Barrier island beaches (see Fig. 2) can be divided into nearshore, foreshore and backshore zones. The nearshore zone is the environment of high wave action and extends from deep water to mean low tide (MLT) and commonly includes one or more longshore bars, the shallowest of which may be just awash at low tide. The foreshore and backshore zones are the beach proper. The foreshore slopes more steeply seaward and may have a beach ridge, runnel (beach-parallel depression at the base of the beach face) and plunge step developed in the lower beachface (portion exposed to direct overwash by the surf). The backshore zone may be nearly horizontal or contain a concave-upward storm beach separated from the foreshore by a raised berm. Behind (landward of) the backshore are dunes backed by gently landward-sloping marshes and washover sediment fans, eventually giving way to the bay or sound waters.

In the breaker, surf and swash zones, surficial sediment is in almost constant motion as the environment reacts to the addition of energy in the form of incoming waves. The offshore-onshore profile of the ideal beach represents an equilibrium between the sediment supplied to the shore and the incident waves. A change in either the nature or amount of the supplied sediment or the character or direction of the waves will lead to a change in the beach. Most (if not all) observable beaches are non-ideal in that the sediment/wave conditions appear to change before the beach achieves equilibrium. This leaves the beach in a constant state of flux as it reacts to daily, seasonal and longer term effects. This constant reworking yields the excellent sorting characteristic of beach deposits.

Measurement of Beach Profile

A precise measurement of beach profile requires the use of surveying equipment. However for those of limited budget, a tape measure (not steel), an inexpensive hand level, two meter sticks and a few pieces of 2"x 2" lumber can be used to do a reasonable job (see Fig. 3).

One 2x2 should be cut to approximately 1.7 m in length and the hand level affixed to one end such that the level is level when the 2x2 is held vertical with the other end on the ground. This completes the "instrument." The other 2X2 is cut to the same length as the first and a meter stick affixed to the side such that the 50 cm mark is even with end of the 2x2. This completes the "rod." If this has been done correctly, one person can hold the instrument at arm's length, with the level reading level, and sight along the top of the level at another person holding the "rod" vertical across a level room and the sight line from the instrument should intersect the rod at the 50 cm mark.

A number of "improvements" are possible. Addition of flat feet to the bottoms of both the instrument and rod (make sure to keep both the same) may help prevent sinking in soft sand. Paper, wood, or cardboard "sights" attached to the top of the level may make sighting easier. Attaching a mirror to the level may help the instrument user read the level while sighting.

Use of this equipment requires a minimum of two people, but works best with 4-6 (instrument operator, level watcher, rod operator, recorder, 2 tape operators).

Ideally you start from a fixed, recognizable position on the beach that will not change over time (usually high on the beach). The foot of the instrument is placed on the reference point and the rod operator is sent a few meters down the beach, at right angles to the shore. The instrument operator sights to the rod and calls out the level of intersection. The tape is used to measure from the rod position to the rod (remember to measure horizontally) and the elevation difference and distance recorded. The rod can be moved farther down the profile for another sight from the same instrument position or the instrument can be relocated on the previous rod position. Keep the sight distances short (<15 m) to enhance precision and don't sight over an elevation difference of more than 50 cm. Be sure to measure the sea level and note the tidal state. Feel free to extend the profile as far into the water as you feel is warranted.

The profile can be plotted on graph paper (vertical exaggeration may be necessary to enhance some of the features). Repeat the profile as often as wave conditions change and compare the changes with the sizes and direction of waves preceding the measurement. Profiles taken immediately following major storms can be dramatically different from fair weather profiles. Changes in beach profiles over short (few days to seasonal) time spans are usually attributed to movement of sand on- or off-shore in response to changing wave conditions.

Profiles collected along the South Carolina coast before and after Hurricane Hugo (see Fig. 4, from Katuna, et al., 1990) reflect the erosion of sand from the dunes and backshore and subsequent deposition in the nearshore. Following the storm, fair weather waves have begun to restore the pre-hurricane profile. Seasonal variation (largely due to seasonal differences in storm size and frequency) also alters the profile as shown in Figure 5, for data taken from Little St. George Island (from Donoghue, et al., 1990).

Sedimentary Structures

Portions of the past history of beach faces are preserved in the immediate subsurface of the beach itself. Obviously,
Figure 2. Environments in the beach and nearshore zone (after Davis, 1983).

Figure 3. Sketch of beach profile measuring apparatus.
Trenching can reveal (1) planar, seaward-dipping crossbeds of the reworked beach face, (2) steeply landward dipping crossbeds of a migrating bar, (3) trough crossbedded sands of the beach runnel, and (4) heavy mineral and shell lag concentrations reflecting dynamic sorting by swash.

**Sediment Peels**

In addition to photographing and sketching the structures observed, it is possible to remove some for later study or display. The technique for doing this is described by Moiola, Clarke, and Phillips (1969) and employs Elvacite, an acrylic resin.

**Sediment Motion**

Sediment motion in the nearshore and swash zones reflects not only an on- and off-shore component but also longshore transport, controlled by interaction between the orientations of the beach and incident waves. Tracer studies can be used to demonstrate short-term (few minutes to few hours) transport dynamics. For longer term studies, the volume of tracer needed becomes prohibitive.

Simple tracer experiments can be carried out using painted sand of the same size common on the beach. Fluorescent paints are most easily traceable; just be sure to disaggregate the sand after the paint is dry to avoid clumping. The sand is dumped in the swash zone (or zone of interest), and the sand a fixed distance (e.g. 10 m) down-transport is watched for the first appearance of a painted grain. The time to arrival of the dispersing "cloud" of painted grains can be visually estimated. For a more quantitative estimate of transport rates, the tracer grain population at the down-transport location can be sampled at regular time intervals using 3"x5" note cards covered with a sticky substance (e.g. vaseline, contact adhesive, etc.) touched face-down on the sand and the number of tracer grains counted. This procedure should yield a "bell shaped" curve of tracer frequency over time.

**HISTORY OF SEDIMENT MOTION**

Long term patterns of sediment motion can be assessed by mapping erosional versus depositional/accretionary portions of the coast. Evidence of erosion includes truncation of previously deposited beach ridges, beachface erosional scarps, and exposed tree stumps and trunks on the beach face. Accretional evidence includes welding of offshore bars to the beachface and construction of beach ridges.

Most of the barrier islands in this area show a predominance of erosional features, especially on the southeast facing portions. Only small segments on the ends of Cape St. George and St. Vincent show evidence of pronounced accretion.

Historical trends in sediment motion can be assessed through map differencing. Comparison of bathymetric charts from 1837 to present (Donoghue, et al., 1990) shows that the southeast facing coasts of the barriers have been retreating under erosion from 1 m/yr (east Little St. George) to 2.5 m/yr (Dog Island). The bay facing coasts of most
islands are also eroding (<1 m/yr) thus making the islands thinner. During the same interval, however, the shoreline west of Cape St. George has been advancing at 2 m/yr.

As a result of eastward erosion and westward deposition, the capes (Cape St. George and Cape San Blas) are migrating westward at rates in excess of 8 m/yr.

STORMS

Storms represent extreme short-term variations in the wave regime to which the shore environments respond. However, because of the enormous energy involved, a large storm may cause changes in the shore environment that may take many years to decades to erase. If storm frequency is high enough, the sedimentary record may be primarily one of storm related deposition.

Historic records from the middle 1800's to the present indicate that major hurricanes have passed within 100 km of St. George Island approximately an average of once every 9 years. Many of these storms were responsible for opening inlets or passes in the islands or producing major overwash fans on them. A major storm in 1837 brought a surge 3.0-4.5 m above normal and opened "New Inlet" (closed by 1900). Sand Island Pass was opened by a storm in 1852, had also closed by 1900. Hurricanes Agnes (1972) and Elena (1985) produced 2.4 m surges and did extensive coastal damage. The most recent major storm to pass the area was Hurricane Kate (1986) and some of the shoreline erosion attributable to that storm can still be seen.

A storm producing a 2.5 m surge is a 20 years storm for these coasts (Ho and Tracy, 1975). A 100 year storm surge is approximately 3.6 m, the approximate elevation of most of the high points on these islands.

SEA LEVEL

According to most data sources for eastern North America, sea level has risen 90+ m in the last 20,000 years, with the most rapid portion (16,000-6,000 years B.P.) averaging over 5.0 mm/year. From 6,000 years B.P. to present, sea level has fluctuated by several meters while rising at 1.2 mm/year (from data in SW Florida, Scholle and Suuver, 1967; cited in Donoghue, et al., 1990). This history has had a pronounced effect on the history and development of the coastal features in the Gulf.

The history of sea level during the last few thousand years is more difficult to discern, but important to the interpretation of the reported historically observable 12-15 cm rise over the past century. One method of estimating sea level history is by looking at preserved beach ridge heights on sections of the coast that have historically remained constructional over the period of interest. Beach ridges are swash-built ridges of sand that are periodically built on the beach face on some constructional coastal areas. They can be seen easily on aerial photos as linear or slightly arcuate ridges and intervening swales.

Tanner and others (e.g. Stapor and Tanner, 1977; Tanner et al., 1989) have carried out considerable work on the beach ridge systems preserved around the Gulf Coast. St.

![Gulf of Mexico Sea Level History](https://example.com/mexico_sea_level.png)

Figure 6. Sea level curve from Tanner, 1990, suggests barrier island formation associated with sea level drops around 2500, 1800, or 1200 B.P.

Vincent Island, in particular, has an exceptionally well-preserved beach ridge record of 180-200 distinct ridges reaching back about 3500 years (based on 14C dating of included material). Using the height of the ridge as an index of sea level at the time the ridge was built, the record indicates that sea level has fluctuated by several meters both above and below current sea level over the past few thousand years (see Fig. 6).

ORIGIN OF BARRIER ISLANDS

The origin of barrier islands has been a much debated topic since the middle 19th century. Three primary mechanisms have been suggested: (a) emergence of shallow coastal bars through wave-generated processes, (b) isolation of coastal ridges by rising sea levels, and (c) breaching and later migration of long spits. In all probability all three processes can (and have) operated to produce islands, as evidenced by the relative variety of barrier island sizes, shapes, and relationships to the mainland. Evidence along the Gulf Coast of Florida strongly suggests that these barrier islands formed through wave action on shoals during a drop in sea level (Tanner, 1990).

Comparison of historical bathymetric charts indicates that the islands are currently thinning and moving both onshore and to the west. During this time, two major bodies of submerged sand (Cape St. George Shoal and Cape San Blas Shoal), have changed very little below the 6 m (18 ft) isobath (approximately 5-8 km offshore). Analysis of grain size frequency information indicates that the sediments of the islands are essentially similar to those of the shoals. Given that beach ridge evidence suggests that some of islands were in existence 3000-3500 years ago, they probably originated during a time when sea level was either falling or oscillating by no more than a few meters. Grain size similarities
between the sand in the earliest beach ridges and in the immediately offshore sand shoals suggests that a slight drop in sealevel exposed portions of the shoals and wave reworking completed the process (Tanner, 1990).

COASTAL DEVELOPMENT

Due to the dynamic nature of the coastal environment and the length of Florida's coastline, the Florida legislature enacted the Beach and Shore Preservation Act in 1971. One of the goals of this legislation was to "preserve and protect [beaches of the State] from imprudent construction which can jeopardize the stability of the beach dune system ...." Accordingly, the Florida Department of Natural Resources, Division of Beaches and Shores, was directed to establish Coastal Construction Setback Lines (CCSBL). The CCSBL's were replaced in 1978 with Coastal Construction Control Lines (CCCL) which "define that portion of the beach-dune system which is subject to severe fluctuations based on a 100-year storm surge or other predictable weather conditions, and so define the area within which special structural design consideration is required to insure protection of beach-dune system...." The intent of this legislation is to regulate coastal construction in order to protect the coast. It does not prohibit construction, but rather requires the developer to obtain permits from DNR for alteration or construction of any structure below mean high tide or seaward of the CCCL. Because sea level is rising and the barrier islands are narrowing through erosion, these limits will need to be re-evaluated in the future.

Quite often individual rights to repair or improve personal property clash with the state's desires to protect its coasts. Variances are allowed and construction continues. By a 1984 estimate (Doyle, et al), almost 400 km (250 miles) of beach in Florida are experiencing "critical erosion", most with structures placed too close to the water. As a result, many communities have requested state and federally subsidized beach nourishment projects, which are temporary solutions, at best.
<table>
<thead>
<tr>
<th>MILEAGE</th>
<th>DESCRIPTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>00.0</td>
<td>Antarctic Circle, north side of Carraway Building (Geology Dept.), FSU campus. Turn right (south) on Woodward Street.</td>
</tr>
<tr>
<td>00.7</td>
<td>Turn left (east) onto Gaines Street.</td>
</tr>
<tr>
<td>01.4</td>
<td>Turn right (south) onto Bronough Street.</td>
</tr>
<tr>
<td>02.0</td>
<td>Pass entrance to FAMU on right.</td>
</tr>
<tr>
<td>04.0</td>
<td>Junction with US 319 at &quot;Four Points&quot; intersection. Bear right and continue south on US 319 (Crawfordville Highway).</td>
</tr>
<tr>
<td>11.3</td>
<td>Pass entrance to Leon Sinks on right.</td>
</tr>
<tr>
<td>11.6</td>
<td>Enter Wakulla County.</td>
</tr>
<tr>
<td>20.5</td>
<td>Pass through Crawfordville.</td>
</tr>
<tr>
<td>30.9</td>
<td>Junction with US 98. Turn right (south) on 98.</td>
</tr>
<tr>
<td>32.3</td>
<td>Split of US 319 and US 98. Turn left (south) on US 98 (toward Panacea).</td>
</tr>
<tr>
<td>39.2</td>
<td>Enter Franklin County and cross bridge Ochlockonee Bay.</td>
</tr>
<tr>
<td>43.7</td>
<td>Road turns to west. View to left (south) is Alligator Point Spit, connected with the coast to the east.</td>
</tr>
<tr>
<td>48.4</td>
<td>Pass turnoff to FSU Turkey Point Marine Laboratory Facility on left.</td>
</tr>
<tr>
<td>57.8</td>
<td>View to left (south) is Dog Island.</td>
</tr>
<tr>
<td>61.3</td>
<td>Pass through Carabelle.</td>
</tr>
<tr>
<td>72.5</td>
<td>View to left (south) is St. George Island.</td>
</tr>
<tr>
<td>75.8</td>
<td>Junction with road G1A to St. George Island Bridge. Turn left (south) and cross St. George Sound.</td>
</tr>
<tr>
<td>81.3</td>
<td>St. George Island. Turn right and continue to lodging.</td>
</tr>
</tbody>
</table>

DAY 2.
Leave lodging and drive east to entrance of St. George Island State Park.

Park Entrance. Continue into park to beach site.

STOP 1. Beach profile and sediment transport measurement.

The Gulf of Mexico beaches along this portion of St. George Island are primarily erosional because they face into the predominantly southeastern incident wave direction. Note the wave-cut benches and eroded tree stumps. Approximately 1.5 km east of the Park entrance is a portion of the island designated as a potential new tidal inlet (Doyle, et al., 1984). Carry out beach profile measurements and swash transport experiment. The prevalent direction of swash transport in this area is to the southwest, but may vary with varying incident wave conditions. When finished, drive east toward eastern end of the island.

The east end of St. George has built outward approximately .7 km (1 mile) between 1855 and 1935, while the beach retreated landward 100 m (Doyle, et al., 1984). Looking northeast you will see Dog Island across East Pass (approx. 2.5 km). This pass is primarily flood current, bringing water into Apalachicola Bay. The inlets on the western end of the Bay are primarily ebb current, giving the Bay an east-to-west circulation. Drive west toward western end of island.

West end of St. George Island and Sikes Cut. This end of the island advanced toward the Gulf approximately 30 m between 1855 and 1935. Sikes Cut was opened in 1954 by the U.S. Army Corps of Engineers and is maintained by regular dredging. To the southwest, across Sikes Cut lies Little St. George Island. Approximately 3 km from the eastern end of Little St. George is the location of New Inlet, which was opened by a storm in 1837, and remained open until approximately 1900. Drive east on St. George Island.

Junction with G1A. Turn left on G1A and cross bridge to Eastpoint.

Junction with State Road 65. Turn left (west) on 65.

Junction with U.S. 98. Turn left (west) on 98. Cross East Bay on John Gorrie Memorial Bridge.

Drive through Apalachicola on U.S. 98. To your left (south) is Apalachicola Bay. The Apalachicola River enters East Bay immediately north of the city.

Junction with State Road 30A. Stay left on 30A.

Gulf-Franklin County line. Apalachicola Bay (really St. Vincent Sound) is narrowing. Immediately south of the county line is Indian Pass, the westernmost inlet to the Bay.

Junction with State Road 30E. Turn left (west) onto 30E.

Intersection with road to Cape San Blas Lighthouse. Turn right (north) on State Road 30E. This is St. Joseph Peninsula, a spit connected to the mainland near Cape San Blas. The westward facing nature of this spit protects it from the more forceful waves.

Enter St. Joseph Peninsula State Park. Drive as far north as possible.

STOP 2. Beach sedimentary structures and sand peels.

The northernmost portion of the peninsula have grown northwestward while the southern portion of the peninsula have retreated. The southernmost portion of Cape San Blas holds the record as the most rapidly eroding portion of the Florida shore, averaging 8 m per year for the 150 year recorded history of the area. Cape San Blas lighthouse has been moved 3 times in the past 100 years (Doyle, et. al., 1984).
Sediment transport along the beach is obviously to the north, from the eroding Cape in the south, to the accreting end of the spit. Carry out beach trenching and sediment peel demonstrations on a suitable stretch of beach. Note the different sedimentary layers visible in the trench, their attitudes, and the different nature of the sediments composing them. Dominant sedimentary components visible are quartz sand, shell fragments, and dark heavy mineral layers. The pattern can be sketched, photographed, and sampled with a sediment peel. If time permits, measure a beach profile for comparison with the sedimentary record. The peninsula has an excellent set of beach ridges illustrating its accretion history and providing a data set useful for tracking past sea level. Retrace route and return to lodging on St. George Island.

Day 3

00.0 Leave St. George Island on G1A. and retrace first days route northeast along U.S. 98.

5.3 Intersection with U.S. 98 north. Turn right (east) on 98.

40.0 Intersection with State Road 370. Turn right (east) onto 370.

43.9 Split of 370. Turn left.

44.2 Rejoin old 370 along the beach.

44.2 STOP 3. Beach erosion

A decade ago a section of Route 370 ran between the water and the lounge at this point. Since this portion of the shore faces southeast, it is extremely susceptible to erosion. Hurricane Kate, 1986, completed demolition of the road and threatened the lounge. North of here efforts have been made to reduce erosion by maintaining the natural beach dune system. Cutting or removal of sea oats (a natural dune stabilizer) is prohibited. Dune fences have been installed where the sea oats are not sufficient. Vehicles are prohibited from certain areas and most paths onto the beach are circuitous to reduce the damage of channeled storm surge. South of here, the area is highly developed, with houses built on the beach or dunes, jeopardizing the environment and the development.

Drive north on 370 to Bald Point turn around and observe beach protection efforts. Retrace route to split in 370 and turn left (south) to Alligator Point. Drive west along 370 and note the houses built on the beach and dunes. Retrace route to U.S. 98 and back to Tallahassee.
REFERENCES


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MODERN FLUVIAL PROCESSES IN A SAND-BEDDED MEANDERING STREAM: FLOW STRUCTURE, SEDIMENT TRANSPORT, BED FORMS AND BEND MIGRATION

Stephen D. Thorne and Valerie Croup

Department of Geology, B-160, Florida State University, Tallahassee, Florida 32306-3026

David Jon Furbish

Department of Geology and Geophysical Fluid Dynamics Institute, B-160, Florida State University, Tallahassee, Florida 32306-3026

INTRODUCTION

Significant effort over the past two decades has been given to clarifying the mechanics of flow and sediment transport in meandering rivers, and the interactions among flow and bedforms during river evolution (for example, Yen and Yen, 1971; Engelund, 1974; Ikeda and others, 1981; Dietrich and Smith, 1983, 1984; Smith and McLean, 1984; Odgaard, 1986; Odgaard and Bergh, 1988; Ikeda and Parker, 1989). The first objective of this field trip is to examine, with reference to this recent work, features of flow and sediment transport operating within Ochlawaha Creek, a sand-bedded meandering stream in Gadsen County, northwest Florida (Fig. 1). In this regard, we will examine how bedforms, bend migration, and related floodplain deposits are systematically related to flow conditions. We will then outline the results of two field studies on the Ochlawaha. The first project involved an experiment (Thorne, 1993) whereby the naturally roughened bank of a bend was replaced with a smooth wall to document how roughness due to riparian vegetation influenced flow in the bend. The second involved sampling bed sediment to document how the flow and related boundary-stress field lead to systematic spatial sorting of particle sizes over the extent of the bend (for example, Dietrich and Smith, 1984; Parker and Andrews, 1985). A theme of the trip is to examine how river reaches—straight and meandering—are mechanically interdependent, and how this influences sediment sorting, patterns of bend migration, and rates of reworking of floodplain sediments. The Ochlawaha is particularly well suited for field study due to the clarity of its water and presence of well-sorted sediment.

FLOW STRUCTURE AND SEDIMENT TRANSPORT IN A BEND

The studies of Dietrich and Smith (1984, 1985) and Smith and McLean (1984), together with the monograph edited by Ikeda and Parker (1989), represent a compact compilation of recent field, laboratory and theoretical studies of flow and sediment transport in river bends. The following summary, taken mainly from these sources, is intended as a context for our field observations, and pertains to rivers having relatively uniformly erodible floodplain material with no significant bank roughness.

In a river with erodible bed and banks, the channel bottom exhibits a characteristic cross-stream geometry: it is relatively flat at the crossings, and steeply sloping at the bend apexes. Flow depth decreases rapidly away from the concave outer bank toward the convex inner bank due to the presence of the point bar on the inside bank of the bend. As water enters a bend it begins to shoal over the point bar. The streamwise and transverse slopes of the bar induce convective accelerations near the bed such that flow is steered around the point bar toward the outside bank. Resistance to flow over the bar increases, and associated with this and the topographic steering of flow are a decrease in the streamwise water-surface slope over the bar, and a steepening over the pool.

The curvature of the bend similarly induces a centrifugal acceleration of flow as it is forced to change its direction through the bend. This is accompanied by a cross-stream pressure gradient manifest as a slope in the water surface, with superelevation of flow near the outside bank. If the bend curvature is sufficiently large, this transverse water-surface slope can equal or exceed the streamwise slope (Smith and Mclean, 1984). This pressure gradient tends to counter the outwardly directed convective accelerations produced by the point bar. In addition, shoaling of flow over the point bar has the effect of reducing the cross-stream water-surface slope below that which would occur in absence of convective accelerations produced by the point bar.

Bend curvature and associated superelevation lead to a secondary flow which is directed inward near the channel bottom, while shoaling over the point bar generally leads to a small outward component of flow throughout the entire water column, over the upstream portion of the bar. In addition, curvature and bed topography cause boundary shear stresses to be high near the inside bank at the entrance of a
bend, and sediment is transported toward the outer bank. Downstream, however, inward flow associated with the curvature-induced secondary current, and associated boundary shear stresses, direct sediment inward.

The shoaling of flow associated with the point bar together with the curvature-induced secondary current combine to shift the core of high streamwise velocity across the channel. Associated with this high-velocity core is a zone of maximum boundary shear stress, and a zone of maximum bed load transport; all tend to gradually shift outward through the bend (Dietrich and Smith, 1984). Then the curvature and point bar downstream tend to drive them back across the channel.

Early theoretical descriptions of flow, bed topography and sediment transport in river bends were largely based on the assumption of an equilibrium bed topography whose transverse bed slope reflected a balance of transverse forces acting on the bed sediment: inwardly directed tractive forces due to the near-bed flow of the secondary current, and the outwardly directed component of the weight of particles residing on the transversely sloping point bar. Under these conditions, the force balance results in sediment transport parallel to the channel banks (for example, Allen, 1970; Kikkawa and others, 1976; Bridge, 1977; Zimmermann and Kennedy, 1978; Odgaard, 1981). Subsequent work from the Muddy Creek experiments (Dietrich and Smith, 1984) indicates that this assumption may be too simplistic. The Muddy Creek data reveal a net inward transport of sediment in the upstream part of the bend, and a net outward transport near the apex and in the downstream part. Near the bend apex, transport is inward from the pool and outward over the slope of the point bar; farther downstream the transport is outward over the entire width of the channel. Analysis of grain sizes further reveals that fine-grained sediment is transported inward near the bend apex, while predominately coarse-grained particles move outward across the bar. These paths of transport are determined by the cross-stream flow induced by the point bar, and by the secondary flow arising from the bend geometry. These flows transport coarser particles across the upstream part of the bar and into the pool, whereas finer particles are transported inward from the pool across the point bar downstream of the bend apex. These inwardly and outwardly directed paths of transport tend to cross near the bend apex, and are then mimicked in a transposed sense in the downstream bend.

Floodplains in humid environments often support unevenly distributed vegetation whose size, for sufficiently small channels, is on the order of the depth of flow or larger. Channel banks roughened by elements of this size can significantly influence the flow field, particularly in bends. Assuming in the simplest case that local migration rates are proportional to the near-bank depth-averaged flow velocity (Ikeda and others, 1981; Pizzuto and Meckelnburg, 1989), an understanding of how flow in bends is influenced by rough banks is therefore essential for describing bend migration and the meandering process in these environments. This topic is examined in the next section, within the following context.

The evolution of a river channel involves a feedback between the flow field and the channel banks and bottom. The extant flow field in a channel at any instant alters the channel by eroding it in some areas and depositing sediment in others. These changes in the morphology of the channel then alter the flow which, in turn, further alters the channel. In this way a constant feedback between the channel morphology and flow occurs, such that channel evolution is a nonlinear (autocatalytic) process. In addition, the shape of a river bend evolves in direct response to upstream bend geometries and the flow conditions delivered from them (Leopold and Wolman, 1960; Yen and Yen, 1971; Englund, 1974; Ikeda and others, 1981; Dietrich and Smith, 1983). Further, evolution of an individual bend is not independent of conditions downstream, as backwater effects from a neighboring bend can influence upstream flow conditions (Smith and Mclean, 1984). Bends are therefore mechanically interdependent during the development of a meander train. As a consequence, a diverse assortment of meander shapes may develop in absence of factors such as variable bank erodibility, changes in sediment size, and differences in channel bank vegetation (Furbish, 1991). How bank roughness due to vegetation figures into this process is of general interest to river scientists, and forms the motivation for the field studies described below.

FIELD EXPERIMENTS

An experiment was designed to assess the importance of coarse bank roughness on flow through bends. The experiment involved measuring flow velocities and water-surface topography through a bend under natural conditions where the outer bank was roughened by vegetation, then again after vegetation on the outer bank was removed and replaced with a smooth wall constructed of plastic sheeting overlaid on a wood and wire framework. The experiment was designed so that variations in discharge and channel dimensions negligibly influenced the results. The resulting changes in the flow field were attributable to the bank vegetation. The Ocklawaha Creek near Tallahassee Florida provided an ideal location for the experiment. It is a sharply meandering sand-bedded creek flowing through a densely vegetated floodplain.

Measurements with and without the smooth artificial bank were taken along section lines whose positions were constant (Fig. 1). Each was made at a specified map coordinate. Velocity measurements were made every 0.75 meters across the width of the channel (unless measurements at those sites were prohibited by bottom or bank vegetation), and at positions of depth whose spacing decreased exponentially toward the channel bottom. A Marsh-McBery current meter was used to measure flow speeds and directions. Ribbons tied to the current-meter rod were used
in conjunction with a Brunton compass to measure the azimuth direction of the current. In addition, a device was constructed to fit the rod which allowed the meter to swivel and align itself with the local flow. The velocity readings (magnitudes and directions) were then used to compute streamwise and cross-stream components relative to the known orientation of each section line. Water-surface elevations were measured with a transit and stadia rod using the procedure described by Dietrich and Smith (1983).

A second set of measurements involved sampling bed sediment to document the extent to which flow systematically sorts particle sizes through the bend (for example, Dietrich and Smith, 1984; Parker and Andrews, 1985). Sediment samples were collected from several transverse locations at each of the sections, then sieved into nine size classes (-1.0 φ to > 2.5 φ). Two key ingredients of this work involved the fact that flow into the bend is delivered by a long straight reach, and that the sediment as a whole possesses a relatively restricted distribution of sizes, predominately meandrum-to-coarse sand. Because the straight reach delivers sediment to the bend that is unsorted in a transverse sense, subsequent sorting downstream in the bend reveals how quickly (spatially) this process occurs.

RESULTS

From measurements made at the upstream section 4.5, discharge was calculated to be 1.83 m³s⁻¹ in the presence of the natural vegetation, and 1.72 m³s⁻¹ when the wall was in place. The small difference in discharge (due to decreasing stage between measurements), and the fact that channel dimensions remained essentially the same with the addition of the wall, suggest that adjustments in the flow field described below were due primarily to absence of bank roughness. In fact, an increase in reach-averaged velocity associated with the decrease in discharge emphasizes the importance of the influence of bank vegetation on bend flow.

The data reflect that bank vegetation significantly alters the near-bank velocities. These roughness elements offer resistance to flow and produce turbulence which extracts energy from the main streamwise flow. When this resistance was removed by replacing the natural bank with a smooth wall, flow velocities increased substantially (Fig. 2). These profiles reflect not only a velocity increase in the absence of the vegetation, but also a smoothing of the velocity distributions as the turbulence is reduced. In addition, the bank roughness has the effect of deflecting surface flow downstream. This occurs for several reasons. The turbulence produced by this roughness strengthens the lateral boundary layer and extracts momentum from the flow which would otherwise be available to carry it into the outer bank. This results in surface currents flowing more directly downstream. There is less superelevation, and the resulting secondary current is weaker. In this way, bank roughness effectively decreases local migration rates (above and beyond the fact it also serves to enhance the cohesion of the bank sediments). In conjunction with this, there was an increase in the transverse water-surface slope in the absence of the vegetation (Fig. 3). All sections except 05 show this increase, with that at section 5.5 being the largest. The transverse water-surface slope decreases between the bend apex at section 4.5 to the inflection point at section 06.

As the transverse water-surface slope increases with the removal of the vegetation, the strength of the secondary current also increases, as represented by the angle between the surface and bottom currents (Fig. 4). The increased secondary current strength, and associated reduction in turbulence throughout the fluid column, also were manifest in profiles of cross-stream velocities (Fig. 5). When the vegetation was removed, the inwardly directed bottom currents were better developed and the velocity profiles became smoother.

Vegetation along the banks tended to delay, in a downstream sense, the full development of the secondary current, and ultimately reduced its strength. Flow did not impinge directly onto the bank as it entered the bend, and as a result, the secondary current could not develop at this location. In the case of the smooth outer bank, surface water in the bend began to superelevate just downstream from the bend entrance, initiating development of the secondary current at a position farther upstream than in the case of the rough bank. The potential effect of this is to increase migration rates nearer the bend entrance. Moreover, the transverse boundary layer is compressed such that the high-velocity filament shifts toward the outer bank throughout the bend (Fig. 6). This suggests that migration rates will be higher at all locations in the bend in absence of the vegetation, aside from the fact that removal of the vegetation eliminates this component of the strength of bank materials.

The transverse distribution of sediment sizes is relatively uniform at the bend entrance. Sorting then begins rapidly, as reflected by the path of coarse sediment (Fig. 7). Coarser sediment (in a proportional sense) moves onto the head of the point bar, then into the pool downstream of the bend apex. It then moves out of the tail of the pool near the right bank onto the head of the point bar downstream. A more detailed description involving other size classes will be provided on the field trip.

IMPLICATIONS

These changes in the flow field have several implications for long-term river evolution. In absence of the vegetation, superelation occurred early in the bend, was of greater magnitude, and extended farther downstream. These results suggest, first, that because the magnitude of the near-bank velocity together with the strength of the secondary current were increased throughout most of the length of the bend, sediment transport should be enhanced (compared with the rough bank) throughout the bend. Second, lateral migration may begin farther upstream than in the presence of the
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IMPLICATIONS

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vegetation. Third, the strengthened secondary current could enhance bar formation, which would contribute to increased topographic steering of flow by the point bar, thereby further shifting the filament of high velocity toward the outer bank. This, in turn, could enhance lateral migration.

The high-velocity filament was mapped in the presence and absence of the bank vegetation (Fig. 8). Changes in this filament suggest that two initially identical bends with different bank roughnesses may evolve differently. With the smooth outer bank, the high-velocity core impinges on the outer bank at section 5.5. The likelihood for lateral migration at this location is high. In principle, the high-velocity filament remains near the bank until it exits the bend, and then enters the downstream bend near the head of the downstream point bar. The high velocity filament in the presence of the vegetation, however, does not impinge on the bank and, as described above, development of the secondary current is delayed. As a result, erosion of bank sediment would not begin as far upstream, and would be of less magnitude than in the case without bank vegetation. The high-velocity filament exits the upstream bend closer to the centerline. The effect of vegetation is therefore twofold: in addition to providing cohesive strength to the bank, it lowers near-bank velocities, and inferably migration rates.

Superimposed on this is a difference in the phase shift between the high-velocity filament and the bend curvature. The position where the high-velocity filament reaches its maximum displacement from the centerline (and is closest to the outer bank) is farther downstream in the case of the smooth bank (Fig. 8). Aside from differences in rates of migration, the effect of this is to produce a greater asymmetry in bend shape following a finite period of migration. Bends in this case would tend to migrate downstream rather than growing in amplitude (Fig. 9a). Conversely, the position where the high-velocity filament reaches its maximum displacement from the centerline is closer to the bend apex in the case of the rough bank (Fig. 8). The effect of this is to produce less asymmetry in bend shape, growth in bend amplitude (albeit at a lower rate than would occur without vegetation), and widening of the meander belt (Fig. 9b).

Sorting of sediment sizes within the Ochlawaha bend is similar to the Muddy Creek results. Of particular significance is that sorting in the Ochlawaha, as reflected by the coarse-sediment path, begins early in the first bend just downstream from a straight reach where sediment sizes are unsorted in a transverse sense. In addition, this sorting occurs with a sediment mixture whose overall size distribution is not as heterogeneous as the Muddy Creek sediments. This surprising degree of sorting is likely due to the steepness of the transverse slope of the point bar, and the strength of the secondary current induced by the sharpness of the bend.

ACKNOWLEDGEMENTS
This work was partly supported by the National Science Foundation (EAR-90-04646). William Parker and Sharon Reeves helped in mapping the Ochlawaha bends. We are grateful to Mr. Lee for allowing access to the site.
ROAD LOG

The following log is provided with the caveat that access to the field trip site may involve crossing privately owned land. Permission from landowners is required. Please contact the authors if you wish to visit the field site.

Mileage     Description
00.0        Gunter Building, Florida Geological Survey, Tallahassee, Florida. Proceed west on Tennessee Street.
07.3        Cross Ochlockonee River.
24.0        Cross two branches of Rocky Comfort Creek.
25.2        Cross Bear Creek.
27.6        Stop at Ochlawaha Creek and park near bridge. The bridge provides an excellent bird’s-eye view of the sand-bedded channel of Ochlawaha Creek. A well-developed, but very dynamic, bar with bedforms ranging from ripples to dunes persists next to the south bank immediately upstream from the bridge. The field trip site described in the accompanying article is approximately 0.5 miles upstream (west) from the bridge. Proceed on foot.
28.1        STOP. Bends of Ochlawaha Creek (Fig. 1 in the accompanying article). At this site we will examine bedforms, evidence of bend migration, and related floodplain deposits, and discuss how these are systematically related to flow conditions. We will then outline the results of two field studies described in the accompanying article, the first involving an experiment whereby the naturally roughened bank of a bend was replaced with a smooth wall to document how roughness due to riparian vegetation influenced flow in the bend, the second involving sampling of bed sediment to document how the flow and related boundary-stress field lead to systematic spatial sorting of particle sizes over the extent of the bend. The site is well suited for examining how river reaches—straight and meandering—are mechanically interdependent, and how this influences sediment sorting and patterns of bend migration. The clarity of the water provides an unusually good opportunity to observe sediment transport and migration of bedforms.
Figure 1. Map of bends on Ocklawaha Creek showing locations of section lines along which velocity measurements were made.
Figure 2. Streamwise velocity profiles at different transverse positions x from right bank, for natural bank (filled circles) and smooth wall (circles). Flow is from left to right.
Figure 3. Transverse water-surface slopes at each section with smooth wall and natural bank (NB). Vertical exaggeration 49x.
Figure 4. Vector diagrams of section 5.5 with natural bank (A) and with smooth wall (B). Plots show maximum flow velocity in ms⁻¹. Numbers indicate relative depth within the fluid column, increasing toward the channel bottom.
Figure 5. Transverse velocity profiles of section 5.5 with natural bank (A) and smooth wall (B). Points to the right of each vertical line represent flow toward the right (outer) bank, and those to the left represent flow toward the left bank. Scale given as 0.10 ms⁻¹.
Figure 6. Contour maps of streamwise velocity of section 5.5 with natural bank (A) and smooth wall (B). Contour interval is 0.02 ms\(^{-1}\). Vertical exaggeration is 2.5x.
Figure 7. Transport path (stippled) of coarse sediment.
Figure 8. Path of high velocity filament with natural bank (A) and with smooth wall (B); phase shifts defined in terms of maximum displacement of high velocity filament from centerline are indicated.
Figure 9. Schematic diagram of three instances during bend migration (short dashed lines are early state, solid lines are latest state) for large phase shift (A) and small phase shift (B) in high-velocity filament. Flow is from bottom to top. Note predominant downvalley migration (A) versus widening (B) of meander belt.
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HYDROGEOLOGY OF THE WESTERN SANTA FE RIVER BASIN

Katherine K. Ellins, Richard A. Hisert, and Todd Kincaid
Department of Geology, University of Florida, 1112 Turlington Hall, Gainesville, Florida  32611

Few locations in the United States are characterized by groundwater/surface water discharge relationships more intricate than those found in the complex karst of the Santa Fe River basin in central-northern peninsular Florida. The objectives of the field trip and its companion guidebook are to (1) describe the karst hydrogeology of the Santa Fe River Basin, and (2) familiarize geologists with the results of research efforts carried out at the University of Florida to advance the knowledge about subterranean flow patterns and groundwater/surface water discharge interactions in the western Santa Fe River basin. The investigations described involve the application of natural radon-222 ($^{222}$Rn) and the man-made sulfur hexafluoride (SF$_6$) as hydrologic tracers. One important finding of these studies is that substantial groundwater movement occurs along laterally extensive cavernous zones within the karst of the Santa Fe River basin at average velocities as rapid as 2.6 kilometers/day (0.03 m/s). Such rapid groundwater movement renders the water resources of the Santa Fe River basin extremely vulnerable to contamination and underscores the urgent need for careful management practices.

FIELD TRIP STOPS

**Stop 1.** O'Leno State Park (Route 441, 6.75 miles north of High Springs). O'Leno Sink. **Discussion:** (1) General description of the hydrogeology of the Santa Fe River basin, and (2) results of an SF$_6$ tracer experiment that tracked the underground flow path of the Santa Fe River between O'Leno Sink and River Rise.

**Stop 2.** New Sink (O'Leno State Park). **Discussion:** Sinkhole formation.

**Stop 3.** Santa Fe River Rise (O'Leno State Park - key to enter this area must be obtained from park personnel at the entrance). **Discussion:** (1) Cave development; and (2) the use of $^{222}$Rn and SF$_6$ together to determine groundwater influx and measure gas exchange.

**Stop 4.** Canoe Outpost (Route 441, 2.25 miles north of High Springs). Pick up canoes and paddle downstream to Rum Island. **LUNCH EN ROUTE.** While paddling downstream, note the syphons through which stream flow is diverted underground to karst conduits in the Upper Floridan aquifer, Columbia Spring, the riparian wetland ecosystem (Blackwater Floodplain Forest), Poe Spring, Lily Spring, and Blue Spring.

**Stop 5.** Rum Island (15.5 kilometers below the Santa Fe River Rise). **Discussion:** (1) The application of $^{222}$Rn and SF$_6$ in estimating groundwater inputs and stream flow losses in the Santa Fe River between Rum Island and Ginnie Springs; and (2) the use of SF$_6$ to identify point-to-point connections of groundwater recharge and spring discharge.

**Stop 6.** Ginnie Springs (17.75 kilometers below the Santa Fe River Rise). **Discussion:** Cave formation in the Florida karst with emphasis on the Devil's Ear Cave system.
INTRODUCTION
A description of the regional geologic framework of the Florida peninsula is reported by Scott (1991). Briefly, the basement rocks of Florida are composed of igneous and metamorphic rocks of Precambrian to Cambrian age, which were once part of the African plate. Overlying the basement rocks is a thick sequence of Mesozoic sediments. Above these are primarily Paleocene and Eocene carbonates, which comprise the Floridan aquifer system. Throughout the western Santa Fe River basin, these carbonate rocks are exposed and a well-developed karst topography has developed.

STOP 1. O'LENO SINK
Hydrogeologic Setting

The Santa Fe River basin (Fig. 1) is located in central-northern peninsular Florida. The Santa Fe River originates in Lakes Santa Fe and Altho in the Northern Highlands physiographic province of Florida and flows generally westwards to its confluence with Florida's second largest river, theSuwannee River. At O'Leno State Park the Santa Fe River crosses the Cody Escarpment (Scott, 1991), a topographic feature which delineates the boundary between the Northern Highlands and Gulf Coastal Lowlands physiographic provinces. Erosional processes along the Cody Escarpment have given rise to a broad transitional zone where the Santa Fe River encounters soluble carbonate rocks of the Ocala Limestone (Upper Eocene). The Santa Fe River is diverted underground at O'Leno Sink, located within the boundaries of O'Leno State Park. The river is believed to resurface about 5 kilometers to the south at a resurgence, named the Santa Fe River Rise, where brown water with a tanin content similar to the upper Santa Fe River is discharged. After resurfacing, the Santa Fe River traverses the Western Valley, a sinkhole plain where the Tertiary limestones comprising the Floridan aquifer system are exposed, towards its confluence with the Suwannee River.

Numerous karst features can be recognized in O'Leno State Park including solution sinkholes, collapse sinkholes, sinkhole ponds, karst windows, springs, and water-filled fractures. Many of the sinkhole ponds intersect subsurface drainage conduits and are in reality karst windows. In some of these, a discrete point of resurgence is clearly visible and the flow can be traced downstream to a sinking point. In others, however, the water appears stagnant and it is difficult to discern which sinkholes are karst windows by visual inspection alone.

In June 1991 a rapidly collapsing sinkhole developed in the park about 2.5 kilometers from O'Leno Sink. The new feature was quickly named New Sink. As the overburden slumped into the newly forming sinkhole, the resurgence of cloudy, sediment-laden flow was observed at the Santa Fe River Rise, indicating that New Sink may intersect the subterranean flow path of the Santa Fe River. At the request of O'Leno State Park managers, we were invited to conduct a tracer experiment to explore the possible connection of New Sink to surrounding sinks and to track the subterranean pathway of the Santa Fe River between O'Leno Sink and the Santa Fe River Rise.

The Underground Flow Path of the Santa Fe River between O'Leno Sink and the Santa Fe River Rise.

The application of fluorescent dyes is the best established technique in tracing subterranean flow routes in karst areas. Fluorescent dyes do not always perform satisfactorily, however, as they are susceptible to chemical breakdown and biodegradation. Furthermore, they may be adsorbed by clay or organics present in the water or the aquifer matrix (Gaspar, 1987; Mull et al., 1988; and Smart and Smith, 1976). In an effort to delineate the subterranean pathway of the Santa Fe River between O'Leno Sink and the Santa Fe River Rise, the man-made volatile tracer, sulfur hexafluoride (SF₆), was utilized instead of a fluorescent dye. SF₆ is a non-biodegradable, man-made, non-toxic, inert gas (Brown, 1989; and Lovelock and Ferber, 1982). Analysis on a gas chromatograph permits the measurement of concentrations of SF₆ in the femtomolar range (10⁻¹⁵ mol/L). The extremely low detection limit of SF₆ makes it possible to use small quantities of the tracer in water tracing studies.

A tracer solution of SF₆ was delivered to O'Leno Sink in a slug injection lasting two hours. At the end of two hours all remaining tracer solution was dumped into the sink. The discharge of the Santa Fe River 50 meters above the injection site (42 m³/s) was obtained to provide an estimate of the amount of water flowing to the rise via the subsurface route and to determine the optimum injection rate for the tracer. Eight ponds located between O'Leno Sink and Santa Fe River Rise (Ogden Pond, Ravine Sink, Parener's Branch Sink, Small Sink, New Sink, Jim's Sink, Two Hole Sink andSweetwater Lake) were monitored in order to determine whether they were hydraulically connected to the subsurface flow of the Santa Fe River (Fig. 2). Sampling for the tracer was also carried out at the Santa Fe River Rise. The duration of the sampling period and the sampling intervals for each location was determined by taking into consideration the estimated travel time of the tracer through the system. Tracer arrival time estimates for each sink were obtained by using the discharge of the Santa Fe River above O'Leno Sink (42 m³/s) and the straight line distances between sampling locations in accordance with the methods of Ratthann (1979). Samples were collected in glass bottles or 50 ml glass syringes. Analyses were done on a Shimadzu gas chromatograph with an electron capture detector (ECD) following procedures developed by Wanninkhof (1986). Concentrations of SF₆ in the parts per trillion range can be detected with a precision of 3%.

SF₆ was detected in all of the sinkhole ponds monitored but not at the Santa Fe River Rise. The concentration response curves for Ogden Pond, Parener's Branch, Small Sink, New Sink, and Sweetwater Lake are shown in Figures 3 and 4.
Arrival times for the leading edge of the tracer (TL), the travel time of the centroid of the SF₆ mass (T), and the total tracer passage time for those sinks monitored for a sufficiently long interval are set forth in Table 1. This table also shows the average SF₆ travel velocities calculated from the travel times of the centroid of the SF₆ mass and the straight line distance between the sinks.

Table 1. Travel times for the SF₆ tracer cloud at sinks (karst windows) in O’Leno State Park.

<table>
<thead>
<tr>
<th>Site</th>
<th>Distance (m)</th>
<th>TL (min)</th>
<th>TP (min)</th>
<th>T (min)</th>
<th>Passage (min)</th>
<th>Ave. Velocity (m/min)</th>
<th>Ave. Velocity (m/s).</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ogden</td>
<td>100</td>
<td>16</td>
<td>110</td>
<td>108</td>
<td>134</td>
<td>0.9</td>
<td>0.02</td>
</tr>
<tr>
<td>Ravine</td>
<td>700</td>
<td>193</td>
<td>560</td>
<td>540</td>
<td>600</td>
<td>1.76</td>
<td>0.03</td>
</tr>
<tr>
<td>Parenthood</td>
<td>950</td>
<td>300</td>
<td>560</td>
<td>540</td>
<td>600</td>
<td>1.76</td>
<td>0.03</td>
</tr>
<tr>
<td>Small</td>
<td>1000</td>
<td>366</td>
<td>590</td>
<td>570</td>
<td>434</td>
<td>1.77</td>
<td>0.03</td>
</tr>
<tr>
<td>New</td>
<td>1100</td>
<td>370</td>
<td>600</td>
<td>620</td>
<td>480</td>
<td>1.77</td>
<td>0.03</td>
</tr>
<tr>
<td>Jim’s</td>
<td>1200</td>
<td>400</td>
<td>600</td>
<td>620</td>
<td>480</td>
<td>1.77</td>
<td>0.03</td>
</tr>
<tr>
<td>Two Hole</td>
<td>1900</td>
<td>44</td>
<td>600</td>
<td>620</td>
<td>480</td>
<td>1.77</td>
<td>0.03</td>
</tr>
<tr>
<td>Sweetwater</td>
<td>2800</td>
<td>67</td>
<td>1167</td>
<td>1650</td>
<td>*</td>
<td>1.70</td>
<td>0.03</td>
</tr>
<tr>
<td>River Rise</td>
<td>4750</td>
<td>119</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>**</td>
<td>**</td>
</tr>
</tbody>
</table>

TL = time to leading edge; TP = time to peak concentration;
T = travel time of centroid of the SF₆ mass.
* period of sampling was too short
** No SF₆ was detected

In order to determine whether the failure to detect SF₆ at the Santa Fe River Rise was due to an underestimation of the sampling period required to monitor the passage of the tracer cloud, a subsequent tracer injection was performed at Sweetwater Lake (Fig. 2). This tracer established that there is a link between Sweetwater Lake and the Santa Fe River Rise and confirmed that the Santa Fe River Rise is the primary point of resurgence for the subsurface flow of the Santa Fe River. The results of the tracer also revealed two additional points of resurgence approximately 50 and 250 meters downstream of the rise which correspond to boils observed in the river. The approximate underground flow path for the Santa Fe River, determined through the use of SF₆, is shown by arrows on Figure 2. The average travel velocity calculated is 2.6 kilometers/day.

The results of the tracer study confirmed that SF₆, which is safe, chemically and biologically inert, and capable of being detected in extremely low concentrations, is an excellent alternative to fluorescent tracers in karst terranes. SF₆ is a volatile tracer, however, so gas exchange occurs whenever subsurface flow is exposed to the atmosphere, as is the case in karst fensters. For this reason, SF₆ is best suited as a qualitative tracer for establishing point-to-point connections and determining flow rates.

STOP 2. NEW SINK (O’Leno State Park).

New Sink formed rapidly in June 1991 about 2.5 kilometers from O’Leno Sink. The resurgence of cloudy, sediment-laden flow from the collapse of the overburden was observed at the Santa Fe River Rise, a distance of about 4.6 kilometers (Fig. 2), suggesting that New Sink intersected the subterranean flow path of the Santa Fe River.

Sinkhole Formation.

Sinkholes (dolines) develop in karst terranes as a result of the complicated interaction between processes of bedrock dissolution, the piping of unconsolidated material, and bedrock collapse. Often, solution chimneys and vertical shafts occur in association with sinkholes and may serve as drains. Sinkholes may be classified according to origin. The different types recognized by White (1988) are briefly described in this work.

Solution sinks are typically bowl-shaped depressions formed from the continual dissolution of soluble bedrock by runoff that collects in the feature. Sinkhole ponds may develop when the solution features that drain the solution sinks become plugged with debris. Eventually, the sinkhole may coalesce with others as it enlarges by deepening and widening.

The continual dissolution of bedrock along joints, fractures and solution cavities at depth may lead to the development of collapse sinks. These are usually shallow features that form when the roof of a solution cavity or cave, which can no longer support the overlying bedrock, collapses. Sinkholes that provide access to underground rivers are karst windows (fensters). Underground streams function as efficient sediment transport pathways and aid in the removal of overburden that slumps into the sinkholes.
Cover collapse sinkholes develop where a thick layer of soil or other unconsolidated material overlies soluble bedrock. Dissolution of the bedrock along high permeability pathways at or near the surface of the bedrock is initiated by surface runoff diverted to the subsurface via vertical solution pathways. In addition, an upwardly domed cavity may be carved out of the overlying unconsolidated material if the cover is thick and cohesive. The stability of an arch composed of unconsolidated material above a cavity will eventually be undermined as a consequence of flushing and cavity enlargement following storm events. When the soil arch can no longer support its own weight, the structure collapses. Often, collapse is triggered by a high runoff event. While the process of carbonate dissolution is a slow one, the failure of an arch composed of unconsolidated material is a catastrophic event.

In areas where the soil or sediment cover is not sufficiently thick to support a stable arch, the ground layer sags as the underlying bedrock is dissolved giving rise to cover subsidence sinks. These may appear as circular puddles on the landscape after rain events.

New Sink probably formed as a result of the combination of the mechanisms responsible for the development of both collapse sinks and cover collapse sinks. The results of our tracer study confirmed that New Sink is directly connected to Small Sink and Jim's Sink, and is hydraulically linked to the subterranean flow path of the Santa Fe River. It is apparent as you walk over the land surface in this area, that Small Sink, New Sink and Jim's Sink are coalescing and will soon form a single, sinuous karst window (Fig. 2).

STOP 3, SANTA FE RIVER RISE (OLeno State Park)

This spot marks the primary resurgence for the subsurface flow of the Santa Fe River. Two additional points of resurgence approximately 50 and 250 meters downstream of the rise, which correspond to boils in the river, may be observed. Several large springs deliver groundwater to the Santa Fe River below the Santa Fe River Rise. Springs of the Ginnie Group (Fig. 1), located on the Santa Fe River about 13 kilometers west of the town of High Springs, collectively deliver about 8.0 m³/s to the main stem of the river (Wilson and Skiles, 1988). At least four of the eight springs of this group (July, Devil's Ear, Devil's Eye and Little Devil) are physically connected by a flooded anastomotic maze cave system. The Devil's Ear Cave, located beneath the Santa Fe River, is a laterally extensive conduit system with over 11 kilometers of mapped passages (Wilson and Skiles, 1988). It is developed in the Eocene Ocala Limestone formation, one of the most productive water-bearing units of the Floridan aquifer system, at a depth of about 35 meters.

Another major input of groundwater is delivered to the Santa Fe River by Ichetucknee Trace (Fig. 1), one of Florida's 27 first order magnitude springs. The Ichetucknee Trace originates at Ichetucknee Head Spring and flows approximately 11 kilometers south/southwest until it joins the Santa Fe River (Hunn and Slack, 1983). Numerous other springs, including Cedar Head Spring and Blue Spring (known also as Jug Hole), contribute to the total discharge of about 10.2 m³/s of the Ichetucknee Trace (Hunn and Slack, 1983).

Groundwater/surface water discharge relationships are extremely complex in the Santa Fe River. As the river flows towards its confluence with the Suwannee River it not only gains groundwater, but also loses stream flow to the Upper Floridan aquifer via solution conduits known as syphons or sucks. Furthermore, surface discharge diverted underground may subsequently be discharged back into the river.

Using ²²²Rn and SF₆ to Determine Groundwater Gains and Stream Flow Losses in the Santa Fe River

Two tracer experiments, involving ²²²Rn and SF₆ were carried out during the summer of 1991 along two short reaches of the Santa Fe River below its resurgence in order to locate and quantify groundwater contributions, obtain estimates of stream losses from the river, identify points of resurgence for surface discharge diverted underground, and determine the gas transfer velocity. The first experiment was conducted between the Santa Fe River Rise and Columbia Spring (4.75 kilometers) (Fig. 5). Along this reach, Hornsby Spring is the only major documented contributor of groundwater to the Santa Fe River. Two large sucks where surface flow is diverted underground are clearly visible at 2.75 kilometers and 4.15 kilometers. It is believed that the surface flow that is diverted underground at 4.15 kilometers resurfaces at Columbia Spring.

²²²Rn (half life= 3.8 days) is derived from radium-226 (half life= 1600 years), which is common in aquifer material. Concentrations in groundwater are usually higher than concentrations measured in stream flow, which are depressed due to gas exchange with the atmosphere. In the Santa Fe River basin the ratio of ²²²Rn concentrations in groundwater to surface discharge ranges between 10 and 15 to one. Any immediate increase in the concentration of ²²²Rn observed in the Santa Fe River is indicative of groundwater influx.

As noted previously, the Santa Fe River not only gains groundwater but also loses stream flow to the Floridan aquifer via sucks. SF₆ was introduced to the Santa Fe River in a continuous injection and measurements were used to identify locations where surface discharge that had been diverted underground returned to the main stream channel. This is possible as gas exchange ceases in underground flow because it is sequestered from the atmosphere. Thus, SF₆ concentrations will be elevated above stream concentrations, which will have declined due to gas exchange, at a point in the stream where the resurgence of underground flow occurs.
The magnitude of groundwater influx to the Santa Fe River is proportional to the difference between the observed and background $^{222}$Rn concentrations at any point in the river. Similarly, at locations where surface discharge that was diverted underground resurfaces in the stream, the magnitude of the return flow component is proportional to the difference between SF$_6$ measured in the stream and the estimated background concentration. Thus, $^{222}$Rn and SF$_6$ measurements provided data that could be used with discharge measurements in a mass-balance equation to determine the magnitude of groundwater influx and return flow. Background $^{222}$Rn and SF$_6$ concentrations were estimated by employing a correction for losses due to gas exchange between sampling locations. Since both SF$_6$ and $^{222}$Rn have approximately the same Schmidt numbers, theoretically their gas transfer velocities in a stream should be same. For this reason, SF$_6$ provided an independent means of obtaining the gas transfer velocity for $^{222}$Rn, which is necessary in order to accurately determine ground-water influxes. The gas transfer velocity can also be related to the rate of oxygen uptake and expressed as a reaeration coefficient (K$_2$), which is a determining factor in assessing water quality in a stream. The reaeration coefficient obtained (10.6 d$^{-1}$) is in good agreement with published K$_2$ values for other streams and the results of various predictive models.

Measurements of $^{222}$Rn were made using Lucas-type cells with alpha scintillation counters. A small portable extraction system was used to strip the $^{222}$Rn from water samples collected in the field into the Lucas-type cells. The minimum detectable $^{222}$Rn activity is 0.2 dpm/L for a 250 ml sample. The precision associated with the technique is 7%.

Sampling locations between the Santa Fe River Rise and Columbia Spring is shown in Figures 5. Figure 6 shows the $^{222}$Rn and SF$_6$ profiles. The sharp peaks in the $^{222}$Rn profiles are indicative of groundwater influx. The SF$_6$ profile shows a general decline due to gas exchange. Increases in the SF$_6$ concentration noted at 2.75 kilometers and 4.75 kilometers (Columbia Spring) indicate the resurgence of flow that was diverted underground at a point farther upstream.

During the experiments, detailed stream gauging of each transect was carried out in tandem with the $^{222}$Rn and SF$_6$ sample collection effort, using a Weathertronics flow meter. In order to show the dynamics of discharge along the reach a cumulative discharge profile (Fig.7) was generated by utilizing the hydrometric data and by balancing the estimated inputs against the estimated losses. Stream flow losses were calculated by difference from the $^{222}$Rn and SF$_6$ derived groundwater influx and return flow estimates, and the measured discharge data.

A summary of estimated stream parameters and flow components obtained through the measurement of natural $^{222}$Rn and deliberately introduced SF$_6$ is shown in Table 2.

**TABLE 2. Summary of stream parameters and flow components (Santa Fe river Rise - Columbia Spring).**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gas transfer velocity</td>
<td>31.0 m/d</td>
</tr>
<tr>
<td>Reaeration coefficient</td>
<td>10.6 d$^{-1}$</td>
</tr>
<tr>
<td>Total groundwater influx</td>
<td>20 m$^3$/s</td>
</tr>
<tr>
<td>Total stream flow losses</td>
<td>26 m$^3$/s</td>
</tr>
<tr>
<td>Total return flow</td>
<td>27 m$^3$/s</td>
</tr>
<tr>
<td>Overland runoff</td>
<td>14 m$^3$/s</td>
</tr>
<tr>
<td><strong>NET GAIN</strong></td>
<td>35 m$^3$/s</td>
</tr>
</tbody>
</table>

Although the estimates of groundwater influx, stream flow loss, and the resurgent underground flow must be treated as preliminary, the results clearly demonstrate that the measurement of natural $^{222}$Rn and deliberately introduced SF$_6$ in a stream provided important new information about the dynamics of groundwater and surface interaction in the Santa Fe River Basin which can be used in conjunction with other types of data to design resource management strategies and to guide further research. Confidence in estimated stream flow losses can be improved in future studies through the addition of a conservative tracer, such as chloride or bromide, which will permit recognition of zones where surface discharge is lost and allow direct measurement of the magnitude of the loss.

**STOP 4. CANOE OUTPOST ON ROUTE 27**

We will paddle from the canoe outpost downstream to Ginnie Springs, a distance of 13 kilometers. LUNCH EN ROUTE. While paddling downstream, note the following features:
Syphon

There are many syphons or suck holes in the Santa Fe River, through which stream flow is diverted underground to karst conduits in the Floridan aquifer. Note two located at 4.25 kilometers and 7.5 kilometers below the Santa Fe River Rise on the north bank of the river.

Columbia Spring, located 4.75 kilometers downstream of the Santa Fe River Rise, is believed to represent the resurgence of stream flow diverted underground via syphons in the stream bed farther upstream.

Blackwater Floodplain Forest.

The wetlands bordering the Santa Fe River represent a type of river swamp classified as a blackwater floodplain forest. River swamps, which constitute about one third of Florida’s swampland, are among the most diverse and productive swamps found in Florida. The most common wetland tree in the floodplain forest is cypress (Taxodium), a deciduous conifer. Mature cypress trees are the most flood-tolerant of all tree species in Florida. Their seedlings, however, grow best in saturated but unflooded soils. Cypress trees are easily recognized by their distinctive "knees", contorted growths that develop when lateral roots bend downwards. Knees may help the plant tolerate low oxygen levels in the floodplain soil by obtaining oxygen from the atmosphere. Other common tree species found in river swamps include black gum (Nyssa sylvatica), water tupelo (N. aquatica), overcup oak (Quercus lyrata), swamp laurel oak (Q. laurifolia), black willow (Salix nigra). Other plants include shrubs from the family Ericaceae, poison ivy (Toxicodendron radicans), epiphytes, and insectivorous plants (Ewel, 1991).

Swamps provide food, cover, nesting sites, and hibernating places for a variety of animals. There are leeches, worms, mites, spiders, beetles, dragonflies, snails, clams, crayfish, and approximately 18 species of mosquitoes! Among the diverse fish populations two species, the cypress darter (Etheostoma proleiare) and the cypress minnow (Hybognathus hanyi), are considered rare or endangered in Florida. Be on the lookout for alligators (Alligator mississippiensis), crayfish snakes, corn snakes, the decidedly unfriendly cottonmouth snakes, and a variety of salamanders. Among the birds inhabiting the floodplain forest of the Santa Fe River are raptors, water birds, warblers, woodpeckers, and wild turkeys. The most notable mammals are the river otter (Lutra canadensis), mink (Mustela vison), and the beaver (Castor canadensis) (Ewel, 1991).

Springs

Numerous springs including Poe (12.0), Lily (13.25 kilometers), Rum Island (15.5 kilometers) Blue Spring (16.0), and the Ginnie Springs Group (17.0 - 19.0 kilometers) discharge groundwater to the Santa Fe River below the Canoe Outpost.

STOP 5. RUM ISLAND.

Using ²²²Rn and SF₆ to estimate groundwater inputs and stream flow losses in the Santa Fe River between Rum Island and Ginnie Springs (2.5 kilometers).

A second tracer experiment was carried out to determine the location and quantification of groundwater influx, estimate stream flow losses from the river, identify points of resurgence for surface discharge diverted underground, and determine the gas transfer velocity between Rum Island and Ginnie Springs (Fig.8). Both the SF₆ and ²²²Rn profiles for this reach (Fig.9) are characterized by many spikes, indicating both groundwater delivery and the resurgence of stream flow. Figure 10 shows the discharge dynamics along this reach. It is likely that the Devil’s Ear Cave System, an extensive cave network which extends beneath the Santa Fe River along a portion of this reach, exerts a dominant influence over interactions between groundwater and surface flow in this part of the Santa Fe River. Cave diving excursions as far back as 1200 meters into the Devil’s Ear System have provided first-hand observations of the complexity of the conduits that characterize the Upper Floridan aquifer.

²²²Rn measurements made within the cave have permitted the location of groundwater inputs and points of surface water intrusion to the Devil’s Ear System. A summary of stream parameters and flow components estimated for this reach obtained through the measurement of natural ²²²Rn and deliberately introduced SF₆ is shown in Table 3.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gas transfer velocity</td>
<td>2.8 m/d</td>
</tr>
<tr>
<td>Reaeration coefficient</td>
<td>1.2 d⁻¹</td>
</tr>
<tr>
<td>Total groundwater influx</td>
<td>38 m³/s</td>
</tr>
<tr>
<td>Total stream flow losses</td>
<td>44 m³/s</td>
</tr>
<tr>
<td>Total return flow</td>
<td>14 m³/s</td>
</tr>
<tr>
<td>Overland runoff</td>
<td>3 m³/s</td>
</tr>
<tr>
<td>NET GAIN</td>
<td>11 m³/s</td>
</tr>
</tbody>
</table>

96
Again, while these estimates of groundwater influx, stream flow loss, and resurgent underground flow are preliminary, they demonstrate that measurements of natural $^{222}$Rn and deliberately introduced SF$_6$ provided important new information about the dynamics of groundwater and surface interaction along this complex stretch of the Santa Fe River. The reaeration coefficient obtained for this reach, however, is much lower than models predict. This discrepancy is due to the substantial diversion of surface discharge to the Devil’s Ear Cave System found beneath this highly interactive reach of the Santa Fe River, where gas exchange with the atmosphere cannot occur.

**Using SF$_6$ to identify point-to-point connections of groundwater recharge and spring discharge.**

A water tracing experiment using SF$_6$ was designed to examine two hypotheses associated with the Devil’s Ear Cave system. The first hypothesis, put forth by Wilson and Skiles (1988), is that there is no hydraulic connection between the conduits associated with July Spring and Little Devil’s Spring, located adjacent to each other on the Santa Fe River. (Fig. 8) July Spring occurs on the north side of the Santa Fe River and is physically connected via a passable conduit to Devil’s Ear and Devil’s Eye Springs. These three springs are the major known groundwater discharge points of the Devil’s Ear Cave system. Little Devil’s Spring is located on the south side of the Santa Fe River about 75 meters from Devil’s Ear and Devil’s Eye Springs. The cave system associated with the Little Devil’s Spring, however, is developed at a depth of 18 meters and is not known to be connected via passable conduits to the Devil’s Ear Cave system, which occurs at a depth of 35 meters. The results of a previous water-tracing experiment in which Rhodamine dye was injected into three wells located on the south side of the Santa Fe River near Ginnie Springs provided evidence of groundwater movement from the south side of the Santa Fe River to Little Devil's Spring, but not Devil's Ear, Devil's Eye or July Springs. These findings led Wilson and Skiles (1988) to conclude that July and Little Devil’s Springs are the discharge points of two different conduit systems although both are less than 75 meters apart. Furthermore, they postulated that July, Devil’s Ear and Devil’s Eye are recharged from the north.

A second objective was to test the hypothesis that Rum Island Spring, located on the north bank of the Santa Fe River 1.5 kilometers upstream of July Spring, is hydraulically connected to the Devil’s Ear Cave System (Ellins et al., in review).

SF$_6$ was injected into an abandoned well and a sinkhole, both located on the north side of the Santa Fe River (Fig. 8). Sampling was carried out at Rum Island, July, and Little Devil Sprngs over a period of 55 hours. The three springs monitored produced positive confirmation of SF$_6$, providing evidence of a definite hydraulic connection between the injection points and the springs. Multiple peaks on the concentration response curve for each spring reflect the use of two different injection sites on the same day. They may also be indicative of a complicated conduit network in the karst through which groundwater is delivered to the Santa Fe River. Since the injection of SF$_6$ at the well and sinkhole was closely spaced in time, it was not possible to calculate average groundwater flow velocities.

The results of the experiment show that July, Rum Island and some portion of the discharge at Little Devil Spring draw their recharge from the north side of the Santa Fe River. These data suggest that the initial hypothesis that Little Devil Spring draws its entire recharge from the the south side of the Santa Fe River and is isolated from the Devil’s Ear Cave is invalid. Wilson and Skiles (1988) do show, however, that some portion of the discharge at Little Devil’s Spring is derived from the south side of the Santa Fe River. The detection of SF$_6$ first at Rum Island Spring and then at July and Little Devil’s Spring supports our hypothesis that Rum Island Spring is connected to the extensive Devil’s Ear Cave network. Groundwater flows from the injection points on the north side of the Santa Fe River to Rum Island, into the Devil’s Ear Cave System, and out to Little Devil and July Springs.

**STOP 6. GINNIE SPRINGS.**

**Cave Formation**

Substantial groundwater movement occurs along cavernous zones within the Upper Floridan aquifer. The development of cavernous zones within the Tertiary limestones of the Floridan Aquifer begins with dissolution by carbonic acid-rich soil water and circulating groundwater along zones of weakness such as cracks, joints or bedding plane partitions. Carbonic acid ($\text{H}_2\text{CO}_3$) forms when CO$_2$ gas is dissolved in water. CO$_2$ is derived from the atmosphere and soil horizons, where it is produced by respiration of plant root systems and microorganisms. Organic acids produced during the partial decomposition of organic matter by microbes also play a key role in limestone dissolution.

The overall chemical reaction describing the dissolution of limestone by carbonic and organic acids, which can also be represented as carbonic acid, is shown below.

$$\text{H}^+ + \text{HCO}_3^- + \text{CaCO}_3 \rightarrow \text{Ca}^{++} + 2\text{HCO}_3^-$$

And can also be expressed as:

$$\text{CO}_2(\text{gas}) + \text{H}_2\text{O} + \text{CaCO}_3 \rightarrow \text{Ca}^{++} + 2\text{HCO}_3^-$$
As CaCO$_3$ becomes dissolved in the groundwater system, the water becomes enriched in bicarbonate (HCO$_3^-$) and the pH increases. If equilibrium is achieved in the system, further dissolution of CaCO$_3$ is halted. In aquifers where cave systems are actively forming, flow velocities are sufficiently rapid to circulate groundwater that is undersaturated with respect to Ca$^{++}$ and HCO$_3^-$. The dissolution process provides a pathway along which water can flow. As dissolution proceeds, the cavity enlarges increasing permeability, which permits more vigorous circulation of groundwater. In an aquifer, many channels may develop providing alternate flow routes for groundwater. A particular avenue of flow, however, may be preferentially selected due to some controlling mechanism which is related to the particular hydrogeologic setting. Water that follows existing joint patterns tends to create deep narrow canyons while dissolution along bedding planes gives rise to broad shallow passages. As long as a conduit or assemblage of conduits remains filled with water, dissolution will continue and a cave system will ultimately form. Limestone dissolution is an exceedingly slow process that takes place on the order of hundreds of thousands to millions of years.

Any opening created by dissolution that is large enough to accommodate a human is loosely defined as a cave. Caves vary in length, passage morphology and layout. In plan view, a cave represents an assemblage of accessible passages whose geometry reflects the complex interaction between the structural and stratigraphic relationships that govern the flow path of subsurface water with the solution chemistry and fluid mechanics of the hydrogeologic setting. Efforts to explain the origin of caves have produced a variety of complex models. Interested readers are referred to White (1983) for an excellent review.

Caves can generally be classified into two broad classes: (A) Single conduit caves and (B) Maze caves. Each class may be further subdivided on the basis of the configuration of the passage elements.

(A) Single conduit caves.
1. Linear passages are primarily developed in regions where a single major joint or fracture trace maintains control of the flow path.
2. Angulate passages often develop in flat-bedded limestones where the hydraulic gradient is oriented diagonally to the regional joint set.
3. Sinuous passages resembling the channel configuration of a meandering surface stream are usually associated with flat-bedded limestones.

(B) Maze caves.
1. Network mazes occur in highly jointed regions where all joints dissolve at a uniform rate.
2. Anastomotic maze passages generally follow highly permeable bedding plane partings.
3. Spagework mazes form in young porous and poorly jointed limestones resulting from random solution of massive bedrock without either structural or hydraulic control.

The development of single conduit caves is governed by the dominance of allogenic recharge. Allogenic recharge refers to runoff that flows off of an impervious catchment onto permeable bedrock, such as a soluble karst surface, where it may become concentrated at a discrete point and sink into the underlying aquifer. This type of recharge is usually undersaturated with respect to carbonate minerals and thus is important in the aggressive dissolution of primarily single-conduit cave systems. In the tropical karst of the Caribbean, extensive cave systems that once served as subsurface stream flow routes are abundant. Good examples are the Quashies and Cave River systems of Jamaica, and the Rio Camuy system of Puerto Rico.

In central-northern Florida a similarly developed corridor-conduit system is believed to serve as the underground flow route for the Santa Fe River between O'Leno Sink and the Santa Fe River Rise. The Santa Fe River transports a large volume of allogenic recharge from the classic Northern Highlands to the Transition Zone, located between the Highlands and the Gulf Coastal Lowlands. In the transition zone, which has been created by erosional processes along the Cody Escarpment, all surface drainage flowing off the highlands, including the Santa Fe River, sinks underground. Work carried out at the University of Florida (Elling et al., in review) has confirmed that several sinkholes in O'Leno State Park are connected to the sinking point of the Santa Fe River by a continuous conduit.

Unlike single conduit caves, the formation of maze caves is determined by the dominance of internal or authigenic (autogenic) flow. Precipitation that falls on a karst surface is funnelled into the underlying aquifer via sinkholes creating multiple discrete concentrated inputs. At some down-gradient location within the aquifer, the flow from these sinkholes will converge. At such a point, a network of cave passages of varying dimensions may develop. Many of Florida's springs discharge from conduits that are accessible to cave divers and are associated with a convergence of diffuse flow indicated by circular depressions in the potentiometric surface of the aquifer. Examples include Blue Springs located in Volusia County, Silver Springs, Rock Springs, and Wekiva Springs (Beck et al., 1985).

The Devil's Ear Cave System.

The best example of a maze cave in the karst of Florida is the Devil's Ear Cave System, shown in Figure 11, which occurs beneath the Santa Fe River part way between the towns of High Springs and Fort White. The Devil's Ear Cave system is an anastomotic maze that has formed within the Ocala Limestone. Most of the conduits are established at a depth of about 33 meters below the stream bed. Numerous side tunnels of varying dimensions intersect the main tunnel, which is large enough to fit a truck! Flow is added or diverted from the cave system by these peripheral passages which cave divers refer to as upstream and
syphon tunnels, respectively. Ripplemarks that have formed in the sediment on the bottom of the cave indicate the primary current direction much as they do in the bed of a stream.

Intense mixing corrosion that takes place at the convergence of diffuse flow from multiple drainage areas is responsible for the development of the labyrinthine Devil's Ear Cave system. Intrusion of acidic tannin-stained Santa Fe River water along fractures and syphons also promotes accelerated excavation of the system. Back flooding, which refers to a reversal in the direction of flow that is associated with flood stage conditions in a river (Thraillkill, 1968), is rarely observed in the Devil's Ear System. For the first time in over twelve years, two near reversals of flow were reported by cave divers during the flood stage conditions of the Santa Fe River during the Spring of 1992.

In addition to the chemical corrosion that occurs in the Devil's Ear Cave System, there is evidence of mechanical erosion within the cave. Turbulent flow is responsible for moving sand and other abrasive material across easily eroded limestone surfaces. Sedimentary features such as ripplemarks, flutes and scallops, and concretions provide evidence of high velocity turbulent flow.

Caves form under water table conditions in karstic terranes. Until recently, flooded cave systems have been largely inaccessible. Thus, most of the concepts developed regarding cave formation have been based on the exploration of dry cave systems. Over the past 20-30 years, however, technological advances in scuba diving have facilitated safe diving excursions into flooded cave systems, which are common in Florida. From a hydrologic standpoint, it is useful to study a flooded cave system such as Devil's Ear in order to gain a better understanding of the behavior of karst aquifers and to advance our knowledge about the formation of caves.

ACKNOWLEDGEMENTS

The guidance and support of the O'Leno and Ichetucknee State Parks staff is gratefully acknowledged. In addition, our sincere appreciation is due the many volunteers who assisted in the collection of samples during all the experiments and to the homeowners who allowed us to use their well for the injection of SF6. Thanks also to Wes Skiles of Karst Environmental Services, Inc. for sharing his knowledge about the Devil's Ear Cave System. Thanks to Guerry McClellan and Brian Katz for reviewing the manuscript. This research was supported by the National Science Foundation (EAR-9004614).
Figure 1. Physiographic provinces and hydrogeologic units of the Santa Fe River basin.
Figure 2. Map showing the karst region where the Santa Fe River flows underground through O'Leno State Park. Arrows indicate the approximate flow path of the river between O'Leno Sink and the Santa Fe River Rise as determined through the use of SF$_6$. 

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Figure 3. $\text{SF}_6$ concentration as a function of time at Ogden Pond. July 1991. Travel time of centroid of $\text{SF}_6$ mass = $T$. 
Figure 4. $\text{SF}_6$ concentration as a function of time at Parener's Branch Sink, Small Sink, New Sink, and Sweetwater Lake. July 1991. Travel time of centroid of $\text{SF}_6$ mass = T.
Figure 5. Sampling locations between the Santa Fe River Rise and Columbia Spring.
Figure 6. $^{222}$Rn and SF$_6$ profiles between the Santa Fe River Rise and Columbia Spring.

Figure 7. Measured and cumulative discharge profiles for the reach between the Santa Fe River Rise and Columbia Spring.
Figure 8. Map showing the locations of Rum Island Spring, Blue Spring and the Ginnie Springs Group (Devil's Ear, Devil's Eye, Little Devil, July, Ginnie, Dogwood, Sawdust, Twin, and Deer Springs) on the Santa Fe River. The abandoned well and sinkhole which served as SF₆ injection sites are also shown.
Figure 9. $^{222}\text{Rn}$ and SF$_6$ profiles between Rum Island and Ginnie Spring.

Figure 10. Measured and cumulative discharge profiles for the reach between Rum Island and Ginnie Spring.
Figure 11. Sketch of the Devil's Ear Cave System. Based on surveys and exploration carried out by Wes Skiles.
REFERENCES


