# Southeastern Geology: Volume 18, No. 4
# August 1977

Edited by: S. Duncan Heron, Jr.

**Abstract**

Academic journal published quarterly by the Department of Geology, Duke University.

# SOUTHEASTERN GEOLOGY

1977
INQUA

**INQUA ISSUE**
**Dedicated To The**
**Tenth International Congress**
**16 - 24 August 1977**

PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

**VOL. 18 NO. 4** **AUGUST, 1977**

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY



This journal welcomes original papers on all phases of geology, geophysics, and geochemistry as related to the Southeast. Transmit manuscripts to S. DUNCAN HERON, JR., BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA. Please observe the following:

(1) Type the manuscript with double space lines and submit in duplicate.
(2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
(3) Submit line drawings and complex tables as finished copy.
(4) Make certain that all photographs are sharp, clear, and of good contrast.
(5) Stratigraphic terminology should abide by the Code of Stratigraphic Nomenclature (AAPG, v. 45, 1961).

**Proofs will be sent authors.**

**Reprints must be ordered prior to publication. Prices are available upon request.**

**Subscriptions to Southeastern Geology are $6.00 per volume Inquiries should be addressed to: SOUTHEASTERN GEOLOGY, BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA. Make checks payable to Southeastern Geology.**

SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 18, No. 4

1977

1. Primary and Secondary Rims of Carolina Bays
Erling E. Gamble
Raymond B. Daniels
Walter H. Wheeler . . . . . . 199

2. Glacial, Periglacial, and Pseudo-Glacial Features in the Grandfather Mountain Area, North Carolina
Loren A. Raymond . . . . . . 213

3. The Arapahoe Ridge--A Pleistocene Storm Beach
R. B. Daniels
E. E. Gamble
W. H. Wheeler
C. S. Holzhey . . . . . . . . . 231

4. Variations in South Carolina Coastal Morphology
P. Jeffrey Brown. . . . . . . 249

5. A Preliminary Study of the Sediment Hydrology of a Georgia Salt Marsh Using Rhodamine Wt as a Tracer
John Nester . . . . . . . . . 265

# PRIMARY AND SECONDARY RIMS OF CAROLINA BAYS[1]

By

Erling E. Gamble
Raymond B. Daniels
Soil Conservation Service and
Soil Science Department
North Carolina State University
Raleigh, North Carolina

and

Walter H. Wheeler
Geology Department
University of North Carolina
Chapel Hill, North Carolina

## ABSTRACT

The Carolina Bays of the Coastal Plain of North Carolina are surface features formed during deposition of the surficial sediments. Power auger drilling indicates that the bedding and sediments underlying a bay are undisturbed. The drill hole data and sections from deep drainage ditches lead to the conclusion that there are two types of bay rims. One is the primary rim or edge of the original depression whose sandy nature is pedogenic, the result of the development of a thick sandy A2 soil horizon on the better drained edge of the bay. The other is a secondary rim, formed primarily from eolian sand deposited after the depression was formed. The source of the secondary rim sand probably was the limited beach of a water-filled bay. One secondary rim was found overlying a buried soil developed on the primary rim of the bay. This indicates a significant time interval can occur between formation of the depression and deposition of the secondary rim sand. In another bay there was no evidence of a soil under the secondary rim sand, which suggests that the primary and secondary rims were nearly contemporaneous.

[1]Paper Number 4728 of the Journal Series of the North Carolina Agricultural Experiment Station, Raleigh, North Carolina. A joint contribution from the Soil Conservation Service, USDA, and the Department of Soil Science.

## INTRODUCTION

Carolina Bays are oval depressions with a sandy rim. Their interiors may be water filled, with or without an accumulation of peat, or may simply be an area of poorly drained soils. The long axes of these depressions are oriented northwest-southeast. They are numerous in the southeastern Coastal Plain especially in North and South Carolina. Prouty (1952) has estimated that there are about half a million bays and 140,000 of them are more than 500 feet (150 m) in length. The largest bay is about 7 miles (11 km) long. Their oriented oval shape makes them such prominent features that they have excited considerable interest and speculation as to their origin.

Theories of origin include meteorite impact, sinks, and wind and water current interactions. Many of the early ideas as to bay origin are discussed by Prouty (1952). More recent ideas involving wind and water mechanism for developing bays and other oriented oval basins have been presented by Thom (1970), Price (1972), and Killigrew and Gilkes (1974). Any mechanism for developing Carolina Bays must account for their large numbers, oval oriented shape, presence on Coastal Plain surfaces of several ages, large variation in size, bays within bays, and bays with multiple rims.

The rims of many bays appear to be accumulations of sand that stand slightly above the surrounding area. In air photos, the rims are narrow light-colored arcuate areas outlining the darker bay interior and are most prominent along the east and southeast sides. The purpose of this paper is to present the results and interpretations of some detailed investigations of three bays in the middle Coastal Plain of North Carolina. One bay is located just east of Goldsboro on the Goldsboro Ridge (Daniels, Gamble, and Wheeler, 1971). A second is located along State Rt. 111 in Wayne County, about 1.6 km south of the Cliffs of the Neuse State Park. The third (The Rayfield-Lee Bay) is in the extreme south corner of Johnston County along State Rt. 55 (Figure 1). The sediment textures given in the text and illustrations are field estimates. The names are from the USDA Textural Triangle (Soil Survey Staff, 1951).

## CAROLINA BAYS AS SURFACE FEATURES

Carolina Bays are surface features that apparently have no effect on subsurface materials. Figure 2 shows a longitudinal section and a cross section of the Goldsboro Ridge located just east of Goldsboro, North Carolina. The characteristics and possible origin of this ridge have been discussed by Daniels, Gamble, and Wheeler (1971). Figure 3 shows the mapped extent of the ridge and the outlines of the associated bays, including two bays that are actually within or on the ridge proper. The larger of the two bays on the Ridge is shown in section from 2-1 to 2-3 on the long axis in Figure 2.

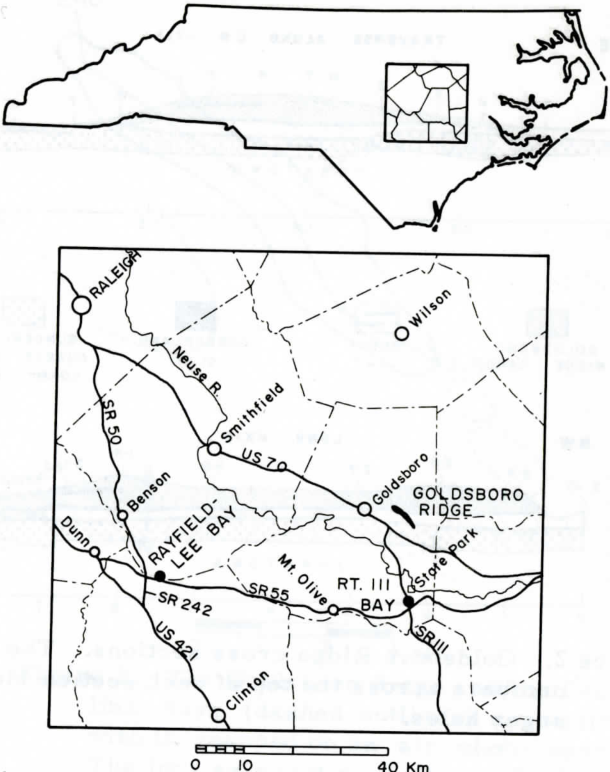


Figure 1. North Carolina index map and area map showing location of bays and other features.

The larger bay is formed on the sand body of the Goldsboro Ridge. The sand is about 4. 3 m thick at the rims and 2. 6 m thick in the bay. The sand has an abrupt contact with the underlying Sunderland clay in all drill holes. The Sunderland clay bed continues under the bay without disturbance or interruption. The relief on this clay surface is no greater under the bay than along the other parts of the section. Whatever processes formed this bay apparently operated at the surface and/or within the Ridge sand and had no detectable effect below the sand.

Dark-colored poorly and very poorly drained moderately fine-textured mineral sils now occupy the floor of the bay. In the eastern end there is 30 to 120 cm of silty clay to silty clay loam at the surface. The eastern rim is sand to a depth of 2. 3 m, and sandy loam to 4. 3 m, the base of the Ridge sand. The soil consists of 30 cm of yellowish brown loamy sand A1 and 60 cm of reddish yellow B horizon over 90 cm of very pale brown C or A'2 horizon. This is a normal sort of pedogenic sequence to be expected in sands in this area.

Further evidence of the surficial character of Carolina Bays is

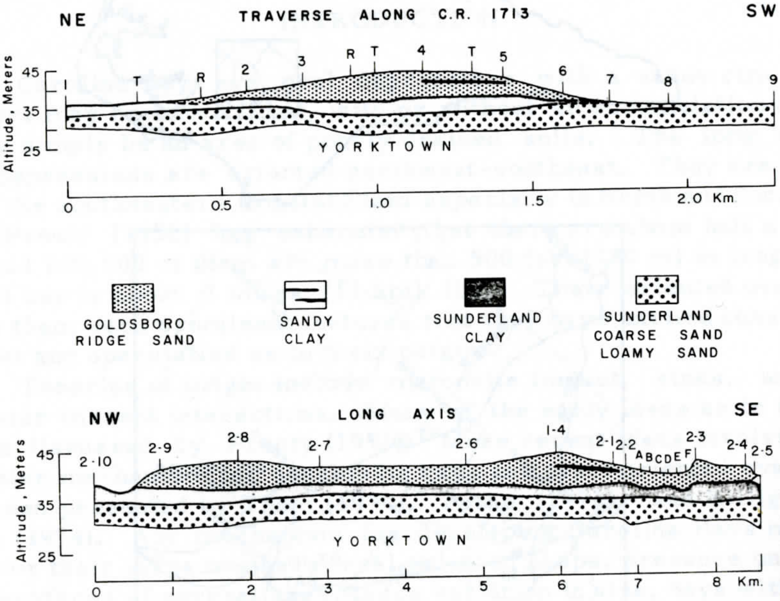


Figure 2. Goldsboro Ridge cross sections. The letters and numbers across the top of each section identify power auger holes.

given by Bryant (1964), who studied the soils of bay and interbay areas in Scotland County, North Carolina. He reported that the bays are in surficial sediments 6.1 m or less thick, with a basal sand to sandy loam that is continuous beneath bay and interbay areas, with no evidence of disruption of material beneath the bays. Preston and Brown (1964) reached the same general conclusion on the basis of a series of power auger drill holes across a bay in Sumter County, South Carolina. They comment that the bay-forming mechanism must produce the bay without deforming the underlying strata and that a surficial mechanism is most consistent with observed data. Drill traverses reported by Thom (1970) contain additional evidence of the surficial character of bays. In Horry and Marion counties, South Carolina, there was no evidence of solution and subsidence in spite of the presence of carbonate-rich strata in the subsurface and some localized sink holes.

## CHARACTERISTICS OF BAY RIMS

Our field studies suggest that there are two types of bay rims. One is the primary rim that is simply the edge of the original oval depression formed as part of the deposition of the surficial sediment. The

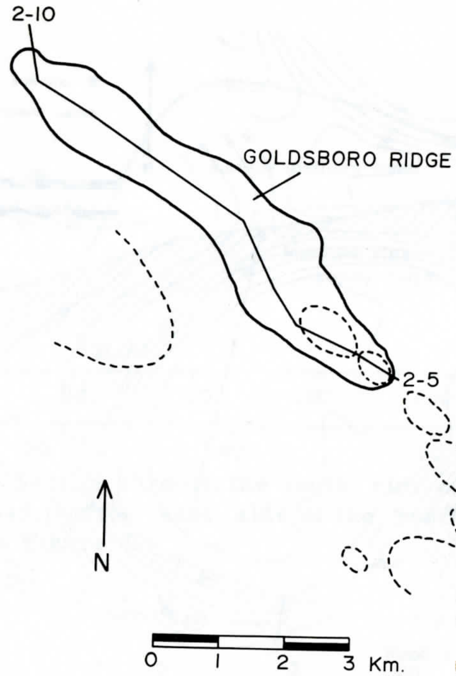


Figure 3. The Goldsboro Ridge and the Carolina Bays (dashed outlines) associated with it, mapped on an air photo base. The long axis section in Figure 2 follows the line, 2-10 to 2-5.

other is a secondary rim, deposited at some later time. The secondary rims have a characteristic arcuate shape. Bays may have only a primary rim or they may have some combination of primary and secondary rim.

## State Rt. 111 Bay

This bay is located in the north-central part of the Seven Springs 15-minute Quadrangle, about 1.6 km south of the Cliffs of the Neuse State Park in Wayne County. It was studied because of the availability of a road cut where Rt. 111 crosses the south rim. Figure 4 shows the topographic details. The bay has an irregular shape. The interior is under cultivation and contains no peat. The rim is well developed on the north, east and south sides but is somewhat flattened and merges with the surrounding sandy upland along the west edge. The rim is lowest along the southwest sector. There is a partial development of two rims on the southeast end where two crests are separated by a well-expressed swale. The north rim separates this bay from one that is even larger

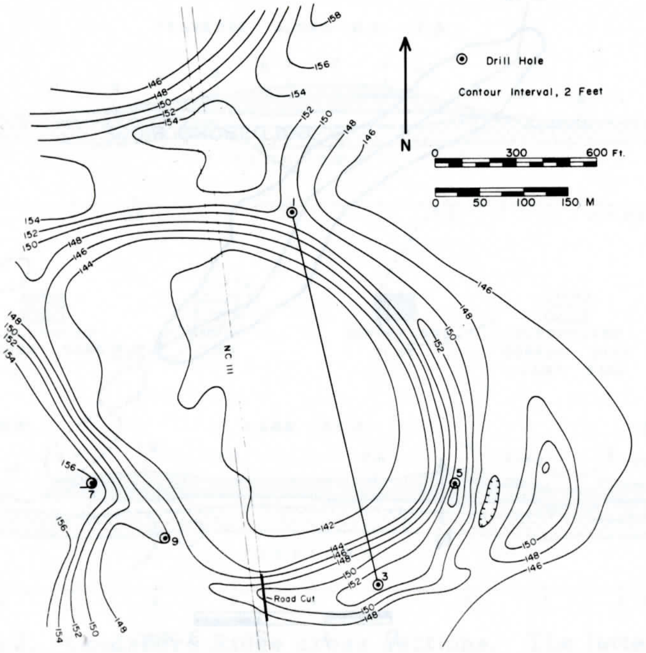


Figure 4. Topographic map of State Route 111 Bay, located about 1.6 km south of Cliffs of the Neuse State Park. Various drill hole locations are shown.

and more poorly defined.

Figure 5 is the section exposed where State Route 111 cuts through the south rim of the bay. It is shown on Figure 4 as "Road Cut." The sand rim is loamy medium sand with a soil profile, consisting of an A1, A2, a banded or lamellar B horizon and an abrupt contact to a buried surface and soil. The upper part of the buried soil is marked by an abrupt change in texture, color, and sand size. Its B horizon is sandy loam and the sands are finer than those in the overlying rim sand. There is some weak subangular blocky structure and some suggestion of Be bodies in the buried soil. Be bodies are areas of clay loss in a B horizon and are common in soils of this area of the Coastal Plain (Daniels and others, 1968).

The buried surface can be traced from under the south edge of the loamy sand rim to where it becomes the general upland level just to the south of the bay. It can also be traced toward the bay interior, to the north, where it becomes the present surface inside the bay. Thus, the loamy sand rim rests on the crest of a slope into the bay that had a soil before the overlying sand was deposited. This buried slope and crest is the primary rim and the loamy sand deposit overlying it is the secondary rim of this bay.

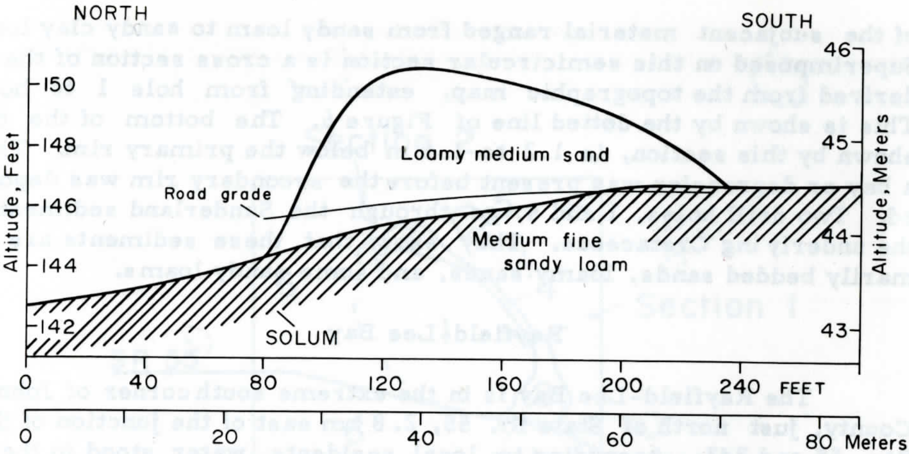


Figure 5. Section through the south rim of State Route 111 Bay, on the east side of the road. See "Road Cut" in Figure 4.

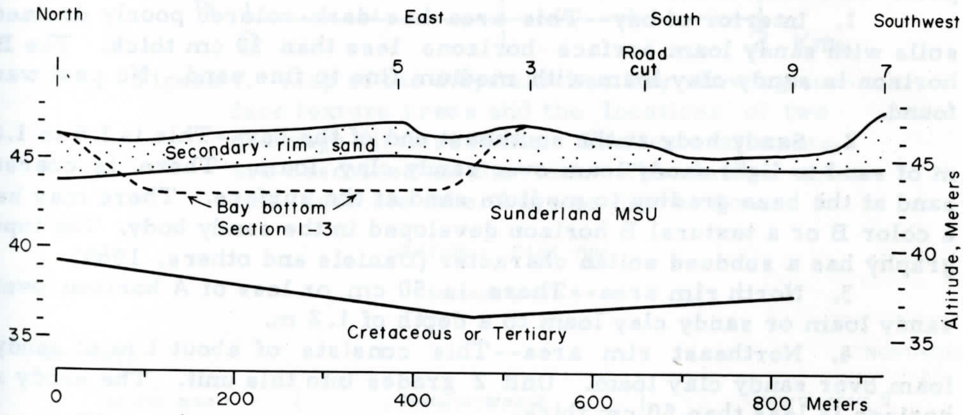


Figure 6. Section around the east rim of the State Route 111 Bay, as seen by looking north, east, and south from a point in the center of the bay. MSU = Morphostratigraphic Unit.

The contact between the loamy sand and the underlying primary rim can be traced for some distance around the bay. This is illustrated in Figure 6 which shows from left to right what one would see standing in the middle of the bay looking in succession to the north, to the east, and to the south. The section is based on drill holes 1, 5, 3, "Road Cut," 9, and 7 (in that order) and on the topographic map. At each drill hole there is an easily recognized sharp contact separating the sand or loamy sand of the secondary rim from the underlying material. Textures of the subjacent material ranged from sandy loam to sandy clay loam. Superimposed on this semicircular section is a cross section of the bay, derived from the topographic map, extending from hole 1 to hole 3. This is shown by the dotted line of Figure 6. The bottom of the bay, shown by this section, is 1.2 to 1.8 m below the primary rim. Thus, a bay or depression was present before the secondary rim was deposited. Two drill holes, 1 and 9, go through the Sunderland sediments to the underlying Cretaceous. They show that these sediments are primarily bedded sands, loamy sands, and some sandy loams.

## Rayfield-Lee Bay

The Rayfield-Lee Bay is in the extreme south corner of Johnston County, just north of State Rt. 55, 2.8 km east of the junction of State Rts. 55 and 242. According to local residents, water stood in the bay before it was ditched. Its general shape, the surface textures of the immediate area, and the locations of drainage ditch sections across the rim are shown in Figure 7. The six textural areas mapped on an airphoto base are described as follows:

1. Interior of bay--This area has dark-colored poorly drained soils with sandy loam surface horizons less than 30 cm thick. The B horizon is sandy clay loam with medium fine to fine sand. No peat was found.

2. Sandy body at the southeast end of the bay--This is 1.5 to 1.8 m of sand to light sandy loam over sandy clay loam. There is coarse sand at the base grading to medium sand at the surface. There may be a color B or a textural B horizon developed in the sandy body. The topography has a subdued eolian character (Daniels and others, 1969).

3. North rim area--There is 50 cm or less of A horizon over sandy loam or sandy clay loam to a depth of 1.2 m.

4. Northeast rim area--This consists of about 1 m of sandy loam over sandy clay loam. Unit 2 grades into this unit. The sandy A horizon is less than 50 cm thick.

5. A small isolated slightly higher area--This is sandy loam about 1.6 m thick. The sandy A horizon is about 45 cm thick.

6. General area surrounding bay--All A horizons are much less than 50 cm thick and are usually not much more than the plow horizon. The B horizons are sandy clay loam with medium sands.

The southwest rim of the Rayfield-Lee Bay is obscure and the bay interior merges almost imperceptibly with the surrounding area. The edge of the bay is marked only by the transition, easily seen on airphotos, from poorly drained to moderately well drained soils. This part of the rim, shown as a dotted line in Figure 7, is in textural area 6.

Figure 8 is a cross section along a deep drainage ditch that cuts through the east rim of the bay. The various sand size and textural units are shown. There is a bed of medium coarse loamy sand and

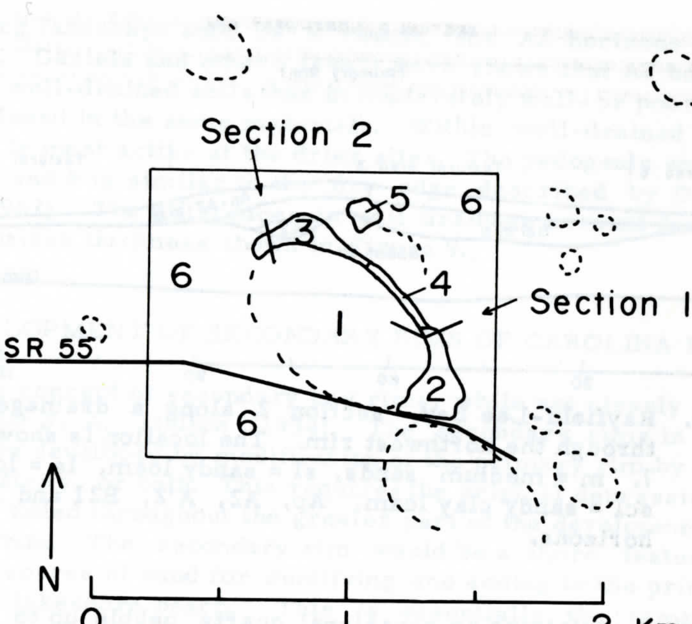


Figure 7. Map of the Rayfield-Lee Bay showing surface texture areas and the locations of two drainage ditch sections. The textural areas are described in the text. The dashed outlines indicate other bays or depressions.

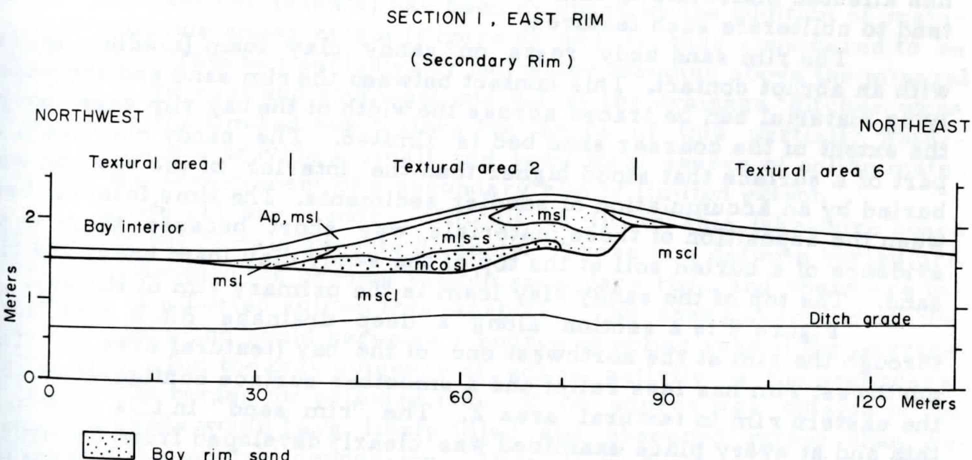


Figure 8. Rayfield-Lee Bay, section 1 along a deep drainage ditch cutting through the east rim. The location is shown in Figure 7. m = medium sands, mco = medium coarse sands, s = sand, ls = loamy sand, sl = sandy loam, scl = sandy clay loam, Ap = plow layer.

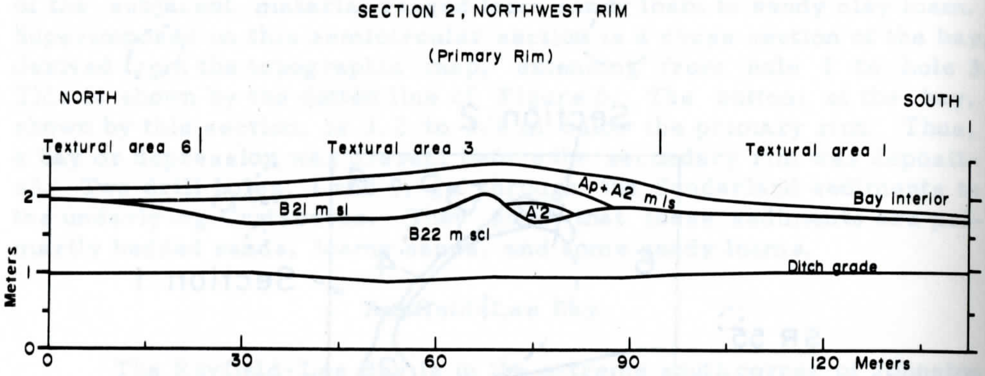


Figure 9. Rayfield-Lee Bay, section 2 along a drainage ditch cut through the northwest rim. The location is shown in Figure 7. m = medium sands, sl = sandy loam, ls = loamy sand, scl = sandy clay loam, Ap, A2, A'2, B21 and B22 are soil horizons.

sandy loam, containing an occasional quartz pebble up to 0.5 cm diameter, at the base of the rim sand body. In the vicinity of the 60 meter station there is some small textural and sand size variation in the lower part of this coarser bed, that is suggestive of bedding. No other evidence of primary sedimentary structures, other than the gross size and textural variations shown in Figure 8, was seen. Soil formation, which has affected materials in this area to a depth of 2 m or more, would tend to obliterate such features.

The rim sand body rests on sandy clay loam (medium sands) with an abrupt contact. This contact between the rim sand and the underlying material can be traced across the width of the bay rim even though the extent of the coarser sand bed is limited. The sandy clay loam is part of a surface that stood higher than the interior of the bay and was buried by an accumulation of sandier sediments. The time interval between the deposition of these materials was short because there is no evidence of a buried soil at the top of the sandy clay loam under the rim sand. The top of the sandy clay loam is the primary rim of the bay.

Figure 9 is a section along a deep drainage ditch that cuts through the rim at the northwest end of the bay (textural area 3). The northwest rim has less relief and a smoother surface configuration than the eastern rim in textural area 2. The "rim sand" in this section is thin and at every place examined was clearly developed from the underlying material by soil formation. Medium sands occur throughout the section, and there is no evidence of a lithologic discontinuity at any point. The loamy sand surface horizon is the A2 horizon (A1 + A2) of a soil, and overlies a B21 and a B22 horizon in the normal sequence.

The A2 horizons are thicker in textural area 3 (and also 5) than in area 6 because area's 3 and 5 are higher and better drained than the surrounding landscape and, as a result, the A2 horizons are better developed. Daniels and others (1967) have shown that A2 horizons are thicker in well-drained soils than in moderately well- or poorly-drained ones developed in the same materials. Within well-drained soils, A2 formation is most active at the drier sites. The pedogenic environment in areas 3 and 5 is similar to the dry edge described by Daniels and Gamble (1967). The differences in soil drainage account for the variable A2 horizon thickness shown in Figure 9.

## DEVELOPMENT OF SECONDARY RIMS OF CAROLINA BAYS

The concept of secondary bay rims, while not clearly stated, is alluded to by F. A. Melton (1934). The secondary rims in this study appear to be developed by modification of the primary rim by eolian or aqueous activity, or both. This requires the original depression to have been water filled throughout the greater part of the development of the secondary rim. The secondary rim would be a shore feature of a bay lake. The source of sand for modifying and adding to the primary rim is from the lakeshore beach. This is essentially the same sort of mechanism as that suggested by Thom (1970) for the development of bay rims in South Carolina and is part of the "artesian-solution-lacustrine-aeolian hypothesis" suggested by Johnson (1942).

The Rayfield-Lee Bay fits the lakeshore beach hypothesis. The northwest rim (area 3) is a primary rim with soils that have thicker A2 horizons than adjacent wetter areas, giving a sandy character to the rim. The east rim (area 2) has been modified by the addition of material of probable eolian origin (Figure 8) and is thus considered to be secondary.. This bay periodically had water standing above the mineral surface in the east and southeast end before the drainage ditches were dug. Thus the limited beach along the shore of this partially water-filled bay would provide a small and very local source of eolian material for the development of a secondary rim of limited extent.

Route 111 bay is more difficult to fit to this lakeshore beach hypothesis but certain aspects do apply. There is a definite buried primary rim beneath the present visible secondary rim, and there are indications of a buried soil on this surface. The areas surrounding the bay are sandy and could serve as a source of eolian sand if the surface were bare. But the only evidence of eolian activity is the secondary bay rim that buries the primary rim. If the source was outside of the bay, the sands would have likely filled and obscured rather than accentuated the original depression. There is no historical evidence that this bay was water filled. It seems probable that it may have been at some time before water tables were lowered as a consequence of stream dissection (Daniels, Gamble, and Nelson, 1971). The permeable Sunderland sediments under the bay would permit ready drainage once an outlet for subsurface drainage was established.

Formation of secondary bay rims as a consequence of wind and water action along the shores of a shallow body of water can account for the development of multiple bay rims and bays within bays. Receding water levels could alter the shape of the shoreline and cause one or more subsequent secondary rims to develop inside the confines of the first one. The altered shape of the water body could cause the new secondary rim to truncate and obliterate part of the old secondary rim.

A secondary rim can develop some time after the original depression and its primary rim is formed. The buried soil under the secondary rim of State Route 111 bay shows that the depression did not have a secondary rim for some time after it was formed. Thus, in this bay, the primary rim and the secondary rim are significantly different in age. In the Rayfield-Lee Bay, there is no buried soil beneath the secondary portion of the rim. This suggests only a short interval between the origin of the primary rim, and the deposition of the secondary rim sand. The two may be nearly contemporaneous. It would seem possible for the present form of the bays and rims on any one Coastal Plain surface to have considerable variation in age, depending on the time between the formation of the primary depression and the development of a secondary rim.

## PRIMARY DEPRESSIONS

The origin of bays appears, in part, related to the textural characteristics of the sediment. In general, we have found that bays are common where the surficial Coastal Plain sediments are sandy (sandy loams, sandy clay loams) and are absent or few in number in areas of silty or clayey materials. Thom (1970) has noted a similar relationship in South Carolina as have Mixon and Pilkey (1976) in North Carolina. Bays do not occur north and west of the Goldsboro Ridge where silty and silty clay soils developed in a clay bed that forms the top of the Sunderland "formation" (Daniels, Gamble, and Wheeler, 1971). There are many bays in sandy materials on the Sunderland surface to the southeast of the Goldsboro Ridge. There are no bays where silty soils are dominant, as in Wilson County, in Wayne County near Mt. Olive, and in southern Wayne and northern Duplin County. The many bays in southern Johnston County are associated with relatively sandy sediments.

Development of the secondary rims of Carolina Bays requires a depression deep enough or in a wet enough site for it to be partially water-filled. Therefore the basic problem in the study of Carolina Bays is the origin of this primary depression. There is a range in shape of depressions on the depositional surfaces of the Coastal Plain from irregular to circular to well-shaped elliptical. This range in shape can be seen in the vicinity of the Rayfield-Lee Bay and at the Goldsboro Ridge. Usually the large depressions are more or less elliptical and small ones are not. The present form of many bays, i. e., those with secondary rims or multiple rims, appears to result from a modification of an original depression by the eolian lake shore process. It would seem in these cases as if the original shape of the depression and its mode of origin (fluvial, marine, or eolian) would be immaterial. The modification would seem to have been responsible for the final oval oriented shape. However, it is apparent from our observations that not all bays have an identifiable secondary rim and yet they have the characteristic oval shape. This suggests that some primary depressions were originally formed with an oval oriented shape.

The bays examined in this study and those examined by Bryant (1964) Preston and Brown (1964), and Thom (1970) are clearly surficial features without subsurface expression. This suggests that the primary depression, regardless of its original shape, was probably formed as a part of the final phase of the process of deposition of the surficial sediments. But the fact remains that the exact mechanism of origin of primary depressions is not known.

## SUMMARY AND CONCLUSIONS

Carolina Bays are surface features formed when the surficial sediments of the Coastal Plain were deposited. These primary depressions have, in some cases, been modified by the development of secondary rims that may bury the original rim or edge of the depression. The secondary rim is apparently a product of wind and water action around the shore of a water-filled bay. Rims within rims and other kinds of bay morphology resulted as water levels changed. The sandy character of many primary bay rims is simply a consequence of the development of a slightly thicker soil A2 horizon, caused by the better drainage on the edge of the primary bay depression. The rims of some bays are a combination of primary and secondary rim with the secondary accumulation of sand occurring on the southeast end.

## LITERATURE CITED

Bryant, J. P., 1964, Soils of the Carolina Bays and interbay areas in Scotland County, North Carolina: Ph. D. Dissertation: North Carolina State University, Raleigh, North Carolina, 256 pp.

Daniels, R. B. and Gamble, E. E., 1967, The edge effect in some Ultisols in the North Carolina Coastal Plain: Geoderma, v. 1, p. 117-124.

__________, and Bartelli, L. J., 1968, Eluvial bodies in B horizons of some Ultisols: Soil Science, v. 106, p. 200-206.

__________, and Buol, S. W., 1969, Eolian sands associated with Coastal Plain river valleys--some problems in their age and source: Southeastern Geology, v. 11, p. 97-110.

__________, and Nelson, L. A., 1967, Relation between A2 horizon characteristics and drainage in some fine loamy Ultisols: Soil Science, v. 104, p. 364-369.

__________, and Nelson, L. A., 1971, Relations between soil morphology and water-table levels on a dissected North Carolina Coastal Plain surface: Soil Sci. Amer. Proc., v. 35, p. 781-784.

__________, and Wheeler, W. H., 1971, The Goldsboro Ridge, an enigma: Southeastern Geology, v. 12, p. 151-158.

Johnson, D., 1942, The origin of the Carolina Bays: Columbia Univ. Press, New York, 341 pp.

Killigrew, L. P., and Gilkes, R. J., 1974, Development of playa lakes in southwestern Australia: Nature, v. 247, p. 454-455.

Melton, F. A., 1934, The origin of the supposed meteorite scars a reply: Jour. Geology, v. 42, p. 97-104.

Mixon, R. B. and Pilkey, O. H., 1976, Reconnaissance geology of the submerged and emerged Coastal Plain province, Cape Lookout area, North Carolina: Geological Survey Prof. Paper 859.

Preston, C. D., and Brown, C. Q., 1964, Geologic section along a Carolina Bay, Sumter County, South Carolina: Southeastern Geology, v. 6, p. 21-29.

Price, W. A., 1972, Oriented lakes: Origin, classification, and developmental histories: Playa Lake Symposium, ed. by C. C. Reeves, Jr., International Center for Arid and Semi-Arid Land Studies Publication No. 4, Lubbock, Texas.

Prouty, W. F., 1952, Carolina Bays and their origin: Geol. Soc. America, Bull., v. 63, p. 167-224.

Soil Survey Staff, 1951, U. S. Dept. Agriculture Handbook No. 18 - Soil Survey Manual: USDA, Government Printing Office, 502 pp.

Thom, B. G., 1970, Carolina Bays in Horry and Marion Counties, South Carolina: Geol. Soc. America Bull., v. 81, p. 783-813.

# GLACIAL, PERIGLACIAL, AND PSEUDO-GLACIAL FEATURES IN THE GRANDFATHER MOUNTAIN AREA, NORTH CAROLINA

By

Loren A. Raymond
Department of Geology
Southern Oregon State College
Ashland, Oregon 97520

## ABSTRACT

A wide variety of pseudo-glacial, periglacial, and glacial features are present in the Grandfather Mountain area of North Carolina. Pseudo-glacial features include outcrops shaped like whaleback forms, grooved and polished bedrock, striated cobbles, and till-like deposits. Periglacial features include tors, blockfields, blockstreams, carpedoliths, and a possible solifluction lobe. Probable relict glacial features include cirques and a U-shaped valley. The longitudinal and cross profiles of the U-shaped valley compare favorably with those of known glacial valleys in the Tatoosh Range, the Rocky Mountains, and the northern Appalachian Mountains. The presence of these physical features (when properly distinguished from periglacial and pseudo-glacial features), supported by temperature calculations for the Pleistocene, studies of Pleistocene climate, and evidence of a southerly displacement of biotopes during the ice ages, supports the hypothesis that alpine glaciation occurred in the southern Appalachian Mountains.

## INTRODUCTION

The hypothesis of alpine glaciation in the southern Appalachian Mountains has been given inadequate attention in studies of the climatic and geomorphic history of the southeastern United States. Although Wentworth (1928) considered the glacial hypothesis as an explanation for striated cobbles found in terraces of streams that drain the higher Appalachians, he discarded it in favor of a river-ice hypothesis. Wentworth's tentative conclusion, that alpine glaciation had not occurred in the southeast, was coupled with the admonition that "much detailed work with accurate maps is needed in higher areas of the southern Appalachians before this (glacial) hypothesis can be discarded." In spite of this admonition, Wentworth's conclusion became dogma (e. g. Flint, 1957; 1971, Ch. 18, 19) before the detailed work which he suggested was completed.

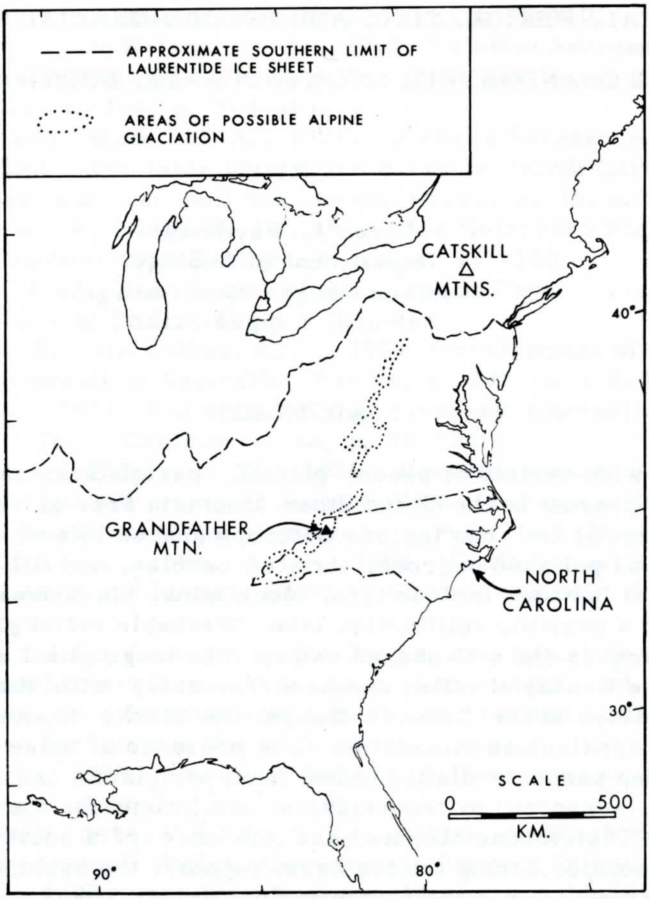


Figure 1. Map showing the location of the Grandfather Mountain area, North Carolina.

In 1973 Berkland and Raymond resurrected the glacial hypothesis as a result of the discovery of a U-shaped valley, a cirque-like feature, and grooved and polished bedrock on Grandfather Mountain, North Carolina (Figures 1 and 2). We later concluded (Raymond and Berkland, 1974) that some of the grooves and polish were produced by steel cables used in logging operations during the early part of this century. This conclusion was confirmed by several other workers (Carson and others, 1974; Hack and Newell, 1974; McKeon, 1974; McKeon and others, 1974), leading them to reject the glacial hypothesis. Some of these workers ignored the U-shaped valley and cirque-like feature, leaving them unexplained, but Carson and others (1974) attempted to attribute the origin of the latter features to mass wasting processes acting under contemporary climatic conditions.

Figure 2. View northwest across Boone Fork valley. Note trough shoulder on south (left). (Photo by J. O. Berkland).

Because controversy surrounding the glacial hypothesis resulted in part from the recognition of the pseudo-glacial features, reported by Raymond and Berkland (1974), Berkland and Raymond (1974), and others (e. g. Carson and others, 1974), it seems critical to establish which features are pseudo-glacial, which features are periglacial, and which features are best explained in terms of a glacial origin. In this paper, I attempt to distinguish between these various types of features and to summarize some of the evidence which supports the hypothesis of alpine glaciation in the southeast.


## Acknowledgements

It is a pleasure to express my gratitude to those who helped make this study possible. Mr. James O. Berkland provided constant encouragement and photographs. Appalachian State University and Mr. Hugh Morton provided support. Technical assistance and (or) criticism were provided by Drs. Jack Callahan, F. K. McKinney, C. W. Myers, Darryll Pederson, Fred Webb, Jr., Mr. Mike Murray, Ms. Carole Muirhead and Mrs. Barbara Winkler. Jay Fleisher reviewed the manuscript and offered several helpful suggestions.

## LOCATION AND GENERAL GEOLOGY

Grandfather Mountain is an 11.5 km long, northeast trending ridge located in northwestern North Carolina (81°48'-50'W, 36°07'N) (Figure 1). The general geology of the area has been summarized by Bryant and Reed (1970). The late Precambrian bedrock, which crops out throughout the area, consists of low grade metamorphic rocks including phyllite, metasandstone, metaconglomerate, felsic and mafic metavolcanic rocks, and metadiabase. The structure of these rocks is poorly understood, but two or more periods of folding have occurred. Although regional strike in the area is approximately N50°E and the dip is generally southeasterly, superposed folding has made local attitudes highly variable. This structurally complex Precambrian bedrock is overlain locally by Late Cenozoic colluvium, alluvium, blockfields and blockstreams.

## PSEUDO-GLACIAL FEATURES

Several types of pseudo-glacial features have been recognized in the Grandfather Mountain area including grooved and polished bedrock, outcrops shaped like whaleback forms, striated cobbles, and moraine-like deposits. Berkland and Raymond (1973) first reported and figured grooved and polish bedrock, believing it to be of glacial origin. Additional work around the north end of Grandfather Mountain revealed intersecting grooves along Moody Mill Creek (a V-shaped valley) and a single groove in a frost crack near the original groove site. The latter find proved that some grooves were not of glacial origin and suggested that they may have been man-made (Raymond and Berkland, 1974; Berkland and Raymond, 1974).

The pseudo-glacial grooves are anomalously uniform in width. However, measurement of more than two dozen grooves does show that groove widths range from 1.5 to 5.5 cm and that the grooves show both circular and parabolic cross sections (Figure 3). Because grooves and polish occur together, it is reasonable to assume that they have a common origin. Thus, as some of the grooves are man-made, it is probable, especially considering the degree of weathering of man-made grooves, that all of the grooves and polish are artificial. This, however, has not yet been demonstrated.

Outcrops shaped like whaleback forms are common in the Grandfather Mountain area (Figure 4). Carson and others (1974) recognized these features, noting that they are pseudo-glacial. These features usually are developed where foliation in exposed bedrock is inclined at a low angle to the slope. Stream erosion, slope wash, and (or) decomposition act on such outcrops to produce the rounded forms observed.

Striated cobbles were reported by Wentworth (1928) in the terraces of several streams which drain the higher Appalachians.

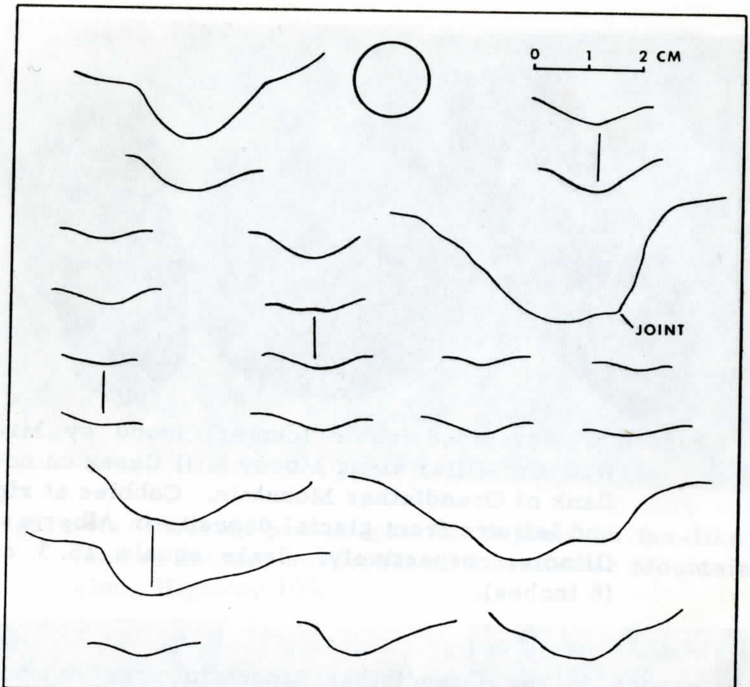


Figure 3. Cross sections of typical pseudo-glacial(?) grooves from Boone Fork locality of Berkland and Raymond (1973). Sections connected by lines represent sections of the same groove at various points along its length. Circle represents approximate cross section of steel cables used in logging operations.

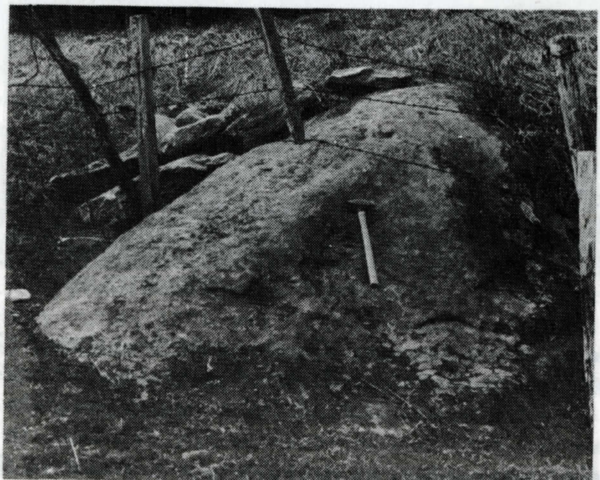

Figure 4. Outcrop shaped like a whaleback form along Highway 105 north of Grandfather Mountain. (Photo by Mike Murray).

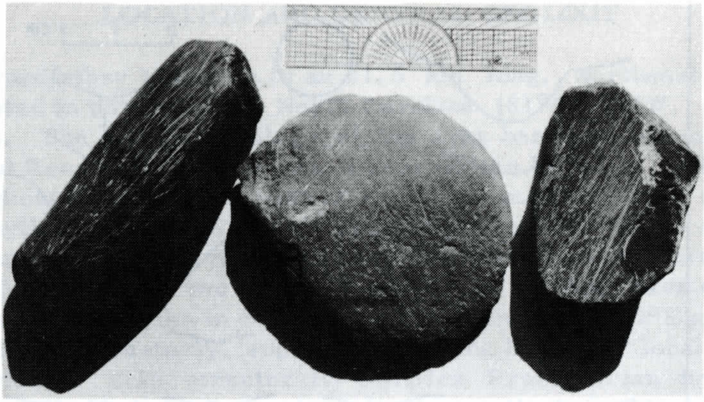

Figure 5. Scratched cobble (center) found by Mr. William Miller along Moody Mill Creek on north flank of Grandfather Mountain. Cobbles at right and left are from glacial deposits in Alberta and Illinois, respectively. Scale equals 15. 3 cm (6 inches).

Reconnaissance in the Grandfather Mountain area has not revealed striated cobbles to which an unequivocably natural origin can be assigned. Some striated cobbles, probably scratched by grading, have been observed (Figure 5). The latter are generally quite similar to glacially striated cobbles and would probably be indistinguishable from them if water worn.

Till-like deposits which commonly have a fan-like topographic form are widely distributed, not only in the Grandfather Mountain area, but in the southern Appalachian Mountains as a whole (see Michalek, 1969, especially p. 19 and 29, and Plate II). In the Grandfather Mountain area, these are typically unsorted, non-bedded masses containing angular boulders (Figure 6), but locally, poorly bedded, moderately sorted to unsorted deposits of similar appearance are also present. These deposits have been assigned a periglacial origin by Michalek (1969, p. 135).

## PERIGLACIAL FEATURES

In addition to the till-like deposits mentioned above, periglacial features present in the Grandfather Mountain area include tors, blockfields and blockstreams, carpedoliths, and a possible solifluction lobe. As noted above, the till-like deposits are fan-like in plan. Michalek (1969) studied the nature, distribution, and origin of these fan-like features, and the blockfields and blockstreams associated with them, and concluded that they originated under periglacial conditions. Similar

Figure 6. Unsorted pseudo-glacial deposit in fan-like feature located northwest of Grandfather Mountain along Highway 105.

Figure 7. Typical blockfield. Blockfield is located in a tributary valley of Boone Fork southeast of the cirque-like feature. Blocks shown generally range from 1. 2 to 2. 5 meters (four to eight feet) in length. (Photo by J. O. Berkland).

conclusions about these and like deposits elsewhere in the southern Appalachian Mountains have been reached by others (e. g. Kerr, 1881; Brunnschweiler, 1962; Clark, 1968), although Hack and Goodlett (1960) concluded that such deposits might have formed under contemporary climatic conditions. Thus, the presence of stratified, moderately sorted deposits in several fan-like features in the Grandfather Mountain area suggests the need for additional study of these landforms.

Blockfields and blockstreams (Figure 7) are very common on Grandfather Mountain. Where well developed, entire slopes are covered with a network of contiguous blocks that are up to 15 meters in length. The origin of blockfields and blockstreams was discussed by Washburn (1973, p. 191-193) who noted that blockfields "of truly angular blocks are certainly reasonable evidence of former frost wedging if located in an environment where such blocks are not accumulating today...." Michalek (1969) summarized evidence which indicates that the Grandfather Mountain blockfields and blockstreams are presently inactive and he concluded that these features formed under former periglacial conditions.

Carpedoliths, tabular layers of stones within soil, were named by Parizek and Woodruff (1957). The term is used here to designate layers of stone in till-like deposits. These features were formerly called stone-lines because of their appearance in cross-section (Figure 8). Carpedoliths are common in the till-like deposits of the Grandfather Mountain area. Parizek and Woodruff (1957) suggested that carpedoliths may originate at the surface in a variety of ways, including through extensive frost action. Because these features occur within the till-like deposits which have been assigned a periglacial origin, they are tentatively considered here to be buried stone layers of possible periglacial origin.

A possible solifluction lobe is present in the valley of Boone Fork (Number 1, Figure 9). This lobate mass of material is covered by a blockfield. Therefore, the internal structure and composition are unknown and determination of the nature and origin of this deposit awaits further study in artificial outcrops.

## GLACIAL FEATURES

At the northeast end of Grandfather Mountain there are at least six valley heads at elevations above 1372m (4500') that have cirque-like forms (Figure 9). In addition, one cirque-like feature heads a U-shaped valley, the valley of Boone Fork, in which pseudo-glacial grooves and polish were originally discovered (Number 1, Figure 9). All of these valleys become V-shaped downstream (Figure 10). They all trend between N50$^{\circ}$W and N80$^{\circ}$E, and are located on the northwestern and eastern slopes of the mountain. Valleys which face other directions (southeast, south, or west) are V-shaped, even at elevations of 1525 m (5000 feet) (e. g. compare valley Number 6, Figure 9 with the next two valleys to the southwest).

The origin of these cirque-like features and the U-shaped valley

Figure 8. Stone line (trace of carpedolith) in fan-like feature located west of Grandfather Mountain along Highway 105. Hammer handle at center is 46 cm (18 inches) in length.

has been attributed to glacial erosion (Berkland and Raymond, 1973; Raymond, in preparation) and to mass wasting (Carson and others, 1974). Similar features, present in southern North Carolina, were described by Haselton (1973), who suggested that they were formed by glacial erosion or nivation.

Mass wasting does not seem to account for the origin of these cirque-like features or the U-shaped valley. Bedding and foliation, which are often subparallel, are highly variable in orientation and do not seem gentle enough to have acted as slip planes for landslides. Because the lithologies which underlie Grandfather Mountain are equivalent throughout the length of the mountain, lithologic and stratigraphic factors explain neither the localized distribution of cirque-like features on the northwestern and eastern slopes, nor the presence of these features only at valley heads. Thus, bedrock factors apparently do not explain the localization or the presence of the cirque-like features or the U-shaped valley. Topographic factors also fail to account for the distribution of these features, as slopes equally steep and favorable for landsliding occur elsewhere on the mountain where no cirque-like features are present. Therefore, regardless of the presence of processes which might initiate a slide, the conditions on Grandfather Mountain do not seem favorable for the development of rockslides of the dimensions necessary to create the cirque-like features or U-shaped valley - nor do these conditions account for the localization of the features.

Both the glacial hypothesis and the nivation hypothesis could satisfactorily explain the location of cirque-like features at valley heads on the northwestern and eastern slopes of the mountain. There masses

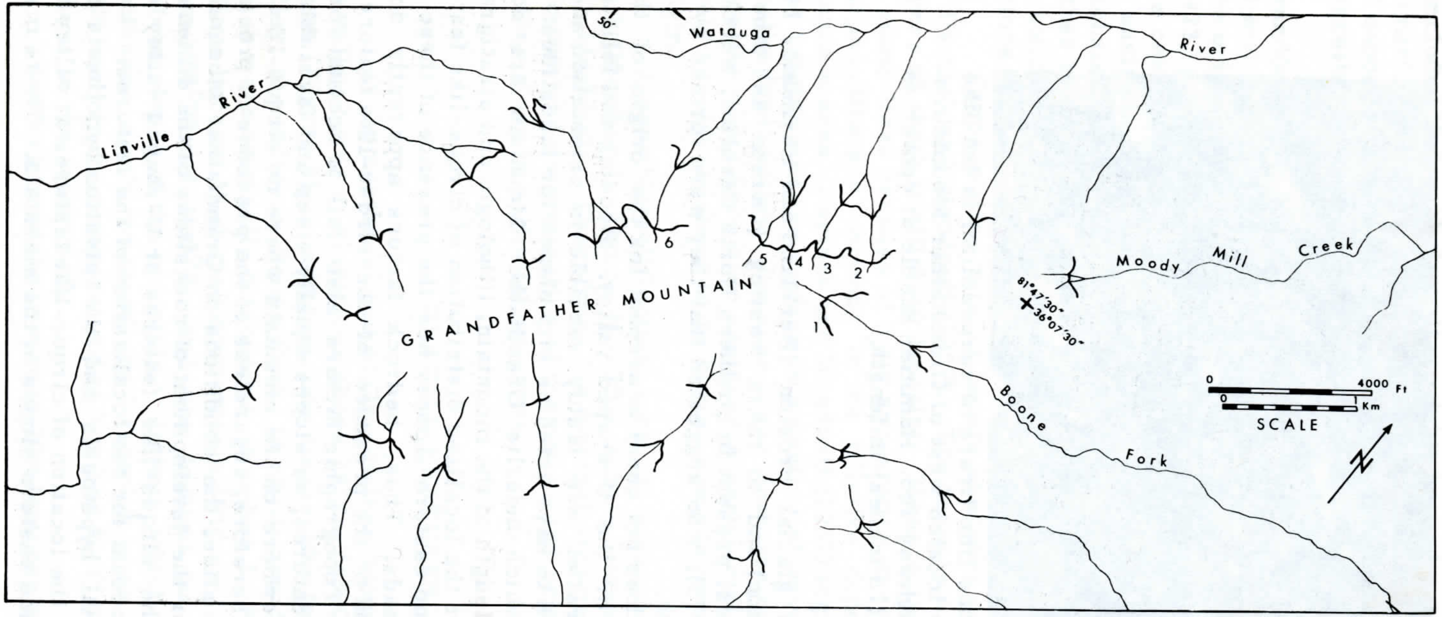


Figure 9. Map showing location of cirque-like features on Grandfather Mountain, N. C. Possible cirques are numbered 1-6. Boone Fork = 1. Fine lines represent drainage lines. Heavy lines represent 1341 m (4400 foot) and, where present, 1524 m (5000 foot) contour segments. Note that U-shaped valley heads are present only in the northwest and east.

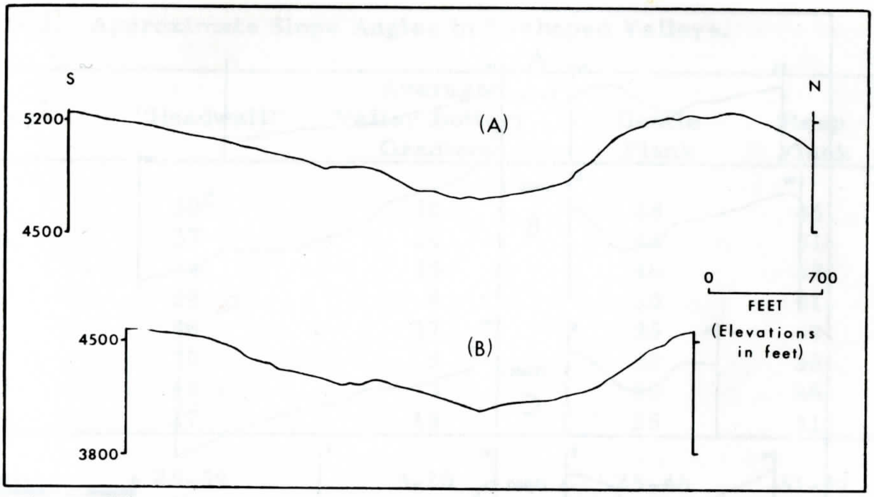


Figure 10. Topographic cross sections across Boone Fort valley above (A) and below (B) the point where the valley changes from a U- to a V-shape. Elevations in feet. Profile natural.

of ice or perennial snow fields would have minimum exposure to the sun's rays. These hypotheses also explain the change in valley form from U- to V-shapes, as such a change would occur downstream where water replaces ice as the dominant agent of erosion. To differentiate between the latter two hypotheses would require discrimination between two degrees of freeze-thaw activity and ice erosion.

If the cirque-like features are true cirques and if the U-shaped valley is a glacial feature, one would expect these features to resemble glacial features in known alpine glacial terranes. Figure 11 compares the longitudinal and cross sectional profiles of Boone Fork valley, (the U-shaped valley - designated number 1 on Figure 9) with those of known glacial valleys in the Tatoosh Range of Washington, the Rocky Mountains of Colorado, and the northern Appalachian Mountains in the New England states.

There are several similarities between the Boone Fork valley and other known glacial valleys. First, Boone Fork has a valley shoulder (Schulterflache) and flat upland surface (Schliffbord) on the south side of the valley (Figure 11A; also see Figure 2) which perhaps parallels the former fluvial valley bottom (Machatschek, 1969, p. 130). A similar feature is revealed in the profile from Unicorn Creek (Figure 11B). Valley shoulders of this type are especially common in some glacial terrains (e. g. in the Tatoosh Range and on Mt. Rainier). Second, the longitudinal profile of Boone Fork does not differ significantly from those of the known glacial valleys (Figure 11, Table 1). Third, the slopes of the Boone Fork valley walls compare favorably with those

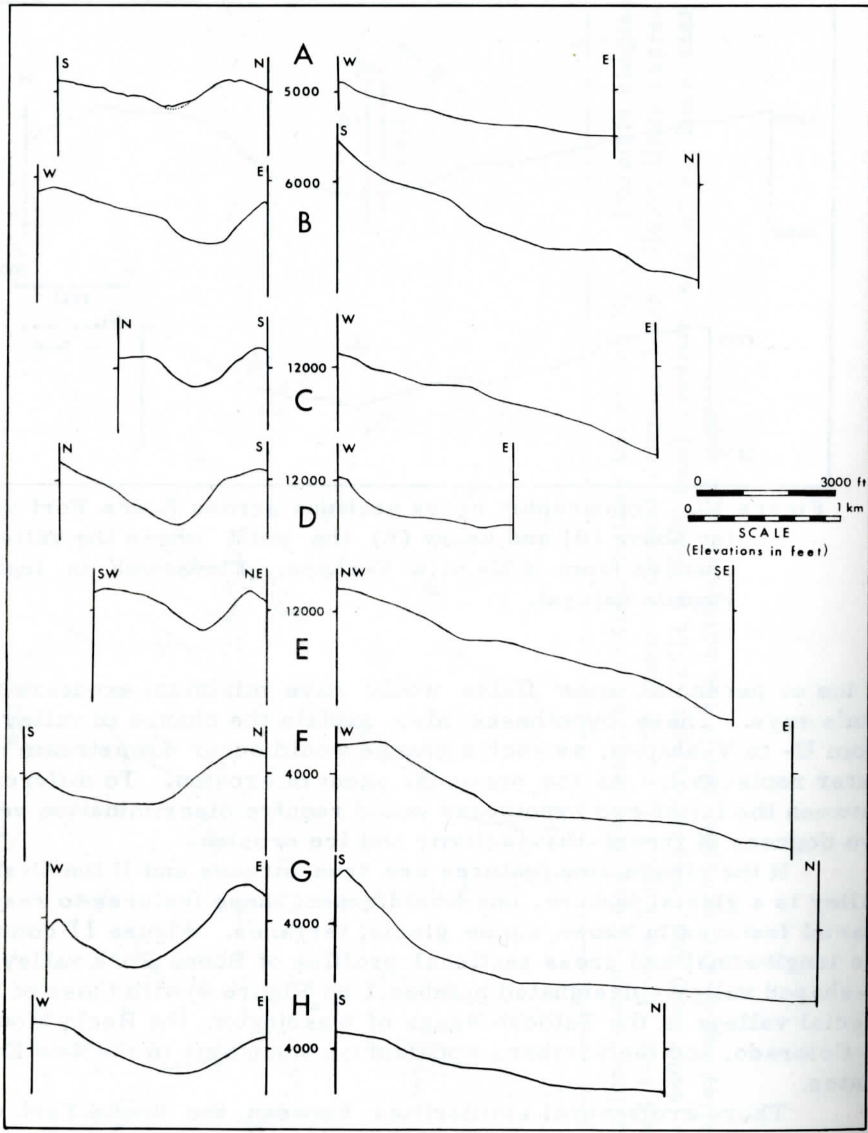


Figure 11. A comparison of the cross section and longitudinal profile of Boone Fork with similar sections from known glacial valleys. A = Boone Fork, Grandfather Mountain, N. C. B = Unicorn Creek, Tatoosh Range, Washington. C = Tyndall Gorge, Rocky Mountain National Park, Colorado. D = Chaos Canyon, Rocky Mountain National Park, Colorado. E = Hidden River Valley, Rocky Mountain National Park, Colorado. F = North Basin, Mt. Katadin, Maine. G = South Basin, Mt. Katadin, Maine. H = Little River Valley, Mt. Guyot, White Mountains, N. H.

Table 1. Approximate Slope Angles in U-shaped Valleys.

| Valley[1] | "Headwall" | Average Valley Bottom Gradient | Gentle Flank | Steep Flank |
|---|---|---|---|---|
| A | 30[2] | 10 | 28 | 45 |
| B | 37 | 20 | 42 | 51 |
| C | 34 | 15 | 46 | 48 |
| D | 59 | 9 | 30 | 51 |
| E | 25 | 17 | 35 | 49 |
| F | 30 | 5 | 30 | 38 |
| G | 44 | 3 | 30 | 36 |
| H | 27 | 13 | 25 | 31 |
| Range | 25-59 | 3-20 | 25-46 | 31-51 |
| Boone Fork | 30 | 10 | 28 | 45 |

[1] Valleys are the same as those listed in Figure 11.
[2] Values in degrees.

of known glacial valleys, as revealed by these arbitrarily selected cross profiles (Table 1). Thus, the location of the Boone Fork valley favors a glacial origin and the valley shape permits such an interpretation. If Boone Fork is a glacial valley, it is reasonable to consider that the five cirque-like features located on the northwestern slope of Grandfather Mountain may be glacial in origin.

The presence of saprolite in valley bottoms at lower elevations indicates that deep erosion has not occurred. In the upper Boone Fork valley, saprolite has not been observed. However, whether this is a function of glacial or periglacial erosion is unknown.

## DISCUSSION

The presence of the physical evidence, i.e. the cirques and the U-shaped valley, when properly distinguished from periglacial and pseudo-glacial features, supports the hypothesis of glaciation in the southern Appalachian Mountains. However, the physical evidence alone is inconclusive, as it has been subject to alternate interpretations. Temperature studies and calculations for the Pleistocene, paleoclimatic studies, and a southerly displacement of biotopes are all consistent with the glacial hypothesis.

Berkland (ms) has deduced Pleistocene temperatures for the highest peaks in the southern Appalachian Mountains using the modern latitudinal and altitudinal temperature gradients. He suggests a Wisconsinan cooling of 20°C for elevations of 2000 m (6560 feet). At an elevation of 1524 m (5000 feet), an elevation lying within the range of elevations for Grandfather Mountain cirques, Berkland's temperature calculations suggest a mean annual temperature of -4.7°C.

Berkland's study (ms) yields temperatures slightly lower than those derived in some other studies of Pleistocene temperatures and significantly lower than those reported by Flint (1971, p. 72) and Whitehead (1973). Values similar to those of Berkland were reported by Goldthwait (1959), who suggested a mean annual cooling of 16.5°C for Ohio, and Watts (1970), who proffered a mean January cooling of 18°C based on his studies of fossil vegetation in northwestern Georgia. Also, Schroeder and Bada (1973) (using the data of Mitterer) reported a minimum cooling of 15°C for the southeastern states. All of the latter values yield mean annual temperatures within the range 10°C to -10°C reported for contemporary snowlines (see Charlesworth, 1957, p. 9) - temperatures appropriate for the development of alpine glaciers.

The results of paleoclimatic studies are controversial (Bryan and Cady, 1934; Lamb and Woodroffe, 1970; Barry and others, 1971; Williams and others, 1973). However, some studies have delineated possible storm tracks for full glacial conditions consistent with the glacial hypothesis. For example, Bryan and Cady (1934) depicted generalized storm tracks which were displaced southward so that they passed through North Carolina. Similarly, for a computer simulated test case for full glacial January, Williams and others (1973) found that zones of cyclone activity were displaced southward. Although these models are based, in part, on moot assumptions, some of the models suggest that precipitation may have been adequate to sustain a positive net mass budget for alpine glaciers in the southeast during glacial maxima.

Paleontological studies also support the glacial hypothesis. Whitehead (1973) recently reviewed the controversy concerning the impact of full - glacial conditions on vegetation zones in North America and concluded that the data for the southeast are consistent with a 1000 km southerly displacement of boreal forest elements. Whitehead (1973) assigned the higher Appalachians in North Carolina to the Tundra - Taiga zone.

A wide variety of vertebrate fossils also suggests substantially colder conditions in the southeast. Fossil species and genera with boreal affinities found in the Appalachian Mountains south of the continental ice margin, include such animals as musk ox (Ovibos moschatus), caribou (Rangifer tarandus), a boreal-type vole (Phenacomys intermedius), collared lemming (Dicronstonyx hudsonius) and ptarmigan (Lagopus) (e.g. Guilday, 1971; Guilday and Parmalee, 1972; also see reviews in Charlesworth, 1957, p. 789 ff.; Berkland, ms; and Voorhies, 1974). In addition, Vorhies (1974) recently discovered a Pleistocene vertebrate fauna in the Georgia Piedmont which "reveals that a cool, moist climate suitable for boreally adapted mammals once extended rather far south of the mountains, nearly to the coastal plain."

In the southern Appalachians, no endemic vertebrate species of boreal affinities are known which owe their distribution to "stranding" during glacial withdrawal (Guilday, 1971). However, two scorpionflies (Boreus nivoriundus and B. brumalis), found at higher elevations in the Great Smokey Mountains, are believed to be "stranded" species (Cooper, 1972; Cooper, pers. comm., 1974).

The physical, biological, and paleontological evidence considered together with paleotemperature studies and studies of Pleistocene climates suggests that Alpine glaciation accompanied marked cooling of the southeastern states during Pleistocene glacial maxima. Perhaps this new evidence supporting the glacial hypothesis will encourage geomorphologists to undertake the detailed studies suggested nearly half a century ago by Wentworth (1928).

## REFERENCES CITED

Barry, R. G., Ives, J. D., and Andrews, J. T., 1971, A discussion of atmospheric circulation during the last ice age: Quaternary Research, v. 1, p. 415-418.

Berkland, J. O., ms (in preparation), Glaciation of the southern Appalachian Mountains and its significance: 44 p.

Berkland, J. O., and Raymond, L. A., 1973, Pleistocene glaciation in the Blue Ridge Province, southern Appalachian Mountains, North Carolina: Science, v. 181, p. 651-653.

__________, 1974, North Carolina glacier: Evidence disputed (Reply): Science, v. 184, p. 89-91.

Brunnschweiler, Dieter, 1962, The periglacial realm in North America during the Wisconsin glaciation: Biuletyn Peryglacjalny, v. 11, p. 15-27.

Bryant, Bruce and Reed, J. C., Jr., 1970, Geology of the Grandfather Mountain window and vicinity, North Carolina and Tennessee: U. S. Geol. Survey Prof. Paper 615, 190 p.

Bryan, Kirk, and Cady, R. C., 1934, The Pleistocene climate of Bermuda: American Journ. Sci., 5th Series, v. 27, p. 241-264.

Carson, R. J., Beck, R. U., Chappelear, J. W., Lanier, M. D., McCurry, G. W., Neel, R. H., Stanley, L. G., Walker, R. G., Wilson, J. R., 1974, Pseudo-glaciation of Grandfather Mountain, North Carolina: Geol. Soc. America, Abs. with Programs, v. 6, p. 340.

Charlesworth, J. K., 1957, The Quaternary Era: Edward Arnold LTD., London, 1700 p.

Clark, G. M., 1968, Sorted patterned ground: New Appalachian localities south of the glacial border: Science, v. 161, p. 355-356.

Cooper, K. W., 1972, A southern California Boreus, B. notoperates n. sp. I. Comparative morphology and systematics. (Mecoptera:

Boreidae).: Psyche, v. 79, p. 269-283.

Flint, R. F., 1957, Glacial and Pleistocene geology: John Wiley and Sons, Inc., New York, 553 p.

__________, 1971, Glacial and Quaternary geology: John Wiley and Sons, Inc., New York, 892 p.

Goldthwait, R. P., 1959, Scenes in Ohio during the last ice age: Ohio Journ. Sci., v. 59, p. 193-216.

Guilday, J. E., 1971, The Pleistocene history of the Appalachian mammal fauna: in The distributional history of the biota of the southern Appalachians: Part III: Vertebrates: ed. P. C. Holt, R. A. Patterson, and J. P. Hubbard: Virginia Polytech. Inst. and State Univ. Research Div. Mono. 4, p. 233-262.

Guilday, J. E., and Parmalee, P. W., 1972, Quaternary periglacial records of voles in the genus Phenacomys Merriam (Cricetidae: Rodentia): Quat. Res., v. 2, p. 170-175.

Hack, J. T., and Goodlett, J. C., 1960, Geomorphology and forest ecology of a mountain region in the central Appalachians: U. S. Geol. Survey Prof. Paper 347, 66 p.

Hack, J. T., and Newell, W. L., 1974, North Carolina glacier: Evidence disputed: Science, p. 88-89.

Haselton, G. M., 1973, Possible relict glacial features in the Black Balsam Knob and Richland Balsam area, North Carolina: Southeastern Geol., v. 15, p. 119-125.

Kerr, W. C., 1881, On the action of frost in the arrangement of superficial earthy material: American Jour. Sci., v. 21, p. 345-358.

Lamb, H. H., and Woodroffe, A., 1970, Atmospheric circulation during the last ice age: Quat. Res., v. 1, p. 29-58.

Machatschek, Fritz, 1969, Geomorphology: American Elsevier Publ. Co., Inc., New York, 212 p.

McKeon, J. B., 1974, North Carolina glacier: Evidence disputed: Science, v. 184, p. 88.

McKeon, J. B., Derksen, S. J., McSaveney, E. R., and McSaveney, M. J., 1974, Non-glacial origin of "glacial grooves" on Grandfather Mountain, North Carolina: Geol. Soc. America, Abs. with Programs, v. 6, p. 529-530.

Michalek, D., 1969, Fanlike features and related periglacial phenomena of the southern Blue Ridge: Univ. North Carolina, Chapel Hill, Ph. D. dissertation, 198 p.

Parizek, E. J., and Woodruff, J. F., 1957, Description and origin of stone layers in soils of the southeastern states: Jour. Geology, v. 65, p. 24-34.

Raymond, L. A., and Berkland, J. O., 1974, Reply on glaciers: in Letters: Geotimes, v. 19, no. 1, p. 9.

Schroeder, R. A., and Bada, J. L., 1973, Glacial - post-glacial temperature difference deduced from aspartic acid racemization in fossil bones: Science, v. 182, p. 479-482.

Voorhies, M. R., 1974, Pleistocene vertebrates with boreal affinities

in the Georgia Piedmont: Quat. Res., v. 4, p. 85-93.
Washburn, A. L., 1973, Periglacial processes and environments: St. Martins Press, New York, 320 p.
Watts, W. A., 1970, The full-glacial vegetation of northwestern Georgia: Ecology, v. 51, p. 17-31.
Wentworth, C. K., 1928, Striated cobbles in southern states: Geol. Soc. America Bull., v. 39, p. 941-954.
Whitehead, D. R., 1973, Late-Wisconsin vegetational changes in unglaciated eastern North America: Quat. Res., v. 3, p. 621-631.
Williams, Jill, Barry, R. G., and Washington, W. M., 1973, Simulation of the climate at the last glacial maximum using the NCAR global circulation model: Nat'l. Tech. Info. Serv. (Inst. Arctic and Alpine Res. Occas. Paper No. 5), 39 p.

# THE ARAPAHOE RIDGE - A PLEISTOCENE STORM BEACH[1]

By

R. B. Daniels
E. E. Gamble
Soil Conservation Service, USDA and
Soil Science Department
North Carolina State University
Raleigh, North Carolina

W. H. Wheeler
Geology Department
University of North Carolina
Chapel Hill, North Carolina

and

C. S. Holzhey
National Soil Survey Laboratory
Lincoln, Nebraska

## ABSTRACT

The Arapahoe ridge is a sand ridge nearly 30 miles long between the Pamlico and Neuse River estuaries. It is 1/2 to 1 mile wide and stands 15 to 20 feet above the Talbot plain to the west. The ridge sands rest abruptly upon sediments of the Talbot morphostratigraphic unit with the contact being marked by peat in more than 1/3 of the bore holes. The sands of the ridge apparently merge with the Pamlico sediments to the east and form part of the Suffolk scarp in the area between the Pamlico and Neuse Rivers. The Arapahoe ridge is believed to be a storm beach, modified by local eolian activity, formed during the maximum transgression of the Pamlico sea.

Two small sand ridges about 4 miles west of the Arapahoe are believed to be its estuarine equivalents. An eolian origin is suggested for these small ridges.

1 Joint contribution from the Soil Conservation Service, USDA, and the Soil Science Department, North Carolina Agricultural Experiment Station. Paper No. 5078 of the Journal Series of the North Carolina Agricultural Experiment Station, Raleigh, North Carolina.

## INTRODUCTION

Exposures along the estuaries of North Carolina have been studied by geologists since the 19th century, and it is not uncommon for strikingly different interpretations to be made about the same section or series of sections. For example, at least five formational names, the Pamlico, Talbot, Flanner Beach, Neuse, and Core Creek sand have been given the exposures Pleistocene surficial sediments along the Neuse estuary (Stephenson, 1912; Richards, 1950; DuBar and Solliday, 1963; Fallaw and Wheeler, 1969; Wheeler et al., in press; Daniels et al., 1972; Mixon and Pilkey, 1976). Stratigraphic studies during the last 15 years have depended less upon one or two exposed sections and more upon drill samples. The development of drilling equipment has given the later worker the distinct advantage of being able to trace the areal extent of his units, but this additional information has not eliminated the different interpretations of essentially the same stratigraphic unit. The differences in interpretation brought out in this paper concern the origin of nearly linear sandy bodies north of the Neuse estuary between New Bern and Morehead City, North Carolina (Figure 1). DuBar et al., (1974) mapped these sand bodies as parts of barrier systems and have named the western sand body the Reelsboro barrier and the eastern body the Arapahoe barrier. Daniels et al. (1972) considered the Arapahoe barrier (their Minnesott ridge) a storm beach ridge associated with the Pamlico sea. Mixon and Pilkey (1976) called the Arapahoe barrier the Minnesott sand.

In the following paragraphs, we will detail our stratigraphic and geomorphic evidence in an attempt to resolve these differing interpretation. We will use the term "Arapahoe ridge" instead of "Minnesott ridge" because DuBar et al. (1974) probably had their paper describing these features written long before we published our findings in a quickly compiled field trip guide book (Daniels et al., 1972). We will use the term "Reelsboro ridge" for only part of the sand body complex near Reelsboro mapped by DuBar et al., and the term "Cayton ridge" (new term) for a sand ridge south of the hamlet of Reelsboro, but west of the Reelsboro barrier as mapped by DuBar. The name Cayton is from the Cayton Cemetery near the center of the ridge.

### Background on Terminology Used

Because three groups of workers have recently published on the geology of the area, and each group used a different set of terms to describe the same feature, it is necessary to define how and why we use certain terms.

The Pamlico and Chowan "terrace formations" were mapped and explained by Stephenson (1912) (Table 1). Stephenson and others assumed that these units were deposited during a time of generally rising sea level and that their upper surfaces were shaped during a stillstand.

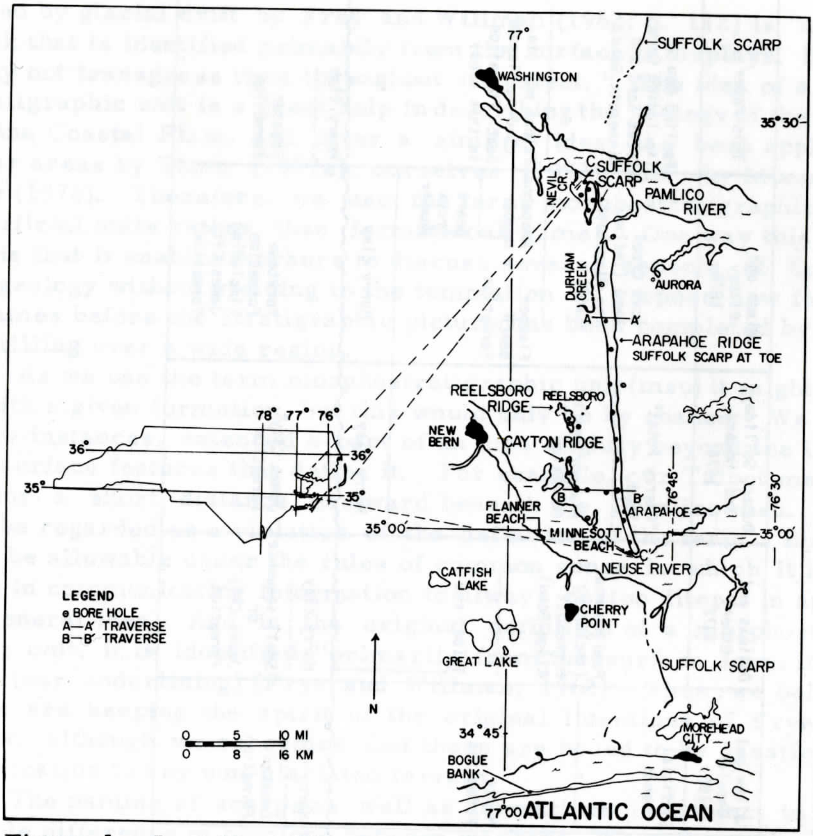


Figure 1. Location of linear sandy ridges between the Pamlico and Neuse estuaries.

A scarp formed along the temporary shores. The "terrace formations" were essentially mapped by the altitude of their surfaces on the interfluves. Later workers (DuBar and Solliday, 1963; Fallaw and Wheeler, 1969; Mixon and Pilkey (1976) have redefined the surficial units mapped as the Pamlico and Talbot by earlier workers, but these earlier names are firmly entrenched in geologic literature. Our approach has been opposite to that of most workers since 1960. We prefer to retain the earlier names given these stratigraphic units because we believe this leads to less confusion than giving a new formational name to each mappable unit. However, we have need of a term to use for all the surficial sediments that underly the Pamlico or Talbot, or whatever terrace plain, without regard to facies changes or interpreted environments, or whether or not the surface is erosional or depositional at any one place, or whether or not stratigraphic units do or do not terminate along the line of the scarp.

The term "morphostratigraphic unit" proposed for use in areas

Table 1. Review of terminology of Plio-Pleistocene units of the lower Coastal Plain in the Neuse-Tar area.

| | Shattuck 1901 | Stephenson 1912 | Richards 1950 | DuBar and Solliday 1963 | Fallaw and Wheeler 1969 | Daniels and Others 1972 | DuBar, Solliday and Howard 1974 | Mixon and Pilkey 1976 | This Paper |
|---|---|---|---|---|---|---|---|---|---|
| Outcropping Upper Pleistocene Units | Talbot Formation | Pamlico Formation | Pamlico unit<br>?<br>Talbot unit in some localities | Late Pleistocene | Surficial sands<br>?<br>Neuse Formation | Pamlico msu (including Minnesott Ridge Sand) | "Cherry Point" Sand<br>----------------<br>Flanner Beach Formation | Core Creek Sand (including Minnesott Sand) | Pamlico msu (including Core Creek Sand and Minnesott Sand) |
| (lower Pleistocene does not outcrop on lower Coastal Plain) | | Chowan Formation | Horry Clay | Flanner Beach Formation<br>Horry clay | | Talbot msu | Horry clay | Flanner Beach Formation: Arapahoe Member<br>-------<br>Beard Creek Member<br>Horry clay | Talbot msu (including Flanner Beach Formation in this area) |
| Pliocene (perhaps lowermost Pleistocene) | | Waccamaw Formation | Croatan Formation | James City Formation | Horry clay<br>----------<br>Croatan Formation | Horry clay<br>----------<br>Small sequence or James City Formation | James City Formation | Yorktown Formation (including James City Formation) | Horry clay<br>----------<br>Croatan Formation |

covered by glacial drift by Frey and Willman (1962; p. 112) is "a body of rock that is identified primarily from the surface it displays. It may or may not transgress time throughout its extent." The idea of a morphostatigraphic unit is a great help in describing the geology of the North Carolina Coastal Plain, and it or a similar idea has been applied to similar areas by Thom (1967a), ourselves (1972), and by Mixon and Pilkey (1976). Therefore, we use the term morphostratigraphic unit for surficial units rather than formational names. One way this term helps is that it enables workers to discuss several aspects of Coastal Plain geology without yielding to the temptation to propose new formation names before the stratigraphic picture has been completed by intensive drilling over a wide region.

As we use the term morphostratigraphic unit (msu) it might coincide with a given formation, but this would only be by chance. We have, in a few instances, extended a part of an msu slightly beyond the limits of the surface features that define it. For example, our Talbot msu extends for a short distance eastward beneath the Pamlico msu. This might be regarded as a violation of the definition of the term. But this should be allowable under the rules of common sense in which it is not helpful in communicating information to always be too literal in applying a general idea. And in the original definition of a morphostratigraphic unit, it is identified "primarily from the surface form it displays" (our underlining) (Frye and Willman, 1962). Thus, we believe that we are keeping the spirit of the original intentions of Frye and Willman, although we recognize that there are bound to be questions in its application to any non-glaciated terrain.

The naming of scarps as well as formations is subject to considerable difference of opinions between workers. We will use the term Suffolk scarp instead of the Grantsboro scarp as proposed by Mixon and Pilkey (1976). The Suffolk scarp as defined by Wentworth (1930) has a toe altitude of about 20 to 25 feet in the type area. This scarp has been traced southward by Oaks et al. (1974) as the Suffolk sand ridge and by Mixon and Pilkey as the Suffolk scarp. The toe altitude of the scarp where it forms the west border of the Dismal Swamp is between 20 and 25 feet, but the toe altitude in this area is controlled by the organic surface of Dismal Swamp, which is Holocene according to Whitehead (1972). South of the organic area of Dismal Swamp the Suffolk scarp has a toe altitude of about +20 feet, and remains at this altitude on south to near Edenton, North Carolina where it is interrupted by the Chowan River and Albemarle Sound. At the Virginia-North Carolina line, the Suffolk scarp truncates the Hazelton scarp of Oaks et al., (1974). The Hazelton is the equivalent of our Walterboro scarp that has a toe altitude of about +45 feet.

Mixon and Pilkey (1976) believe that south of Albemarle Sound their Pinetown scarp with a toe altitude of about +30 feet is the equivalent of the Suffolk and that their Union Chapel scarp with a toe altitude of about +20 feet is the equivalent of the Grantsboro south of the Pamlico

River. They believe that the Union Chapel and Hickory and Big Bethel scarps of the Norfolk area are equivalent. This correlation requires a sharp westward swing of the Hickory scarp along what is now Albemarle Sound.

We must disagree with this interpretation. The southward extension of the Suffolk scarp to include the Chapel and Grantsboro scarps appears to be much more logical than correlating the Suffolk with the Pinetown. First, the toe altitude of the Suffolk scarp north of Albemarle Sound is very close to +20 feet (except at the organic area of Dismal Swamp), as are the toe altitudes of the Chapel and Gransboro scarp. Secondly, to suddenly change the toe altitude of the Suffolk from +20 to +30 at the Pinetown scarp and then swing the Hickory scarp 35 miles to the west to again assume a toe altitude of about +20 feet violates both the areal continuity of the trend of the scarp and the toe altitude. Toe altitudes of higher scarps in North Carolina are remarkably uniform and there is little reason to expect the lower scarps to have sharp changes in toe altitudes over short distances. If the Hickory scarp is to extend westward to the Chapel scarp, then it also must go northward along the Suffolk scarp north of Albemarle Sound. The Pinetown scarp can more easily represent a barrier system related to the Talbot sea than to the Pamlico sea because the toe altitude is +30 feet and the area between the Pinetown and Chapel scarps is composed of sandy linear ridges (Mixon and Pilkey, 1974). For the above reasons, we are using the name Suffolk for the scarp on the east side of the Arapahoe ridge that has a toe altitude of +20 feet. If additional work proves Mixon and Pilkey to be correct in their interpretations, then our Suffolk scarp at the Arapahoe ridge should be renamed the Hickory if it is truly correlative.

## MORPHOLOGY

The Arapahoe ridge (the Arapahoe barrier of DuBar et al. (1974) is a linear sand ridge unique to the area between the Pamlico and Neuse Rivers in Beaufort and Craven Counties, North Carolina (Figure 1). The ridge stands from 10 to 20 feet aobve the Talbot plain to the west and 20 to 40 feet above the Pamlico plain to the east. The ridge is from 1/2 to 1 mile wide and is continuous for 27 miles from Durham Creek, 4 miles west of Aurora, south to the Neuse estuary (Figure 1). The Arapahoe ridge has a maximum altitude of slightly over +60 feet southwest of the settlement of Small in Beaufort County, North Carolina and minimum altitude of +35 feet near Minnesott Beach. Between the Pamlico River and Durham Creek is a 3 1/2 mile northward extension of the Arapahoe ridge that is bordered on the west by Nevil Creek and on the east by the Suffolk scarp (Figure 1). The ridge has a smooth, nearly level summit in most areas. A minor amount of subdued dune topography, less than 5 percent of the ridge, is found near the northern end of the ridge just southwest of the Sandy Grove Church (see the Aurora Quadrangle). The east slopes of the ridge are smooth and the contour lines are relatively straight for considerable distances, but the west slopes are more irregular (Daniels et al., 1972; DuBar et al., 1974). In the nearly level lower Coastal Plain (the Coastal Plain east of Surry scarp), the Arapahoe ridge is a distinctive feature on the ground, topographic sheets, and satellite photos.

West of the Arapahoe ridge are two linear sand bodies that we call the Reelsboro and Cayton ridges (Figure 1). Our Reelsboro ridge is only the northern part of the Reelsboro barrier of DuBar et al. (1974). The Cayton ridge, which is west of Reelsboro barrier as mapped by DuBar et al., 1974, is described here for the first time. It is very similar to the Arapahoe ridge in general morphology and altitude above the surrounding landscape. The major difference is that the larger and more conspicuous slope of Cayton Ridge faces to the west. The Pamlico plain is to the west and the Talbot plain to the east and north of the Cayton and Reelsboro ridges. The Reelsboro is less distinct than the Arapahoe or Cayton ridges, partly because it has a number of ephemeral and intermittent streams heading on its flanks. The Reelsboro and Cayton ridges have a subdued dune topography only on their crests.

All three ridges are sand to loamy sand with the fine sand fraction being dominant. Bh horizons, the humate of Thom (1967b) and Swanson and Palacas (1965), 2 to 15 feet thick are found in the sands of these ridges. See Figures 2 for the typical distribution of humate in Arapahoe ridge. Quartz and iron-rich opaques, magnetite, and ilmenite dominate the very fine sand fraction of Arapahoe ridge (Table 2) with feldspars and zircon being second in abundance. Vermiculite-chlorite and micas dominate the clay minerals.

There are very few exposures on the Arapahoe ridge where bedding can be studied. The sand of the ridge was massive in a sand pit north of NC highway 33 in Beaufort County, and some faint horizontal bedding was found in the 1975 sanitary landfill pit for the village of Arapahoe.

The sands of the Arapahoe ridge have an abrupt contact with the underlying sediments of the Talbot morphostratigraphic unit (msu) (Figures 2, 3, and 4). Peat 1 to 3 feet thick overlies the Talbot in many areas (Figure 2). The peat is part of the post-Talbot weathering that preceded the deposition of the sands of the Arapahoe ridge. Where peat or buried Al horizons are absent, the ridge sands have an abrupt contact with the clays, silts, and micaceous sands of the underlying Talbot. In about 1/3 of the drill holes, sandy Talbot sediments underlie the ridge sands. The contact between the sandy Talbot and ridge sands was identified by the downward increase in mica, decrease in opaques, and commonly a slight increase in clay content in the Talbot sands. In traverse B-B' (Figure 3), the Talbot is dominantly sand, but under Cayton ridge an organic zone overlies the Talbot (Figure 3, D-D'). Two drill holes through the Arapahoe ridge in the same traverse penetrated a 1-

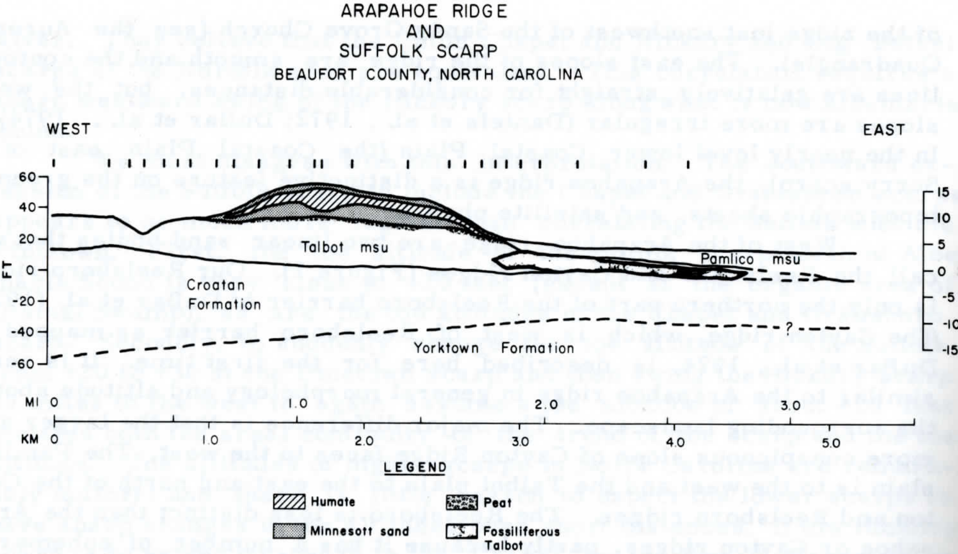


Figure 2. Cross section of Arapahoe ridge shows the relations of sediments of the Talbot and Pamlico morphostratigraphic units. See A-A' in Figure 1. The Suffolk scarp is the east slope of the ridge. The tick marks above the drawing are the locations of power auger holes.

foot bed of sandy clay loam over the micaceous sands of the Talbot.

Thirty-two drill holes penetrated 5 feet or more of ridge sand before reaching the underlying Talbot. Thirteen drill holes penetrated peat or buried A-1 horizons at the contact, 9 penetrated 5GY or 5Y colors indicating reducing conditions in the ridge sand and the underlying Talbot. Ten drill holes penetrated 10 YR colors, indicating oxidizing conditions, in both units. Nearly two-thirds of the Talbot surface buried under ridge sands either had organic deposits or poorly drained soils before the ridge sands were deposited, and about one-third of the soils were well to moderately-well drained. This is about the same ratio presently found on the Talbot surface between the Neuse and Pamlico rivers (unpublished statistical study, Soil Conservation Service, Raleigh, North Carolina).

Sands of the Arapahoe ridge are about 15 to 20 feet thick along the ridge axis (Figure 4). The somewhat uneven surface altitudes of the ridge are the result of the sands mantling a surface with 15 feet of relief, not large differences in ridge sand thickness. The base of the ridge sands (the top of the Talbot msu) ranges from +25 to +40 feet throughout the traverse except at the extreme south end near Minnesott Beach where the altitudes decrease to about +15 feet. The ridge sands bury the Talbot geomorphic surface (the constructional Talbot plain) in most areas, but near the south end of the ridge the altitudes of the

Table 2. Mineralogy of Arapahoe Ridge Sediments.

Very Fine Sand (0.50-0.10 mm)
Mineral[1]

| Stratigraphic Unit | Depth ft. | ZR | SP | TM | ST | KY | GN | CL | MS | HN | AU | EN | Fe | FD | QZ |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | Number % | | | | | | | | | | | | | |
| Arapahoe Ridge | 1- 2 | 7 | <1 | | | <1 | <1 | | <1 | <1 | | <1 | 42 | 14 | 35 |
| | 9-10 | 7 | <1 | | <1 | 1 | 1 | | <1 | 1 | 1 | <1 | 36 | 13 | 39 |
| | 19-20 | 10 | 1 | <1 | | | 1 | | | 2 | | <1 | 19 | 14 | 52 |
| Talbot msu[2] | 34-35 | 9 | <1 | 1 | | | <1 | <1 | 1 | 1 | 1 | 1 | 22 | 17 | 46 |
| | 35-42 | 4 | 1 | <1 | | | 1 | | <1 | 1 | 1 | <1 | 6 | 24 | 61 |

Clay Mineralogy
< 0.002 mm
Minerals[3]

| | | MT | VC | VM | MI | KK | QZ | FD | AS |
|---|---|---|---|---|---|---|---|---|---|
| Arapahoe Ridge | 1- 2 | | 5[4] | | | | 2 | | |
| | 9-10 | | | * | | | 4 | * | * |
| | 19-20 | 2 | 4 | | 2 | 2 | 2 | 2 | |
| Talbot msu | 34-35 | 1 | 3 | 2 | 4 | 2 | 2 | 2 | |
| | 35-42 | 2 | 2 | | 4 | 3 | | | |

[1]Mineral code: AU = Augite, Cl = Chlorite, EN = Enstatite, Fe = Iron (magnetite, ilmenite), FD = Feldspar, Gn - Garnet, HN = Hornblende, KY = Kyanite, MS = Muscovite, QZ = Quartz, SP = Sphene, ST = Staurolite, TM = Tourmaline, ZR = Zircon.

[2]msu = morphostratigraphic unit.

[3]Mineral code: As = amorphous, FD = feldspar, KK = kaolinite, MI = mica, MT = montmorillonite, VC = vermiculite chlorite, VM = vermiculite-mica, QZ = quartz.

[4]Approximate weight fractions: >5 = 1/2; 4 = 1/2 - 1/3; 3 = 1/3 - 1/5; 2 = 1/5 - 1/20; 1 = <1/20; * = quantity uncertain.

The procedures used are described by the Soil Survey Staff (1972). Feldspars were not separated by species, and the approximate weight fractions of the clay minerals are based on the x-ray peak heights.

buried surface are the same as those of the Pamlico plain east of the ridge.

On the south end of the Arapahoe ridge near Minnesott Beach, the sands of the ridge may bury part of the Pamlico sediments. But in the two traverses across the Arapahoe ridge (Figures 2 and 3) the ridge sands appear to slope down to and merge with the sediments of the Pamlico morphostratigraphic unit. If our interpretation of drill hole data is correct, the sands of the Arapahoe ridge south of Durham Creek (Figure 1) are contemporaneous with the sediments in the upper part of the Pamlico msu. This suggests that the buried surface at the south end of the ridge with an altitude of about +15 feet may be related to the erosion that preceeded deposition of the Pamlico msu (Figures 2 and 3).

The sands of Cayton ridge are 8 to 19 feet thick and those of the Reelsboro ridge are 15 to 20 feet thick. The altitudes at the base of the sands are +25 to +30 feet, the same as the adjacent level parts of the Talbot plain. The west slope of Cayton ridge grades downward to the Pamlico surface or Plain but sediments of the Talbot msu crop out on the lower part of this slope. The sands of the Reelsboro ridge have surface altitudes of less than +20 feet in small areas and appear to

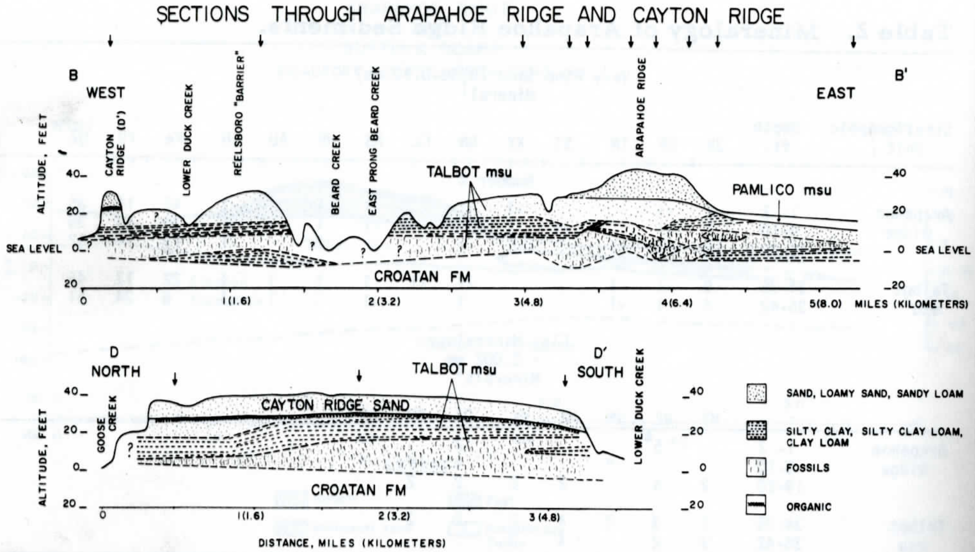


Figure 3. Relation of Cayton and Arapahoe ridges to the Talbot and Pamlico morphostratigraphic units. See B-B' in Figure 1. D-D' is a north-south section through Cayton ridge. The east slope of Arapahoe ridge in B-B' is the Suffolk scarp.

merge with the sediments of the Pamlico msu to the south.

The Suffolk scarp is a regional shoreline feature of the Pamlico sea that can be traced inland along the Neuse and several other major rivers. The toe altitude of the scarp in the area we have mapped (Figure 1) is about +20 feet. Between Durham Creek and the Neuse River, the east slope of the Arapahoe ridge is what we call the Suffolk scarp. The scarp is discontinuous in the Neuse valley because it is truncated in many places by the Neuse River and some of its tributaries. Where the scarp is present, a contour at +15 or +20 feet marks the zone of distinct flattening of slopes leading down from flats at +30 feet to flats at +15 feet or less.

Figure 5A shows the areal extent of the +20 foot contour lines in areas that we believe have been little modified by post-Pamlico erosion. This figure also helps establish the relation of the south end of the Arapahoe ridge to the Cayton and Reelsboro ridges. The toe of the Suffolk scarp and the +20 foot contour lines near Cayton and west of the Reelsboro ridge are nearly the same line. The +20 foot contour line in the headwaters of Goose Creek and similar streams normally would not be considered as part of the original shoreline because these are areas subject to considerable post-Pamlico entrenchment. However, the 20 foot contour line parallel to Beard Creek is the Suffolk scarp and apparently marks a narrow nearly continuous body of water during the high stand of the Pamlico sea (Figure 5B).

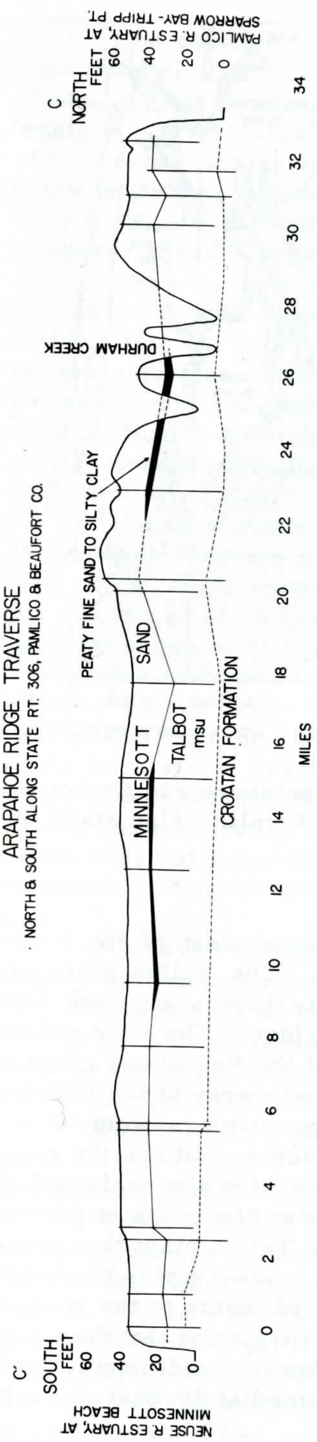


Figure 4. Thickness and distribution of the sediments along the longitudinal axis of Arapahoe ridge. See C-C' in Figure 1.

If sea level were at +20 feet today, the west side of Cayton ridge and most of the south side of the Reelsboro ridge would be at the edge of the Neuse estuary or sound. A small area along the southwest margin of the Reelsboro ridge would be awash (Figure 5B) and Goose and Beard Creeks would be flooded throughout much of their lower reaches. With sea level at +20 feet, the Talbot plain would be broken by salt or brackish water creeks and long reaches of salt marsh much like the Pamlico plain next to the Pamlico Sound is today.

## INTERPRETATION

The Arapahoe ridge is a distinct topographic feature, and the mapping of the ridge by DuBar et al., (1974), Mixon and Pilkey (1976), and ourselves between the Neuse and Pamlico Rivers is in close agreement. But we do not map the Arapahoe ridge south of the Neuse River (Figure 1) nor do we agree with DuBar et al., that the Arapahoe ridge is a barrier in the sense the term was used by Hoyt (1967), nor is it a secondary barrier as used by Colquhoun (1974). According to Hoyt's concept of barrier island formations and Colquhoun's concept of a primary barrier, sand is initially deposited on a mainland surface by a transgressing sea. Thus, peats or older mainland surfaces can be found under both storm beaches and barrier islands as they are under the Arapahoe ridge. Field and Duane (1976) state that it is erroneous to assume a genetic relation between the barrier and underlying sediments.

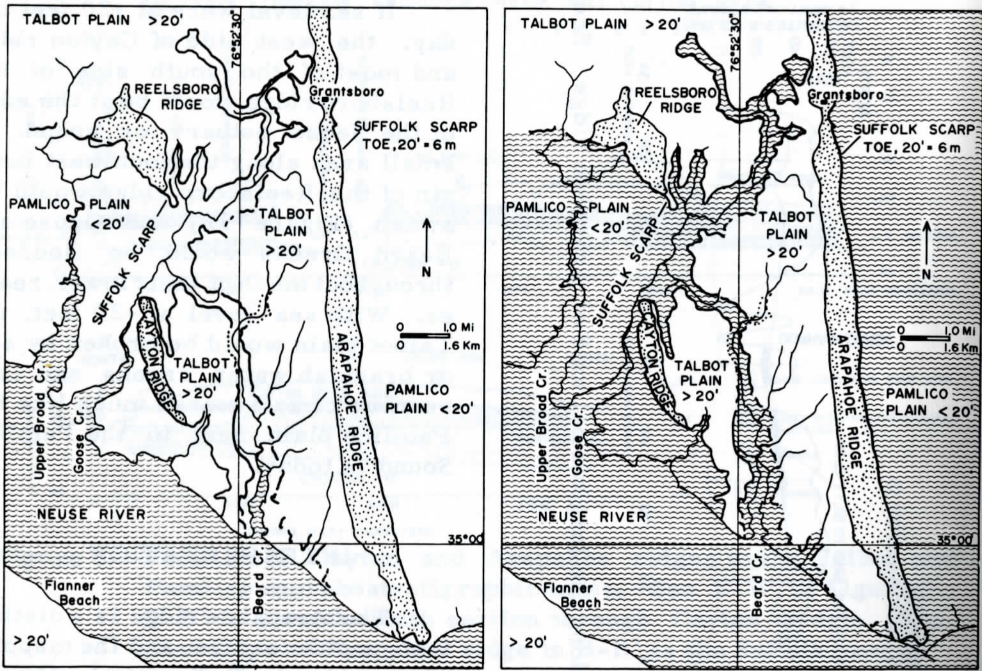


Figure 5. Left: Location of the Reelsboro, Cayton, and Arapahoe ridges, the Neuse estuary and the up-valley equivalent of the Suffolk scarp.
Right: Areal distribution of the Reelsboro and Cayton ridges and the Neuse estuary during the Pamlico high stand of sea level at +20 feet.

There is no evidence of a lagoon or sound west of the Arapahoe ridge that is contemporaneous with the ridge. The Talbot plain can be traced continuously from the toe of the Walterboro scarp, toe altitude of +45 feet, to and underneath the Arapahoe ridge. The silty sediments found west of the Arapahoe ridge and north of the Reelsboro ridge interlayer vertically and horizontally with other sediments of the Talbot msu. The peats that overlie the Talbot in some pocosins (swamp-on-a-hill) west of the Arapahoe ridge probably are Holocene, and are the result of surface-water retention on these flat surfaces. But the peats and other thin well- and poorly drained buried soils under the sands of the Arapahoe ridge are part of the weathering of the Talbot plain that predates deposition of the ridge sand. These lines of evidence plus the apparent merging of the ridge sand with the Pamlico sediments to the east (Figures 2 and 3) are strong support for the interpretation that the Arapahoe ridge sands are contemporaneous with the Pamlico sediments, the Core Creek sands of Mixon and Pilkey (1976), immediately east of Arapahoe ridge.

The relation of the Arapahoe ridge sands to the Pamlico sediments and plain indicates that the ridge was formed during a high stand of sea level at about +20 feet. The shape, sediments, thickness, and length of the Arapahoe ridge are similar to the Trail ridge of Georgia and Florida (Pirkle, 1972). The smooth ridge crest of the Arapahoe ridge is similar to that found on chenirs of the Gulf Coast (Byrne et al., 1959; Hoyt, 1969), and is not distinctly different in shape from modern washover areas along the present shoreline.

The evidence for the mechanisms or mechanism responsible for deposition of the Arapahoe ridge sands are inconclusive. The relatively uniform fine and medium sands, very poor preservation of bedding in the few exposures available for study, the large number of opaques, and the width and height of the ridge can be used as evidence for an eolian origin. But these features are not exclusively eolian, and the very limited dune topography argues against an eolian origin for the present surface of the ridge.

The dominantly smooth nearly level ridge crest (about 95 percent of the area) suggests a washover-swash origin. Features that do not deny nor strongly support this origin are the nearly horizontal bedding and abundance of opaque heavy minerals. The fine to medium sand size does not argue against this origin because these sand sizes are dominant in cheniers that are formed by similar processes (Hoyt, 1969). The absence of gravel in the ridge sands should be expected because gravel is rare in the Pamlico sediments east of the ridge. The strongest evidence for a storm beach origin for Arapahoe ridge is its smooth surface form, discordant relation to the underlying Talbot sediments, and apparent contemporanity with the Pamlico sediments to the east. This origin does not preclude significant additions of eolian material, but washover and swash processes had to dominate to form the final surface.

We have retained the term "Reelsboro ridge" for a small sand ridge near the hamlet of Reelsboro. Our Reelsboro ridge is only the very northern part of what DuBar et al., (1974) called their Reelsboro barrier. In the following discussion the term "Reelsboro barrier" will refer to the sandy area mapped by DuBar et al., (1974) south of what we call the Reelsboro ridge, but east of our Cayton ridge.

DuBar et al., (1974) mapped the Reelsboro "barrier" on the north side of the Neuse River N-NW from Whisk Point to the vicinity of Reelsboro. Whisk Point is the type location for the Neuse Formation of Fallaw and Wheeler (1969). The surface sediments at Whisk Point and north toward Reelsboro are sandy (the Arapahoe sand member of the Flanner Beach Formation, Mixon and Pilkey (1976), but we consider them to be sandy facies of the Talbot msu. Along traverse B-B' (Figure 3) from Cayton ridge, across the Reelsboro "barrier" of DuBar et al., and on east to the Arapahoe ridge, the sediments show little change. The vertical and horizontal changes in texture and fossil content are similar to the Talbot msu at Flanner Beach on the south side of the

Neuse (Daniels et al., 1972). The top of the Talbot sediments under the north part of Cayton ridge have the same altitude as most of the Reelsboro barrier of DuBar et al., (Figure 3, D-D' and B-B').

The area mapped as the Reelsboro "barrier" is dissected by south-flowing creeks. Although we lack enough bore holes to be positive, we suggest that this area is a remnant of the Talbot plain that has sandy sediments at or near the surface. This interpretation is strengthened by topographic maps that show this area merging with the flat Talbot plain to the north.

DuBar et al. (1974) implied that the low area between the Arapahoe and Reelsboro "barrier" with an altitude of near +25 feet is a back barrier flat associated with the Arapahoe ridge. Topographic maps show that this swale or low rises gradually toward the north until it has altitudes of +40 feet or more before decreasing to +30 feet farther north toward the Pamlico River. South-flowing Beard Creek occupies the axis of the low near the Neuse estuary. Beard Creek and its tributaries are incised into sediments of the Talbot msu (Figure 3, B-B'). Therefore we suggest that the topographic low is related to the incision of Beard Creek, not to the deposition of the Arapahoe ridge.

South of the Neuse River, DuBar et al., (1974) admitted that their Reelsboro "barrier" is a very indistinct feature. An inspection of topographic maps suggests the interfluve mapped as a barrier is an erosional remnant of the Talbot Plain, not a depositional feature. Several drill holes (Daniels et al., 1972) suggest that the sediments in this area are merely a sandy facies of the Talbot (probably similar to the Arapahoe sand member of Mixon and Pilkey, 1974) similar to that in parts of the Croatan National Forest to the southwest.

Based on the above lines of evidence, we must modify the concept of the Reelsboro "barrier" as presented by DuBar et al. We agree that at the community of Reelsboro there is a sand ridge that has an origin different from that of the surrounding sediments underlying the Talbot Plain. But the best interpretation for the remainder of the Reelsboro barrier is that post-Talbot, probably largely post-Pamlico, erosion by north- and south-flowing streams has left part of the Talbot plain as a N-S interfluve. The sandy sediments at or near the surface are part of the complex facies found in the marine Talbot msu of the area (Daniels et al., 1972; Mixon and Pilkey, 1976). The sandy character of these interfluves is enhanced by the formation of sandy A2 horizons in well-drained topographic positions (Daniels et al., 1967; Gamble et al., 1970).

We interpret the Reelsboro and Cayton ridges as features associated with sound or estuarine conditions during the Pamlico high stand of sea level. The subdued dune topography of these ridges and the apparent burial of part of the Suffolk scarp by the sand of the Reelsboro ridge (Figure 5A) suggests an eolian origin. However, interpretations of origin based largely on surface form, grain size, and relation to other sediments are subject to change or modification when sedimentary structures can be examined in detail. A compound origin for these ridges certainly cannot be excluded.

## SUMMARY

The Arapahoe ridge is believed to be a storm beach related to the high stand of the transgressing Pamlico sea. The ridge sand rests upon sediments of the Talbot msu that were exposed to subaerial weathering long enough to develop soils three feet thick. The storm beach origin at the peak of a transgressing sea for the ridge explains its height, smooth crest, length, continuity, abundance of opaque minerals, and relation to the sediments of the Talbot msu. White (1970) favors this explanation for Trail ridge and the Atlantic Coastal ridge of Florida as does Pirkle (1972) for Trail ridge. The Arapahoe ridge is confined to the area between the Pamlico and Neuse Rivers (Figure 1). North or south of these rivers the Suffolk scarp is a more subdued feature and does not have a sandy ridge similar to the Arapahoe (see Mixon and Pilkey, 1976), at its crest. The crest of the scarp commonly is sandy, but the sandy character in many areas is a distinct soil A2 horizon formed by the better drained conditions (Daniels et al., 1967; Gamble et al., 1970) at the Suffolk scarp in contrast to adjacent less well-drained Talbot and Pamlico Plains. We suggest an open ocean east of the Arapahoe is required to build this large and continuous beach ridge. Why the ridge is not found north of the Pamlico and south of the Neuse River is subject to speculation. We have found no recognizable Pamlico barriers east of the Arapahoe ridge that would protect the shoreline and it is risky to postulate the presence of barrier and have them destroyed by a regressing sea. However, the very smooth character of the Pamlico plain east of the Arapahoe ridge suggests either a very fast withdrawal of the Pamlico sea as suggested by Barry et al. (1975) so smaller beach ridges were not built (also see White, 1970, p. 89-90; Hoyt, 1967) or the barrier was east of the present limits of the Pamlico.

The Cayton and Reelsboro ridges are believed to be nearly contemporaneous estuarine equivalents of the Arapahoe. Their location on the northeast and east side of a long SSE reach of the Neuse estuary during the high stand of the Pamlico sea favor an eolian origin, but a compound origin cannot be excluded in the absence of detailed sedimentary bedding studies.

## LIST OF REFERENCES

Barry, R. G., Andrews, J. T., and Mahoffy, M. A., 1975, Continental ice sheets: condition for growth: Sci., v. 190, p. 979-981.

Byrne, J. V., LeRoy, D. O., and Riley, C. M., 1959, The chenir plain and its stratigraphy, southwestern Louisiana: Trans. Gulf Coast Assoc. Geol. Soc., v. 9, p. 237-260.

Colquhoun, D. J., 1974, Cyclic surficial stratigraphic units of the middle and lower Coastal Plain, central South Carolina, In Post-Miocene stratigraphy, central and southern Atlantic Coastal Plain, Oaks, R. Q., and DuBar, J. R., Eds., Utah State University Press, p. 179-190.

Daniels, R. B., Gamble, E. E., and Nelson, L. A., 1967, Relation between A2 horizon characteristics and drainage in some fine loamy Ultisols: Soil Sci., v. 104, p. 364-369.

Daniels, R. B., Gamble, E. E., Wheeler, W. H., and Holzhey, C. S., 1972, Some details of the surficial stratigraphy and geomorphology of the Coastal Plain between New Bern and Coats, North Carolina: Carolina Geol. Soc. and Atl. Coastal Plain Geol. Assoc. Field Trip Guidebook, 36 p.

DuBar, J. R., and Solliday, J. R., 1963, Stratigraphy of the Neogene deposits, lower Neuse Estuary, North Carolina: Southeastern Geol. v. 4, p. 213-233.

DuBar, J. R., Solliday, J. R., and Howard, J. F., 1974, Stratigraphy and morphology of Neogene deposits, Neuse River Estuary, North Carolina, In Post-Miocene stratigraphy central and southern Atlantic Coastal Plain, Oaks, R. W., and DuBar, J. R., Eds., Utah State University Press, p. 102-122.

Fallaw, W. and Wheeler, W. H., 1969, Marine fossiliferous Pleistocene deposits in southeastern North Carolina: Southeastern Geol., v. 10, p. 35-54.

Field, M. E., and Duane, D. B., 1976, Post-Pleistocene history of the United States inner continental shelf: Significance to origin of barrier islands: Geol. Soc. Amer. Bull., v. 87, p. 691-702.

Frye, J. C. and Willman, H. B., 1962, Morphostratigraphic units in Pleistocene stratigraphy: Am. Assoc. Petroleum Geologists Bull., v. 46, p. 112-113.

Gamble, E. E., Daniels, R. B., and McCracken, R. J., 1970, A2 horizons, Pedogenic or geologic: Southeastern Geol., v. 11, p. 137-152.

Hoyt, J. H., 1967, Barrier island formation: Geol. Soc. Amer. Bull., v. 78, p. 1125-1136.

__________, 1969, Cenier versus barrier, genetic and stratigraphic distinction: Am. Assoc. Pet. Geologists, v. 52, p. 299-306.

Mixon, R. B., and Pilkey, O. H., 1976, Reconnaissance geology of the submerged and emerged Coastal Plain province, Cape Lookout area, North Carolina: U. S. Geol. Survey Prof. Paper 859, 45 p.

Oaks, R. Q., Jr., Cah, N. K., Sanders, J. E. and Flint, R. F., 1974, Post-Miocene shorelines and sea levels, southeastern Virginia, In, Post-Miocene stratigraphy central and southern Atlantic Coastal Plain, Oaks, R. Q., and DuBar, J. R., Eds., Utah State Univ. Press, p. 53-87.

Pirkle, W. A., 1972, Trail ridge, a relic shoreline feature of Florida and Georgia: Unpublished Ph.D. thesis, University of North

Carolina, Chapel Hill, N. C., 85 p.
Richards, H. G., 1950, Geology of the Coastal Plain of North Carolina: Trans. Am. Phil. Soc., v. 40, 83 p.
Shattock, G. B., 1901, The Pleistocene problem of the north Atlantic Coastal Plain: John Hopkins Univ. Circ., v. 20, p. 67-75. (Reprinted in Am. Geologist, v. 28, 87-107).
Soil Survey Staff, 1972, Soil survey laboratory methods and procedures for collecting soil samples: USDA-SCS, U. S. Govt. Printing Office, 63 p.
Stephenson, L. W., 1912, The Quaternary formations, In, Clark, W. B., Miller, B. L., and Parker, H. N., The Coastal Plain of North Carolina: N. C. Geol. and Economic Survey, v. 3, 372 p.
Swanson, V. E. and Palacas, J. G., 1965, Humate in coastal sands of northwest Florida: U. S. Geol. Survey Bull. 1214 B., 29 p.
Thom, B. G., 1967a, Coastal and fluvial landforms - Horry and Marion Counties, South Carolina: LA State Univ. Coastal Studies Inst., Coastal Studies Series 19, Tech. Rept. 44, 75 p.
__________, 1967b, Humate and Coastal geomorphology: Coastal Studies Bull., La. State University, v. 1, p. 15-17.
Wentworth, C. K., 1930, Sand and gravel resources of the Coastal Plain of Virginia: Va. Geol. Survey, Bull. 32, 146 p.
Wheeler, W. H., Daniels, R. B., and Gamble, E. E., In press, The post-Yorktown stratigraphy and geomorphology of the Neuse-Pamlico area, Eastern North Carolina, In, The geology and paleontology of the Lee Creek Mine of North Carolina, C. Ray, Ed., Remington Kellogg Mem. Vol., US Nat. Museum.
White, W. A., 1970, The geomorphology of the Florida Peninsula: Florida Geol. Bull. No. 51, 164 p.
Whitehead, D. R., 1972, Developmental and environmental history of the Dismal Swamp: Ecol. Mongr., v. 42, p. 301-315.

# VARIATIONS IN SOUTH CAROLINA COASTAL MORPHOLOGY

By

P. Jeffrey Brown
University of South Carolina
Columbia, South Carolina 29208

## ABSTRACT

A classification of the South Carolina coast is proposed that is based on a statewide beach erosion study begun July 1, 1974. The coast is divided into four major geomorphological zones: arcuate strand, cuspate delta, transgressive barrier, and beach-ridge barrier, each of which has its own characteristic sediment type, bathymetry, and erosional-depositional history.

The arcuate strand, a relatively straight coast incised by few tidal inlets, extends approximately 100 km from the North Carolina border to Winyah Bay. Although presently the most stable portion of South Carolina's coast, the area is backed by mid-Pleistocene beach-ridge deposits. These deposits are apparently one major present source of beach sediment for the area. The continental shelf off the arcuate strand is very steep nearshore, then it levels off to a flat featureless surface seaward of the 8 m depth contour. The cuspate delta area, which lies between Winyah and Bull Bays, is composed principally of sediments supplied by the Santee River system. A decrease in the sediment supply owing to damming and diversion of the river brought about extensive shoreline erosion. Beaches in the area are steep and narrow and are composed of relatively coarse, texturally immature sediments. The continental shelf off the cuspate delta is gently sloping, but it is very irregular due to the presence of cape-associated shoals.

The southern 160 km of South Carolina's coast is made up of barrier island and tidal inlet systems. A distinction is made between barrier islands that have beach ridges (beach-ridge barriers) and those that do not (transgressive barriers). Beach-ridge barrier islands are the predominant type of barrier in South Carolina. They normally have a bulbous updrift (northern) end a straight to crescentic central portion, and a downdrift portion made up of recurved spits. The morphology of the islands, as well as their erosional-depositional nature, is controlled by proximity to tidal inlets and the stability of the inlets and their associated shoals. Beach-ridge barrier sediments are better sorted and finer grained that those of the cuspate foreland and arcuate strand. Bathymetry offshore of the barriers is controlled by the sand trapping mechanisms of tidal inlets.

Transgressive barrier islands are the erosional end-product of beach-ridge barriers; that is, they are islands whose beach ridges have eroded away. These areas are made up of a thin packet of sand that is

rapidly retreating over back-barrier salt marshes as a series of washovers and washover terraces. Transgressive barrier sediments are also relatively fine grained and well sorted.

## INTRODUCTION

A statewide beach erosion study, sponsored by the South Carolina Sea Grant program of N. O. A. A., was begun in July, 1974, to delineate present areas of critical erosion and to gain insights concerning future shoreline changes for land management purposes. The data and ideas which follow represent preliminary results of the first year of a three-year study.

The coast of South Carolina lies between 32° and 34° north latitude and extends approximately 255 km (Figure 1). Major portions of the coast (predominantly barrier islands in the southern and central portions) remain undeveloped. Large salt marshes supply nutrients to the sea, and estuaries serve as natural harbors and recreation centers.

### Climate

South Carolina's climate is mild, with an average temperature for the coastal region of 18.7°C, ranging between 10.0°C (December) and 27.2°C (July). The Atlantic Coastal Plain, which makes up 40 percent of the state, receives an average of 118.4 cm of precipitation annually (Landers, 1970). An average of 1.4 hurricanes and tropical storms affect South Carolina's coast annually (Crutcher and Quayle, 1974), although none occurred during the study period.

On an annual scale, no predominant direction of wind approach is apparent along South Carolina's coast (Figure 2). However, seasonal trends can be seen. During the spring and summer, winds from the south and southwest prevail, whereas in the fall and winter, most winds come from northerly directions. Summary of Synoptic Meteorological Observations (U. S. Naval Weather Service Command, 1970), hereafter referred to as SSMO data, were used to derive wind frequency distributions for the Charleston data square. Wave energy flux values computed from these SSMO data show the same seasonal trends as the wind (Figure 3). The wave energy flux for a given sea station is defined as total wave energy per unit area of the sea surface times the velocity of propagation of this energy. For a more complete discussion of wave energy flux, see Nummedal and Stephen (1976). The average annual wave energy flux values for those directions affecting the coast of South Carolina (NE to SW) are approximately $1.7 \times 10^6$ g-m/s$^3$, with a maximum of $3.1 \times 10^6$ g-m/s$^3$ from the NE. This compares with average annual values of $1.1 \times 10^6$ g-m/s$^3$ and $2.1 \times 10^6$ g-m/s$^3$ for Pensacola, Florida and Atlantic City, New Jersey, respectively.

Numerous rivers drain the Coastal Plain of South Carolina, but

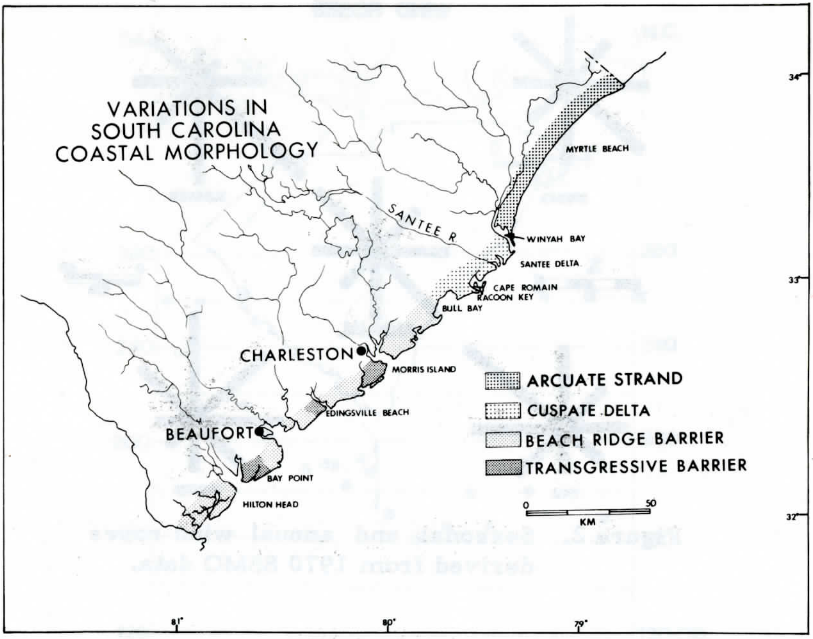


Figure 1. Location map.

the Santee River (70m$^3$/sec) (which actually originates in the Piedmont) and the Edisto River (47.6 m$^3$/sec) are presently the only rivers with any appreciable discharge (South Carolina Water Resources Commission, 1975). Coastal Plain rivers currently empty into the Atlantic only in the central and southern portions of the coast; no river outlets are found north of Winyah Bay.

## Tidal Regime

The morphology of the coast of South Carolina is, in essence, a transition zone between that of North Carolina and Georgia. North Carolina's coast is predominantly made up of long, thin barrier islands broken by few tidal inlets. Characteristic of the microtidal classification of Hayes et al. (1973), the morphology of the North Carolina coast is controlled mainly by wind-generated waves and currents. Georgia's coast to the south is comprised of short, stubby barriers separated by many large tidal inlets; it is a tidal-current controlled coast. South Carolina has a mixed (wind and tidal) energy coast with tidal influence increasing to the south.

Tidal range increases considerably from north to south, from approximately 170 cm to 270 cm (U. S. Dept. of Commerce, 1974) (Figure 4). Increasing tidal range has several effects: 1) tidal inlets become more frequent and are larger to accommodate greater tidal flow;

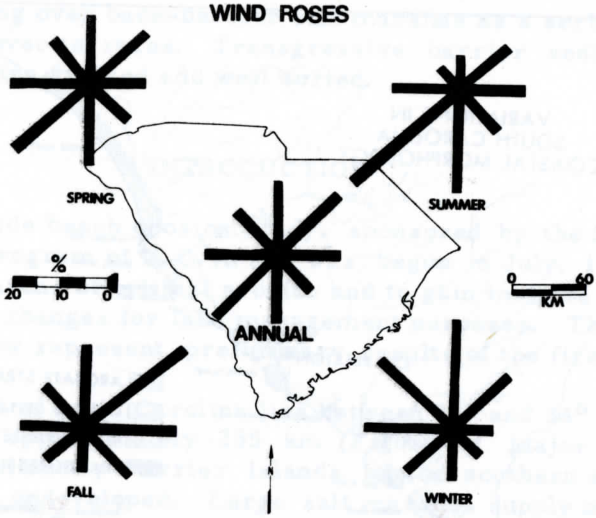


Figure 2. Seasonal and annual wind roses derived from 1970 SSMO data.

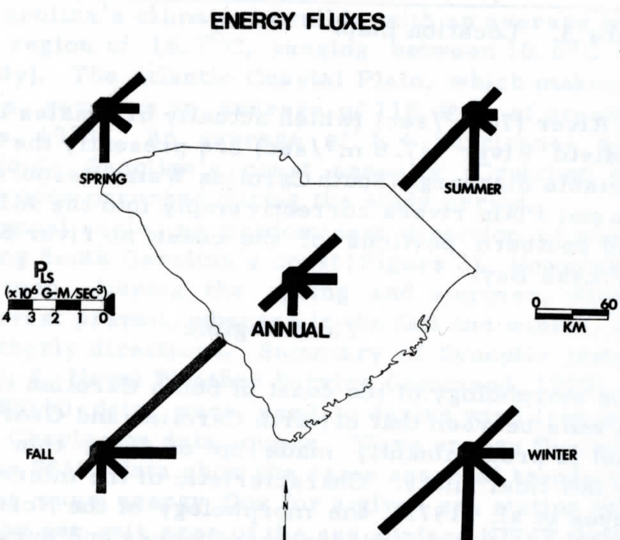


Figure 3. Seasonal and annual wave energy flux values computed for alongshore and onshore winds from the Charleston (S. C.) SSMO data square. The length of any bar is a relative measure of the wave energy coming from that direction.

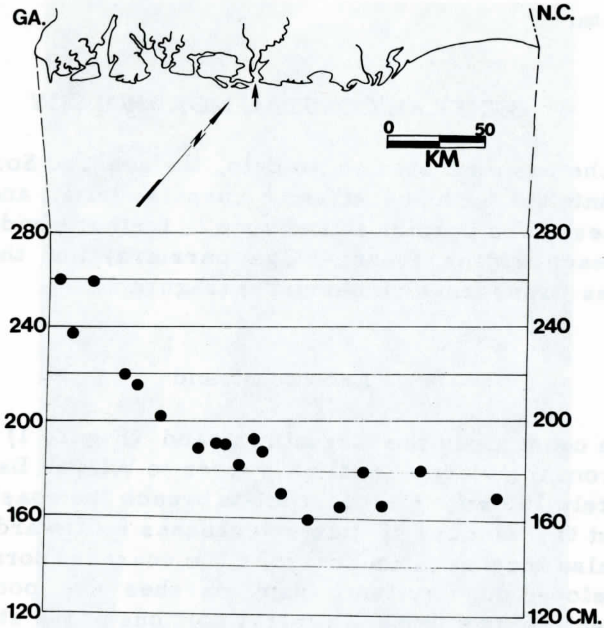


Figure 4. Spring tidal range variation along the shoreline of South Carolina. Note increase in southerly direction. Based on Tide Tables (U. S. Dept. of Commerce, 1974-1975).

2) salt marshes are more extensive; and 3) the ebb-tidal deltas are much larger off the inlets and estuaries in the southern portions of the state.

Acknowledgments

The majority of the data presented in this paper was collected while working on a statewide beach erosion inventory sponsored by the National Sea Grant office of the National Oceanographic and Atmospheric Administration, grant #R-CZ-1, Miles O. Hayes, Principal Investigator. Michael F. Stephen and Duncan M. FitzGerald were of great assistance. Mr. Stephen supplied a great deal of his time reviewing this paper and furnished several of the photographs used.

The vertical aerial photographs used in the short term historical study were supplied by the Department of Agriculture, whose cooperation was greatly appreciated. The sediment size analysis was conducted by Dr. Frank Stapor of the South Carolina Wildlife and Marine Resources Laboratory, Charleston. Frank Lesesne and Mike Johnson aided in

data collection, and Mr. Lesesne was of great assistance in the organization of data.

## GENERAL COASTAL MORPHOLOGY

On the basis of studies to date, the coast of South Carolina can be divided into the arcuate strand, cuspate delta, and barrier island coastal zones. The barrier island zone is further divided into islands that have beach ridges (beach-ridge barriers) and those that have no beach ridges (transgressive barriers) (Figure 1).

### Arcuate Strand

The coast along the arcuate strand (Figure 1) forms a gentle crescent from the North Carolina border to Winyah Bay, a distance of approximately 100 km. Few tidal inlets breach the coast in the northern section, but the number of inlets increases southward to Winyah Bay. Inlet size also increases southward. The coast is normally backed by a well-developed dune system. Salt marshes are poorly developed or totally absent in the north and central portions of the strand, becoming more prominent in the southern section (Figure 5).

The arcuate configuration and poor salt marsh development are the result of the developmental history of the arcuate strand. Although the area is currently the most stable portion of South Carolina's coast, it has historically undergone extensive erosion. The shoreline is presently backed by barrier sands of the Myrtle Beach Formation which was formed before the Wisconsin glacial period, approximately 100,000 years B. P. (Thom, 1970). The present shoreline configuration generally parallels the orientation of these Pleistocene beach ridges.

### Cuspate Delta

The Santee River delta, which extends 30 km along the South Carolina coast, is the largest deltaic complex on the east coast of the U. S. It is classified as a cuspate delta according to the scheme of Price (1955). The shoreline components of the delta include: 1) a classic cuspate foreland, Cape Romain; 2) an eroding beach-barrier complex, Raccoon Key; and 3) distributary mouth sand bars and mud flats. The lower deltaic plain is covered by an extensive salt marsh.

The presence of washover terraces and truncated beach ridges all along the shoreline of the delta is evidence of its rapid retreat. Studies by Aburawi (1972) and others have shown that decreased sediment supply, owing to extensive damming of the river and the diversion of a major portion of the river's flow into the Cooper River in 1942, has had a pronouned effect upon the Santee delta complex. Prior to the 1930's, the delta was in a stable or constructional phase. After that

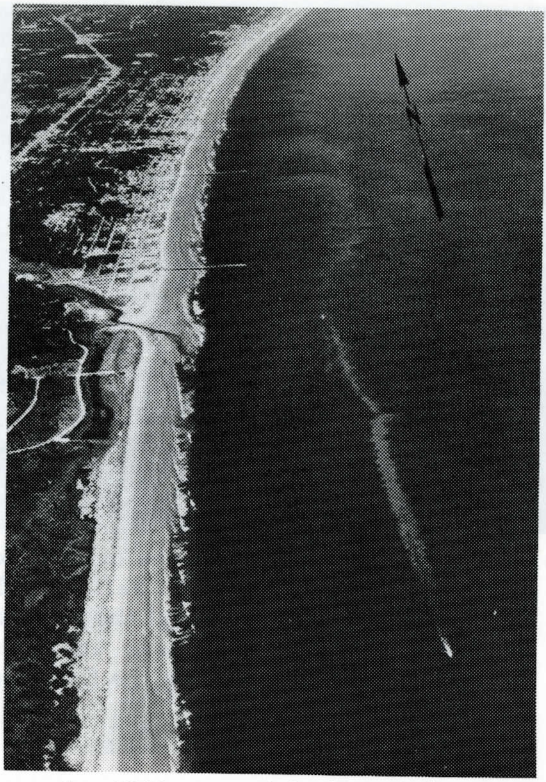

Figure 5. Oblique aerial view of the arcuate strand near Garden City. The coast is straight and interrupted by few tidal inlets. Note dune development and the absence of salt marshes. Photo taken October, 1974.

time, the delta entered a destructive phase which has continued to the present. Sequential vertical aerial photographs reveal that the area has been severely eroded since 1941, with erosion of over 215 m occurring at several points on Cape Romain. Raccoon Key, to the south of the cape, has eroded up to 275 m at some localities (Stephen et al., 1975). Although plans are now underway to redivert a major portion of the fresh water discharge back into the Santee River, it is expected that the area will continue to erode at a rapid rate owing to upstream dams impeding the oceanward transport of sand-sized material.

The formation of cuspate forelands, such as Cape Romain, has been the subject of much discussion (Gulliver, 1896; Johnson, 1919; White, 1966; Shepard, 1973; Rosen, 1975). South Carolina's cuspate foreland owes its origin in part to the sediments supplied by the Santee

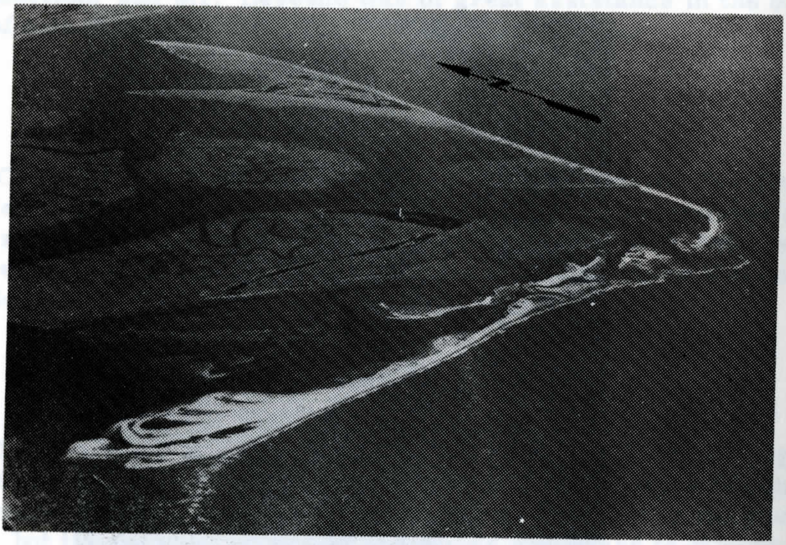

Figure 6. Oblique aerial view of the cuspate foreland, Cape Romain. The headland is being eroded while the flanks are elongating through the formation of recurved spits. Note the extensive salt marsh behind the cape, and the washovers on the flanks near the cape. Photo taken October, 1974.

River. These sediments have been eroded and shaped into the present form of the cape (Figure 6) by waves generated by the predominant northeasterly winds that blow in the winter and the prevailing southwesterly winds that blow in the summer. Erosion of the cape headland has caused its northern flank to change orientation dramatically during the past century, shifting from N-S in 1886 to its present NNE-SSW orientation. Since 1886, the northern arm of the cape has elongated approximately 1.8 km, while the westward arm grew a length of 3.7 km (Figure 6) (Shepard and Wanless, 1971; Stephen et al., 1975).

## Barrier Islands

Between Bull Bay and the Georgia border, a distance of approximately 160 km, a series of barrier islands front the coast (Figure 7). South Carolina's barrier islands average about 7 km in length and are separated from the shoreline by a zone of salt marsh that generally increases in width toward the south. Numerous tidal inlets separate the islands.

Transgressive Barriers. The areas classified as transgressive barrier islands are, from north to south, Morris Island, Edingsville Beach, and the Bay Point area (Figure 1). Transgressive barrier

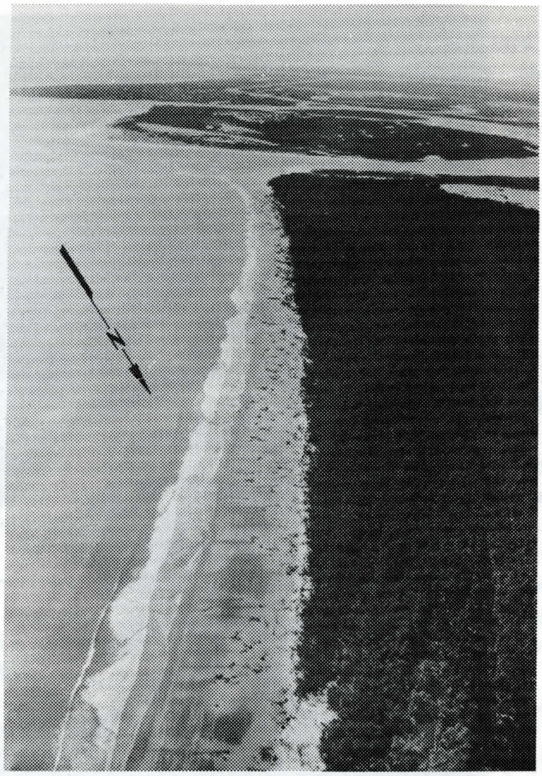

Figure 7. Oblique southerly aerial view of a series of beach-ridge barrier islands and associated tidal inlets. Capers Island, in the foreground, is presently undergoing extensive erosion, as is evidenced by uprooted trees in the surf zone. Note the downdrift (southern) offset configurations and the salt marsh backing the islands. Photo taken in February, 1975.

shorelines are characteristically straight and are made up of a thin layer of sand that is retreating landward as a succession of washovers. Dunes and beach ridges are normally absent with washover terraces backing most beaches (Figure 8).

Studies of sequential charts and aerial photographs show that transgressive barriers are simply the erosional end-product of beach-ridge barriers. Coastal charts of Morris Island and Edingsville Beach dating back to 1779 show that these areas originally possessed beach ridges which have since eroded away.

Lacking the local relief and storm protection provided by dunes

Figure 8. Oblique aerial view of a transgressive barrier, Edingsville Beach. Note that the beaches are straight and no beach ridges are present. A continuous washover terrace complex backs the beach and marsh peat is exposed on the beach face, giving evidence of continuing erosion. Photo taken October, 1974.

and beach ridges, transgressive barriers are presently eroding at an extremely rapid rate. Average erosion rates of up to 15 m/yr have been documented at Morris Island, while Edingsville Beach is eroding at approximately 3 m/yr (Stephen et al., 1975).

Beach-Ridge Barriers. Beach-ridge barriers comprise the majority of the central and southern portions of South Caroina's coast. The morphology of beach-ridge barriers has been discussed by Hayes et al. (1974). They are characterized by a bulbous updrift (northern) end, a straight to crescentic central portion, and a downdrift end which elongates and progrades through the formation of recurved spits.

Beach-ridge barrier morphology is greatly affected by the presence of tidal inlets. Wave refraction and storm protection provided by the ebb-tidal delta cause accretion on the updrift end of the barrier. Slight changes in inlet configuration and position can cause shoreline reorientations up to 6.4 km from the inlet (Stephen et al., 1975). The central portions of the smaller beach-ridge barriers are subject to re-orientations, while the central sections of the larger barriers are relatively stable. The southern (downdrift) ends are generally accretional.

## STATEWIDE BATHYMETRIC TRENDS

In an attempt to delineate statewide trends in nearshore bathymetry, seventy offshore profiles were constructed to a depth of 15 m (Figure 9). The 1970 version of the Coast and Geodetic Survey sheets (1:80,000 scale) were used.

From the North Carolina border to the area just north of Winyah Bay (arcuate strand zone), the sea bottom immediately offshore is characteristically concave upward, very steep near the shore and leveling off at a depth of approximately 8 m to a flat featureless surface. Based upon 17 profiles in the area, the average slope is 7.4 m/km for the first 0.8 km offshore and 1.3 m/km for the second 0.8 km. The bottom then levels off to a very gently inclined surface with an average slope of 0.74 m/km to a depth of 15 m.

From the entrance to Winyah Bay south along the rest of the coast, there is a dramatic change in nearshore topography; the slope directly offshore is much gentler than it is to the north (Figure 9). The reason for this is twofold: 1) in the vicinity of Winyah Bay and south to the Cape Romain area, the Santee River has deposited a tremendous amount of sediment on the inner continental shelf; and 2) further to the south, where there is presently little fluvial sediment input, the nearshore bathymetry is controlled by the sand trapping mechanism of tidal inlets (Hayes et al., 1974).

In the cuspate delta zone, the average slope is only 2.0 m/km for the first 0.8 km and 1.8 m/km from 0.8 to 1.6 km offshore, with an average slope of 0.8 m/km out to a depth of 15 m (based upon 14 profiles). The offshore bathymetry is much more irregular in this zone (especially near Cape Romain) owing to a succession of cape-associated shoals (Figure 9).

South of Bull Bay, where the coast is comprised primarily of barrier island-tidal inlet systems, the offshore topography is highly variable. Profiles run in the vicinity of the inlets are characteristically flat and irregular to a depth of approximately 9 m where a sharp drop-off occurs. This is due to the accumulation of sand in the ebb-tidal delta platform. Profiles off the central portions of the barriers are typically flat and featureless with a fairly uniform slope (Figure 9). Overall, the slope of the shelf off the barriers averages 1.6 m/km for the first 0.8 km, 2.2 m/km for the second 0.8 km, and 0.85 m/km to a depth of 15 m. No significant difference could be seen between the topography

## VARIATIONS IN BATHYMETRY

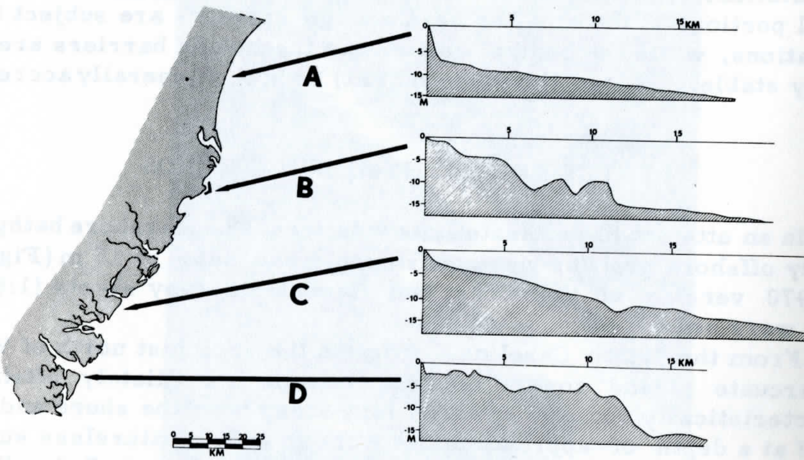


Figure 9. Statewide trends in bathymetry from the shoreline to a depth of 15 meters. Vertical exaggeration 330:1. Profile A is off the arcuate strand, B is off Cape Romain (cuspate delta). Profile C is off the central portion of a barrier island (Edisto Island) while D is in the vicinity of a large tidal inlet and its associated ebb-tidal delta (Port Royal Sound).

offshore of islands where beach ridges are present and those where beach ridges are absent.

In summary, there is a significant change in bathymetry along South Carolina's coast. North of Winyah Bay, typical profiles are concave and very steep in the nearshore. South of Winyah Bay, the influx of fluvial sediments and the storage of sand in ebb-tidal delta complexes causes a decrease in nearshore slope, although the overall slope to a depth of 15 m is very uniform along the entire coast (0. 74 - 0. 85 m/km). The area off Cape Romain has the most irregular bathymetry due to the presence of cape-associated shoals. Bathymetry in the barrier island portion of the coast is governed by proximity to tidal inlets. No appreciable difference is apparent in the bathymetry between beach-ridge and transgressive barriers.

## STATEWIDE SEDIMENT CHARACTERISTICS

In an attempt to delineate general statewide trends in beach sand size and sorting, a total of 97 samples representing all beach environments were analyzed by sieving, and statistical parameters were computed using the methods of Folk (1968). The carbonate fraction was not removed.

## SAND SIZE AND SORTING VARIATION

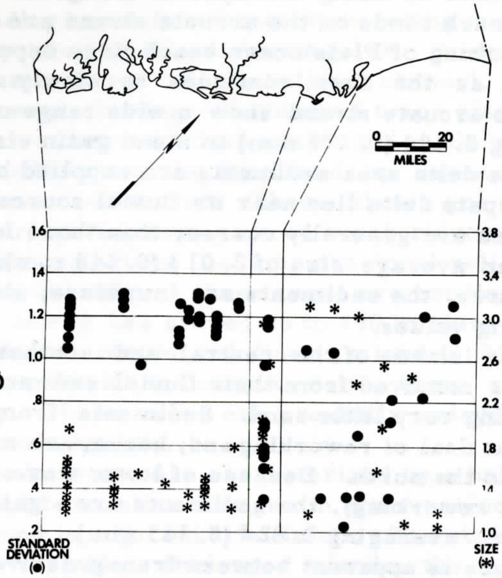


Figure 10. Statewide trends in beach sediment size and sorting based on statistical parameters of Folk (1968). Arcuate strand sediments, apparently derived from reworked Pleistocene beach ridge deposits, show a wide range of size and sorting values. Cuspate delta sediments, which are near their fluvial source, are also immature. Barrier island sediments are better sorted and finer grained.

It was found that beach sediments tend to coarsen seaward along the S. C. shore. The finest-grained sediments were found in the dunes and berm, with average values of 2.81Φ (0.144 mm) and 2.64Φ (0.158 mm) respectively. The coarsest sediments were found in the lower beach face, with an average size of 1.58Φ (0.331 mm). Upper beach face samples average 2.37Φ (0.185 mm). The only exceptions to the seaward coarsening trend were the samples from washovers, which had an average size of 1.86Φ (0.265 mm).

When sediment size and sorting parameters are plotted on a statewide basis, three distinct sediment populations can be seen, corresponding to the arcuate strand, cuspate delta and barrier island zones (Figure 10). It is felt that the sediment types reflect three different populations of beach material. There is presently no direct source of fluvial sediments on the arcuate strand, although several rivers emptied into the ocean in this area during the Pleistocene (Shepard and Wanless, 1971). While some sediment is being transported alongshore and onshore into the area, most beach sands on the arcuate strand are apparently derived from the reworking of Pleistocene beach ridge deposits lying directly behind the shore, as the area continues to slowly recede. Sediment samples from the arcuate strand show a wide range of size and sorting values, averaging 2.48$\phi$ (0.175 mm) in mean grain size.

The cuspate delta area sediments are supplied by the Santee River, and since the cuspate delta lies near its fluvial source, sediment samples from the area are generally coarser than those found elsewhere on the coast, with an average size of 2.01$\phi$ (0.248 mm). Owing to proximity to their source, the sediments are immature, showing a wide range of size and sorting values.

The barrier islands of the central and southern portions of the coast are farther removed from their fluvial sediment sources and are presently receiving very little sand. Sediments from this area have undergone a great deal of reworking and, hence, are much better sorted than sediments to the north. Because of lower wave energy in the area (and the constant reworking), the sediments are significantly finer than those to the north, averaging 2.82$\Phi$ (0.143 mm).

No difference is apparent between transgressive and beach-ridge barrier sediments. The limited number of samples which were analyzed might have failed to reveal any such differences if they exist.

## SUMMARY

Using overall geomorphology as a basis, the coast of South Carolina has been classified into four units: arcuate strand, cuspate delta, transgressive barrier, and beach-ridge barrier. Each morphological unit has its own characteristic offshore bathymetry, erosional-depositional history, and sediment characteristics.

The arcuate strand is characterized by a straight coast which is cut by few, small tidal inlets. Although fairly stable at present, it has undergone extensive erosion since the late Pleistocene and is now backed by mid-Pleistocene beach ridges, approximately 100,000 years old. Sediments in the area are apparently derived from reworked beach ridge deposits, and show a wide range of size and sorting values, with an average size of 2.48$\phi$ (0.175 mm). The surface offshore of the arcuate strand is concave upward, with a steep nearshore area gradually leveling off to a surface with little local relief. Because this area is being increasingly developed by mankind, erosion problems in this area are expected to increase in the future.

The cuspate delta area is characterized by a complex of eroding features, including an eroding cuspate foreland, Cape Romain. Erosion exceeding 275 m since 1941 has been documented. Beaches are steep and narrow and are composed of coarse (2.01$\Phi$; 0.248 mm), immature, fluvially-derived sediments whose size and sorting characteristics are highly variable. A recent decrease in sediment supply to the area has changed a prograding coast into a rapidly receding one. Bathymetry off the delta is gently sloping but very irregular. Despite proposed plans to redivert the Santee River into its original channel, the area will probably continue to erode because of sediment traps (dams) upstream.

Transgressive barrier island coasts are made up of a thin veneer of sand that is rapidly retreating landward by successive washovers, burying back-barrier salt marsh. These islands are the erosional end-product of beach-ridge barriers, whose beach ridges have eroded away during the past few centuries. Lacking the local relief and storm protection afforded by the ridges, these areas are eroding at rates of up to 15 m/yr; Morris Island has eroded up to 490 m since 1939. Sediments in this area are better sorted and finer grained (2.8$\Phi$; 0.143 mm) than those of the arcuate strand and cuspate delta, being subject to a great deal of reworking. These areas are presently undeveloped by man and will continue to erode rapidly.

Beach-ridge barriers are characterized by a bulbous updrift end that undergoes rapid shoreline changes in short periods of time, in response to fluctuations in adjacent tidal inlets. The straight to crescentic central portion is variable. On larger islands, the majority of the central portion is beyond the range of tidal inlet influence, and is an area of stability or accretion. Smaller islands, where tidal inlet influence affects the whole island, are unstable along their entire length. Tidal inlet effects extending over 4 km are documented. Beach-ridge barrier sediments are fine-grained (2.8$\Phi$; 0.143 mm) and well sorted. Offshore bathymetry is gentle, due to the trapping of sand on ebb-tidal deltas.

## REFERENCES CITED


Aburawi, R., 1972, Sediment facies of the Holocene Santee River delta: Unpb. M. Sc. Thesis, University of South Carolina.

Crutcher, H. D., and Quayle, R. G., 1974, Mariners worldwide climate guide to tropical storms at sea: Naval Weather Service Command, U. S. Dept. of Commerce, Washington, D. C.

Folk, R. L., 1968, Petrology of sedimentary rocks: Hemphill Publ. Co., Austin, Texas, 170 p.

Gulliver, F. P., 1896, Cuspate forelands: Geol. Soc. Am. Bull., v. 7, p. 399-422.

Hayes, M. O., Owens, E. H., Hubbard, D. K., and Abele, R. L., 1973, The investigation of form and processes in the coastal zone: in Coates, D. R., ed., Coastal Geomorphology: SUNY, Binghampton, N. Y., p. 11-42.

Hayes, M. O., Hulmes, L. J., and Wilson, S. J., 1974, Importance of tidal deltas in erosional and depositional history of barrier islands: (abs.) in Abstracts with Programs, 1974 Annual Meeting

Geol. Soc. of Am., Miami, Florida.

Johnson, D. W., 1919, Shore processes and shoreline development: John Wiley Co., New York, N. Y., 584 p.

Landers, H., 1970, Climate of South Carolina: in Climates of the states, South Carolina, Climatography of the United States, No. 6038, ESSA Environmental Data Service.

Nummedal, D., and Stephen, M. F., 1976, Coastal dynamics and sediment transportation, northeast Gulf of Alaska: Tech. Dept. No. 9-CRD, Coastal Research Division, Geology Dept., University of South Carolina, 148 p.

Price, W. Armstrong, 1955, Correlation of shoreline type with offshore bottom conditions: Dept. of Oceanography, A & M College of Texas, Project 53.

Rosen, P. S., 1975, Origin and processes of cupate spit shorelines: in Cronin, L. E., ed., Estuarine Research 2, Academic Press, New York, p. 77-92.

Shepard, F. P., 1973, Submarine geology (Third Edition): Harper and Row, New York, N. Y., 517 p.

Shepard, F. P., and Wanless, H. R., 1971, Our changing coastlines: McGraw-Hill, New York, N. Y., 579 p.

South Carolina Water Resources Commission, 1975, personal communication.

Stephen, M. F., Brown, P. J., FitzGerald, D. M., Hubbard, D. K., and Hayes, M. O., 1975, Beach erosion inventory of Charleston County, South Carolina: a preliminary report: S. C. Sea Grant Technical Report No. 4, 79 p.

Thom, B. G., 1970, Carolina bays in Horry and Marion Counties, South Carolina: Geol. Soc. Amer. Bull., v. 81, no. 3, p. 783-813.

U. S. Dept. of Commerce, 1974, Tide Tables--high and low predictors, east coast of North and South America: N. O. A. A., National Ocean Survey, Rockville, Md.

U. S. Naval Weather Service Command, 1970, Summary of synoptic meteorological observations, North American coastal marine areas: National Climatic Center, Asheville, North Carolina.

White, W. A., 1966, Drainage asymmetry and the Carolina capes: Geol. Soc. Am. Bull., v. 77, p. 223-240.

# A PRELIMINARY STUDY OF THE SEDIMENT HYDROLOGY OF A GEORGIA SALT MARSH USING RHODAMINE WT AS A TRACER*

By

John Nestler
Institute of Ecology
University of Georgia
Athens, Georgia 30602

## ABSTRACT

A water tracer experiment using Rhodamine WT was performed on a Sapelo Island, Georgia, salt marsh to determine the nature of water movement through the sediments. Dye detection was infrequent and random in "dye wells" 25, 50 and 100 cm from the point in the sediments in which the dye was introduced. Capillary water movement in the upper 0-20 cm, caused by extensive evaporation on the marsh surface, was implicated as the primary type of water movement. This type of vertical water movement is probably affected by heavy rainfall, extreme dissication, heavy wave action or any other factor which destroys or alters the sediment microstructure that controls capillary water movement.

## INTRODUCTION

The need for estimates of water flow through pore spaces of salt marsh intertidal sediments became apparent during the course of an experiment designed to elucidate interstitial salinity phenomena on a Georgia salt marsh (Nestler, 1977). Also, the amount of water movement should be highly correlated with the exchange rate of nutrients bound in the marsh sediments as well as determining pore water chemistry. Literature review provided little information on this topic. Gardner (1973) suggested that tidal flushing in the natural levees bordering tidal creeks could be a significant factor. Therefore, the initial purpose of this study was to obtain a preliminary description of the rates and direction of interstitial water movement.

*Contribution No. 317 from the University of Georgia Marine Institute, Sapelo Island, Georgia.

Acknowledgments

Sincere gratitude is expressed to L. R. Pomeroy for directing the field work in this study and also to the personnel of the University of Georgia Marine Institute for their valuable assistance. This work was supported by a NSF grant, #DES72-01605-A02, under the direction of Dirk Frankenberg and L. R. Pomeroy.

## MATERIAL AND METHODS

This study was conducted on a Spartina alterniflora dominated salt marsh near Sapelo Island, Georgia. Marsh sediments are primarily silts and clays with some admixed sand and occasional sand lamellae (Frey and Howard, 1969).

Rhodamine WT, a fluorescent dye that can be detected at concentrations less than one part per billion, was used as a water tracer because it does not adsorb to clays as readily as the other Rhodamines (Turner Fluorometry Reviews, 1971).

Six dye movement stations were located along Teal's Boardwalk near the Marine Institute Laboratory. Each station consisted of nine "dye wells" (Figure 1) located on the marsh surface in the shape of a cross. Three stations on the more elevated high marsh, characterized by stands of dwarf Spartina, had wells 1.6 m long, inserted to a depth of 30 cm and spaced 25 cm apart. The other three stations on creek levees, dominated by stands of tall Spartina, had wells 3 m long, inserted to a depth of one m and spaced 50 cm apart. These latter wells were located further apart and deeper because greater water movement was expected on the steeply sloping levees.

The dye wells were constructed of 1.25 cm diameter polyvinyl chloride pipe. Six holes (0.64 cm diameter each) were drilled into the end of the well to be inserted into the marsh sediments. These six holes plus the open bottom of the well provided about 10 $cm^2$ for dye seepage. A solid rubber stopper was placed in the bottom of the well before it was pushed into the marsh. After insertion a ramrod was used to tap out the rubber stopper. A one-hole stopper with a five cm long U-shaped tube in the single hole was placed on top of the well to prevent pressure build-up, rain contamination and perching and fouling by birds. Water samples were withdrawn from the wells with a syringe having a tygon tubing extension.

Blank determinations were made over a one week period before the beginning of the experiment. The blank was 1 ug/l; therefore, a sample value had to exceed 2 ug/l to be considered significant. Initially, eight ml of the dye was deposited into the center well. If this amount of dye were diluted into a volume of interstitial water within one meter of the center well, the resulting dye concentration would be about 100 ug/l or about 50 times the value of the blank. Thereafter, at each

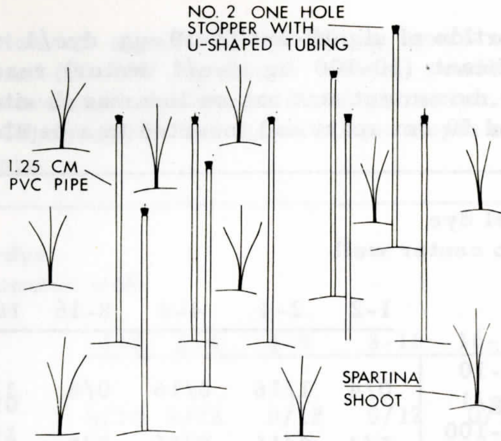


Figure 1. Schematic representation of a high marsh (short Spartina) water movement station.

sampling time the center well was checked to substantiate the presence of the dye. The dye sample was then replaced to prevent lessening the dye concentration. Water movement was followed by withdrawing 15 ml of sample water from the peripheral wells continually at time intervals of one to 50 days after the initial dye introduction. The outside (distal) wells were sampled before the inner (proximal) wells so that the lowest dye concentrations would be collected first. The sampling apparatus was rinsed twice with distilled water before a sample was collected from a well. The sample was then transferred to a plastic bottle and allowed to settle for 24 to 36 hours before being analyzed on a Turner Fluorometer with a Turner #110-832 (546) primary filter, a Turner #110-833 (590) secondary filter and a standard F4T4 BL light source.

## RESULTS

Only slight dye movement was detected by the 22 wells (two of the wells were inadvertently knocked out) of the three creek bank dye movement stations (Table 1). Note that the majority of the significant readings do not occur until at least 32 days after initial dye introduction into the center well. Directional trends were not detected. Remarkably, one year later, the dye concentrations in the center wells of these stations appear to be only slightly less than the initial concentration.

Much of the same results were obtained for the high marsh wells (Table 2). Again, the majority of the significant readings occurred 32-50 days after initial dye introduction into the center well. Two of the 24 wells contributed almost all of the high readings (10-100 ug/l). The

Table 1. Proportion of significant (2-10 ug dye/l water) and highly significant (10-100 ug dye/l water) readings for the three water movement stations on low marsh sites. The wells were spaced 50 cm apart and inserted to a depth of 100 cm.

Days after initial dye introduction into center well

| | | 1-2 | 2-4 | 4-8 | 8-16 | 16-32 | 32-50 | total |
|---|---|---|---|---|---|---|---|---|
| Proximal Wells | 2-10 ug/l | 0/4 | 1/16 | 0/16 | 0/8 | 1/12 | 3/36 | 5/92 |
| | 10-100 ug/l | 0/4 | 0/16 | 0/16 | 0/8 | 0/12 | 1/36 | 1/92 |
| | | | | | | | | 6/92 |
| Distal Wells | 2-10 ug/l | 0/4 | 0/12 | 0/14 | 0/10 | 0/10 | 5/30 | 5/80 |
| | 10-100 ug/l | 0/4 | 0/12 | 0/14 | 0/10 | 0/10 | 0/30 | 0/80 |
| | | | | | | | | 5/80 |

appearance of the dye in the remaining 22 wells was a random and infrequent event and only two of the latter readings were above 10 ug/l.

## DISCUSSION

The dye study reflects the relative importance of each type of soil water. Two of the total 46 wells consistently yielded high dye concentrations (10-100 ug/l) suggesting that these two wells were connected to the center well by a crab burrow or sand lamella. A combination of direct connection and capillary water movement was responsible for the occurrance of the dye in moderate concentrations (2-10 ug/l). Those wells showing no significant dye concentrations (36 of the 46 wells) were imbedded in sediments characterized by stagnation of interstitial water. Apparently, based on the lack of directional trends and scarcity of significant readings, horizontal water movement and percolation does not readily occur.

The following four observations also support the postulate that free water movement through sediment interstices is insignificant. Radioactive phosphorus ($^{32}PO_4$) introduced into the sediments was readily taken up by Spartina (Reimold, 1972). However, when Pomeroy et al. (1967) labeled estuarine water with $^{32}P$ and $^{65}Zn$, significant quantities of the label were not detected in the plants, although it did

Table 2. Proportion of significant (2-10 ug dye/l water) and highly significant (10-100 ug dye/l water) readings for the three water movement stations on the high marsh. The wells were spaced 25 cm apart and inserted to a depth of 30 cm into the sediments.

| Days after initial dye introduction into center well | | 1-2 | 2-4 | 4-8 | 8-16 | 16-32 | 32-50 | total |
|---|---|---|---|---|---|---|---|---|
| Proximal Wells | 2-10 ug/l | 3/18 | 0/12 | 0/12 | 0/12 | 0/12 | 4/29 | 7/83 |
| | 10-100 ug/l | 2/18 | 2/12 | 2/12 | 2/12 | 2/12 | 3/29 | 13/83* |
| | | | | | | | | 20/83 |
| Distal Wells | 2-10 ug/l | 2/20 | 0/12 | 0/12 | 0/12 | 0/12 | 7/25 | 9/81 |
| | 10-100 ug/l | 0/20 | 0/12 | 0/12 | 0/12 | 0/12 | 1/25 | 1/81 |
| | | | | | | | | 10/81 |

*12 of these readings came from two wells which invariably yielded highly significant readings suggesting the wells were connected to the center well by a crab burrow or sand lamella.

**When significant dye concentration was detected in a distal well, it was not detected in the proximal well in line with the center well.

occur in other salt marsh organisms. Apparently, Spartina roots are relatively isolated from the overlying water.

Since the amount of soil water movement is a function of soil particle size (larger particles have larger interstices) and sorting, then the extremely small interstitial spaces between silt and clay particles of Georgia salt marshes should preclude significant gravitational water movement.

The interstitial salinity values at a depth of 30 cm in the high marsh and one m in the low marsh are remarkably stable even when exposed to rain and overlying estuarine water of variable salinity (Nestler, 1977).

The water table in soils can be detected by the presence of a reduced horizon since the water boundary prevents the free movement of oxygen. The oxidized zone in the salt marsh is only one mm deep suggesting that the water table drops insignificantly, and therefore, that gravitational water movement is very slow. The oxidized zones found in the levee sediments (Gardner, 1973) probably represent increased oxygen movement through the Spartina stems (Nestler, 1977).

Other evidence points toward the occurrence of some type of water movement. First, free water occurs on the salt marsh in puddles and small reservoirs such as crab burrows. As the tide recedes the puddles disappear and the water level drops in the crab burrows. Second, if a core is removed from the high marsh, the hole that remains may fill up with water or remain empty.

Microscopic examination of sediment scrapings reveals an abundance of plant fibers, debris and sediment irregularities providing an excellent medium for capillary water movement. As evaporation proceeds on the marsh surface, some capillary water will be lost. Water from puddles and sediment reservoirs travels by capillary movement to the surfaces experiencing the deficit. Thus, an empty hole left by the removal of the core indicates that all free water pools near the surface have been exhausted replacing lost capillary water. Hygroscopic water is not lost since daily flooding occurs 95 percent of the time (Briggs and Gallagher, unpublished report). This drawing-out effect probably does not occur to depths greater than 10-15 cm into the sediments since replenishment by tidal action occurs before extensive evaporation can take place.

Two additional factors come into play on the low marsh (levee). First, the cracks in Georgia salt marsh levees, although appearing superficial, actually extend as deep as 50 to 100 cm and communicate extensively with crab burrows (personal observation). The cracks effectively increase surface area for mass diffusion and allow for greater and deeper drawing out effects by evaporation. Also, puddles will not remain after the tide recedes since the grade is steep on the levee. Therefore, evaporation will immediately result in capillary water movement from sediment reservoirs to the marsh surface. Both of these factors, essentially surface area effects, cause greater sediment water-estuarine water exchange on the levee as evidenced by the variable salinity values obtained from the levee sediments at a depth of 30 cm (Nestler, 1977).

The results of the dye movement experiments, showing little or no horizontal water movement, and the above discussion suggest that water movement other than flushing through animal burrows and sand lamellae is essentially vertical. Water movement probably does not occur to a depth greater than 10-15 cm in the high marsh and 15-20 cm in the low marsh. This interpretation also suggests that there is a potential for large scale exchanges during infrequent events that destroy the sediment microstructure responsible for capillary water movement such as heavy rainfall following marsh dissication or a combination of wave action and rainfall. The intrusion of fresh water into the clays by capillary action and the exposure of new surface areas by wave and wind action would destroy clay structure and release stored materials.

## REFERENCES CITED


Briggs, J. and J. Gallagher, 1972, University of Georgia Marine Institute unpublished report.

Frey, R. W. and J. D. Howard, 1969, A profile of biogenic sedimentary structures in a holocene barrier island-salt marsh complex, Georgia. Transactions of the Gulf Coast Association of Geological Societies 19: 428-444.

Gardner, L. R., 1973, The effects of hydrologic factors on the pore water chemistry of intertidal marsh sediments. Southeastern Geology 15 (1): 17-28.

Nestler, J., 1977, Interstitial salinity as a cause of ecophenic variation in Spartina alterniflora (Loisel). Accepted for publication by Estuarine and Coastal Marine Science.

Pomeroy, L. R., R. E. Johannes, E. P. Odum and B. Hoffman, 1967, The phosphorus and zinc cycles and productivity of a salt marsh. In Proceedings of the Second Radioecology Symposium. Edited by D. Nelson and F. C. Evans. Clearinghouse for Federal Scientific and Technological Information, National Bureau of Standards, U. S. Department of Commerce, Springfield, Va., pp. 412-420.

Reimold, R. J., 1972, The movement of phosphorus through the salt marsh cord grass, Spartina alterniflora. Limnology and Oceanography 17 (4): 606-611

Turner Fluorometry Reviews, 1971, Fluorometry in the studies of pollution and movement in fluids.