Geology and Tectonic History of the Lower Cape Fear River Valley, Southeastern North Carolina

U. S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1466 - A

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By DAVID R. SOLLER

SURFACE AND SHALLOW SUBSURFACE
GEOLOGIC STUDIES OF THE CAROLINA COASTAL PLAINS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1466-A

A study of a major river valley on the outer Atlantic Coastal Plain including mineralogic analysis of the deposits and geomorphic evidence for tectonic uplift

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON: 1988

DEPARTMENT OF THE INTERIOR

DONALD PAUL HODEL, *Secretary*

U.S. GEOLOGICAL SURVEY

Dallas L. Peck, *Director*

Library of Congress Cataloging in Publication Data

Soller, D.R. (David R.)
Geology and tectonic history of the lower Cape Fear River valley, southeastern North Carolina.

(U.S. Geological Survey professional paper ; 1466A)
Bibliography: p.
1. Geology—North Carolina—Cape Fear River valley. 2. Cape Fear River valley (N.C.) I. Title. II. Series.
QE148.C35S65 1988 557.3 s [557.56'2] 86-600245

For sale by the Books and Open-File Reports Section, U.S. Geological Survey
Federal Center, Box 25425, Denver, CO 80225

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GEOLOGY AND TECTONIC HISTORY OF THE LOWER CAPE FEAR RIVER VALLEY, SOUTHEASTERN NORTH CAROLINA

By DAVID R. SOLLER

ABSTRACT

The Cape Fear River is a major Piedmont-draining river system that flows across the Atlantic Coastal Plain in North Carolina. A detailed study of stratigraphy, mineralogy, and geomorphology was undertaken to assess the geology and tectonic history of the valley and the implications for late Cenozoic tectonism in the region. The deposits of these Piedmont-draining river systems cover large areas of the Coastal Plain and are therefore geologically and culturally significant and worthy of further study. The data from the Cape Fear River valley await integration with future studies of other rivers in the area.

Five river terraces are present in the Cape Fear River valley; all lie northeast of the river, with successively older terraces farther from the river. The terraces are correlated with isotopically dated marine and strandline deposits ranging in age from approximately 2.75 to 0.1 Ma. This drainage system has therefore been in existence in some form since at least late Pliocene.

The mineralogies of weathering profiles in the fluvial and upland deposits were compared to assess the variation in profile development on different units, as an aid in estimating the ages of the deposits. The river sands are quartzose with minor feldspar, immature heavy mineral assemblages, and an immature clay mineral suite; sediments beneath older terraces generally contain less hornblende, epidote, and feldspar than younger terraces, as a result of longer exposure to weathering. The abundance of labile minerals and the maturity of the clay-sized mineral suite were found to be useful indicators of the age of the deposit in both valley and upland areas.

Regional uplift and a series of local flexures are proposed to explain the terrace distribution in the valley, based on geomorphic and drill-hole data and on the use of longitudinal profiling with a best fit analysis incorporating time and uplift. A gentle, sustained uplift to the north or northeast of the valley has forced the Cape Fear River to migrate southwestward over time. This migration has allowed the preservation of river terraces and large tributaries only to the northeast of the river. Roughly normal to this uplift (along the trend of both the river and the Cape Fear arch), a complex flexure beginning more than 750 ka uplifted the upper valley near the Piedmont and caused incision of the river, while the lower valley subsided. These minor flexures along the Cape Fear arch were superimposed on a gentle, persistent regional uplift of the region which is largely responsible for preservation of the elevated shorelines. To the south of the Cape Fear River, the Pee Dee River flows along the southeast flank of the uplift; geomorphic and lithologic evidence in the Pee Dee River valley and drainage patterns on the uplands between the two rivers support the regional tectonic model proposed from analysis of the Cape Fear River valley.

INTRODUCTION

In the late 1800's reconnaissance geologic mapping of the Atlantic Coastal Plain established that episodic transgressions and regressions of the ocean had deposited marine sediments and formed a series of marine benches, or "terraces" (McGee, 1886, 1888; Shattuck, 1901, 1906; Johnson, 1907; Stephenson, 1912). Each terrace was assumed to be the product of a single ocean highstand, and the apparently level nature of the terraces was cited as evidence of a passive or epeirogenically active continental margin. The concepts invoked in these early studies persisted relatively unchanged into the modern era; recent detailed geologic investigations (Owens, 1970, in press; Mixon and Newell, 1982; Newell and Rader, 1982; McCartan and others, 1984; Newell, 1985; Owens and Gohn, 1985) have revealed a wealth of stratigraphic and tectonic information and have challenged the longstanding assumptions about the stability of the Atlantic margin.

On the southern Atlantic Coastal Plain, detailed mapping has been largely confined to South Carolina, and most notably to the Charleston area, where a major U.S. Geological Survey study was conducted of the geologic setting around the epicenter of the Charleston 1886 earthquake (Gohn, 1983; McCartan and others, 1984). In an effort to assess regional variations in depositional style and tectonic stability, James P. Owens of the U.S. Geological Survey recently undertook a mapping study of the Coastal Plain astride the Cape Fear arch, encompassing northeastern South Carolina and southeastern North Carolina. Within this complex of Pleistocene and older offshore and marginal marine sediments, two major rivers, the Cape Fear and the Pee Dee, have carved valleys and deposited sediments during past intervals of ocean highstand. In the past, river deposits have rarely been studied in the detail given the marine and strandline sediments on the Coastal Plain, and the significance of mineralogic and stratigraphic data from river valleys was not fully known at the outset of the present study. As one aspect of the regional study directed by Owens, the geologic history of a major river valley was assessed, and is the object of this report. Of the two valleys within the study area, the Cape Fear River valley presented the most opportunity for significant new data; it had never been mapped, has an unusual configuration, and is astride the Cape Fear arch, an area of known tectonism during the Late Cretaceous and Tertiary (Owens, 1970; Owens and Gohn, 1985).

ACKNOWLEDGMENTS

I express sincere thanks to James P. Owens, U.S. Geological Survey, for his encouragement and support of this study, which was done under the auspices of his mapping of the Cape Fear Arch area, North and South Carolina. His advice and suggestions contributed to the formulation of many of the ideas contained herein; the interpretation of weathering beneath the Cape Fear River terraces is based on the weathering sequences used by Owens for the region. Meyer Rubin and his staff at the U.S. Geological Survey Radiocarbon Laboratory supplied the radiocarbon dates, and Leslie Sirkin, Adelphi University, examined the pollen. The author benefited from discussions with Lucy McCartan, Wayne Newell, John Hack, and Helaine Markewich. Finally, the author wishes to thank Earl Lemon and Dennis Duty for field assistance, with the power auger.

GEOGRAPHIC SETTING

The Cape Fear River rises in the Blue Ridge of central North Carolina; in its upper reach it is called the Haw River. In both the past and the present, the river has eroded the Late Proterozoic to Permian granitic, gneissic, and volcanic rocks of the Blue Ridge and Piedmont provinces, delivering this detritus to the coast in a relatively fast moving current compared with nearby rivers that drain only the Coastal Plain sediments. For example, the average discharge of the Cape Fear River at the northwestern end of the study area is 4,956 cubic feet per second, while discharge for nearby Coastal Plain rivers, the Waccamaw and Lumber Rivers, is 1,067 and 3,020 cubic feet per second, respectively (Hendricks, 1961).

The study area lies well to the southeast of the headwaters, along the river's course across the outer Coastal Plain east of the Orangeburg Scarp (fig. 1). The study area is of an irregular shape, covering roughly 1,500 square miles between North latitude 34° and 35° and West longitude 77°50′ and 79°. Figure 2, a low-resolution Landsat image, shows the study area in more detail. Within this area, from just south of Fayetteville, N.C., to the head of the estuary northwest of Wilmington, N.C., the river flows southeast and lies against the southwest wall of the valley. The pre-Holocene flood-plain deposits of the valley, preserved beneath terraces paralleling the Holocene course of the river, are, except for the most recent deposit, each of a roughly uniform width. The terraces number up to five, and the overall width of the valley varies accordingly, from a maximum of 22 miles in the north near Roseboro, N.C., to a narrow feature less than 5 miles wide and covered entirely by the modern flood plain northwest of Wilmington. The river turns to the south into the estuary at Wilmington and exits to the Atlantic Ocean at Cape Fear, N.C.

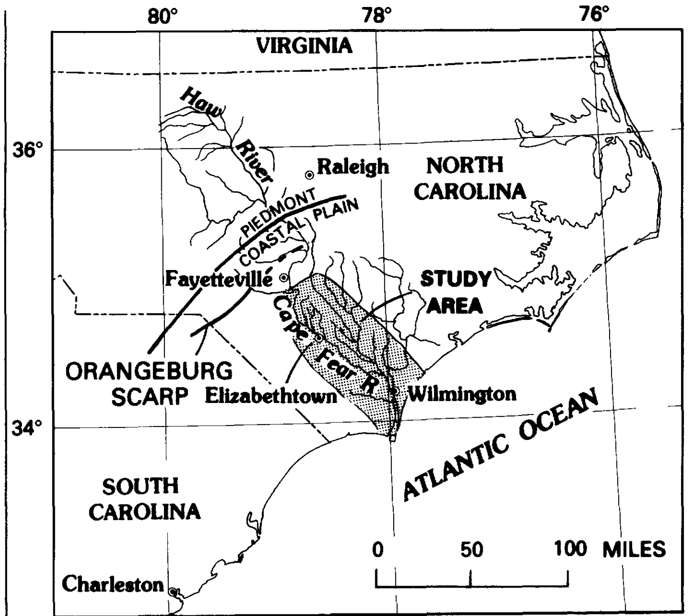


FIGURE 1.—Location of the study area.

The region is rural and is covered mostly by pine forests and small farms. Much of the land in the valley is quite sandy and does not support intensive farming. In addition, the lower reaches of the valley to the south, where the valley constricts, are often swampy in many places.

The sandy nature of the valley is evident from low-level aerial photography (fig. 3). Recent vegetation blankets the valley, but parabolic dune forms, large areas of bare, white sand, and other wind-derived features are common. The most conspicuous of these surface features in the Cape Fear River valley are oriented, elliptical depressions; these features, known as Carolina bays, are common on sandy deposits of certain ages on the Coastal Plain and are perhaps best developed in this valley. Carolina bays are also visible on the low-resolution Landsat imagery shown in figure 2.

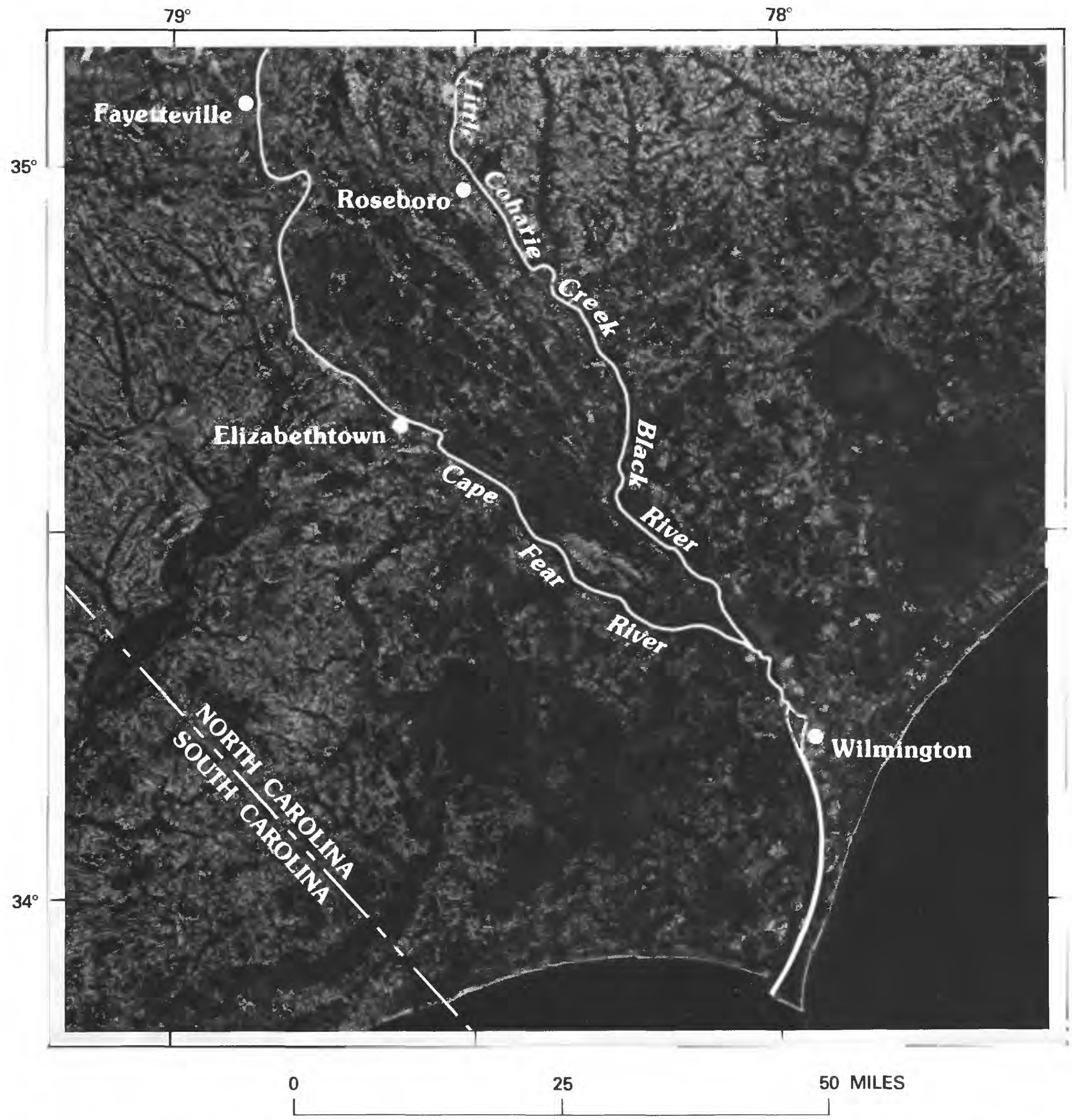


FIGURE 2.—Landsat image of the study area. The river valley is confined between the Cape Fear River and the Black River and attains a maximum width of 35 kilometers just northwest of Elizabethtown, N.C. The elliptical features in the valley are Carolina bays. From National Oceanographic and Atmospheric Administration, 1976.

GEOLOGIC SETTING

The Cape Fear River valley lies astride a broad and temporally persistent tectonic feature, the Cape Fear arch (fig. 4). Tectonic warping of the arch axis has clearly deformed the Cretaceous sediments and influenced sedimentation patterns (see fig. 4*A*, unpublished data from J.P. Owens, U.S. Geological Survey (USGS)); these sediments generally thicken offshore and down the flanks, and are thinner along the arch. The deep-seated nature of this feature is illustrated by structure contours on the top of the basement rock, as shown in figure 4*B* by Owens (written commun., 1986) from Gleason (1981) and Costain and Glover (1982). From a high of -1,500 feet on the arch axis along the coast near Wilmington, N.C., the top of basement slopes

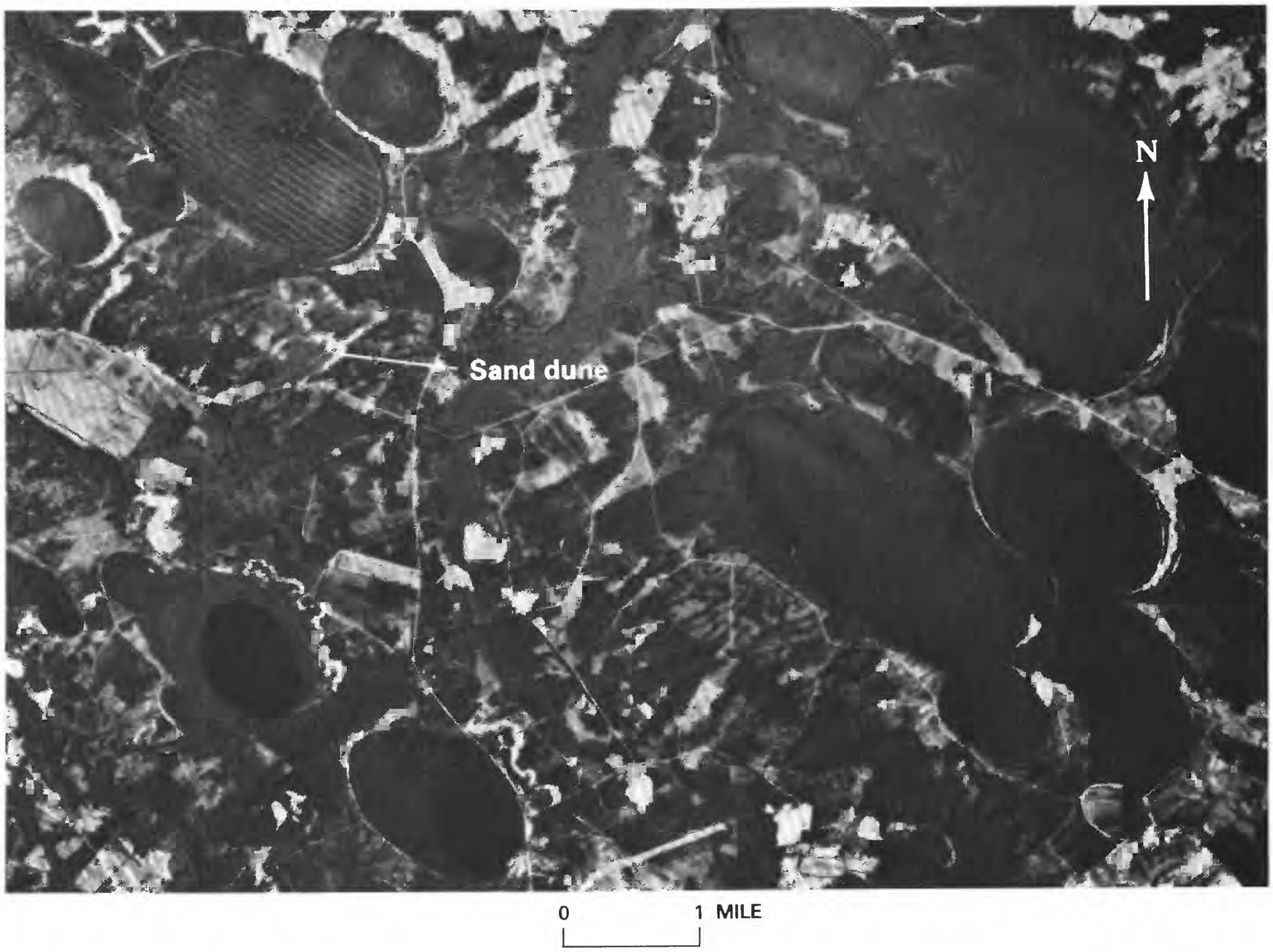


FIGURE 3.—Aerial photograph of the lower Cape Fear River valley. Carolina bays and sand dunes are the most conspicuous features of the valley.

to the south to -2,500 feet near Charleston, S.C., and to the north to below -6,000 feet along the Outer Banks of North Carolina. The Cape Fear arch remained a positive tectonic feature into the Tertiary, affecting middle Eocene and Miocene sedimentation (Owens, 1970; Ward and others, 1978). Quaternary uplift of at least a portion of the arch is discussed in this report. Tectonic movement of the arch during the Holocene has been implied, on the basis of leveling survey data (Brown, 1978). While the arch was active during the Cenozoic, episodes of subsidence may have occurred, alternating with periods of uplift (J.P. Owens, USGS, personal commun., 1985). In addition, the distribution of Cretaceous and Paleogene sediments suggests that the position of the actively uplifting part of the arch has not been stationary (figure 4*A*).

Upper Tertiary and Pleistocene sediments strike across the Cape Fear arch, covering the Upper Cretaceous sediments of the Cape Fear, Black Creek, and Peedee formations, which crop out along the cutbanks of the southwest wall of the Cape Fear River valley. It is the younger sediments that are of concern here, because they document the history of the late Cenozoic drainage now known as the Cape Fear River system.

The upper Cenozoic sediments form a thin blanket over the older sediments, roughly 50 or more feet thick, and show a systematic map pattern. These sediments were deposited during transgressive-regressive cycles caused by glacioeustatic sea level fluctuations. The sequence of deposits in a cycle ideally includes a thin, basal marine unit laid down as the ocean transgressed and a thicker overlying series of deposits preserved as the ocean regressed. The sediments of the regressive phase generally include beach or barrier sands, estuarine-backbarrier sands and clays, and the sands

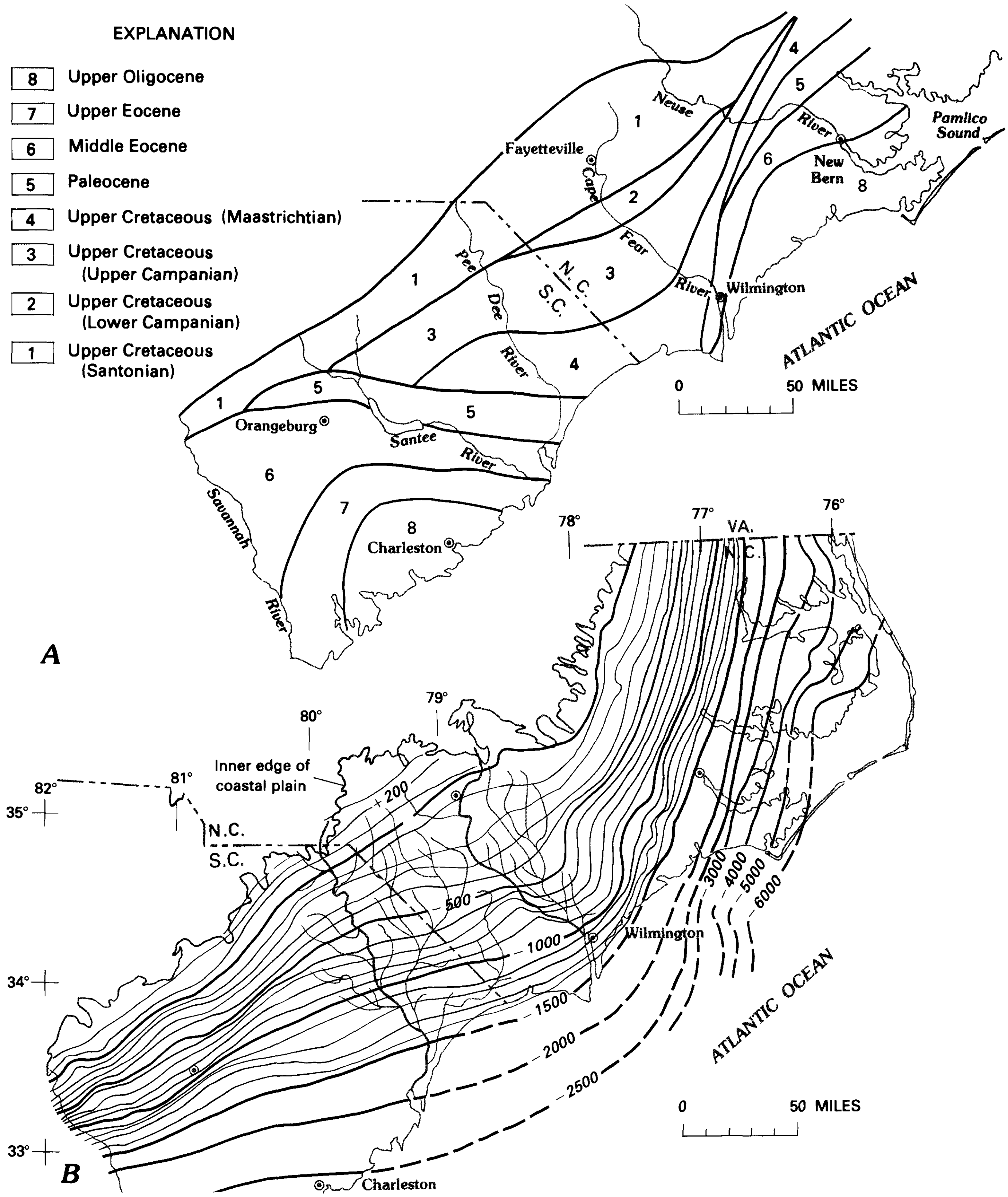


FIGURE 4.—Geologic evidence for the Cape Fear arch as shown by (*A*) the generalized suboutcrop pattern of Santonian to upper Oligocene units as mapped by J.P. Owens, and (*B*) contours on the top of basement as compiled by Owens (written commun., 1986), from Gleason (1981) and Costain and Glover (1982). Contour interval 500 feet, with supplementary 100-foot contours. Contours offshore are shown dashed. The Cape Fear River in North Carolina, the Pee Dee River in South Carolina, and some tributary drainages are also shown.

and clays deposited as flood plains upriver from the backbarrier.

Along the Atlantic Coastal Plain, several transgressive-regressive sequences are evident; older sequences lie some distance inland from the modern shoreline, while younger sequences are present progressively at lower elevations closer to the shoreline. The surface of these sequences is fairly flat, and where several sequences occur nearby, they impart a steplike character to the landscape. In many areas, the break in elevation between deposits of successive cycles is distinct and the various units can be mapped simply by elevation. As discussed in the next section, this mapping tool was commonly applied prior to the use of more sophisticated stratigraphic techniques in recent studies.

Admittedly a crude regional mapping tool, topographic elevations do provide a reasonable way to assess relative ages in a local area, as original, uneroded surfaces of younger deposits occur lower than those of older units. As these sequences are the product of glacioeustatic fluctuations of sea level, it has historically been inferred that the range of glacioeustatic sea level oscillation has progressively decreased since deposition of the older, upper Tertiary sequences. This assumption has been questioned on the basis of the known tectonic history of the Coastal Plain and oxygen isotope data from deep sea cores. An assessment of probable mechanisms for preservation of this series of transgressive-regressive sequences is included in the section entitled "Geomorphic Evidence for Uplift of the Cape Fear River Valley."

PREVIOUS INVESTIGATIONS

The unconsolidated deposits of the Atlantic Coastal Plain in the Carolinas consist of a series of transgressive-regressive sequences, dominated surficially by the backbarrier-barrier sediment complexes. In most areas the deposits of any sequence have a roughly planar surface expression, paralleling the coast. Indeed, it is possible in many places to trace these flat surfaces on topographic maps for many miles. Topographic elevation is a correlation tool of historically wide use on the Coastal Plain and although it is a rough and often misleading tool by modern standards, the construction of topographic maps for any area of the Coastal Plain served to immeasureably advance geologic mapping and interpretation. The evolution of geologic concepts was therefore due in part to availability of adequate topographic base maps, as well as to the geologic skills of the investigator.

In North Carolina prior to topographic mapping of the area, perhaps the first recognition that the major Piedmont-draining valleys had asymmetric cross sections was by local farmers who, when going to markets in South Carolina, observed that low, swampy areas were common to the north of the major rivers and that high bluffs were present on the south banks. Kerr (1875) noted this and correctly assumed that the river had carved an asymmetric valley in response to an external force. He did, however, reject a theory of crustal warping (uplift to the north or subsidence to the south of the valley), because geologic evidence for these events did not yet exist. Kerr instead accounted for valley shape solely by the Coriolis effect, the tendency for the Earth's rotation to cause a moving object, the river, in the Northern Hemisphere to be deflected to the right (in this case, to the south). Although the assumptions on which his argument was based are not valid, Kerr was also an early proponent of the theory that uplift and subsidence have shaped the history of the Coastal Plain. This theory was generally unopposed until the early 1930's.

From 1886 to 1888, W.J. McGee published a series of reports on the middle Atlantic Coastal Plain, mostly on Virginia and the District of Columbia. Without the aid of topographic contour maps, McGee recognized a series of marine terraces and elevated shorelines across the region whose origin was ascribed to periods of submergence and uplift. McGee reasoned that older deposits were not stripped away because later episodes of emergence and subsidence were of lesser magnitude. Of interest here, he relied on an interpretation of topography and stream behavior to infer epeirogenic activity during the late Holocene. In his study of the Chesapeake Bay, a contrast between the upper, Piedmont course and the lower, Coastal Plain course of the rivers was noted. The rivers actively downcut into the Piedmont, yet are at base level or drowned in estuaries on the Coastal Plain. Bluffs in the estuary are talus-free, indicating that the rate of removal by water has outpaced the development of talus by erosion. To McGee, these observations implied differential uplift along the Chesapeake Bay; he suggested that the Piedmont is being uplifted while the Coastal Plain is subsiding. This hypothesis was an early attempt to use fluvial response as a tool of geologic interpretation on the Coastal Plain.

With completion of topographic mapping in eastern Maryland, the first detailed surficial mapping and correlation of a region was done, by Shattuck (1901, 1906). Five major terraces were identified: Lafayette, Sunderland, Wicomico, Talbot, and Recent. These names have become firmly established in the literature and have been correlated by subsequent authors, with varying success, into adjacent States. As with the earlier studies, alternating intervals of emergence and subsidence were proposed as the mechanism preserving the deposits.

Johnson (1907), with the aid of newly completed topographic mapping in northeastern North Carolina, subsequently mapped a series of terraces and noted that terraces having similar elevations are present in Virginia and Maryland. More extensive work on the Coastal Plain terraces of North Carolina was pursued by Stephenson (1912), who adapted the framework of terrace chronology erected by Shattuck (1901, 1906); although somewhat expanded, the terminology remained principally intact, and a correlation of terraces across Maryland, Virginia, and North Carolina was proposed. Clark and others (1912) suggested that the Coastal Plain, hinged at the Piedmont, had alternately tilted up and down to produce the series of terraces. A uniformity of tilt along the Coastal Plain (i.e., epeirogenic movement) was implied for this interstate correlation of terraces.

The preceding studies make reference to a generalized and episodic rise and fall of the Coastal Plain surface in order to explain the terrace pattern. The mechanism driving such epeirogenic motion could not be explained by current knowledge; it was simply assumed to exist. In an abrupt departure from conventional theory, Cooke (1930) proposed that a series of interglacial highstands of the sea, of progressively lesser magnitude, is the sole reason that a series of terraces is preserved. Cooke was skeptical that a periodic, epeirogenic rise and fall of the land surface could have been accomplished without some differential tilting, and he preferred the glacioeustatic mechanism. His study of maps of the Atlantic and Gulf Coast shores from Connecticut to Texas suggested an absence of shoreline tilt; Cooke therefore embraced the theory that glacioeustatic fluctuations have produced and preserved the terraces and suggested a worldwide correlation of terraces on stable coasts solely on the basis of their height above present sea level. While it now seems certain that the terrace sediments were deposited during interglacial highstands of the ocean, the long-held assumption that the Atlantic margin of the United States is stable or uniformly tilting is untenable in light of current research. This topic is discussed in detail in the section entitled "Geomorphic Evidence for Uplift of the Cape Fear River Valley."

In the aforementioned works, two related topics are discussed: correlation and mechanism. The latter is quite speculative and relies on a careful and accurate study of the former, for without a proper correlation of deposits, speculation on how the deposits came to be preserved (i.e., mechanism) is meaningless. The reports of more contemporary researchers indicate that elevations are viewed with trepidation; many of the terraces mapped by earlier workers are complex and include deposits of more than one age. Lithostratigraphic studies in southeast Virginia (Oaks and Coch 1963, 1973) and South Carolina (DuBar, 1971; Colquhoun, 1974; DuBar and others, 1974) have greatly refined the body of earlier work.

Lithostratigraphic correlations without a time constraint have the same limitation as the correlation of terrace surfaces: essentially time-stratigraphic units are being mapped with an independent tool, lithology or elevation. Strictly lithostratigraphic correlations may in places be suspect on the Coastal Plain because the upper Cenozoic deposits, while of widely varying age, were formed under the same conditions and hence are lithologically similar.

The use of biostratigraphy to refine lithostratigraphic correlations is limited by sparse to absent faunal assemblages in marginal marine deposits. Also, the hiatuses between many depositional events are small, and resolution of distinct and useful faunal zones is difficult. Correlation tools such as molluskan and ostracode zonation, uranium-disequilibrium series studies, amino acid racemization, and paleomagnetics have been integrated in some studies (McCartan and others, 1982; Cronin and others, 1984). Cronin (1981) provided a summary of available techniques applied to Coastal Plain stratigraphy, and Szabo (1985) discussed the role and limitations of uranium series and amino acid dating methods in stratigraphic studies of Coastal Plain sediments.

Mineral weathering studies on the Atlantic Coastal Plain have proved useful in differentiating units and determining relative ages. A systematic alteration of immature detrital minerals into a mature, weathered assemblage occurs in the soil and subsoil (i.e., in the weathering profile) of rocks and sediments; these mineral alterations in the soil profile were detailed by Jackson (1965) in his model for weathering sequences as a function of time, environment, and intensity of the weathering processes. In sediments of similar lithology in a given area, the degree of alteration toward the weathered assemblage is a function of the age of the deposit. Owens and others (1983) investigated the mineralogy of various formations of the middle Atlantic Coastal Plain, from Virginia to New Jersey, and concluded that clay and sand mineral assemblages reflect not only the original lithology, but the degree of weathering as well. In Owens' study, both the unweathered mineral assemblages and the ages of the formations were known; these data were used to assess the time required for development of the mature assemblages detected in the weathered zone. These concepts are invaluable both as supporting evidence in stratigraphic studies and in differentiating units when other criteria (e.g., fauna, isotopic dates) are not available. Mineral alteration sequences were found to be a significant stratigraphic guide in a preliminary study of the Cape Fear River valley (Soller, 1984).

GENERAL STRATIGRAPHY

Since the early 1970's, numerous lithostratigraphic studies on the Atlantic Coastal Plain have refined our knowledge of local geologic history. However, the sediments of the various transgressive-regressive sequences are macroscopically quite similar, and detailed mineralogic, biostratigraphic, and isotopic dating studies were required to correctly assign ages to these deposits and to allow regional correlation. From detailed work in the Charleston, S.C., area, McCartan and others (1984) established mineralogic differences between units caused by weathering and hence the relative age of the deposit, and determined both relative and absolute age for units on the basis of molluskan zonation, magnetic polarity, and uranium-series dates on corals. The stratigraphy erected for the Charleston area was compared with that for northeastern South Carolina and northeastern North Carolina, and correlations were suggested (McCartan and others, 1982). The continuing research of J.P. Owens (as cited previously) along the Carolina coast, as well as studies by Szabo (1985), Mixon and others (1982), and Cronin and others (1984), has served to refine the regional stratigraphic framework of these units.

Prior to the Charleston research, J.R. DuBar and others had conducted lithostratigraphic mapping in northern South Carolina, around Myrtle Beach. DuBar and others (1974) subdivided the post- Miocene stratigraphy of northeast South Carolina into 11 units. The oldest unit of concern to this study is the Duplin Formation (DuBar and others, 1974), of Pliocene age, which lies seaward of the Orangeburg scarp at an elevation of roughly 55 meters (180 feet) or more. The next younger unit, the Bear Bluff Formation, of late Pliocene age, lies generally seaward of the Duplin, with a surface elevation of roughly 30 to 37 meters (100 to 120 feet) above sea level. These sediments are predominantly shallow marine sands and appear weathered, with a well-developed soil profile. The next younger deposit mapped was the Waccamaw Formation (lower Pleistocene), occurring at elevations between 20 and 30 meters (65 to 100 feet) above sea level. Lying just seaward of the Waccamaw Formation is DuBar's Canepatch Formation (middle to upper Pleistocene), which has preserved at the surface a barrier-backbarrier complex at 12 to 14 meters (40 to 45 feet) above sea level. The Socastee Formation (upper Pleistocene) is represented by a barrier complex between the Canepatch-age barrier and the modern barrier beach. The Socastee- age barrier occurs at elevations up to 12 meters (40 feet) above sea level.

McCartan and others (1982) compared the Pleistocene stratigraphy around Charleston, S.C., with dated units at Myrtle Beach, S.C. (within DuBar's field area), and at Flanner Beach (on the Neuse River, northeastern North Carolina). McCartan's study identified at least four major cycles of transgression and regression during the Pleistocene. The deposits of individual cycles have been assigned formational status and are characterized by a thin, basal transgressive marine unit overlain by thicker, regressive strandline deposits.

The Waccamaw Formation was the oldest unit mapped by McCartan, at elevations up to 32 meters (105 feet) above sea level. Uranium-series dates on corals indicate that this unit may be at least one million years old. DuBar and others (1974) considered the Waccamaw to be a single, time-transgressive unit of early to middle Pleistocene age on the basis of fauna. In Charleston, S.C., the possibility exists, based on differences in elevation, intensity of weathering, and isotopic dates, that the Waccamaw could be subdivided into an older, topographically higher Waccamaw and a younger, lower unit at least 750,000 years old (McCartan and others, 1984). Although the local units mapped in McCartan and others (1984) were given only numerical designations, these units are correlated regionally in McCartan and others (1982) and are assigned formational names. In the subdivision of the Waccamaw, the upper unit will be assigned to the overlying Penholoway Formation (Owens, in press).

The Socastee Formation is represented by an areally extensive backbarrier-barrier complex around Charleston, at elevations up to 12 meters (40 feet) above sea level. The Socastee tentatively correlates with the Flanner Beach Formation of northeastern North Carolina, on the basis of elevation and uranium-series dates, which cluster between 180 and 240 ka (McCartan and others, 1982). McCartan's work necessitated a reinterpretation of DuBar's type Canepatch and type Socastee (McCartan and others, 1982); Socastee-age deposits unerlie the extensive 12- to 14-meter (40- to 45-feet)-above-sea-level barrier-backbarrier complex that was assigned a Canepatch age by DuBar and others (1974), while Canepatch deposits are not preserved as a barrier complex but exist only as isolated subsurface deposits of marine origin.

The Wando Formation is the youngest Pleistocene unit identified in the Charleston area (McCartan and others, 1980). Near Charleston, the surface of the Wando backbarrier flat does not exceed 5 meters (16 feet) above sea level. McCartan and others (1982) tentatively correlated the Wando with the 5-meter level

TABLE 1.—*Generalized late Cenozoic stratigraphy for the outer Coastal Plain of parts of North Carolina and South Carolina*[1]

Formation	Approximate age (Ma)[2]
Wando	0.1
Socastee	0.2
Canepatch	0.45
Penholoway	>0.75
Waccamaw	1.75
Bear Bluff	2.75
Duplin	3.25

[1]Stratigraphy derived from McCartan and others (1982), DuBar and others (1974), and J.P. Owens (U.S.G.S., pers. commun., 1985).

[2]The ages given here are near the average for dates obtained by isotopic dating and other means. Given the spread in dates and the uncertainty involved in dating these sediments, this table gives a generalized, approximate age for each depositional interval.

on the Neuse River, which is the Core Creek sand described by Mixon and Pilkey (1976). The age of the Wando deposits was determined by uranium-series and amino acid methods to be between 87 and 126 ka. The stratigraphic framework used in this study, derived from McCartan and others (1982) and supported by the numerous cited reports, is shown in table 1. The approximate age for each formation in table 1 is generalized from the range of dates obtained by the various dating methods. These dates are provided largely for purposes of uplift rate calculation (see "Geomorphic Evidence for Uplift of the Cape Fear River Valley").

STUDY APPROACH

To assess the geology and history of the Cape Fear River valley, this study included mineralogic, stratigraphic, and geomorphic analyses. Although geomorphology proved to be a useful tool in mapping terrace surfaces and in analyzing the region's tectonic history, the identification of each terrace as a unit geologically distinct from adjacent terraces was dependent on subsurface analysis. Lithologic and mineralogic study of the fluvial deposits, which are buried beneath a dune cover of variable thickness, was essential for correlation. In fact, these analyses of the subsurface provided the basic geologic data upon which much of the geomorphic analysis was based.

SAMPLE LOCALITIES

In the study area, surface exposures of any great thickness are rare, and exposures of the entire thickness of fluvial sediments beneath any terrace were not found. Therefore, deposits were sampled with a power auger. Sample localities were confined to roads or trails capable of supporting the truck- mounted auger. Each drill site was located to provide some information on important topographic features, such as Carolina bays, dune fields, and terrace scarps, in addition to stratigraphic and lithologic data. During 1982 and 1984, 27 holes were drilled in the Cape Fear River valley and several holes were drilled in the adjacent upland deposits (fig. 5). In most holes, samples were collected from the weathering profile and in the various lithologies encountered downhole. Next to the Cape Fear River, a cutbank exposure of an upland unit was sampled for comparison with the borehole data.

In 1980, a seismic reflection profile of the Cape Fear River channel was recorded from Elizabethtown to just northwest of Wilmington (fig. 1) by Jim Henry, Skidaway Institute. These records show a prominent reflector whose depth agrees closely with the depth to the Cretaceous sediment beneath the flood plain in nearby drill holes. Seismic data were used in the evaluation of the most recent phase of valley entrenchment and filling, which is associated with late Pleistocene and Holocene sea level fluctuations.

METHODS OF ANALYSIS

A weathering and provenance study was conducted for minerals in the fine and very fine sand fraction (between 63 and 250 micrometers) of 142 selected samples. These sands were separated by standard techniques into "heavy" minerals (those having a specific gravity greater than 2.85) and "light" minerals (specific gravity of 2.85 or less). Heavy and light minerals were identified under the petrographic microscope with index oils, according to standard criteria (Krumbein and Pettijohn, 1938). For untwinned feldspars, the species was determined by the x-ray mapping technique and x-ray fluorescence capability (EDAX) of the scanning electron microscope. Paul Hearn, USGS, provided the EDAX analyses.

Clay-sized minerals were identified from oriented, slurry mounts in a Diano x-ray diffractometer. The samples were often treated with ethylene glycol or heated to 350°C for 1 hour to facilitate identification. A clay mineral (e.g., vermiculite) is identified by its basal spacing and its response to chemical or heat treatment; the chemical composition and detailed structure cannot readily be assessed in mixed assemblages, especially when subjected to weathering. The true nature of two vermiculites from different localities or different levels within a weathering profile may be quite different; for example, when heated to 350°C, one sample may lose only a portion of its interlayer water but maintain crystallinity (i.e., decrease in basal spacing) while the other sample may appear to lose crystallinity or show a

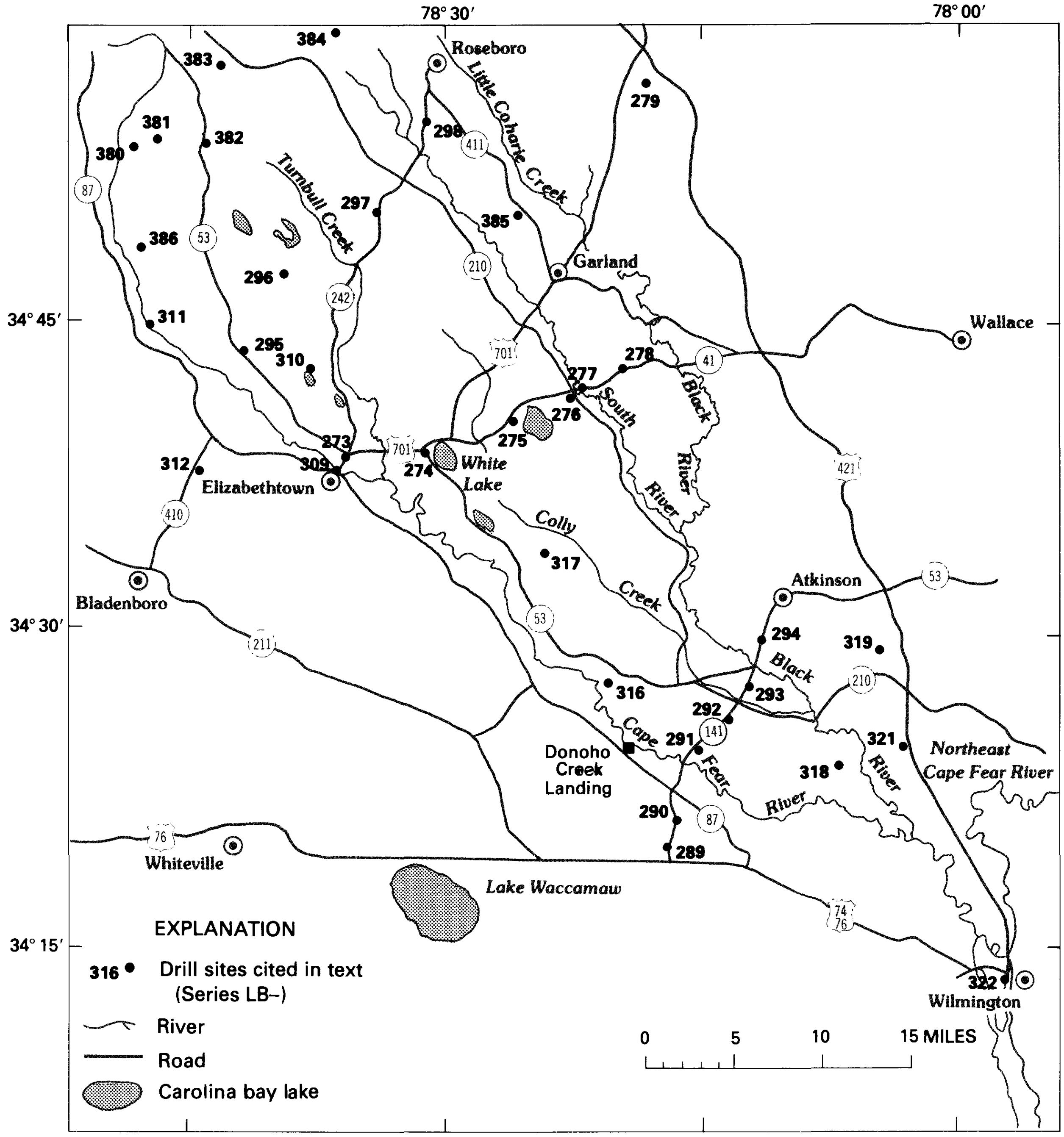


FIGURE 5.—Location of drill sites within the study area. The Cape Fear River valley is located between the Cape Fear River and the Black River and Little Coharie Creek. Carolina bay lakes are also shown for later reference in text.

random loss of interlayer water (i.e., disappearance of peak). Their compositions and responses to treatment are slightly different, but both samples are considered to be vermiculites. Similarly, kaolinite may be highly crystalline and ordered, disordered along the b-axis, or partially hydrated. These conditions are manifested in different x-ray diffraction patterns. Of greater complexity are the mixed-layer clay minerals. A detailed discussion of clay minerals, mixed layering, and identification techniques is given elsewhere (Brown, 1961; Thorez, 1975).

Samples of peaty or organic layers were collected by power auger from beneath the sand rims of Carolina bays, beneath sand dunes, and in the Holocene valley fill. Radiocarbon analyses were provided by the U.S. Geological Survey Radiocarbon Laboratory, and pollen

TABLE 2.—*Radiocarbon data for peat and macerated wood samples*

C^{14} lab number	Sample number	Sample depth (feet)	C^{14} date (years)	Stratigraphic position
W-5141	LB-291	5	7,700 ± 100	Base of dune.
W-5157	LB-292	7	5,720 ± 80	Base of dune.
W-5096	LB-294	9.5	>40,000	Base of dune.
W-5099	LB-296	19	>40,000	Peat at base of bay rim sand, Bushy Bay.
W-5155	LB-309	35	3,540 ±60	Wood layer near base of flood-plain section.
W-5177	LB-317	10.5	>37,000	Peat at base of bay rim sand, Tedder Bay.
W-5171	LB-317	14	>35,000	Same as above.
W-5181	LB-321	30	>36,000	Base of dune.
W-5167	LB-322	33	7,270 ± 90	Base of tidal marsh peat.

analyses were conducted by Leslie Sirken of Adelphi University.

RADIOCARBON DATA

Nine peat and macerated wood samples were dated by the USGS Radiocarbon Laboratory (table 2). Samples were either from the peaty interval beneath dunes or rims of Carolina bays or from within a fluvial interval. The majority of dates beneath dunes, and both dates beneath bay rims, are minimum dates greater than 35 ka. At LB-291 and LB-292, dates indicate another dune forming event, of Holocene age. The dates from the fluvial intervals (LB-309 and LB-322) are Holocene and document post-glacial sea level rise in the area.

POLLEN DATA

Of nine samples analyzed for pollen (table 3), six were also radiocarbon dated. For all samples, the sediment was deposited in freshwater wetlands similar to those of the modern Coastal Plain; nearby forests were dominated by pine, oak, and birch, and a temperate climate existed during deposition of all samples except

TABLE 3.—*Pollen data*

Sample number	Sample depth (feet)	
LB-291	5	Pine dominant, oak, hickory, birch, alder, cedar; nonarboreal pollen well represented with grass, sedge, composites. Looks like pine, oak, hickory forest, open land; freshwater wetland deposition site; temperate climate.
LB-294	9.5	Nonarboreal pollen dominant over arboreal pollen with grass, composites, including abundant ragweed, aquatics; arboreal pollen mainly pine, birch, cedar, alder, oak; *Sphagnum* and fern spores common. Freshwater wetland, pine, oak, birch regional forest; temperate climate.
LB-296	19	Pine, oak, birch, other hardwoods; grass, sedge, composites, minor chenopod and aquatics. May be pine barrens with ericaceous understory. Temperate climate.
LB-309	10.5	Pine, oak, birch, cedar, holly, Ericaceae; nonarboreal pollen-grass composites; moss and fern spores common. Similar to above samples in table but with holly and ericaceae more abundant. Temperate climate.
LB-309	14	Pine, birch, hickory, sweet gum, oak, black gum, alder, and others; composites. Fewer wetland species; some warmer indications (the gums). Pine, birch, sweet gum, hickory dominated forest. Warm temperate climate.
LB-309	30	Pine, oak, birch, hickory, sweet gum, black gum. Trace of nonarboreal pollen. Warm temperate climate.
LB-309	35	Pine, oak, hickory, birch, minor black gum, sweet gum, alder, ash, cedar, grass. Warm conditions began or existed as early as this level.
LB-317	14	Pine, oak, alder, minor birch, cedar; nonarboreal pollen minor with grass, composites, sedge. Freshwater wetland; pine, oak forest; temperate climate.
LB-321	30	Pine, oak, birch, alder, ericaceae, holly; composites, grass; moss and fern spores. Freshwater wetland; pine, oak, birch ericaceae forest, possibly pine barrens type; temperate climate.

at LB-309. From LB-309, four pollen and one radiocarbon analyses were done to assess climatic variations during the Holocene. The radiocarbon age of the lowest sample (35 feet) at LB-309 is 3,540 ± 60 yBP. At that time, and until the flood plain had filled to within 12 feet of the present surface, the climate was warmer than either the present or the recent past (7,700 yBP at LB-291). A warmer climate is indicated by the presence of black and sweet gum in the pollen assemblage. A discussion of Holocene sea level rise and climatic variation is not within the scope of this paper.

IDENTIFICATION AND CORRELATION OF TERRACES

The lower Cape Fear River valley, below Fayetteville, N.C., attains a maximum width of nearly 22 miles and is mostly filled with river terrace deposits; the flood plain is confined to the southwest edge of the valley, between the lowest terrace and the older formations making up the uplands. These river terraces are the fluvial facies of formations discussed earlier in the paper. The terrace deposits were sampled by augering and were found generally to consist of a dune sheet capping sandy fluvial sediments which in turn overlie Cretaceous marine and deltaic sands and clays of a distinctive dark gray or green color.

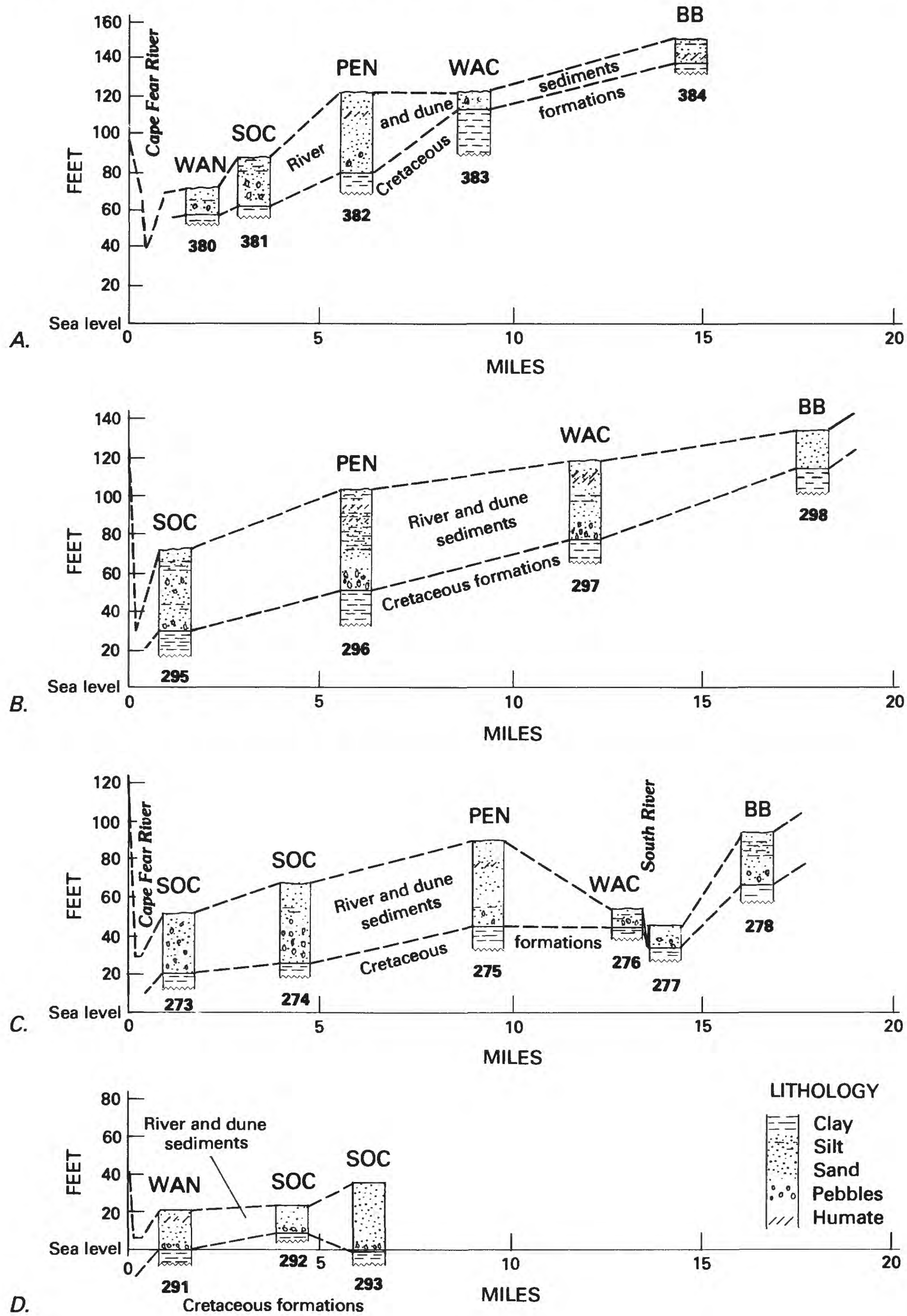
Elevation of the land surface and of the fluvial-Cretaceous contact rises to the northeast away from the flood plain, as shown by valley cross sections (fig. 6). The Cape Fear River clearly has been incising the uplands and migrating to the southwest during the Pleistocene, preserving old fluvial sediments as unpaired terraces to the northeast of the river. Tributaries entering the Cape Fear River from the northeast side are numerous, and many are quite large, while the few tributaries entering from the southwest side are merely small drainages of the upland scarp bordering the river. The lack of tributaries to the southwest is due to stream capture by the Cape Fear River during migration and subsequent erosion of the uplands. If river migration had been a gradual and continuous process, a single slipoff terrace would have formed. However, topographic breaks are apparent and delineate a series of terraces having different elevations that presumably formed at different times and base levels. Mineralogic data support the mapping of several distinct terraces in the valley and provide an estimate of terrace age. These data are discussed in following sections.

Alternating with periods of deposition were intervals when sea level was depressed and erosion and incision occurred in the river valleys of the Coastal Plain. If the incised Holocene channel of the Cape Fear River is representative of the dimensions of past incised channels, the probability of encountering one of these narrow channels in any of the drill holes would be low. Discussion of the history of the Cape Fear River valley, therefore, necessarily deals with the tangible record; erosion and incision during glacial intervals, when sea level was lowered, is assumed, but the extent of these processes cannot readily be assessed.

Six levels (the Holocene flood plain and five terraces) were identified from mineralogic and geomorphic analysis (pl. 1) and were correlated with formations composed of coastal and nearshore facies whose ages are known from isotopic dating and faunal study. Preliminary correlations (Soller, 1984) have been revised by Owens (in press) on the basis of regional mapping. A fluvial equivalent of the Duplin marginal marine unit is not preserved in the Cape Fear River valley. The oldest and highest terrace, of Bear Bluff age, is bounded on the southwest by the South River and on the northeast by Little Coharie Creek and the Black River (fig. 7) and was mapped from northwest of Roseboro southeastward to the north-south stretch of the Black River. The terrace surface appears old relative to the other terraces; it is more dissected, and Carolina bays are poorly preserved. The presence of Carolina bays distinguishes this terrace from the upland deposits immediately to the north and east, which are devoid of bays.

The surfaces of the Waccamaw, Penholoway, and Socastee terraces appear much younger than the surface of the Bear Bluff terrace. Dissection is limited to the terrace borders along the courses of major streams. Elsewhere, the surfaces of these younger units are flat to gently rolling, with numerous well-preserved Carolina bays. All bays in the area are oriented approximately S. 50 E. (Johnson, 1942). Sand ridges on the east and southeast margins of the bays are prominent, and many merge into dunes. Dune fields are common, and the arcuate dune forms (parabolic and longitudinal) are visible on aerial photographs. Dunes tend to be oriented NE.-SW., and the parabolic forms indicate that the wind blew from the southwest. The orientation of Carolina bays, whose long axes tend to lie perpendicular to wind direction (Kaczorowski,

FIGURE 6.—Cross sections of the Cape Fear River valley along four auger hole transects. The uppermost transect in the figure is the farthest upvalley, and each lower transect is farther downvalley. Refer to figure 5 for drill hole locations. (*A*) transect along LB-380 to LB-384; (*B*) transect along LB-295 to LB-298; (*C*) transect along LB-273 to LB-278; (*D*) transect along LB-291 to LB-293. Note the decrease in slope of the valley floor (top of Cretaceous age sediments) from northwest to southeast (downvalley). WAN, Wando Terrace; SOC, Socastee terrace; PEN, Penholoway terrace; WAC, Waccamaw terrace; BB, Bear Bluff terrace.


A.
FEET
MILES
Sea level
Cape Fear River
WAN
SOC
PEN
WAC
BB
River and dune sediments
Cretaceous formations
380
381
382
383
384
B.
SOC
PEN
WAC
BB
River and dune sediments
Cretaceous formations
295
296
297
298
C.
Cape Fear River
South River
SOC
SOC
PEN
WAC
BB
River and dune sediments
Cretaceous formations
273
274
275
276
277
278
D.
WAN
SOC
SOC
River and dune sediments
Cretaceous formations
291
292
293
LITHOLOGY
Clay
Silt
Sand
Pebbles
Humate

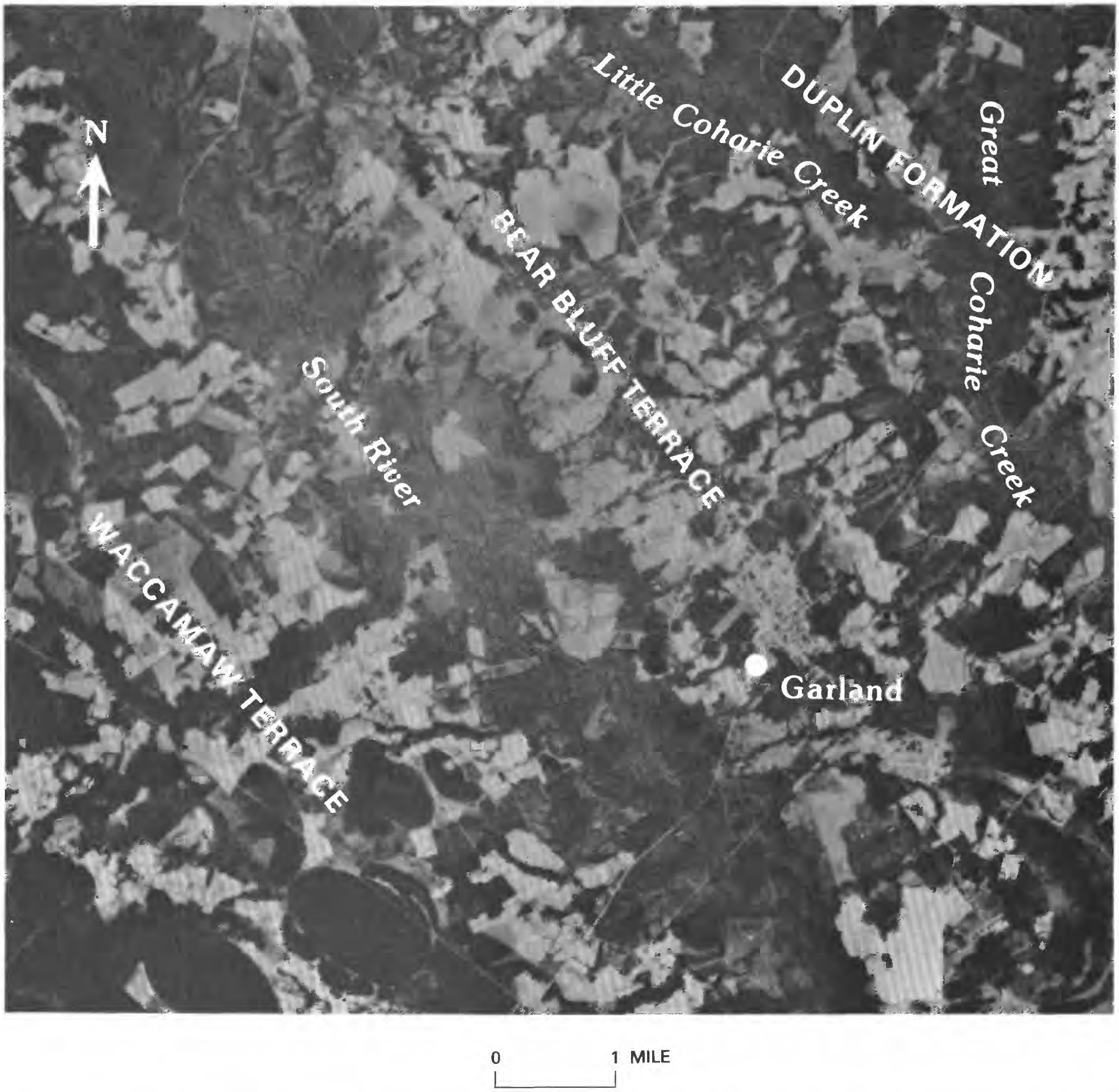


FIGURE 7.—Aerial photograph of the Cape Fear River valley and uplands near Garland, N.C. The Bear Bluff terrace trends from the northwest to the southeast corner of the photograph, between the two flood plains. Upland sediments (Duplin Formation) lie to the northeast, and the Waccamaw terrace lies to the southwest.

1977), also indicates a strong southwesterly wind during the interval of Carolina bay and dune formation.

While the Waccamaw terrace is clearly bounded on the northeast by the South River, the position of the southwest border with the lower, Penholoway terrace is not as apparent. However, the courses of tributary streams may be an aid in delineating terraces because they sometimes flow along the base of scarps (e.g., Little Coharie Creek-Black River, South River). In this manner, the course of Turnbull Creek defines a portion of the Waccamaw-Penholoway terrace contact (pl. 1). In a similar manner near White Lake, Colly Creek is diverted by a dune field into a southeasterly course and follows the scarp between the Penholoway and Socastee terraces. The dunes block drainage south toward the Cape Fear River, forcing Colly Creek into a swampy course to the Black River. This dune migration and stream diversion is relatively recent, as Carolina bays are partially covered by the Colly Creek flood plain (fig. 8). Elevations across Colly Creek differ by 10-20 feet; therefore the creek was selected as the approximate position of the scarp between the Penholoway and Socastee terraces. Northeast of White Lake, the Penholoway-Socastee terrace border is defined by a dune field at the base of the scarp.

The Wando terrace differs markedly from the higher terraces. The topography is irregular owing to the well-preserved meander scars and scrollwork features; the sharpness of these features readily distinguishes the Wando terrace from other surfaces (figs. 8, 9). The Wando surface is also notable for its lack of Carolina bays. Soil surveys (Drake and Belden, 1906; Hearn and others, 1914; Hardison and others, 1915, 1917; Jurney and others, 1926; Perkins and Goldston, 1937) map Wando terrace and flood-plain deposits in the Cape Fear River valley; the Wando terrace surface lies slightly higher than the flood plain and is informally called the "second bottoms."

Upvalley from the confluence of the Cape Fear and Black Rivers, around LB-292 and LB-318 (fig. 5), the lower reaches of the Socastee and Wando terraces lie at about the same elevation; in this region there is a coexistence of the Carolina bay and dune topography characteristic of Socastee and older terraces and the scrollwork features of the Wando terrace. The areas of Carolina bay and dune topography on the Socastee terrace are elongate and are surrounded by areas of scrollwork topography characteristic of the Wando terrace (fig. 9).

The modern flood plain is extremely narrow northwest of Elizabethtown, and the river becomes entrenched upvalley. To the southeast the flood plain widens and occupies the entire valley width of less than 5 miles in the stretch between LB-318 and the confluence with the Black River. The flood plain decreases in extent as the Cape Fear River turns south and approaches the estuary at Wilmington.

GROSS LITHOLOGY AND THICKNESS OF THE FLUVIAL DEPOSITS

During the Holocene sea level rise the Cape Fear River gradually aggraded to the present base level, depositing fine sediments on the flood plain. At two locations on the flood plain, power auger samples were obtained, at Wilmington and Elizabethtown, N.C. The Holocene section at Wilmington is largely tidal marsh peat, deposited since approximately 7,270 yBP in the flood plain adjacent to the Cape Fear River channel as the sea level rose. At Elizabethtown, channel aggradation began about 3,540 yBP, when silt and clay with lesser interbedded sand filled the incised channel. Between 40 and 45 feet of sediment was deposited at both localities.

Fluvial sediments beneath the terraces reflect a river at grade and are therefore unlike those sediments beneath the flood plain, which were deposited as sea level rose and the channel backfilled. Instead of a thickness of fine-grained sediments, the pre-Holocene deposits are coarser and a generally fining upward sequence is preserved. Commonly a pale tan to gray pebbly coarse sand at the base of the section, pebbles decrease in abundance upward while the matrix in many places grades from coarse sand into a silty fine sand or clay. A generalized section consists of channel sands and overbank silts and clays, capped by dune sand. In some cases, the overbank silts and clays are absent or are intercalated with channel sands.

Some dune sands in the Cape Fear River valley are colored dark brown by humate. Swanson and Palacas (1965) reported the impregnation of northwest Florida dune sands by organic compounds and originated the term "humate" to describe this secondary accumulation of organics in the subsurface. Although the mechanism is unclear, soluble and colloidal humic acids are leached from plant litter and accumulate in the subsurface. Daniels and others (1975, 1976) and Holzhey and others (1975) studied the humate (Bh horizon) occurrences in North Carolina. Bulk sediment chemistry in humate zones is dominated by aluminum, and pollen counts are low, indicating postdepositional introduction of acidic organic debris. Well-drained sediments (sandy, with less than 8 percent clay), a high water table, and vertical water flow were found to be associated with humate and presumably are required for formation of humate. These conditions are common in the Cape Fear River valley.

From the Bear Bluff terrace to the Socastee, the thickness of fluvial deposits is rather constant, being generally 22 to 30 feet and ranging up to 44 feet. As an exception, beneath the upper portion of the Bear Bluff terrace from Roseboro to the northwest, the majority of sediment sampled in the two auger holes (LB-298 and LB-384) is dunal. Fluvial sediment was probably eroded during subsequent uplift and tilting of the valley, as discussed in a later section. Other areas of unusually thin fluvial sediments (e.g., at LB-276, 277, and 292) were at one time probably of greater thickness, but erosion by adjacent rivers has since reduced the terrace elevation at these localities.

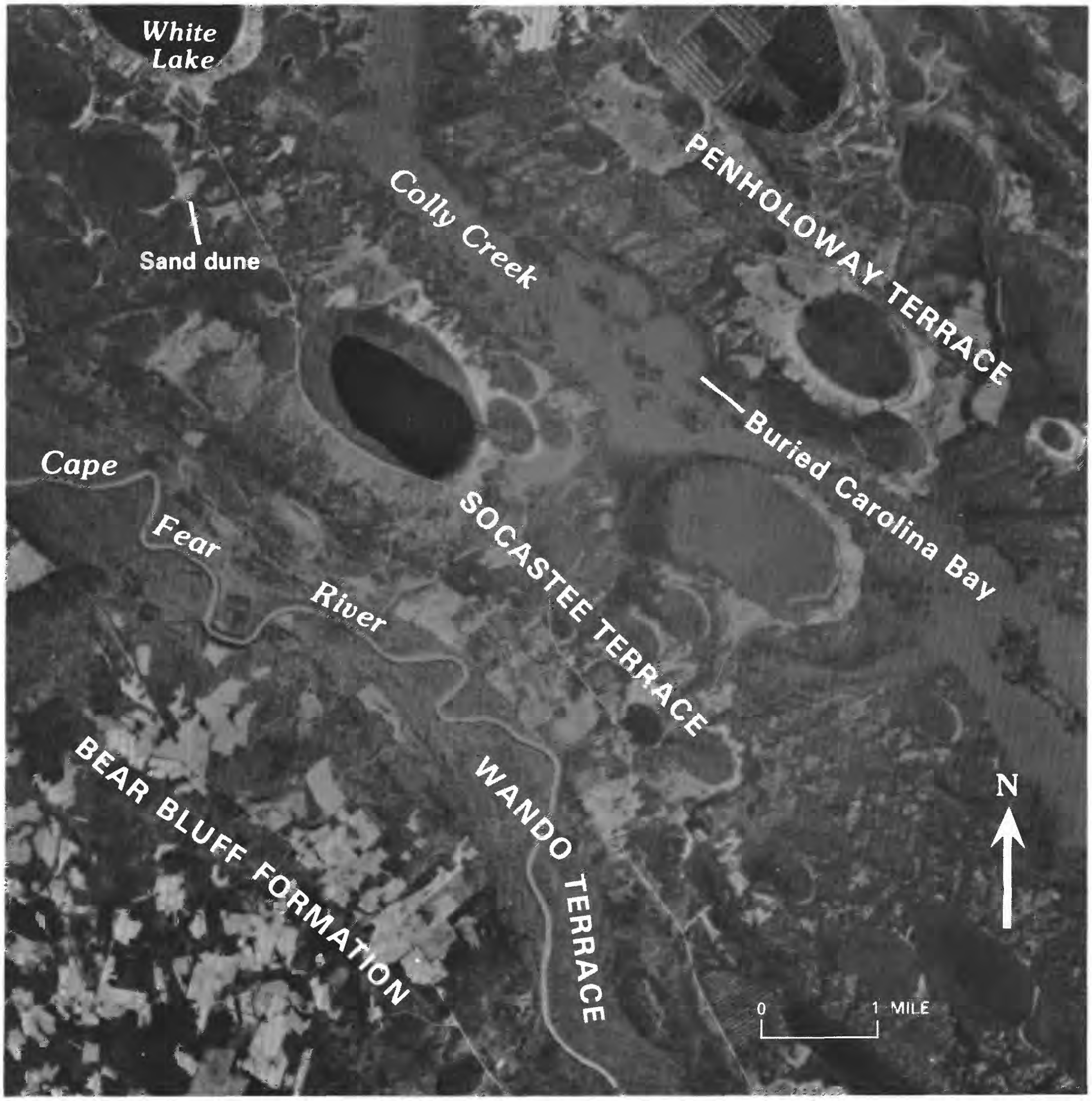


FIGURE 8.—Aerial photograph of the Cape Fear River valley to the northeast of the Cape Fear River; upland sediments (Bear Bluff Formation) lie to the southwest of the river, and the Wando, Socastee, and Penholoway terraces lie to the northeast. Note widespread sand dunes and diversion of Colly Creek from a southerly to a southeasterly course near White Lake by migrating dunes. The modern Colly Creek flood plain has buried several Carolina bays along the scarp between the Socastee and Penholoway terraces.

FIGURE 9.—Aerial photograph of the lower Cape Fear River valley. Upland sediments (Waccamaw and Penholoway Formations) lie southwest of the Cape Fear River, and valley sediments lie to the northeast. Just upriver from the confluence of the Cape Fear and Black Rivers, elongate areas of Carolina bay-covered Socastee terrace are surrounded by scrollwork topography characteristic of the Wando terrace.

Fluvial deposits beneath the Wando terrace are markedly thinner, ranging between 3 and 13 feet in thickness. The relatively thin blanket of Wando sediment may be due to a change in any one of several factors that governed sedimentation in the valley since Bear Bluff time. These include river regime, duration of the depositional interval, and variation in sea level or uplift rate during deposition. While the factors affecting the thickness of fluvial deposits are not explored here, conditions during Wando deposition were somewhat different than those operating during older depositional intervals.

MINERALOGY OF THE FLUVIAL AND UPLAND DEPOSITS

"HEAVY" SAND-SIZED FRACTION

Petrographic analysis of nonopaque heavy minerals revealed a variety of species of two general types, labile and nonlabile. Labile minerals are dominantly garnet, hornblende, and epidote, the first-cycle erosional products of the gneisses, granites, and volcanic rocks drained by the Piedmont tributaries of the Cape Fear River. Nonlabile components (zircon, tourmaline, rutile, staurolite, kyanite, and sillimanite), while present in small quantities in Piedmont rocks, dominate in the shallow marine and strandline sediments of the surficial units on the outer Coastal Plain, bordering the river valley. These resistant minerals have become concentrated over many cycles of erosion and redeposition, owing to the eventual mechanical and chemical destruction of labile minerals. Fresh river sediments might be expected to contain a mixed assemblage of labile minerals washed from the Piedmont rocks and nonlabile minerals from the Coastal Plain sediments.

Study indicated that drill holes are of two basic groups, those whose sediments contain abundant labile minerals and those that do not. Hornblende is especially unstable in well-drained sediments and provides the clearest differentiation between the two groups. Sediments in holes LB-279, 312, 294, 289, 290, and 319, and the Donoho Creek Landing outcrop (fig. 10) have essentially no hornblende and few labile minerals in general, and lie outside the river valley. These drill holes sample shallow marine, backbarrier, or beach sediments that have been reworked and weathered extensively and are in most cases older than the fluvial sediments. Within this group there is a noticeable trend: the abundance of epidote, hornblende, and feldspar (to be mentioned later) decreases in older units. Sediment from the upland drill holes LB-279 and 312 is Pliocene in age and is nearly devoid of these minerals. In younger upland deposits (LB-294, 289, 290, 319, and Donoho Creek Landing) sediments are richer in these labile minerals, but their abundance decreases upward in the weathering profile.

In the drill holes that sample fluvial deposits in the Cape Fear River valley (fig. 11), there is no clear trend of an upward decrease in labile mineral abundance. There is, however, a lower proportion of epidote and hornblende in successively older units. In general, the valley sediments contain much larger proportions of labile minerals than the upland sediments, although the oldest sediments in the valley tend to be mineralogically mature and therefore bear more resemblance to the upland sediments (than do the younger valley deposits). Labile minerals commonly account for more than 40 percent of the heavy minerals in the river terrace sediments, compared with less than 20 percent in the uplands.

For the fluvial interval of each drill hole, an average percentage of hornblende and epidote in the nonopaque heavy mineral fraction was computed from the samples analyzed, which numbered between one and six samples per drill hole (figs. 12, 13). For comparison of weathering in different units, this technique was used instead of depth of weathering because, as already noted, weathering profiles are uncommon in the fluvial deposits. A discussion in support of the average percentage technique is given in the section on "Weathering of the Fluvial Deposits." The sediment beneath the Bear Bluff terrace has the least hornblende of any terrace in the valley (less than 5 percent of the nonopaque heavy mineral fraction). Hornblende generally increases in younger deposits in the valley: less than 5 percent in Waccamaw sediments and 10 percent or more in Penholoway, Socastee, Wando, and floodplain sediments. At any transect across the valley, the abundance of hornblende is less on older terraces than on younger ones. Also, for Penholoway and younger terraces, the abundance of hornblende seems to decrease downvalley within each unit, although this trend is not strong. Of the two trends noted for hornblende distribution, the crossvalley variation is due largely to weathering effects and the downvalley variation to the effects of dilution (i.e., the hornblende-rich sediment carried by the Cape Fear River is gradually diluted downvalley by introduction of sediment rich in resistant minerals from Coastal Plain-draining tributaries).

The crossvalley and downvalley trends apparent for hornblende also seem to characterize epidote distribution. However, epidote is somewhat less susceptible to weathering than hornblende and in many samples is by far the major nonopaque heavy mineral constituent. Although the abundance of hornblende generally decreases gradually across the valley from younger to older units, the amount of epidote sometimes fluctuates greatly between samples in a drill hole. For these reasons epidote is not considered as sensitive a predictor of weathering as hornblende.

Labile minerals are less common upward in several holes, notably LB-293, 296, and 317, but in other holes the trend is less clear or absent. The lack of an obvious upward decrease in labile minerals and the lack of a conventional weathering profile are supported by other mineralogic data and are discussed in the section entitled "Weathering Processes and Patterns."

Figures 14 and 15 show the distribution of hornblende and epidote, respectively, in dune sands. The values expressed are average abundances, computed in the same manner as for the fluvial intervals. The dunes on older terraces and on the uplands were found to have less hornblende and epidote than the dunes nearer the modern Cape Fear River.

In samples of dune sand rich in humate, labile heavy minerals, feldspar, and clay minerals are absent or less common (e.g., LB-297, 5 and 10 feet; LB-321, 3 and 10 feet) than in samples above and below the humate zone. Leaching and destruction of minerals in peat zones is minimal by comparison (e.g., LB-322, 14 and 30 feet; LB-292, 7 feet) because the organics are not in the soluble and colloidal state commonly found in humate. Leaching of labile minerals by the organic acids of humate produces a sample that mineralogically appears more weathered and, therefore, older than samples that have not been impregnated with humate. In figures 14 and 15, the values exclude humate samples and are more reliable than if humate- impregnated samples had been included.

In the study area, ilmenite, weathered (brown) ilmenite, and leucoxene constitute the detrital opaque heavy mineral suite. Authigenic pyrite is common in some places in the Cretaceous formations and a minor constituent in the base of the overlying fluvial deposit, but is disregarded here because it is not a detrital mineral. Although ilmenite is susceptible to weathering, the absolute quantity of opaque heavy minerals is thought to remain somewhat constant during weathering owing to the stability of leucoxene, the weathering product of ilmenite. In contrast, the absolute quantity of nonopaque heavy minerals decreases during weathering, as the labile species are destroyed. The ratio of opaque to nonopaque grains therefore increases with time and should generally reflect the age of the sediment. Beneath the Penholoway and younger terraces the opaque/nonopaque ratio is less than 1, while beneath older terraces and on the uplands the ratio exceeds 1 and is generally more than 1.5. Hornblende and epidote are far more abundant in Penholoway and younger sediments, and their destruction due to weathering is largely responsible for the difference in ratios. The high ratios in all upland deposits are due at least in part to the reworking that occurred in those marginal marine sediments.

Although the effect of weathering on mineral distribution has been stressed here, minor variations over time in the mineralogy of sediment supplied to the valley are expected. For example, sediments beneath the Penholoway and younger terraces are distinctly richer in hornblende and nonopaque heavy minerals in general than sediments beneath the Waccamaw and Bear Bluff terraces. Although much of the difference is due to weathering, the contrast between Waccamaw and Penholoway heavy mineralogy suggests that the mineralogy of the detrital sediments may have varied with time. While this does not diminish the validity of weathering intensity as a relative dating tool, the limitations of this method must be understood.

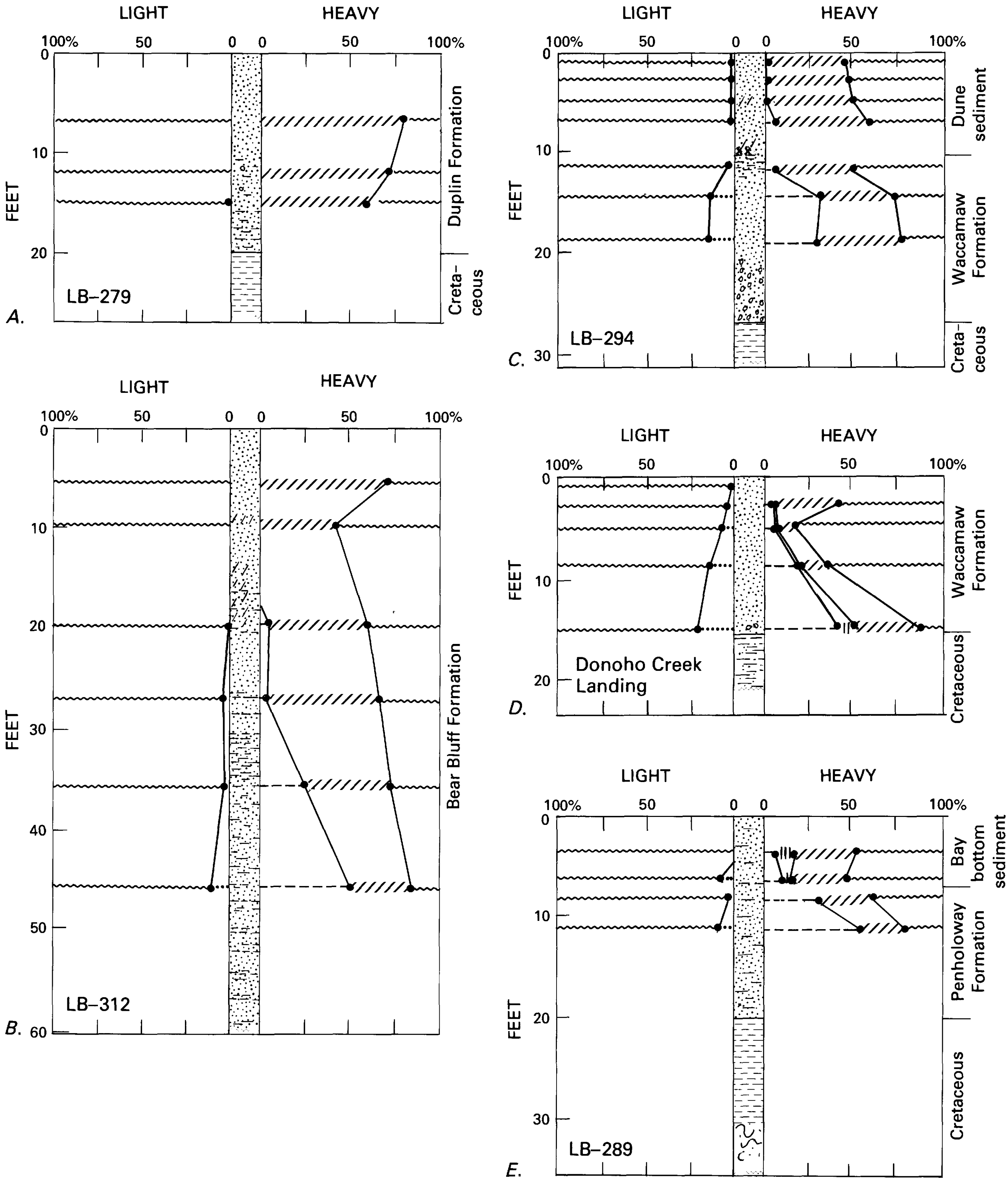
"LIGHT" SAND-SIZED FRACTION

Monocrystalline quartz accounts for the bulk of all the fluvial and upland sediments, with the remainder (generally less than 25 percent) being polycrystalline quartz and feldspar. Quartz grains are moderately spherical, and angular to subrounded. Muscovite is a common accessory mineral. Feldspar is dominantly untwinned, with lesser amounts of microcline and twinned plagioclase. Index oils and x-ray fluorescence were used to identify the twinned plagioclase as albite and oligoclase. In the upper portion of some weathering profiles, clay galls (presumably weathered feldspars, at least in part) and iron oxide coatings and aggregates are common.

The species of untwinned feldspar was determined by the x- ray mapping technique and x-ray fluorescence capability (EDAX) of an ETEC AUTOSCAN scanning electron microscope. This technique was used on a few samples near the base of the fluvial sections, where weathering was minimal and feldspar most abundant. Potassium-rich sand grains were found routinely during the x-ray mapping; EDAX analysis of individual grains indicated a composition appropriate to potassium feldspar. Scanning electron microscopy did not reveal twinning; these common, untwinned species were therefore assumed to be orthoclase. X-ray mapping of sodium atoms revealed a few sodium- rich grains in the samples. The few sodium-rich grains located were determined (by EDAX analysis) to be plagioclase feldspars of approximately oligoclase composition, and they showed twinning. This limited analysis (of three samples from the base of the fluvial intervals) suggests that orthoclase is the only variety of untwinned feldspar in these sediments.

Drill-hole data can be divided into groups on the basis of feldspar content, the grouping being conceptually the same as for the heavy minerals. The first group, all the upland deposits, has little or no feldspar (fig. 10). Feldspar is absent from the upper 20 feet at LB-279 and 312, which sample Pliocene deposits. LB-294, 289, 290, 319, and the outcrop at Donoho Creek Landing sample younger upland deposits; feldspar persists much higher in the section here, although it does decrease in abundance upwards.

The second group consists of the fluvial deposits; the samples from these drill holes do not show a noticeable upward decrease in feldspar content (fig. 11). The average feldspar content for each drill hole is shown in figure 16. There is a somewhat irregular crossvalley variation in feldspar abundance; the contrast between


LIGHT
HEAVY
100%
50
0
0
50
100%
FEET
10
20
Duplin Formation
Creta-
ceous
LB-279
A.
Dune
sediment
Waccamaw
Formation
Creta-
ceous
LB-294
C.
30
LB-312
B.
40
50
60
Bear Bluff Formation
Donoho Creek
Landing
D.
Cretaceous
Bay
bottom
sediment
Penholoway
Formation
Cretaceous
LB-289
E.

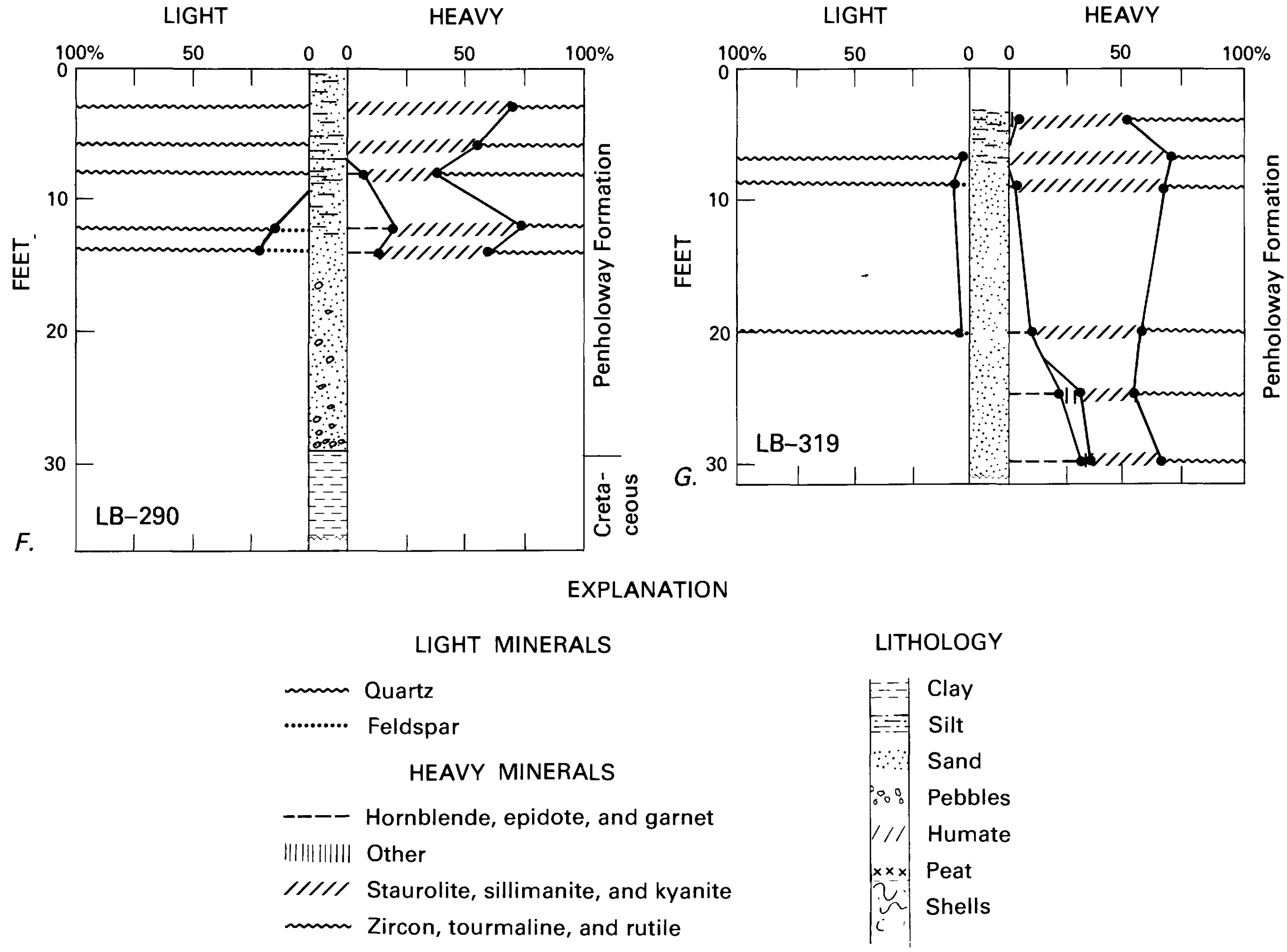


FIGURE 10—Continued.

old and young terraces is not as striking as with hornblende abundance. A downvalley decrease in feldspar is more apparent and indicates that feldspar is derived largely from the Piedmont; feldspar-rich sediments of the Cape Fear River were diluted downvalley by the contribution of reworked, feldspar-depleted sediments from Coastal Plain-draining tributaries.

FIGURE 10.—Comparison of lithology and sand mineralogy of auger hole samples of upland deposits. On the left side of the diagrams, the proportions of the various light minerals in the fine and very fine sand fraction are displayed, and on the right side are shown the proportions of the various nonopaque heavy minerals of the same size fraction. The least resistant minerals are plotted closest to the 0-percent lines; note the increase in abundance of these mineral species with depth and their absence in the upper 20 feet of the older deposits (Duplin (DuBar and others, 1974) and Bear Bluff Formations). The lithology of the auger hole sediments is shown at the center of each diagram. Along the right edge of the diagrams, the depth to the Cretaceous sediment is shown and the overlying units are identified.

In the dune sands, feldspar abundance varies across the valley (fig. 17). Feldspar is sparse or absent in dune sands on the Bear Bluff and Waccamaw terraces, from the uplands to the east of the lower valley (LB-294) and from the thick accumulation of dune sand on the Socastee fluvial or backbarrier surface that lies east of the Cape Fear River and north of Wilmington (LB-321). Humate is well developed at LB-321 and may be responsible for the absence of feldspar at this location. The feldspar content of dune sands increases as the Cape Fear River is approached from the northeast: 8 to 14 percent feldspar on the Penholoway terrace, 4 to 15 percent on the Socastee terrace, and 21 to 35 percent on the Wando terrace.

It is apparent that dune sands were derived from unweathered fluvial sediments to the west of their current position. The Cape Fear River channel is the

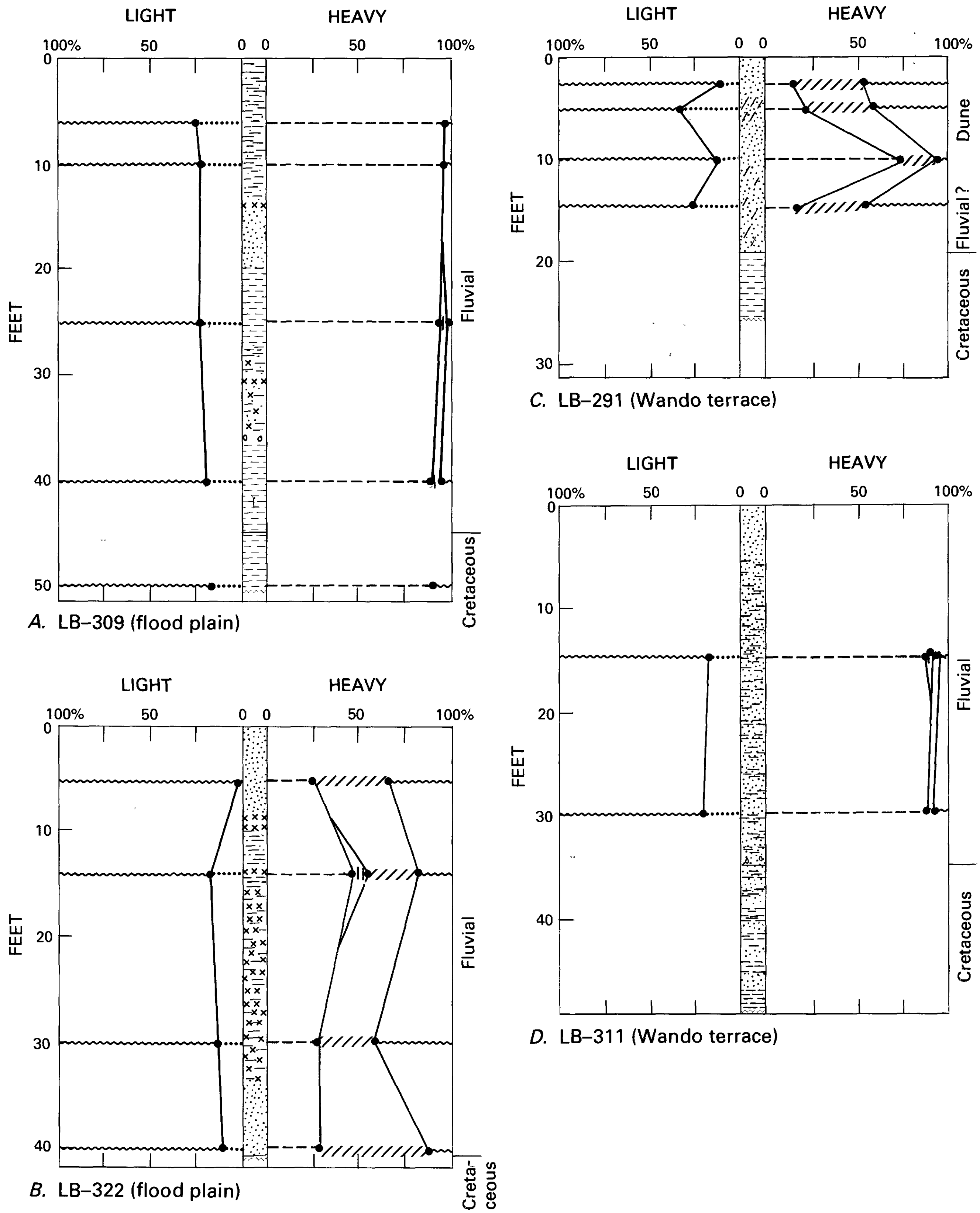
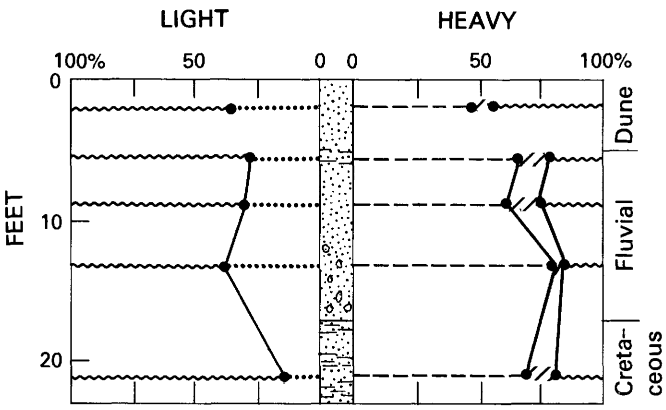
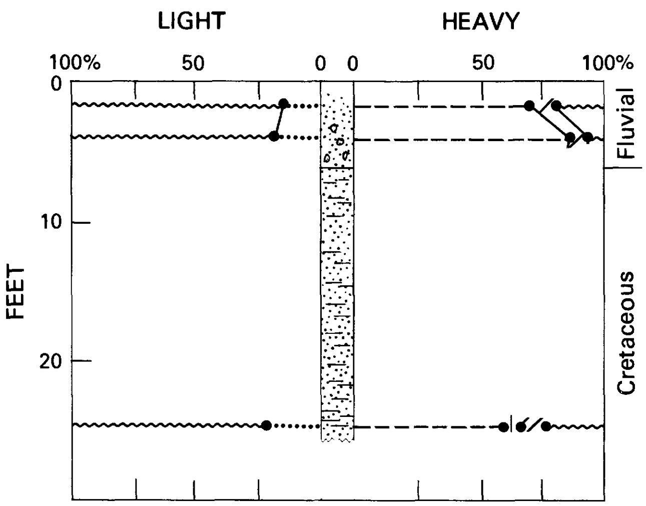


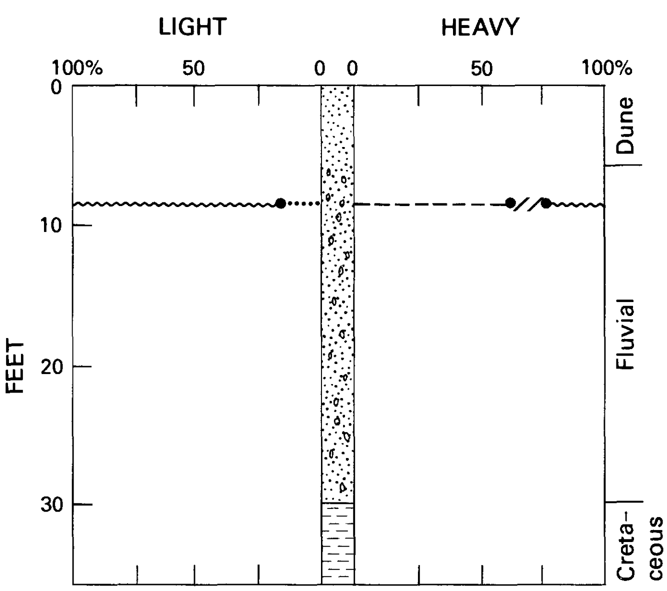
FIGURE 11.—Comparison of lithology and sand mineralogy of auger hole samples of valley deposits. On the left side of the diagrams, the proportions of the various light minerals in the fine and very fine sand fraction are displayed, and on the right side are shown the proportions of the various nonopaque heavy minerals of the same size fraction. The least resistant minerals are plotted closest to the 0-percent lines. The lithology of the auger hole sediments is shown at the center. Along the right edge of each diagram, the depth to the Cretaceous sediment at the floor of the valley and the fluvial and dune portions of the valley sediments are shown.



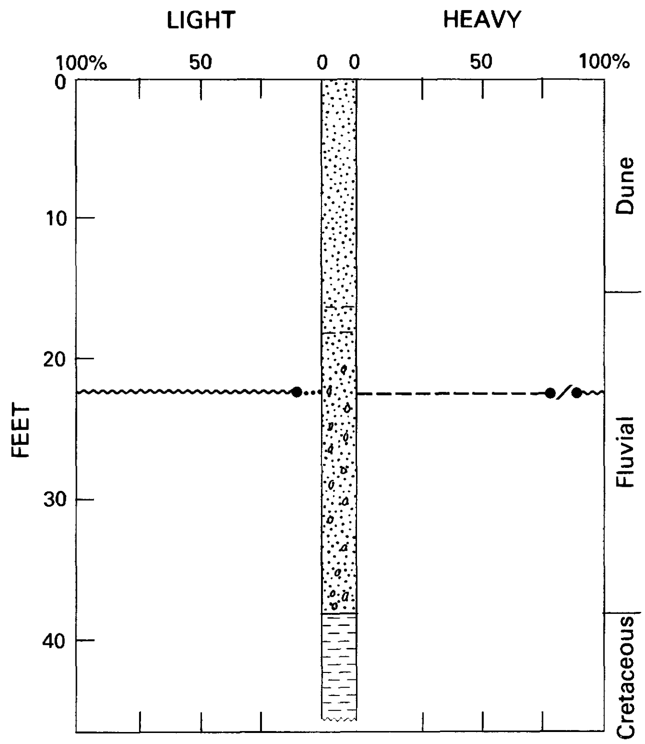
E. LB–380 (Wando terrace)



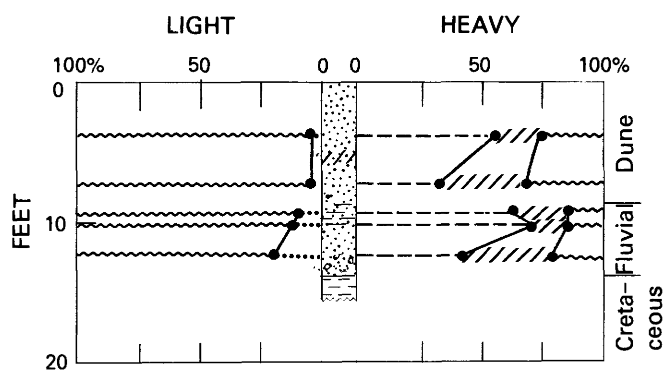
F. LB–386 (Wando terrace)



G. LB–273 (Socastee terrace)



H. LB–274 (Socastee terrace)



I. LB–292 (Socastee terrace)

EXPLANATION

LIGHT MINERALS

- Quartz
- Feldspar

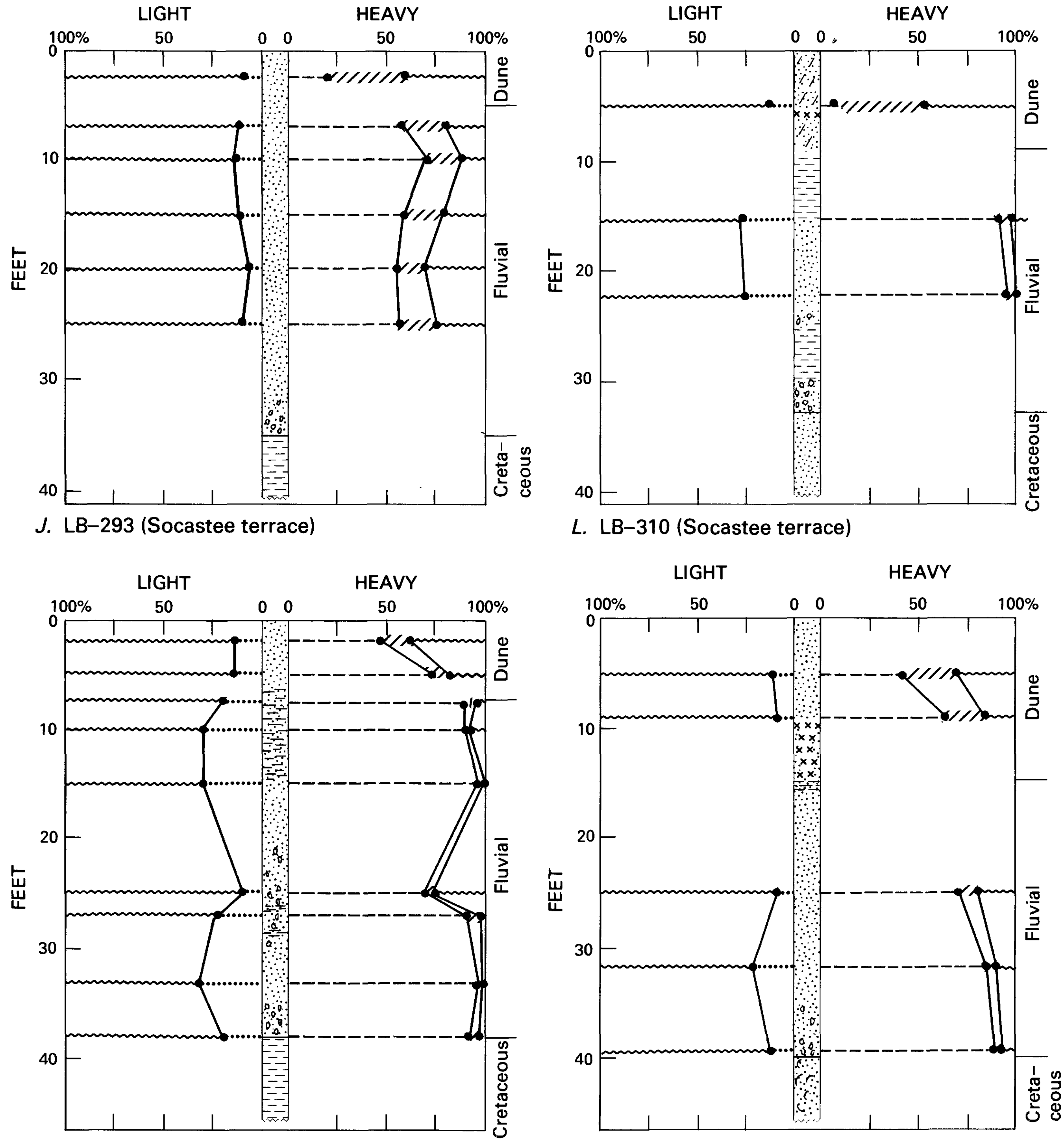
HEAVY MINERALS

- Hornblende, epidote, and garnet
- Other
- Staurolite, sillimanite, and kyanite
- Zircon, tourmaline, and rutile

LITHOLOGY

- Clay
- Silt
- Sand
- Pebbles
- Humate
- Peat
- Shells

FIGURE 11—Continued.



J. LB–293 (Socastee terrace)

L. LB–310 (Socastee terrace)

K. LB–295 (Socastee terrace)

M. LB–317 (Socastee terrace)

FIGURE 11—Continued.

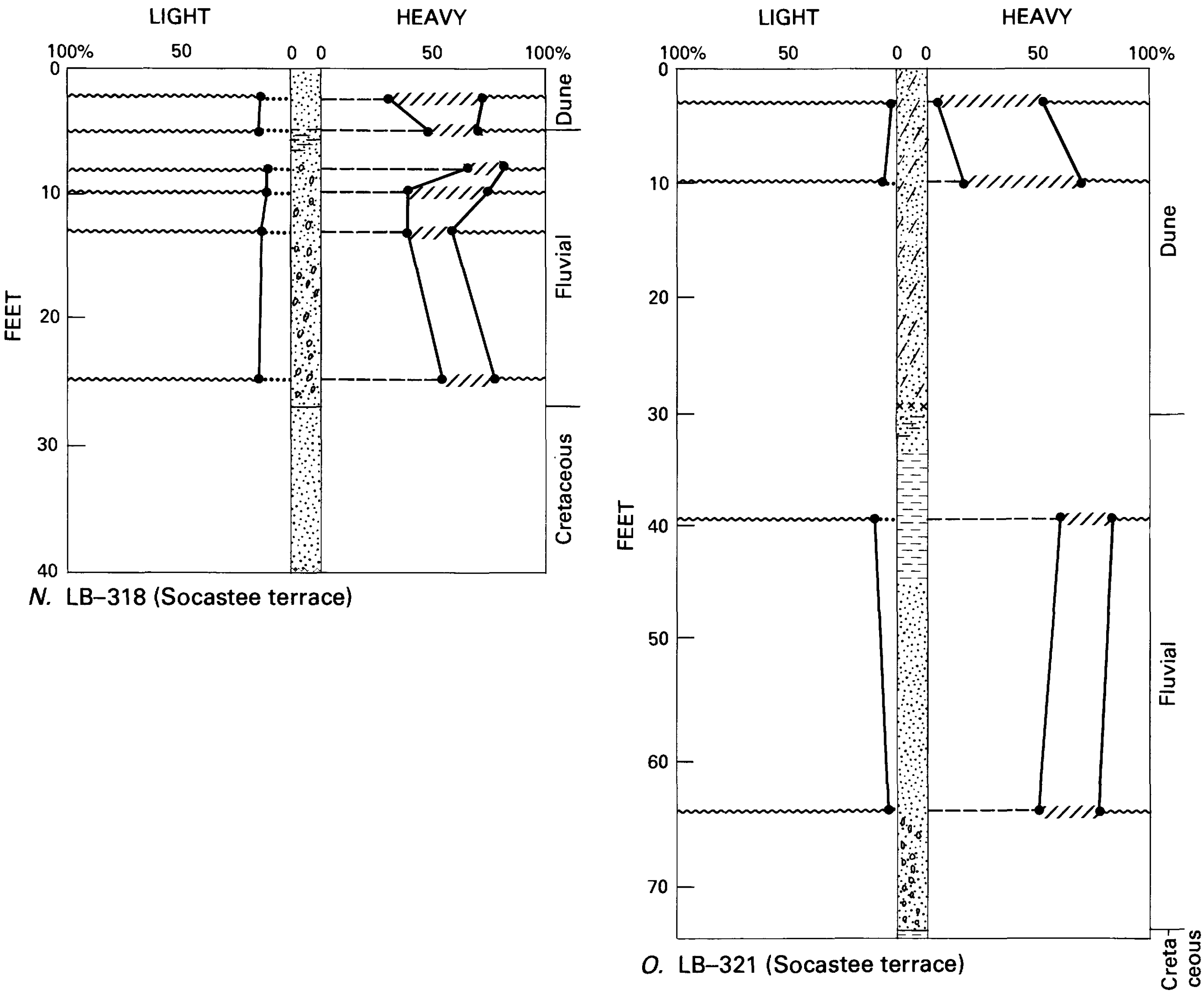


FIGURE 11—Continued.

likely source of the sand; the exposed channel was eroded during sea level lowstand and the sand was transported a short distance to cover the adjacent terraces. Thom (1967) proposed this mechanism for the Pee Dee River valley. In the Cape Fear River valley, the theory is supported by two lines of evidence. First, the orientation of dunes and Carolina bays records a southwesterly wind direction. Second, the abundance of feldspar in dunes near the Cape Fear River is quite similar to that in the younger pre-Holocene (Wando) fluvial sediments. Given this evidence, it is suggested that the decrease in feldspar abundance to the east of the river reflects at least two generations of dunes, with the older dunes on the Bear Bluff and Waccamaw terraces lacking in feldspar owing to weathering. Originally containing as much feldspar as the unweathered fluvial sediment from which they were derived, the dunes on the older terraces were exposed to weathering and most of the feldspar was destroyed. The underlying fluvial sediments, of somewhat greater age, were not as severely weathered because their protective cap of finer grained, overbank sediment inhibits unrestricted vertical movement of water.

CLAY-SIZED FRACTION

In an undisturbed weathering profile in the southern Atlantic Coastal Plain, labile clay and sand-sized minerals are systematically transformed to a weathered, mature suite of clay- sized minerals. Vermiculite, kaolinite, and gibbsite tend to dominate in the upper soil horizons of deeply weathered profiles and represent the mature assemblage for the Cape Fear area. From

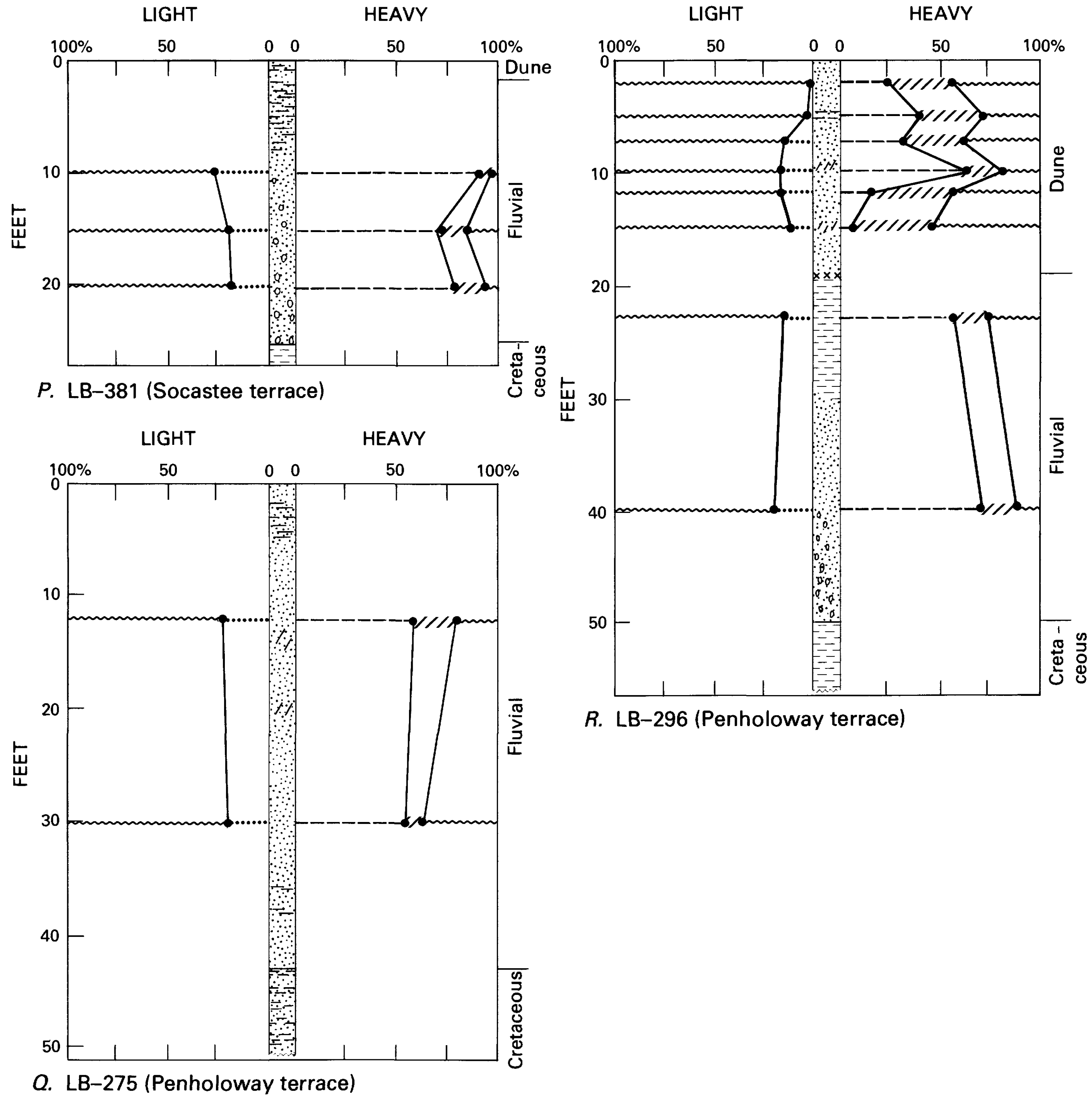
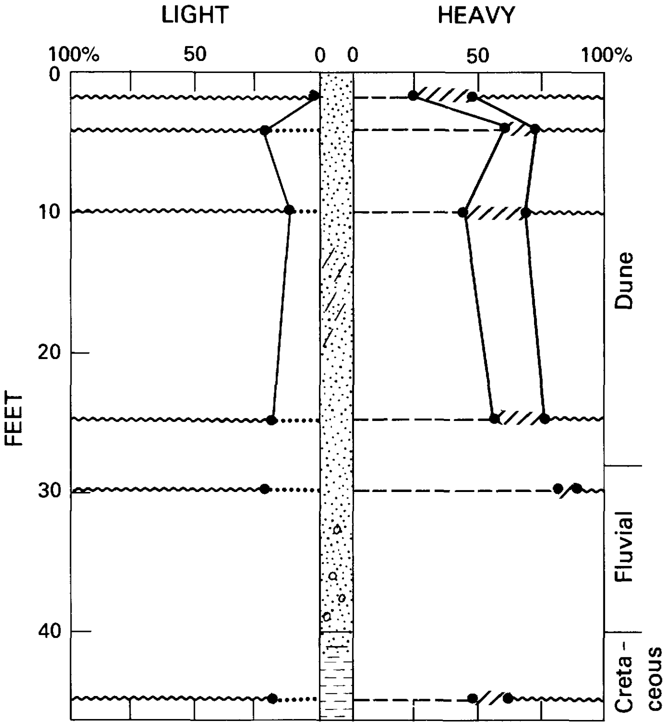
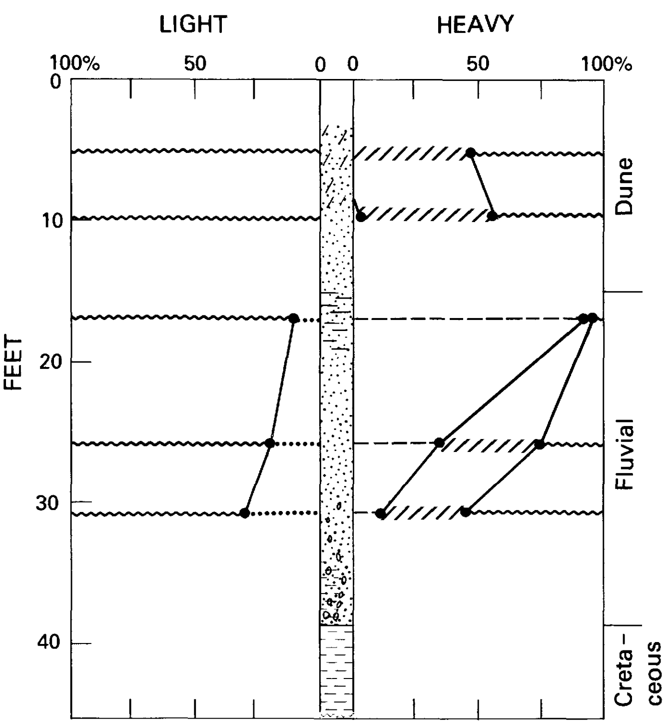


FIGURE 11—Continued.

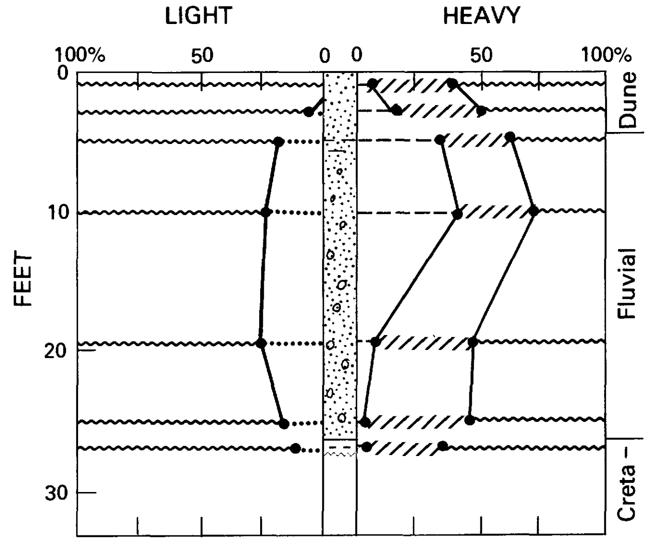
the uplands surrounding the Cape Fear River valley, clay mineralogic profiles from six drill holes and one outcrop are presented (fig. 18). LB-312 (fig. 18*A*) sampled shallow marine sands of the Bear Bluff Formation (Pliocene). Near the base of section, there occur unweathered examples of two facies having different mineralogies. At a depth of 46 feet, in a gray medium-to-fine sand, unaltered kaolinite dominates the clay suite, with minor mixed-layered material (identified by a slightly raised background around 6° 2-theta). At 36 and 33 feet, in gray silty sand and gray clay, illite is abundant, with kaolinite and minor mixed-layered material. Illite is absent from the sand and abundant in the fine sediment, which suggests a detrital, facies-controlled origin for this mineral. At a depth of 27.5 feet, the sediment is mottled and the clay



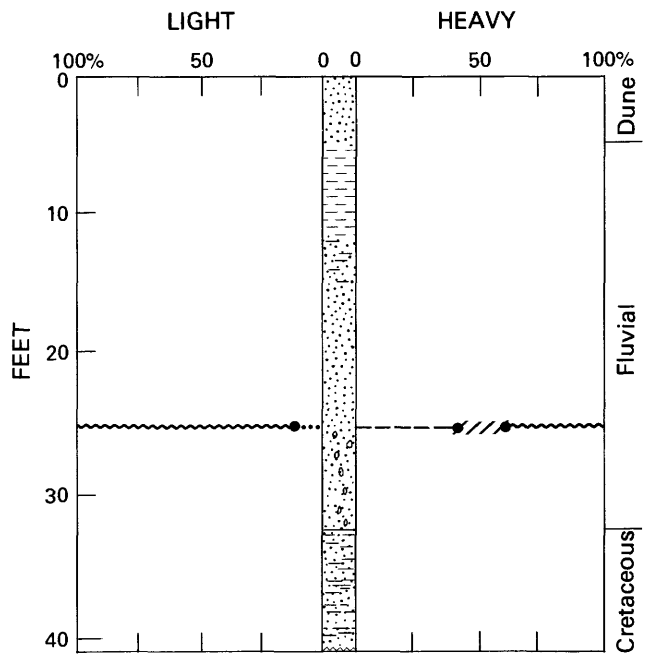
S. LB–382 (Penholoway terrace)



T. LB–297 (Waccamaw terrace)



U. LB–383 (Waccamaw terrace)



V. LB–278 (Bear Bluff terrace)

FIGURE 11—Continued.

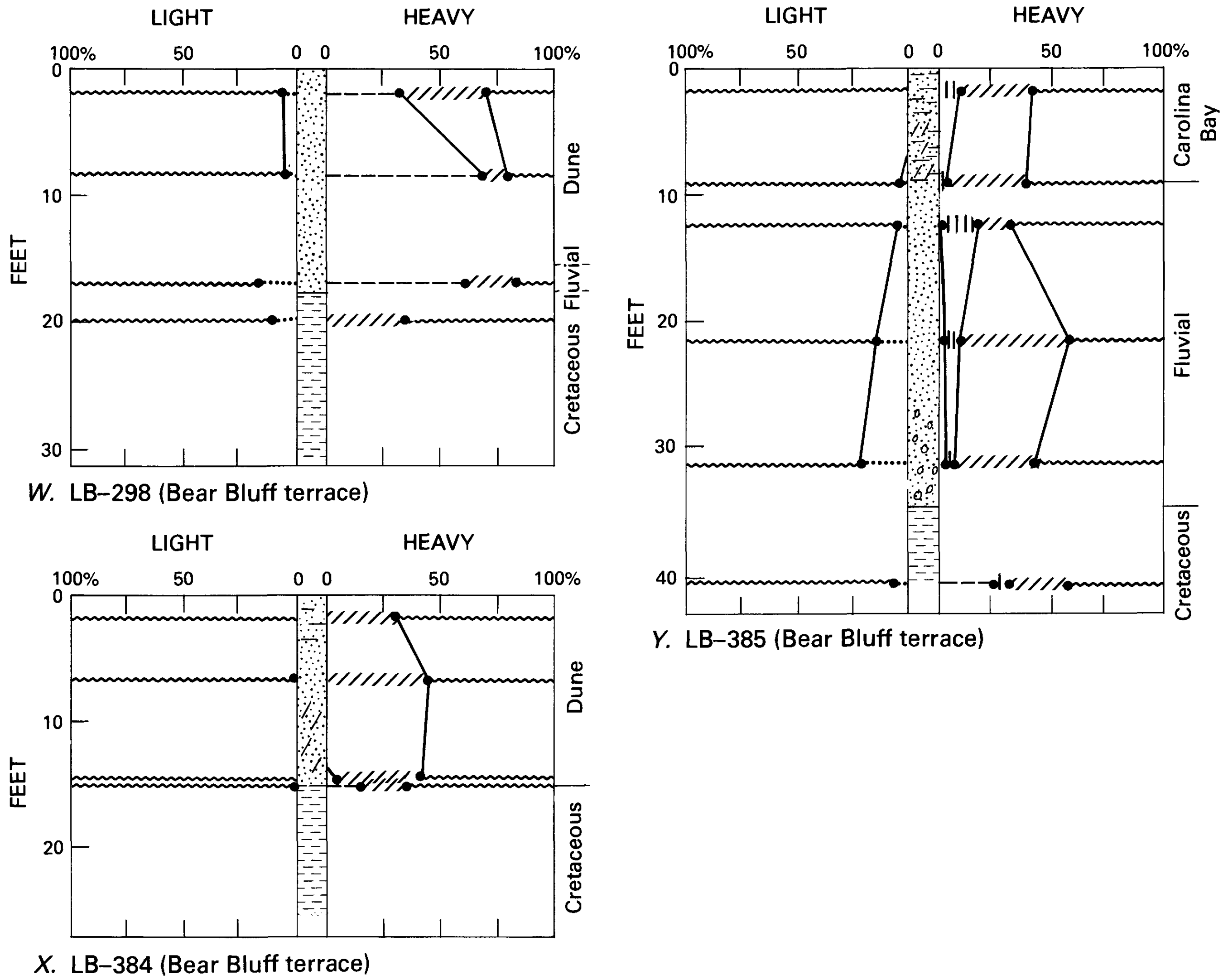


FIGURE 11—Continued.

mineral suite shows some incipient weathering. Disordered kaolinite and halloysite appear to dominate, with minor illite and lepidocrocite.

In the sample at LB-312, 27.5 feet, an oriented clay mount could not be achieved; the 20° 2-theta peak attributed to halloysite is probably a nonbasal kaolinite reflection that is detected in poorly oriented mounts. Poor orientation of clay mounts is likely the cause for detection of the 20° peak in other samples as well, although some well-oriented mounts (e.g., LB-385) also show a 20° peak. The 20° peak is therefore attributed to weathered and possibly disordered kaolinite, and the related mineral halloysite.

Upward, in the section at LB-312, halloysite and lepidocrocite are absent, but disordered kaolinite persists. At 22 and 20 feet, kaolinite dominates with lesser expandable mixed-layering (determined by ethylene glycol treatment) which is poorly ordered. The uppermost sample (6 feet) is dominated by vermiculite, with kaolinite and gibbsite. This drill hole documents a gradual transformation of the unstable clay minerals into a more stable weathered assemblage.

At LB-312, 15 feet, the sample is impregnated with humate. Clay minerals, abundant in all samples lacking humate, are not detected here. The organic acids of the humate must have degraded the clay minerals as well as the labile sand-sized mineral grains of this sample.

Pliocene sediments are also sampled at LB-279, but the mineralogy is much different (fig. 18 *B*). All Pliocene samples in the drill hole (15 feet and above) are orange-red clayey sands, and their mineralogy is constant: kaolinite with minor mixed-layering. Ver-

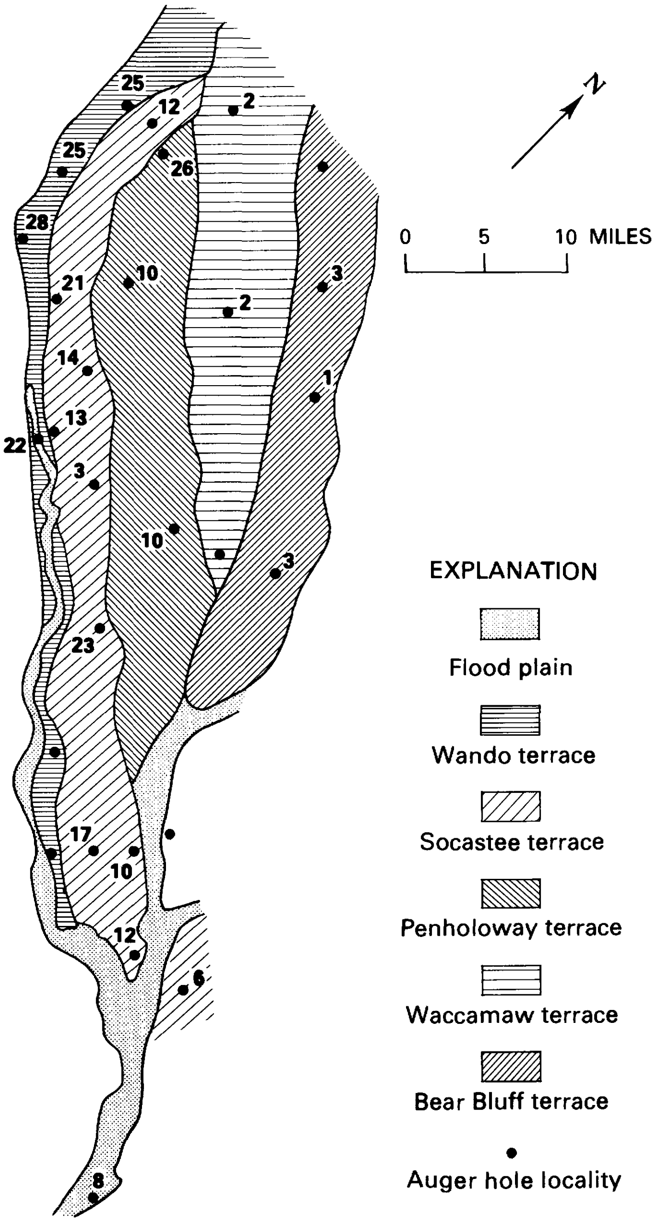


FIGURE 12.—Average percent hornblende in the nonopaque heavy mineral fraction in the fluvial sediments, plotted on a simplified geologic base map of the Cape Fear River valley. For a more detailed map, refer to plate 1. An average value was computed for each auger hole, as shown here. At auger holes without a value, fluvial deposits either were absent or were not sampled. Note the low abundance of hornblende on the Bear Bluff and Waccamaw terraces and the progressive increase in abundance, along any northeast to southwest transect, on younger terraces. There is also a somewhat vague trend on at least two surfaces of a downvalley decrease in hornblende.

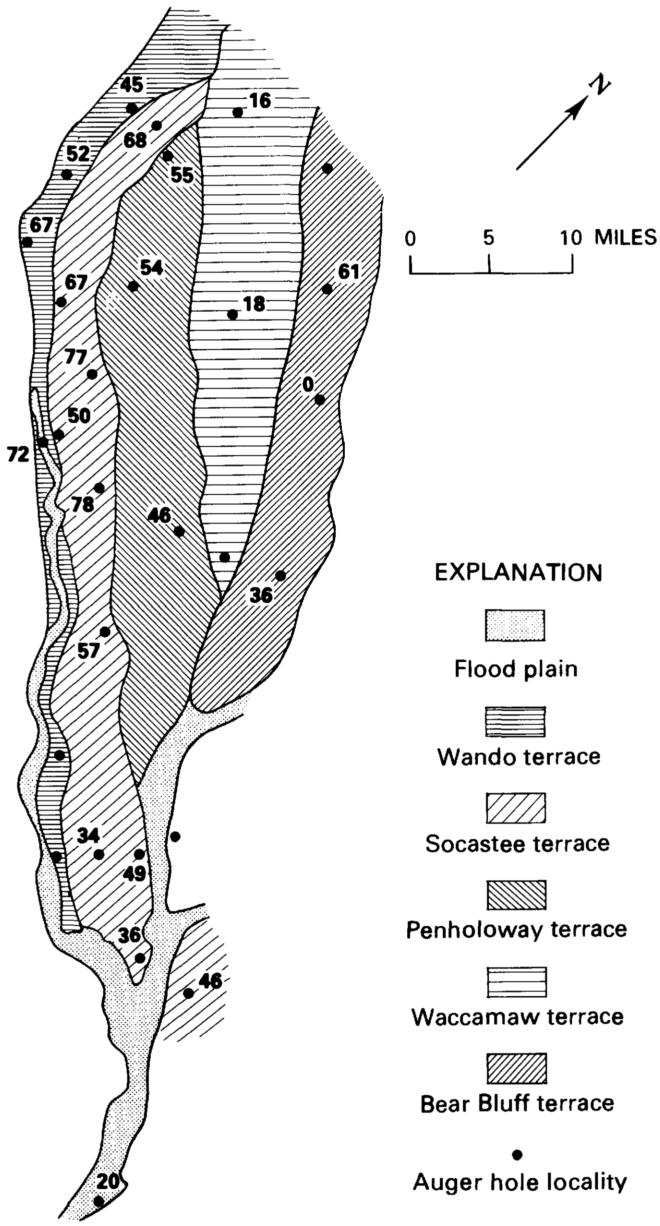


FIGURE 13.—Average percent epidote in the nonopaque heavy mineral fraction in the fluvial sediments, plotted on a simplified geologic base map of the Cape Fear River valley. For a more detailed map, refer to plate 1. An average value was computed for each auger hole, as shown here. At auger holes without a value, fluvial deposits either were absent or were not sampled. Note the general increase in epidote to the southwest, onto younger terraces, and the decrease downvalley on most terraces.

miculite and gibbsite do not occur in the upper profile as they do at LB-312. The orange color indicates that iron oxides have been generated at the expense of detrital minerals, yet the clay mineralogy does not reflect the obviously weathered character of the profile. The upper portion of this weathered deposit has apparently been eroded; this possibility is further discussed in the section on "Weathering Processes and

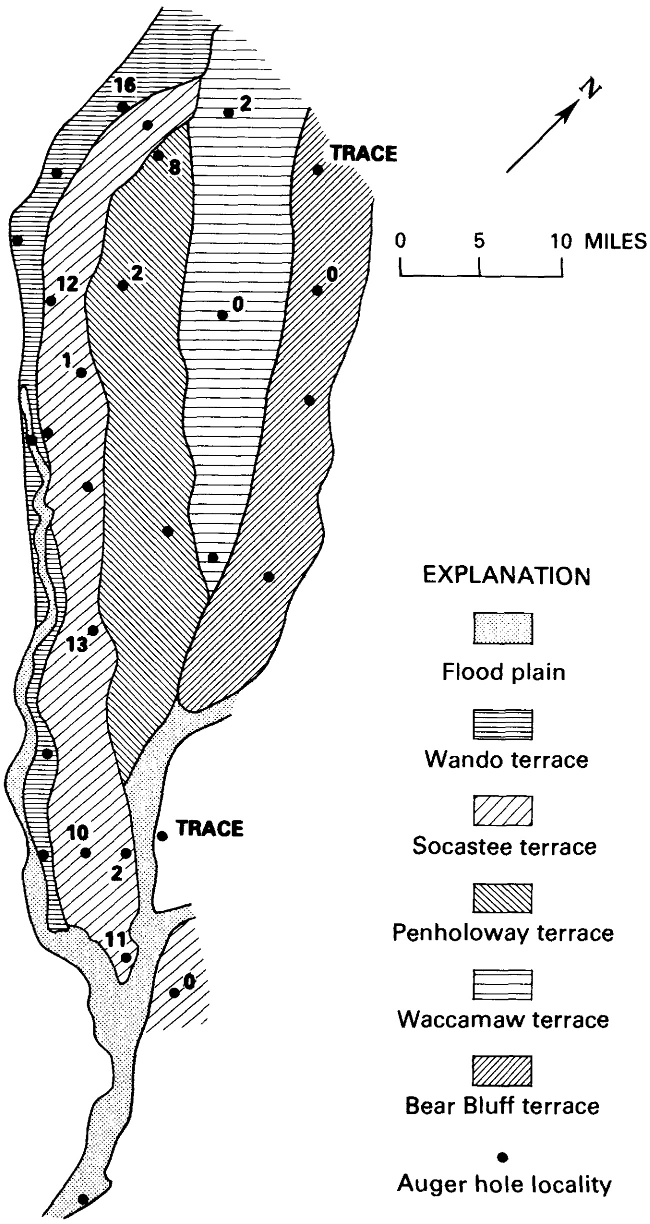


FIGURE 14.—Average percent hornblende in the nonopaque heavy mineral fraction in dune sands, plotted on a simplified geologic base map of the Cape Fear River valley. For a more detailed map, refer to plate 1. An average value was computed for each auger hole, as shown here. At auger holes without a value, fluvial deposits either were absent or were not sampled. Note the increase in hornblende abundance in dune sediments toward the Cape Fear River.

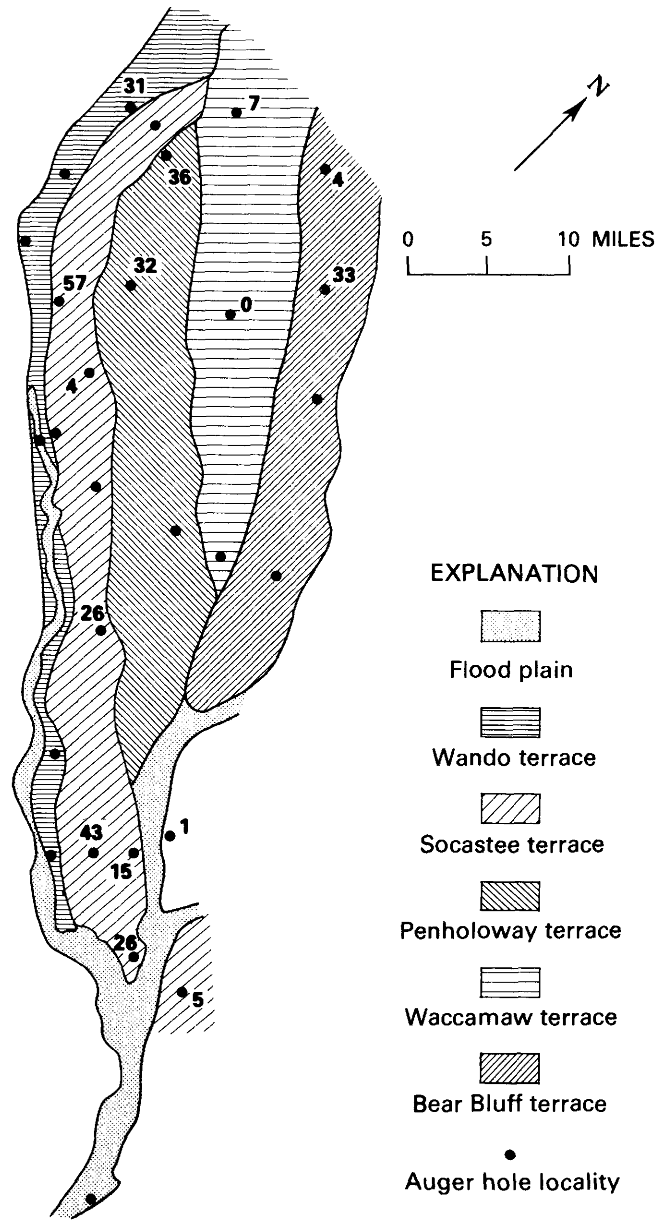


FIGURE 15.—Average percent epidote in the nonopaque heavy mineral fraction in dune sands, plotted on a simplified geologic base map of the Cape Fear River valley. For a more detailed map, refer to plate 1. An average value was computed for each auger hole, as shown here. At auger holes without a value, fluvial deposits either were absent or were not sampled. Note the general tendency for higher abundance of epidote in holes near the Cape Fear River.

Patterns." The lowest sample at LB- 279 (25 feet) is an unweathered, dark-gray clayey sand, Cretaceous in age. Smectite is the sole clay component in this sediment.

Five other profiles in the uplands are presented, LB-290, 294, 319, 289, and Donoho Creek Landing (fig. 18*C* through 18*G*); these sediments are Waccamaw to Penholoway in age. The mineralogy in these profiles generally conforms to the pattern at LB-312. At LB-290 and 319, goethite and lepidocrocite have precipitated, indicating the degradation of ferrous minerals in the weathering profile. At LB-294, weathering occurs in the older sediments (below 12 feet), which are capped

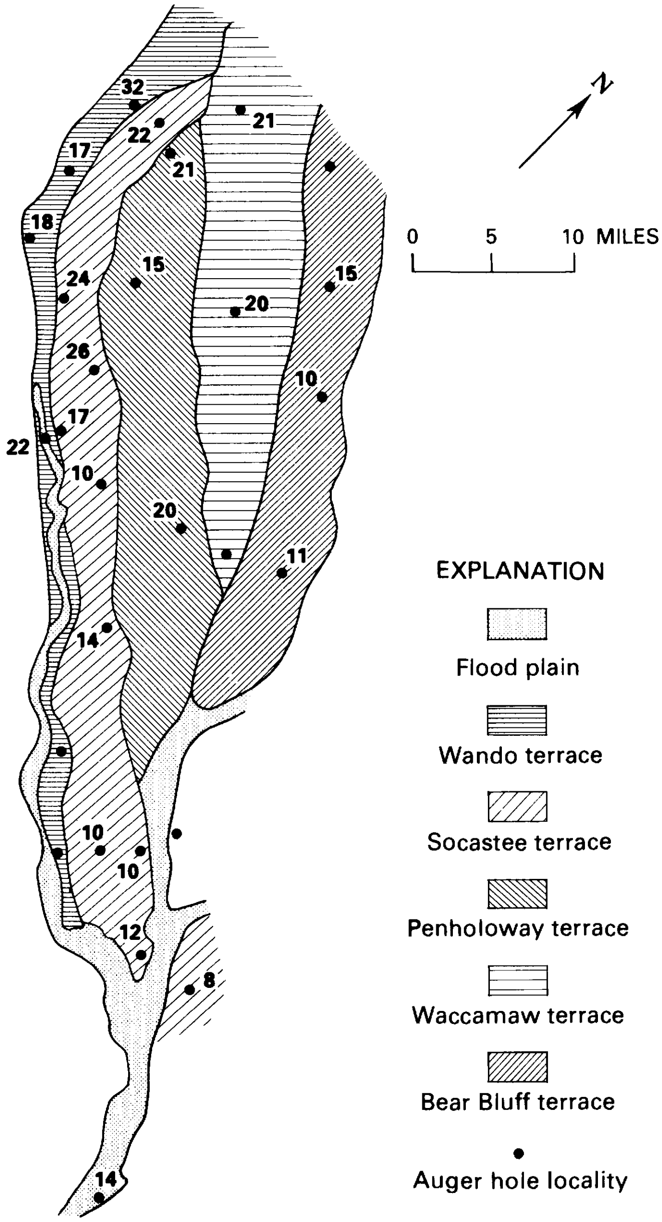


FIGURE 16.—Average percent feldspar in the light mineral fraction in the fluvial sediments, plotted on a simplified geologic base map of the Cape Fear River valley. For a more detailed map, refer to plate 1. An average value was computed for each auger hole, as shown here. At auger holes without a value, fluvial deposits either were absent or were not sampled. Note that the trend showing an increase in abundance nearer the Cape Fear River is less apparent than the trend in hornblende and epidote (figs. 12, 13). A downvalley increase in feldspar abundance on the terraces is more distinct.

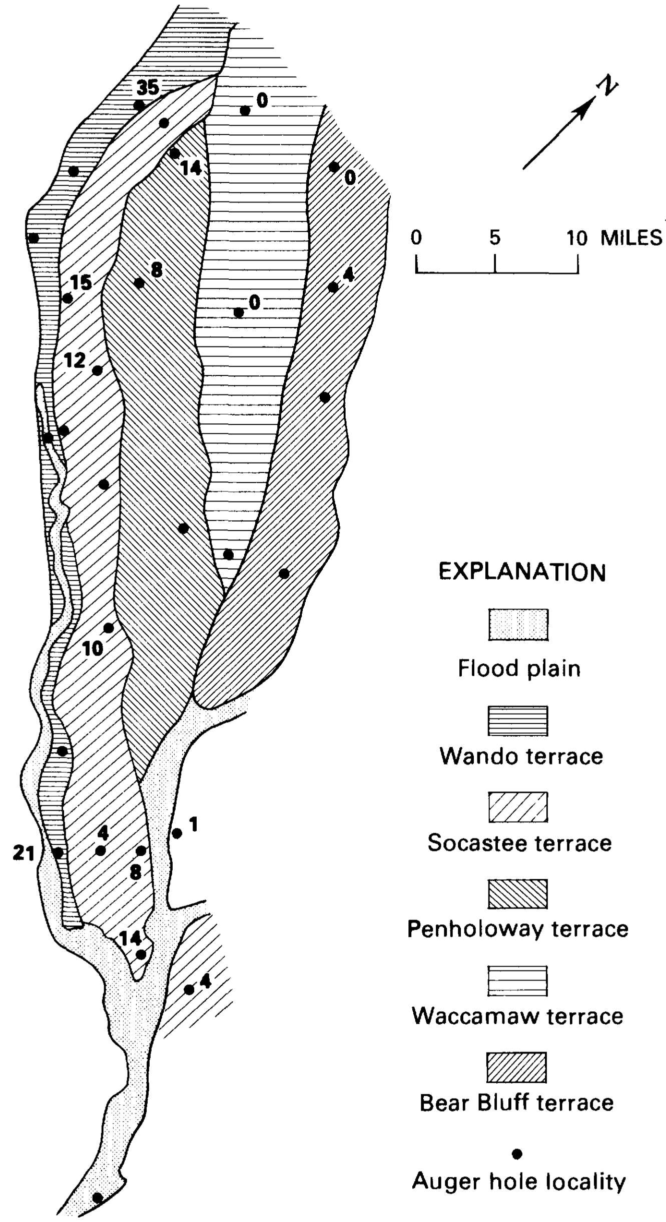


FIGURE 17.—Average percent feldspar in the light mineral fraction in dune sands, plotted on a simplified geologic base map of the Cape Fear River valley. For a more detailed map, refer to plate 1. An average value was computed for each auger hole, as shown here. At auger holes without a value, fluvial deposits either were absent or were not sampled. Note the increase in feldspar abundance in dune sediment along any transect across the valley, toward the Cape Fear River.

by a much younger dune sequence. Selected profiles are discussed in the section on "Weathering Processes and Patterns."

In the fluvial intervals of drill holes in the valley (fig. 19) the most prevalent clay suite is kaolinite and expandable mixed-layer material. The mixed-layer material is random and is composed of illite and smectite, with a poorly developed peak at 15 A and broad shoulders. The mixed-layer clay tends to become more vermiculitic upward in the weathering profile (i.e., the layering becomes less random and the mineral loses its ability to expand after glycolation). There is no

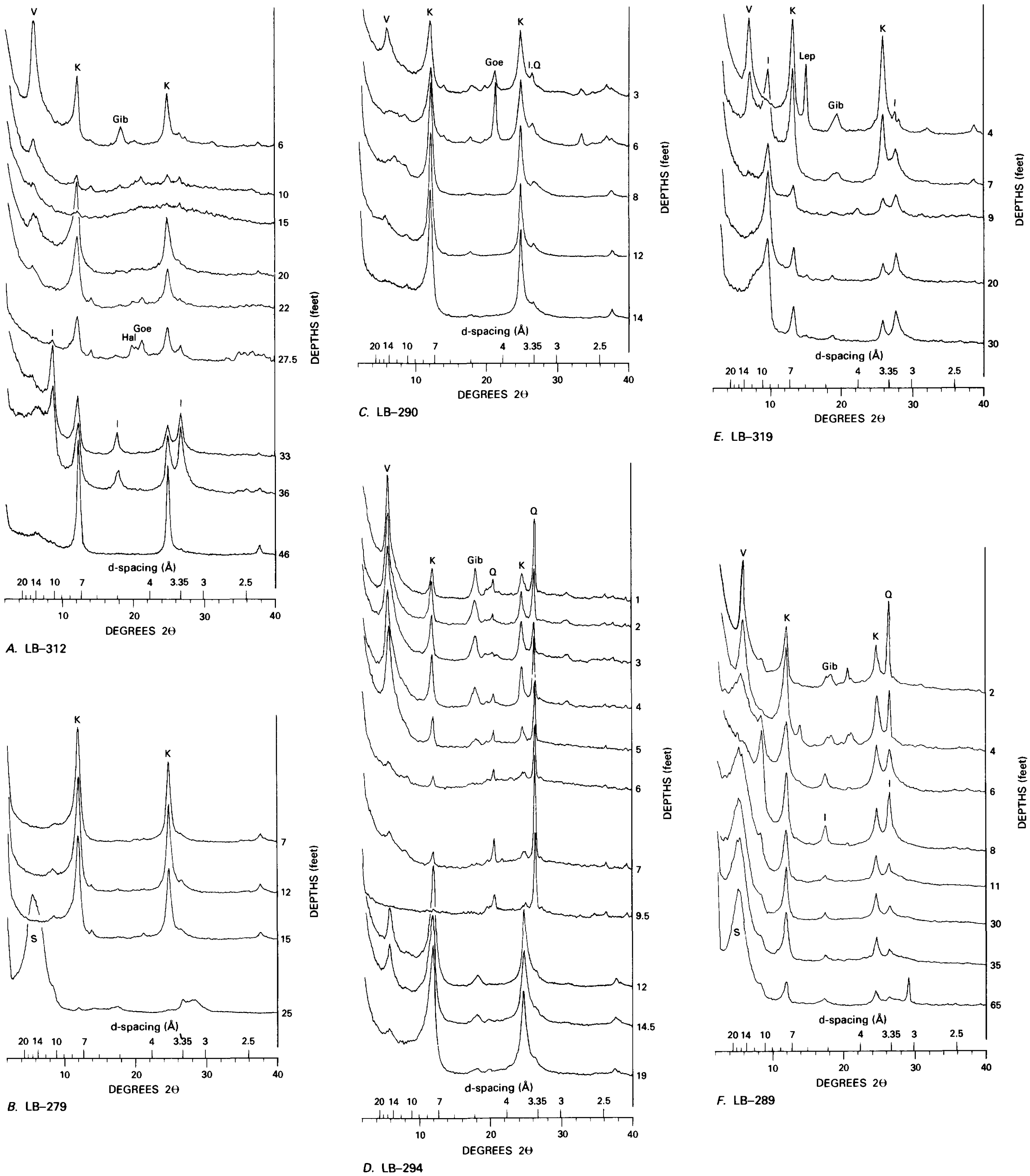
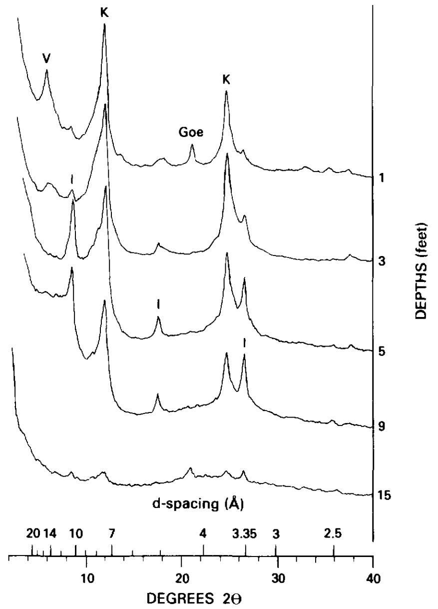
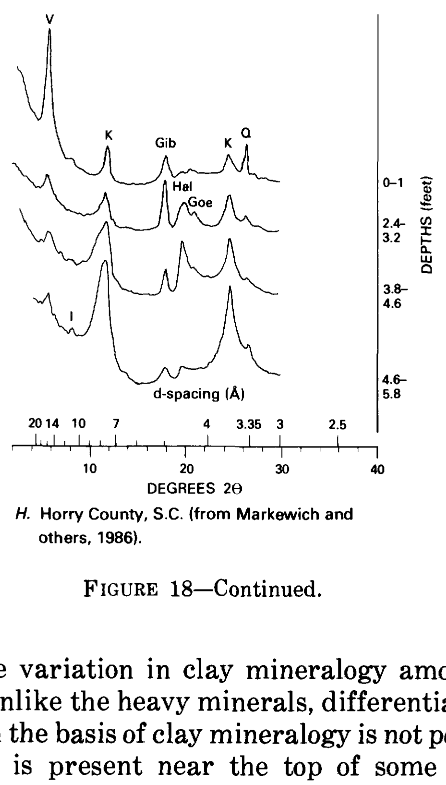


FIGURE 18.—X-ray diffraction traces of the untreated clay-sized fraction in samples of upland deposits: *A*, Bear Bluff Formation; *B*, Duplin Formation (DuBar and others, 1974); *C*, Penholoway Formation; *D*, Waccamaw Formation; *E* and *F*, Penholoway Formation; *G*, Waccamaw Formation; *H*, Bear Bluff Formation. Gib, gibbsite; Goe, goethite; Hal, halloysite or poorly oriented kaolinite; I, illite; S, smectite; K, kaolinite; Lep, lepidocrocite; Q, quartz; V, vermiculite.



G. Donoho Creek Landing



H. Horry County, S.C. (from Markewich and others, 1986).

FIGURE 18—Continued.

perceptible variation in clay mineralogy among the terraces; unlike the heavy minerals, differentiation of terraces on the basis of clay mineralogy is not possible.

Gibbsite is present near the top of some fluvial intervals. Gibbsite develops in sediments that have been extensively weathered (e.g., LB-312; fig. 18*A*); therefore, the older terraces should have more gibbsite than the younger terraces. This expected trend is not apparent in fluvial sediments in the study area. Gibbsite is not routinely detected, even in the oldest terrace sediments, but it is common in the flood plain. When gibbsite does appear in the fluvial sediments, it is usually in small quantities, and it may increase toward the top of the drill hole (as at LB-293, fig. 19*I*) or persist throughout (LB-309, fig. 19*A*; LB-322, fig. 19*B*). A possible trend is observed for gibbsite, but it is unrelated to the terraces: fluvial sediments in the lower valley (fig. 19, LB-309, 322, 292, 293, and 318) have some gibbsite, while those in the upper valley (fig. 19, LB-311, 295, 381, 296, 297, and 383) have none.

The clay-sized mineral suite of the dunes is composed of vermiculite, kaolinite, gibbsite, and quartz. These mature clay- sized minerals are usually the product of deep, protracted weathering, yet they occur in white, nearly clay-free dune sands which appear to be only slightly weathered. The maturity of these clays far exceeds that of the much older fluvial sediments beneath. The 14 A peak of vermiculite in the dune sediments is sharp, indicating good crystallinity; the crystallinity is much better than in the vermiculitic material in the underlying fluvial sediments. Gibbsite occurs in most of those samples and is generally abundant. Perhaps the most significant mineral in this dune suite is quartz; the detection of clay-sized quartz by x-ray diffraction appears to be diagnostic for dune sediments, because quartz occurs in all dune samples but is absent from all fluvial and upland samples.

Humate commonly develops near the base of dunes and tends to destroy all minerals but quartz (LB-294, 9.5 feet, fig. 18*D*, and LB-321, 3 and 10 feet, fig. 19*N* are extreme examples). Mere presence of organics is not, however, sufficient to leach clay minerals; clays are abundant and not severely altered at LB-322 (fig 19*B*), despite a 30- foot-thick peat interval.

WEATHERING PROCESSES AND PATTERNS

The weathering and degradation of minerals and the development of a weathering profile in a sediment are functions of time, the intensity of physical and chemical weathering (i.e., climate), and the lithology. Although weathering intensity within the study area may have varied in the past and occasionally may have been much less (i.e., during glacial maxima), it is reasonable to assume that the terrace sediments were usually exposed to a weathering intensity similar to that of the Holocene temperate climate. Even if this were not the case, climatic variations would induce a similar degree of weathering in all deposits existing at that time. Throughout the many climatic variations since deposition of the oldest unit in the area the mineral alterations induced by periods of warm, wet climate have been cumulative in the weathering profile; therefore, the oldest deposits should appear to be the most weathered. On the Wando, Socastee, and Penholoway terraces, samples were taken from peat zones that are present on

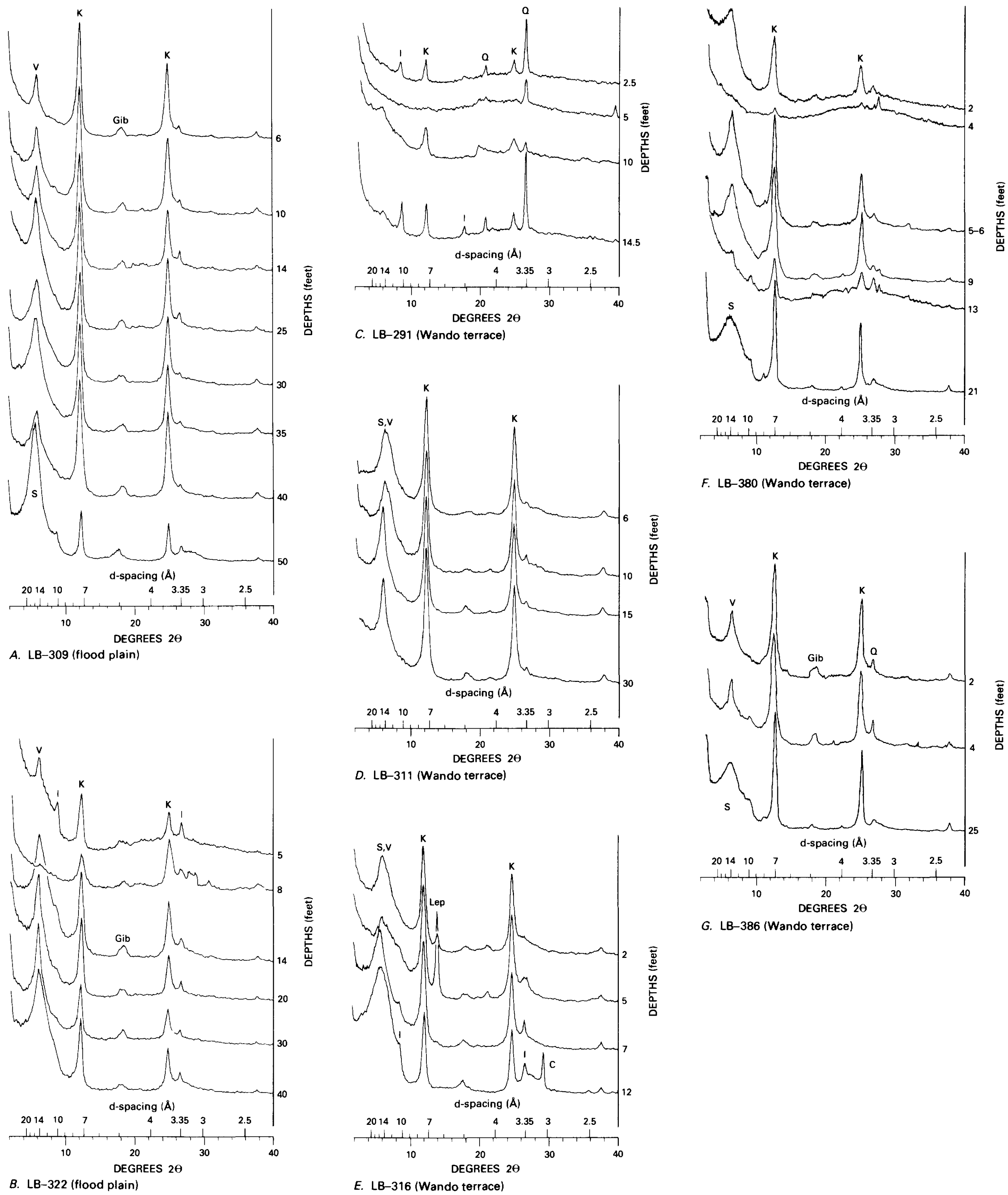
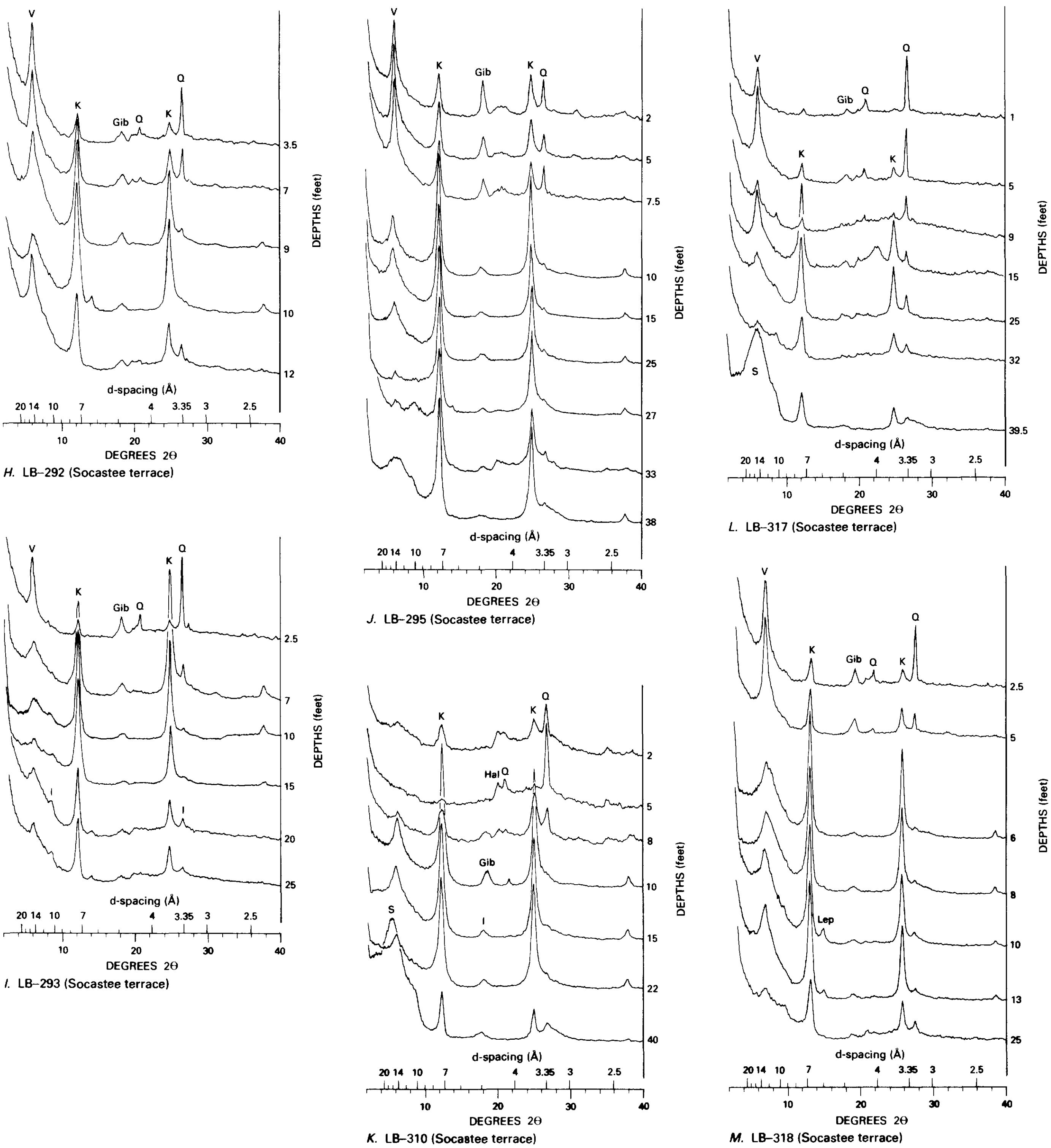


FIGURE 19.—X-ray diffraction traces of the untreated clay-sized fraction in samples of valley deposits. C, calcite; Gib, gibbsite; Hal, halloysite or poorly oriented kaolinite; I, illite; S, smectite; K, kaolinite; Lep, lepidocrocite; Q, quartz; V, vermiculite.



H. LB–292 (Socastee terrace)

I. LB–293 (Socastee terrace)

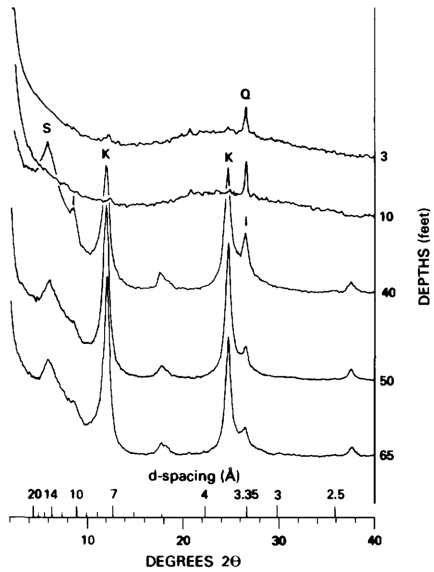
J. LB–295 (Socastee terrace)

K. LB–310 (Socastee terrace)

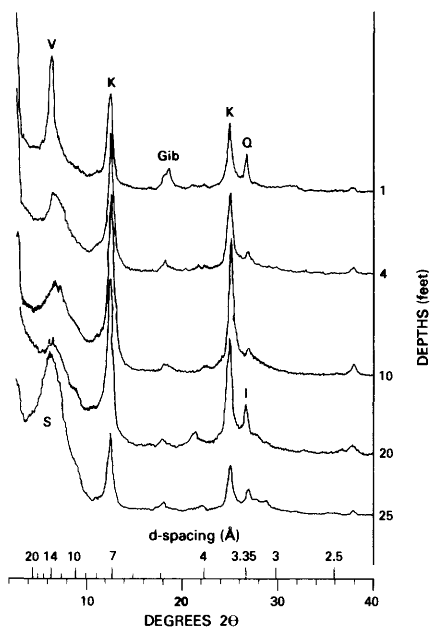
L. LB–317 (Socastee terrace)

M. LB–318 (Socastee terrace)

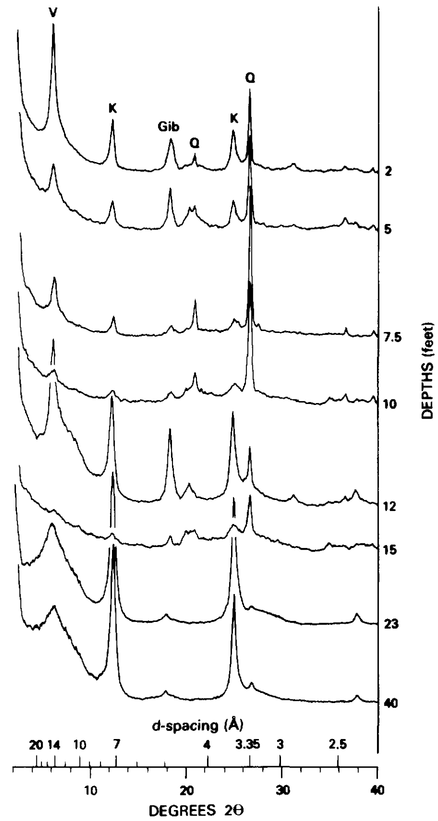
FIGURE 19—Continued.



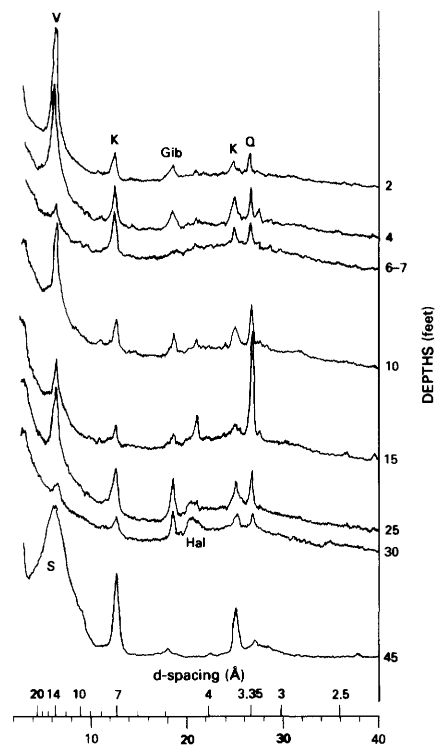
N. LB–321 (Socastee terrace)



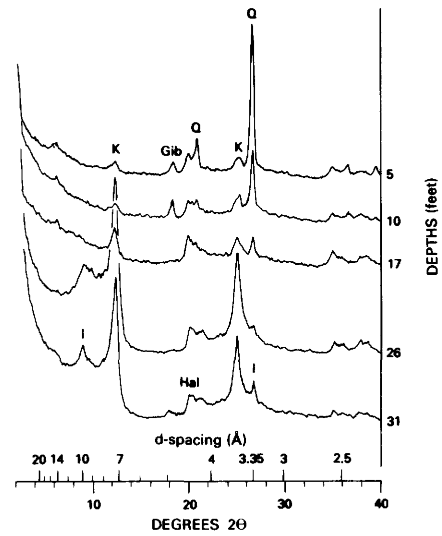
O. LB–381 (Socastee terrace)



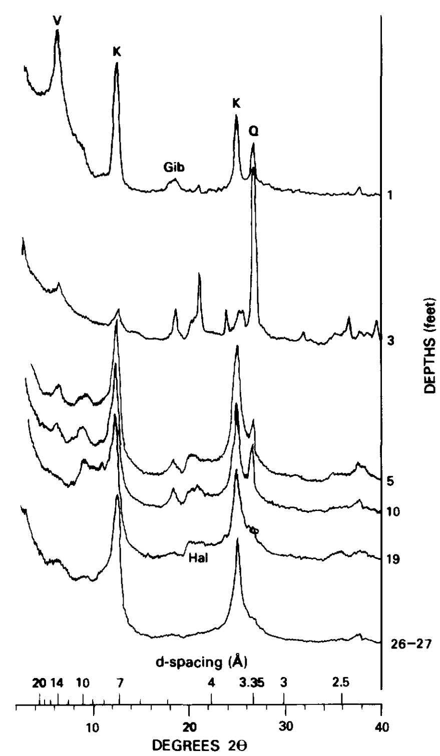
P. LB–296 (Penholoway terrace)



Q. LB–382 (Penholoway terrace)

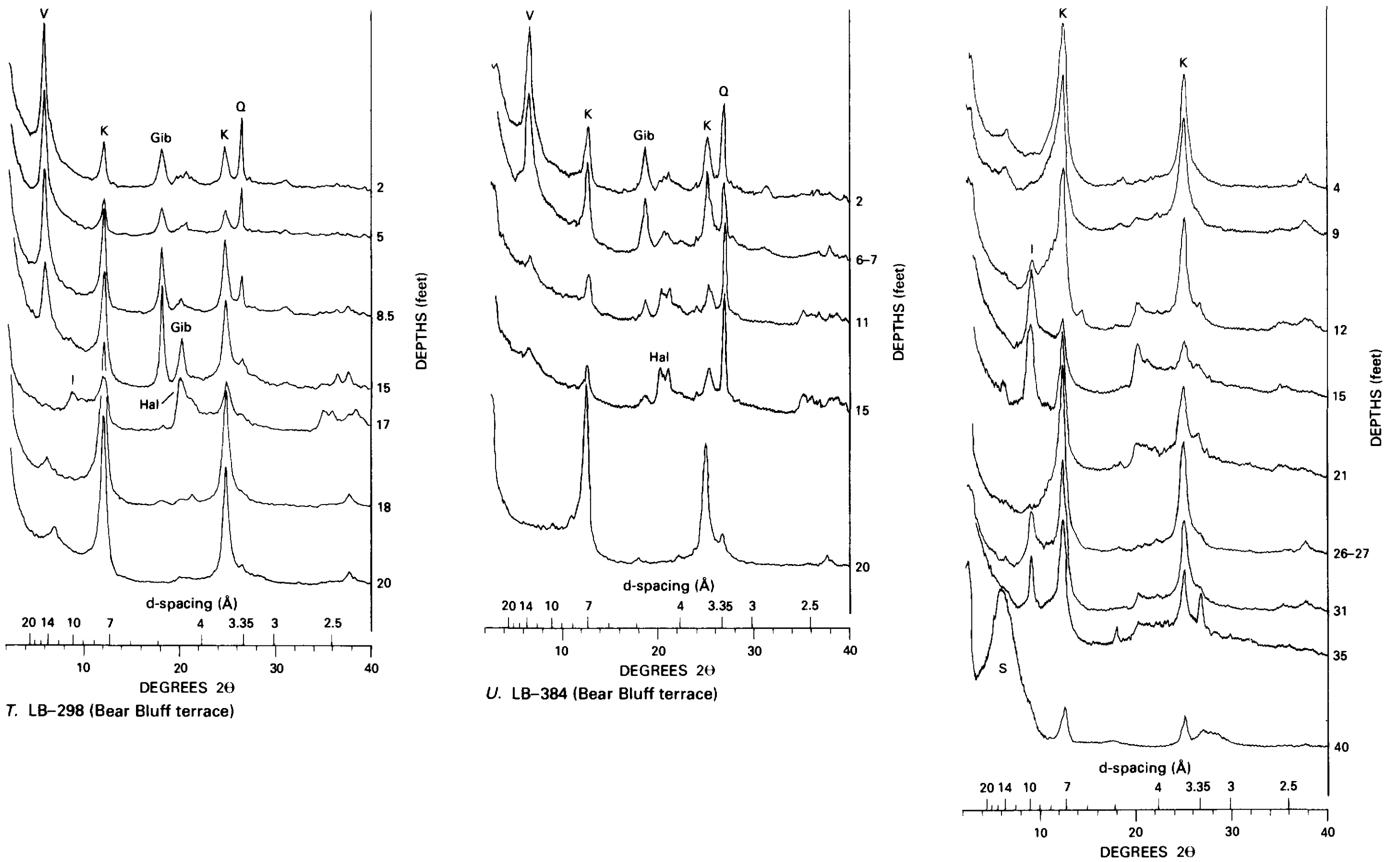


R. LB–297 (Waccamaw terrace)



S. LB–383 (Waccamaw terrace)

FIGURE 19—Continued.



T. LB-298 (Bear Bluff terrace)

U. LB-384 (Bear Bluff terrace)

V. LB-385 (Bear Bluff terrace)

FIGURE 19—Continued.

top of the fluvial deposits. Pollen data from these peats support the assumption stated above; the climate was temperate during formation of these organic layers, which are radiocarbon dated at more than 35,000 years old. Variations in weathering intensity were probably of minor importance in producing the different mineralogies and thicknesses of weathering profiles. Lithologic variations also should not be responsible for major differences between weathering profiles within the study area. These units are generally quartzose sands derived from weathering and transport of older Piedmont and Coastal Plain rocks and sediments, and lithologic variations are small. It is likely that the depth and degree of weathering are determined largely by the length of time these sediments have been exposed to weathering processes.

When sediments are weathered, a logical succession of mineral alteration products can be expected, and in some cases these data can be used as a stratigraphic tool. In the glaciated Eastern and Midwestern United States, the alteration of clay minerals in weathering profiles developed in tills has been given some systematic study (Droste, 1956; Droste and Tharin, 1958; Bhattacharya, 1962; Willman and others, 1966). In these studies, the sampling interval through the weathering profiles was small, and subtle, systematic variations in clay mineralogy were detected in the profiles. These variations in clay mineralogy, both vertically in a weathering profile and between tills of different glacial stages, were sufficient to differentiate tills of different stages. Bhattacharya (1962) sampled profiles in southern Indiana and detected the following alterations with time to a till containing illite, chlorite, and feldspar. Illite alters to illite/montmorillonite mixed layers and eventually to montmorillonite, while chlorite alters to chlorite/vermiculite mixed layers. Feldspars decompose to form kaolinite, possibly through mica as an intermediate product. Chlorite alteration is significant in even the youngest till (Wisconsinan). The weathering of illite proceeds more slowly, and only the Illinoian or older tills contain abundant illite/montmorillonite and discrete montmorillonite. The complete decomposition of feldspar occurs only in the upper part of the oldest till (inferred Kansan age).

These studies of till weathering demonstrate the utility of weathering profiles as a stratigraphic tool when some of the variables of weathering profiles (e.g., lithology, climatic variations) can be fixed for all the deposits under study. Research of this nature has been uncommon in the past, but study of sediments on the Atlantic Coastal Plain (Owens and others, 1983) has demonstrated that given the proper setting, weathering profiles can be used as indicators of age. Owens and others (1983) studied the surficial sediments in the Coastal Plain of the Middle Atlantic States and recognized a pattern of mineral oxidation and alteration similar to a general model of weathering sequences for silicate minerals proposed by Jackson (1965). The leaching of silica, or desilication, during weathering causes transformation of silicate minerals, notably feldspar and detrital clay minerals, into silica-deficient phyllosilicates, allophane, and oxides of iron and aluminum. Jackson's model relates the intensity and duration of weathering to the clay minerals formed. Since the mineral transformations are a function of both weathering intensity and time, if intensity is relatively constant, the process of desilication and the resulting clay minerals are solely a function of time.

According to Jackson's model, any parent material containing feldspar or phyllosilicates will, within some period of time, undergo mild desilication and produce a secondary clay-sized suite of montmorillonite or vermiculite, possibly as weathered edges or rinds on feldspars or micas. Allophane is also common, as a consequence of loss of silica from mineral structures. Upon these minerals are precipitated oxides and hydroxides, most notably Al_2O_3, which in the case of the crystalline clays imparts a choloritic nature to the structure. The clay is considered to be chloritized when interlayer precipitates inhibit a total collapse to 10 A upon heating. Intermediate desilication results in formation of kaolinite, halloysite, and allophane, in the continued development of vermiculite or montmorillonite from parent material, and in minor development of gibbsite. Intensive desilication, or laterization, causes the nearly complete loss of silica and transformation of silicate minerals into oxides. Hematite, goethite, anatase, gibbsite, boehmite, kaolinite, and allophane dominate in these intensely weathered profiles.

On the middle Atlantic Coastal Plain, Owens and others (1983) recognized the following sequences of mineral alteration:

Clay-sized

illite → vermiculite

feldspar → illite/smectite → kaolinite → halloysite → gibbsite

ferromagnesian minerals → lepidocrocite → goethite → hematite

Sand-sized

immature heavy assemblage → more mature assemblage (i.e., more zircon, tourmaline, and rutile)

immature light assemblage (2-feldspar) → more mature (i.e., mostly K-feldspar, grain size reduced near surface)

In addition to these common weathering sequences, there are a host of other possible transformations which are dependent on variables such as the efficiency of drainage. In the fluvial sediments of the Cape Fear River valley, where drainage is somewhat restricted, poorly ordered illite/smectite seems to weather to a somewhat better ordered vermiculute.

According to Owens and others' data, in a deeply weathered profile on the middle Atlantic Coastal Plain the detrital minerals would be altered to a suite perhaps composed of vermiculite, kaolinite, gibbsite, goethite, few labile heavy minerals, and minor feldspar. This suite roughly corresponds to Jackson's "intensely weathered" soil. Younger units, exposed to weathering for less time, should contain an altered mineral suite that is somewhat less mature.

Since the ages of the sediments were known, Owens and others were able to compare mineral assemblages from units of different ages and to assess the utility of mineral weathering studies in relative age determination. The maturity of the mineral suites was found to reflect the age of the sediment and to be a useful predictive tool. Upper late Tertiary and middle and upper Pleistocene deposits were examined, and the older sediments were found to contain a more mature mineral suite than the younger sediments. Development of gibbsite, maturation of the heavy mineral suite, and reduction in size and abundance of feldspar was more pronounced in the deeply weathered upper Tertiary sediments than in the younger units, and more pronounced in the middle Pleistocene than in the upper Quaternary sediments. These three age groups of sediments correspond roughly to Jackson's three weathering classes; mild desilication of the minerals occurs in the youngest sediments, late Quaternary, while intensive desilication was responsible for the mature mineral suite of the upper Tertiary sediments.

Owens and others demonstrated that on the Atlantic Coastal Plain, the silicate minerals are transformed to oxide minerals systematically with time. The presence of mature minerals, and the depth to which they are encountered in the weathering profile, can be used to assign a relative age to the deposits, in contrast to units of similar lithology having greater and lesser quantities of mature minerals.

WEATHERING OF THE UPLAND DEPOSITS

In the well-drained upland deposits dissected by the Cape Fear River, the abundance of epidote, hornblende, and feldspar in the sand-sized fraction was found to decrease in older units. Sediment from the upland drill holes LB-279 and 312 is Pliocene in age and is nearly devoid of these minerals throughout the entire thickness of the unit. Samples from younger upland deposits (LB- 289, 290, 294, 319, and Donoho Creek Landing) are richer in these labile minerals, but their abundance decreases upward in the weathering profile. Since these deposits are marginal marine sediments of similar source lithology, the contrast in sand mineralogy between the older and younger upland sediments is largely a function of time, and therefore supports the conclusions of Owens and others (1983).

In the upland sediments, the clay mineral suite tends to be more mature toward the surface. At depth, the clay suite consists largely of kaolinite. Nearer the surface, kaolinite, vermiculite, and gibbsite dominate the clay mineralogy. In the Horry barrier (Bear Bluff age) in Horry County, S.C., Markewich and others (1986) sampled the soil, which forms the upper 5.8 feet of the weathering profile. The x-ray diffraction data (figure 18*H*) records the progressive destruction of kaolinite (widening of the 7A peak and development of low-angle shoulder, and loss of crystallinity) and formation of gibbsite, halloysite, and goethite upward from the C horizon (sampled at 4.6 to 5.8 feet) to the B21t horizon (sampled at 2.4 to 3.2 feet). The upper x-ray diffraction trace is distinctly different from lower traces and marks a thin dune cap over the weathering profile. This profile is a clear illustration of the mineral transformations suggested by Jackson (1965) and by Owens and others (1983).

The weathering profile of the cutbank at Donoho Creek Landing (fig. 18*G*) is developed in sands of Waccamaw age; the upper 3 feet of the profile is enriched in clay and becomes reddish in color near the surface. Illite, probably detrital, is abundant at depth and weathers to mixed-layered material (dominantly smectite) and eventually to vermiculite in the upper profile. The upper foot of the profile is brick-red in color and contains the mature clay suite of kaolinite, vermiculite, gibbsite, and goethite. At 15 feet, near the contact with impervious Cretaceous clays below, clay minerals and nonopaque heavy minerals have been leached from the sediment.

At LB-289 (fig. 18*F*), sediments of Penholoway age are much less weathered than the sediments of Waccamaw age at Donoho Creek Landing; in the upper profile (upper 6 feet), smectite has begun to weather to a poorly ordered expandable mixed-layer phase, and the more labile nonopaque heavy minerals (e.g., hornblende) are depleted. The upper 4 feet is fine sediment fill in a Carolina bay and is discussed in the section on "Weathering of the Dune Deposits."

In some holes (e.g., LB-279, fig. 18*B*), a mature suite does not exist; kaolinite persists as the sole detectable clay mineral to the top of the profile. While the sediment at LB- 279 is quite old, according to Jackson (1965) and Owens and others (1983), the clay mineralogy is more appropriate to a young, unweathered sediment. Abnormally immature profiles such as this can be due to stripping of the upper weathering profile (Owens and others, 1983); this mechanism would account for the lack of mature clay-sized minerals in some areas and the presence of normal, mature suites in other areas of the same deposit.

WEATHERING OF THE FLUVIAL DEPOSITS

In the drill holes that sample fluvial sediments, there is no clear trend of an upward decrease in labile mineral abundance. There is, however, a lower proportion of epidote and hornblende in successively older units. Two explanations are suggested for this distribution: weathering and a change in supply. It is possible that the rate of erosion of Piedmont rocks has gradually and steadily increased during the Quaternary, delivering more labile minerals at each stage of terrace formation. This would require a gradual and continuous change in climate during the Quaternary, a change in location of the Piedmont headwaters, or an increase in the depth of erosion into unweathered rocks. There is no evidence to support the first two possibilities, and the latter is considered unlikely; the mineral distribution can be explained more simply by the weathering process already demonstrated by Owens and others (1983) to be operative in Coastal Plain deposits.

If weathering is responsible for the lower abundance of labile heavy minerals in the older terraces, a weathering profile would be expected. In holes LB-293, 296, and 317, labile heavy minerals are less abundant upward, but in other sections no trend is apparent. Unrestricted vertical movement of water and development of weathering profiles in the fluvial sediments may be prevented by the clayey or silty overbank deposits that cap most fluvial deposits in the valley, and by the fine-grained Cretaceous sediments flooring the valley. These deposits inhibit percolation of water and somewhat confine the sandy fluvial aquifer; water flow is therefore likely to be subhorizontal, resulting in intrastratal solution of labile minerals throughout the fluvial column. The fairly uniform abundance of labile minerals throughout these fluvial sediments contrasts with the more commonly observed upward decrease in labile minerals found in the weathering profiles of permeable surficial sediments in any area, and specifically near the Cape Fear River valley. Although the relative ages of the fluvial deposits cannot here be determined by the depth of weathering and the progressive loss of immature minerals upward in the section, the relative abundance of labile minerals in a drill hole can serve the same purpose: the older terraces have undergone intrastratal solution for a longer time than younger terraces, and are more depleted of epidote and hornblende. Feldspar abundance data portray a similar, albeit less well defined, trend. Owens and others (1983) noted that although the number of feldspar grains do not significantly decrease with weathering, a decrease in feldspar grain size upward in the weathering profile does occur.

In the fluvial sediments, clay mineralogy is fairly uniform and could not be used to differentiate terraces. Kaolinite and expandable (with ethylene glycol treatment) mixed-layer clays dominate the clay-sized mineralogy. The length of time needed to produce significant weathering of clay-sized minerals (as in the upland sections) in buried units must be greater than the age of the oldest terrace; otherwise a variation in clay mineralogy between terraces would be apparent. As an alternate explanation, or contributing to the lack of weathered clays in older fluvial sediments, the somewhat restricted movement of water through these sediments may have retarded development of mature clay mineral species. The amount of expandability of the mixed-layer clays was in some cases less in the upper fluvial samples than in samples lower in the same drill hole. A complete lack of expandability (upon glycolation) of the 14 A peak is characteristic of vermiculite; the partial loss of expandability in some fluvial samples suggests the presence of a clay mineral intermediate in character between the unweathered expandable mixed-layer variety and the more mature product of weathering, vermiculite. When the water table fluctuates within the fluvial aquifer, oxidation and weathering of minerals is promoted. This may be the mechanism responsible for the decrease in expandability of the mixed-layer clay.

There is a subtle difference in the abundances of labile sand-sized minerals between the fluvial deposits. Older deposits have fewer labile minerals than younger deposits, owing to the relatively longer duration of intrastratal solution. The clay assemblage is not noticeably more mature in older units; the clay-sized minerals detected by x-ray diffraction appear to be less sensitive to weathering than the sand-sized minerals and therefore are a less precise indicator of weathering intensity and age. Clay minerals can, however, be used to distinguish fluvial from older upland deposits.

WEATHERING OF THE DUNE DEPOSITS

Sand-sized mineralogy indicates that sand was blown onto the terraces during at least two separate intervals, and that the relative ages of the dune deposits can be determined by the abundance of feldspar and labile heavy minerals. In the dunes on the Penholoway and older terraces and the dunes on the upland deposits at LB-294, hornblende and feldspar are less common than in dune sands nearer the modern Cape Fear River. These mineralogic variations reflect the presence of at least two generations of dunes: an older, more weathered dune sheet covering the upper terraces and the uplands, and a younger, relatively unweathered dune sheet nearer the Cape Fear River. The differences in mineral abundance between dune sheets were, however, insufficient for mapping purposes.

The clay mineralogy of the dunes is a weathered, mature assemblage consisting of well-crystallized vermiculite, kaolinite, gibbsite, and quartz, but lacking crystalline iron oxides, specifically goethite. This assemblage is far more mature than that in the underlying fluvial sands, and it occurs in all dune deposits, even in those younger than 5,720 yBP (from a radiocarbon date beneath dune sands at LB-292). In addition, these clays are far more mature than the sand-sized mineral assemblage in the dunes. This suggests two possibilities—that a mature assemblage is more rapidly attained in the clay fraction than in the sand fraction, or that weathered eolian clays were deposited with relatively fresh dune sands. The first possibility is unlikely; as shown in this study, labile sand-sized minerals seem to be more sensitive to weathering than the clay minerals. In all upland drill holes, vermiculite and gibbsite occur only at or above the highest occurrence of feldspar and hornblende. In the Bear Bluff Formation at LB- 312, vermiculite and gibbsite occur only above a depth of 8 feet, which is well above the highest occurrence of feldspar or hornblende. In the Waccamaw Formation at Donoho Creek Landing, vermiculite and gibbsite occur only in the upper foot of the profile, while hornblende is virtually absent through the 15- foot thickness of the unit and feldspar has systematically decreased from 20 percent at 15 feet to 1 percent at a depth of 1 foot from the surface. However, in the dune sands vermiculite and gibbsite occur with high concentrations of feldspar and hornblende; this indicates that the dune sands contain a mixed mineral assemblage compared with the in situ weathered deposits of the uplands. The absence of a reddish, deeply weathered soil and iron oxides (i.e., goethite) from a sediment containing an otherwise mature clay-sized mineral suite also indicates a mixed mineral assemblage. In addition, the weathered sequences sampled in the uplands are at least 750,000 to 3.25 million years old, while some of the dune sands are younger than 5,720 years. Development of large amounts of vermiculite and gibbsite in a 5,000- to 6,000-year-old dune sand is unlikely, as none develops in much older fluvial sediments of similar lithology.

The second possibility, eolian deposition of weathered clays into fresh dune sands, is a more likely explanation for the constancy of mineralogy among the dune sands. The quantity and crystallinity of the vermiculite and gibbsite are constant throughout the dune sections; these minerals appear to have been derived from the uplands, where a similar clay mineral suite occurs at the surface, available for erosion and redeposition. At LB-279, and at numerous other drill holes in the upland areas of the region (not shown in this report), the upper portion of the surficial unit contains kaolinite, with little or none of the mature clay-sized minerals that are common in a weathered sediment. The absence of these mature clay minerals from some upland deposits, and their presence in dunes of much younger age, may be explained by stripping of the weathered, upper horizons of upland deposits by winds and incorporation of the resulting sediment (including clay) into the dunes. The coexistence of a mature clay-sized suite and an immature sand- sized suite has also been reported for the Parsonburg Sand, an upper Pleistocene eolian sand on the Delmarva Peninsula (Owens and others, 1983).

The eolian clay-sized mineral suite of vermiculite, kaolinite, gibbsite, and quartz was also detected in the fine grained sediments in a Carolina bay, in the upper 4 feet at LB- 289 (fig. 18*F*). In the weathering profile below the bay sediment, the clay mineral suite is immature, with illite and smectite persisting upward to at least 6 feet below the surface. In contrast, the maturity of the sand mineral suite gradually increases upward in the section; this suggests a mixing of eolian clay-sized minerals into the in situ deposits flooring the Carolina bay, probably when the surface depressions were lakes. The deposition of eolian clays into Carolina bay lakes as well as into actively forming dunes would be expected, since Carolina bays are wind-derived features (Kaczorowski, 1977).

The presence or absence of other clay-sized minerals in the dune sands is readily explained. Kaolinite, ubiquitous regionally, was derived from both upland and fluvial deposits. The expandable mixed-layer clays found in the fluvial sections are not present in the dune sands. Their abundance in fluvial samples is low, and further dilution by other eolian clays would prevent their detection. Also, expandable mixed-layer clays are unstable in the weathering profile and would be degraded relatively rapidly in the well-drained dunal soils.

Clay-sized quartz was found only in dune sand and fine grained sediments within Carolina bays. Other mature clay-sized minerals in the dune and bay bottom sediments (e.g., vermiculite and gibbsite) are eolian particles derived from the weathered upland deposits. The origin of clay-sized quartz is, however, less clear than for vermiculite or gibbsite, since it does not occur in the weathered profiles of upland deposits. Eolian quartz has been identified in Hawaiian soils by using size analysis and oxygen isotope data to correlate the soil quartz with tropospheric quartz (Rex and others, 1969). Rex and others discounted the importance of a pedogenic origin for quartz in the Hawaiian soils. If clay-sized quartz were being produced in soils in the Cape Fear area, quartz should be detected in non-truncated upland sections as well as in the dune sands, since the upland soils are intensely weathered and have contributed other clay-sized minerals to the dunes. Clay-sized quartz is not detected in the upland deposits; therefore, the presence of quartz in the clay fraction of dunes may be due to mechanical breakdown of weathered quartz grains and generation of clay-sized fragments during eolian transport.

AN ESTIMATE OF THE AGES OF DUNES AND CAROLINA BAYS

Dunes and Carolina bays cover much of the outer Coastal Plain in the Carolinas and are particularly striking in the Cape Fear River valley. Carolina bays in particular have been studied in detail; past research has dealt largely with morphology and mode of origin (Johnson, 1942; Buell, 1946a, 1946b; Frey, 1951, 1953, 1955; Prouty, 1952; Kaczorowski, 1977). Thom (1970) inferred the age of formation for Carolina bays in the Pee Dee River valley on the basis of radiocarbon data; his estimate is here considered to be a minimum age because the age as concluded in this report far exceeds the limits of radiocarbon dating. Data gathered for this study provide a unique opportunity to indirectly date the formation of the Carolina bays and dunes.

If it is assumed that all Carolina bays were formed by a single, unique climatic event, then this event must have occurred after deposition of the Socastee Formation and before deposition of the Wando Formation, as Carolina bays are prominent on the Socastee terrace but absent from the Wando. Since these terraces are correlated with isotopically age-dated formations, the date of Carolina bay formation falls somewhere between roughly 100 and 200 ka, roughly equivalent to the Illinoian glaciation.

If there are indeed two or more generations of Carolina bays, the age of formation suggested above would necessarily be true only for those bays on the

Socastee terrace; bays on older terraces could be much older. Absence of Carolina bays from the Wando terrace, and the derivation of only minimum radiocarbon dates from peat beneath bay rims is strong evidence that supports a pre-Wando age for formation of all Carolina bays.

Carolina bays have been shown to be wind-derived features (Kaczorowski, 1977) and therefore likely to have formed at the same time as the dunes. Mineralogic and radiocarbon data indicate at least two periods of dune formation. The most recent episode was Holocene; migrating dunes were dated at 7,700 and 5,720 yBP in the lower valley. These dunes have immature sand mineral assemblages and are present only next to the river. There are no Carolina bays associated with these dunes; conditions apparently were not favorable for the formation of these features. These dunes were a short-lived phenomenon, as sand was available from the incised Cape Fear River flood plain only until approximately 3,540 yBP, when backfilling of the channel began. Farther from the river, and associated with Carolina bays, are dunes with mature sand mineral assemblages that are dated at greater than 35,000 yBP. Assuming these dunes to be the same age as the Carolina bays, they were formed approximately 100 to 200 ka. Whether these older dune sands are of two or more ages could not be resolved in this cursory examination of dunes and Carolina bays; a more detailed geomorphic and mineralogic study would be required.

GEOMORPHIC EVIDENCE FOR UPLIFT OF THE CAPE FEAR RIVER VALLEY

In the study area, and along the Atlantic Coastal Plain in general, a series of transgressive-regressive sequences of late Cenozoic age are preserved, with progressively younger sequences lying nearer the modern coast and topographically below older sequences. Deposits of a typical sequence include a relatively thick section of regressive sediments (estuarine-backbarrier and barrier) overlying a thin transgressive section of marine sediments, or an erosional unconformity. The surfaces of these barrier-backbarrier complexes are level to gently sloping and may be traceable for great distances laterally. This characteristic prompted early workers (McGee, 1886, 1888; Shattuck, 1901, 1906; Stephenson, 1912; Cooke, 1930) to map these surfaces as a series of "terraces" stepping down toward the coast. A broad, regional uplift of the crust was generally assumed to have elevated the deposits beneath a terrace so that the succeeding transgression could not erode them. Glacio-eustatic fluctuations of successively lesser magnitude were invoked by Cooke (1930) as the mechanism responsible for terrace elevation.

The historically persistent assumption that a specific terrace is uniformly uplifted and everywhere the same elevation is too generalized and has led to miscorrelation of transgressive- regressive sequences along the Coastal Plain. The regional uplift suggested by the early workers is known to be punctuated by localized warping, including the Cape Fear arch, and indeed may not even be wholly responsible for preservation of the terraces. Eustatic changes in the sea level, due predominantly to glacial activity, could have contributed to the emergence of old coastlines. It is likely that some combination of eustasy and tectonism has ensured the preservation of these deposits, offering the researcher a challenge to assess the relative contributions of each to the geology and geomorphology of any given area.

Modern studies have been more limited areally than early, regional efforts, in an attempt to understand the local stratigraphy. In the Carolinas these studies include McCartan and others (1984) around Charleston, S.C., Dubar and others (1974) and Thom (1967) in the area of Myrtle Beach, S.C., and the lower Pee Dee River, Soller (1984) in the Cape Fear River valley, and Owens (in press) in the Cape Fear area in general, encompassing the Florence and part of the Georgetown USGS 2-degree topographic sheets.

North of Fayetteville, N.C., the Cape Fear River flows southward, and its valley extends a short distance to the east. Downstream, the river veers to the southeast (approximately S. 50° E.), the valley widens dramatically to the east, and the number of river terraces increases to a maximum of five. The valley maintains a southeasterly trend almost to its estuary at Wilmington, N.C., while the valley narrows gradually until the flood plain covers the entire valley width. Stratigraphic, mineralogic, and geomorphic evidence indicates that the river terraces decrease in elevation and age southwestward toward the river and are unpaired because the river is steadily migrating away from older river deposits and eroding its southwest banks. The youngest terrace, which slopes beneath the flood plain in the river's lower course, lies well above the entrenched Cape Fear River upstream. The Cape Fear River, astride an area of past tectonism (the Cape Fear arch), and with a peculiar course and valley configuration, preserves a geomorphic record of the tectonic and eustatic forces that have shaped the Coastal Plain. The following discussion attempts to identify and quantify localized tectonism so that the broader, regional causes of the emerged shorelines can be assessed.

To quantify the forces shaping the Cape Fear River valley, several techniques were devised which are based on a few simple assumptions. The first is that the Tertiary upland sediments are relatively homogeneous and have not caused the present valley shape by differential erosion. The river's course has been similarly unaffected by the underlying Cretaceous formations; the strata are horizontal to gently sloping and resistant beds are not extensive. The river's behavior and course have not, therefore, been constrained by variations in lithology of the substrate. The second assumption is that during each period of deposition recorded by a fluvial terrace the river had reached equilibrium, with a longitudinal profile identical to the modern Cape Fear River. Third, it is assumed that the time required for deposition of a flood plain is insignificant compared with periods of erosion and nondeposition. This is not strictly correct, but it must be assumed to calculate uplift rates.

Analysis of longitudinal profiles provides the best assessment of the tectonic history of this river system, but another approach gives a faster, albeit generalized, indication of local tectonism. A measure of the river's response to uplift, here loosely termed the "rate of migration," can be determined from the geologic map. The valley in the study area trends roughly N. 50° W. throughout, and therefore several parallel transects were drawn across the valley from points on the Cape Fear River about 25 miles northwest, 4 miles northwest, 10 miles southeast, and 28 miles southeast of Elizabethtown, N.C. Distances between the centers of each terrace along a transect were considered to roughly indicate the distance that the Cape Fear River had migrated laterally during the hiatus in response to uplift. Although the Cape Fear River did not actually migrate from midpoint to midpoint, the relative distances are thought to be compatible.

Uplift cannot be assumed to be continuous from Bear Bluff time to the present, and the technique described above indicates periods of tectonism and quiescence. Where one or more terraces is absent in the sequence along a transect, uplift is presumed to have caused a southwesterly river migration and preservation of terraces until deposition of the now-missing terrace; uplift then halted locally and the youngest terrace was eroded during the next depositional interval. For instance, to the south of Garfield, N.C., along the transect 10 miles southeast of Elizabethtown, uplift occurred from Bear Bluff time until after deposition of the Waccamaw flood plain. Cessation of uplift and of migration in this, the lower course of the river, allowed erosion of the downvalley end of the Waccamaw terrace by the Penholoway-age river. Although the assumptions required here may seem too rigid, the results are intuitively reasonable estimates of the location and timing of uplift.

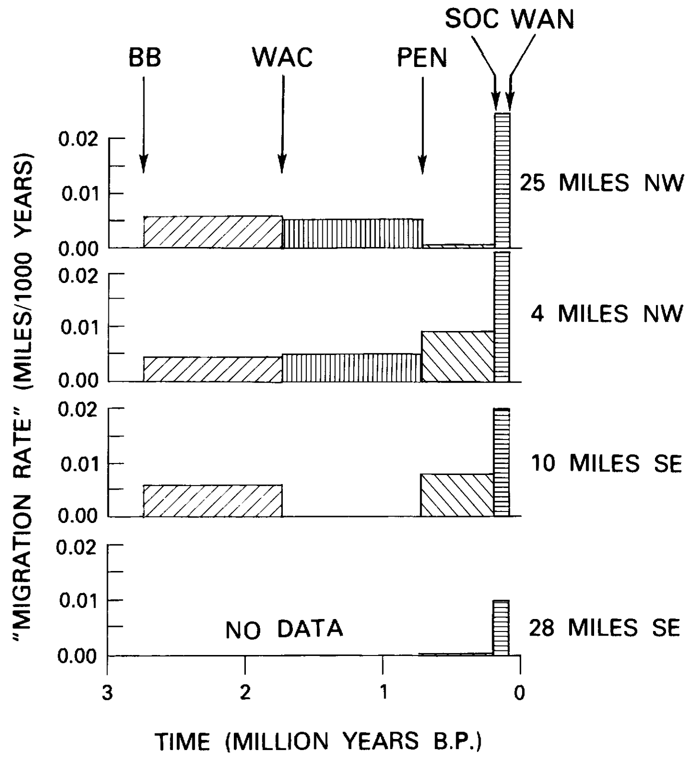


FIGURE 20.—Geomorphic analysis of the configuration of fluvial terraces in the Cape Fear River valley, attempting to show the rate at which the river has migrated to the southwest. The position and dimensions of each terrace along four transects across the valley were studied. These transects are located upriver (to the northwest) and downriver (to the southeast) from Elizabethtown, N.C.; distances are shown. BB, Bear Bluff; WAC, Waccamaw; PEN, Penholoway; SOC, Socastee; WAN, Wando.

Calculated migration rates across the four transects are shown in figure 20. Prior to Penholoway time, the valley above Elizabethtown was gently uplifted, from the north or northeast, causing river migration down the flank of the uplift, to the southwest. Between Penholoway and Socastee time, the uplift increased and became more localized, raising the central part of the valley while stabilizing or depressing both the upper and lower valley. In post-Socastee time, uplift of the upper portion of the valley was extreme and overshadowed the effects of uplift from the north or northeast.

LONGITUDINAL PROFILING

To gain a more precise account of the geomorphic effects of localized uplift, the longitudinal profiles of the five terraces and the Holocene flood plain were analyzed. Profiles for all surfaces were constructed on a trend of N. 50° W.; comparison of profiles and uplift

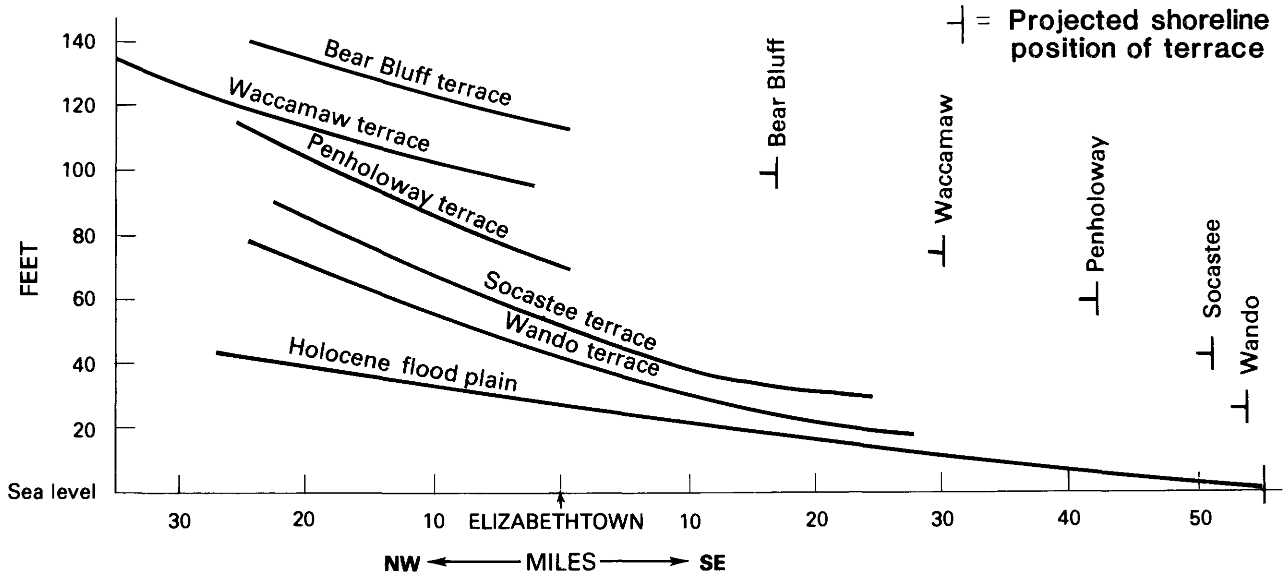


FIGURE 21.—Longitudinal profiles of fluvial terrace upper surfaces in the Cape Fear River valley. The terraces and river trend about N. 50° W.; these profiles were constructed along this trend, and Elizabethtown, N.C., was used as a common point on all profiles. These profiles were constructed from auger hole data and supplementary topographic data; sand dunes cover the terraces and necessitate the use of auger hole data to identify the true upper surface of the fluvial terrace. The projected shoreline position for each terrace is also shown. Note the similarity in profiles of the Bear Bluff and Waccamaw terraces, and their contrast with the profiles of the lower terraces.

analysis is relatively simple because the terraces and river follow essentially the same trend throughout the area studied here, from 25-30 miles upriver (northwest) of Elizabethtown to the estuary at Wilmington. Profiles were constructed along the upper surface of the fluvial deposits, which is masked by the dune sheet covering all terraces in the valley. Mineralogic and lithologic analysis indicated the dune thickness at each location. From these control points and supplementary topographic evidence where borehole data were unavailable, upper fluvial surface profiles were drawn.

Vital to a study of the profiles is an accurately determined paleoshoreline position and sea level elevation for each terrace. The barriers or beach sands along the seaward margin of preserved regressive sequences were used to locate the paleoshoreline. Paleo-sea levels were selected from study of the backbarrier-barrier elevations and data of J.P. Owens (USGS, verbal commun., 1985) and are estimates whose accuracy is constrained by the complex and differing tectonic histories of the areas studied in the Carolinas.

Profiles and shoreline positions along a line N. 50° W. are shown in figure 21. The slope of the Bear Bluff and Waccamaw terraces are clearly similar, and different from younger terraces. Also, the younger terraces, especially the Wando, do not have smoothly dipping profiles. An inflection point southeast of Elizabethtown suggests differing tectonic regimes along the river's course, in at least post-Waccamaw time.

Uplift cannot be calculated by a direct comparison of these profiles. Along a transect 20 miles northwest of Elizabethtown, the elevations of the Bear Bluff and Waccamaw terraces differ, by 21 feet; there has not been, however, 21 feet of uplift in the intervening time. The true uplift can be determined by a best fit procedure that compares the elevation of points on each profile that were at the same distance upriver from their shorelines. To demonstrate the logical basis for this procedure, a simple example is considered that assumes a regional uplift or a eustatically lowered sea level between deposition of a Pleistocene terrace and the modern flood plain (fig. 22*A*). There was an apparent differential (i.e., localized) uplift of the terrace prior to the Holocene; the vertical distance between terrace and flood plain changes along the profiles. The apparent uplift is illusory and is due solely to the differences in shoreline position. To illustrate this point, the factor for the regional uplift or eustatic sea level drop is removed by assuming a common sea level elevation (fig. 22*B*); any residual difference in elevation will have a localized cause. Since both the terrace and the flood plain are products of a graded river, their slopes at any given position along the profiles are identical. When the profiles are overlain with a common shoreline, any difference in elevation at a common point must be a product of local tectonism; as shown here (fig. 22*C*) there has been no local tectonism, contrary to its apparent existence noted in figure 22*A*).

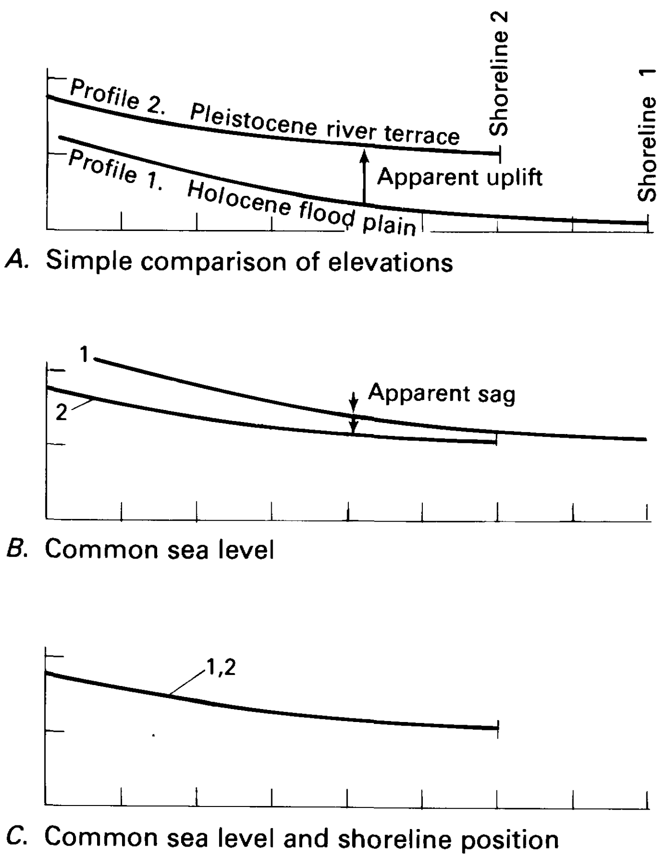


FIGURE 22.—Hypothetical profiles of a graded river along a stable continental margin. In *A*, the profiles of a terrace and a flood plain are compared, and differential uplift along the course of the river is apparent. However, when the effects of eustatic sea level fluctuations or epeirogenic uplift are eliminated by comparing the two profiles using a common sea level, subsidence of the area is suggested (*B*). Part *B* erroneously compares points that were at different distances from their respective shorelines; since these surfaces are the product of a river at grade, the two profiles should be identical when overlain if the terrace has not been warped, as shown in *C*.

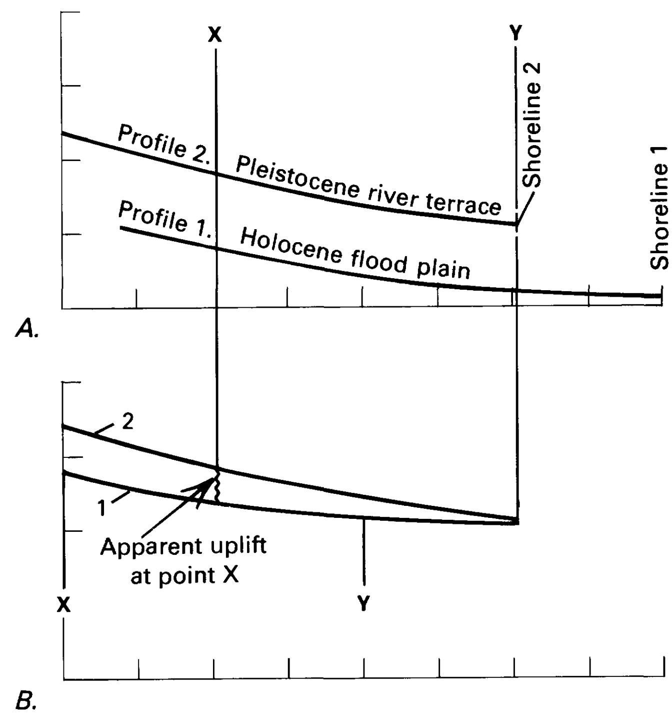


FIGURE 23.—Hypothetical profiles along a tectonically active continental margin. Following the procedure outlined in figure 22, the profiles (part *A*) are compared with identical shoreline positions and sea level (*B*). Uplift of the Pleistocene terrace is shown; however, because the shoreline positions were not originally the same, this procedure involves computing uplift for areas that underwent different degrees of uplift. The location of geographic point x on profiles 1 and 2, or y on profiles 1 and 2, does not coincide in *B* owing to the procedure of overlaying shorelines; therefore, a measure of uplift at any point along the profiles in *B* does not truly measure uplift at a single place, but rather compares elevations at two separate locations (see "apparent uplift at point x").

When local tectonism does occur, and terraces are uplifted or downwarped differentially along their length, the procedure outlined above may have serious limitations. Assuming that the head of the valley is being uplifted, the procedure used in figure 22 is attempted in figure 23: the regional effects are removed, shorelines are overlain, and a localized uplift is measured (fig. 23*B*). Although common points along the profiles are used to measure uplift, these points are not at the same locations on the land surface and did not necessarily undergo the same degree of uplift. This error is progressive when both locations are on the flank of an uplifted or downwarped area, but when they lie in two tectonically distinct areas the utility of this simple approach is limited.

The tectonic and depositional history of the Atlantic Coastal Plain is far more complex than the simple models depicted in figures 22 and 23, yet the elementary logic used in those two examples is at the heart of the best fit procedure used for the Cape Fear River valley. As an example of the best fit procedure, when temporally adjacent profiles are overlain with a common shoreline position and sea level elevation, the local uplift measured for the older terrace is sometimes unrealistic. For instance, between Wando time and the present, 10 feet of uplift was measured along the transect 20 miles northwest of Elizabethtown and 16 feet of subsidence was measured along the transect 25 miles southeast, with the inflection point 5 to 10 miles northwest of Elizabethtown. Topographic evidence

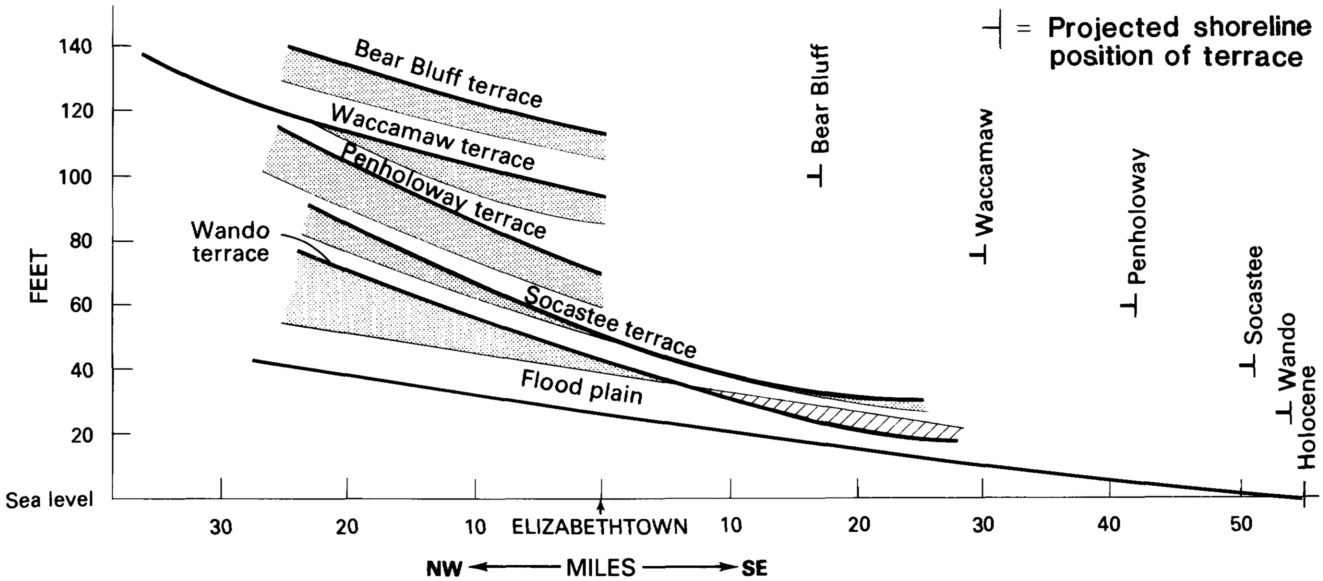


FIGURE 24.—Uplift of the Cape Fear River valley as measured by a best fit procedure on longitudinal profiles. The longitudinal profiles of terraces and corresponding paleoshorelines are shown by thick lines. Shaded areas beneath each terrace profile represent the amount of uplift that occurred prior to deposition of the succeeding terrace. In the lower part of the Wando terrace, 10 miles and more southeast of Elizabethtown, relative subsidence of the Wando surface prior to Holocene deposition is indicated by diagonal lines.

cannot support these values; more uplift and less subsidence appears likely, with an inflection point downriver from Elizabethtown. To reproduce this, the elevation of the Holocene shoreline must be drawn below the longitudinal profile of the Wando terrace and a localized uplift of the Wando shoreline introduced to account for the difference. This correction was also required for the Socastee terrace. This best fit procedure provided values for local uplift between each depositional event along any valley transect, as shown by the shaded areas in figure 24.

In a complexly warped valley such as that of the Cape Fear River, the local uplift as measured here (fig. 24) may be in error when any two profiles cross an area of differential uplift. To assess these errors and the source of the regional uplift/sea level drop, the uplift values were used to recreate the terrace profiles at each stage of flood-plain deposition. The degree of similarity between the series of reconstructed profiles for the Holocene and the actual profiles (fig. 21) should indicate the accuracy of the computed uplift or the need for further best fitting of profiles. The procedure for profile reconstruction is as follows (see also fig. 25). Prior to deposition of the Waccamaw flood plain, the Bear Bluff terrace had been uplifted 11 feet along the transect 20 miles northwest of Elizabethtown and 8 feet along the transect at Elizabethtown. Uplift was projected to increase upvalley and to decrease to 6 feet at the Bear Bluff shoreline. In the first step, the Bear Bluff flood plain is raised by the measured uplift values to the new profile it assumed as a river terrace during Waccamaw time. Next, the Waccamaw shoreline is plotted, 25 feet lower than the uplifted Bear Bluff shoreline (as required by estimates of shoreline position and elevation) and the Waccamaw flood-plain profile (identical to the Holocene profile) is plotted beneath the Bear Bluff terrace. The difference in elevation between terrace and flood plain along any randomly selected transect should equal the difference determined by stratigraphic and topographic analyses (fig. 21). If not, the uplift values, particularly the value projected for the paleoshoreline, must be reassessed until a best fit solution is reached. When the fit is satisfactory, the amount of regional uplift or eustatic sea level drop equals the difference in shoreline elevations minus the projected localized uplift of the older shoreline; in the above example, this would equal either a 19 foot regional uplift or sea level drop between Bear Bluff and Waccamaw time.

LOCAL UPLIFT OF THE CAPE FEAR RIVER VALLEY

Profile reconstruction using the uplift values derived from best fit analysis results in a close match with the real profiles drawn in figure 21. These reconstructed

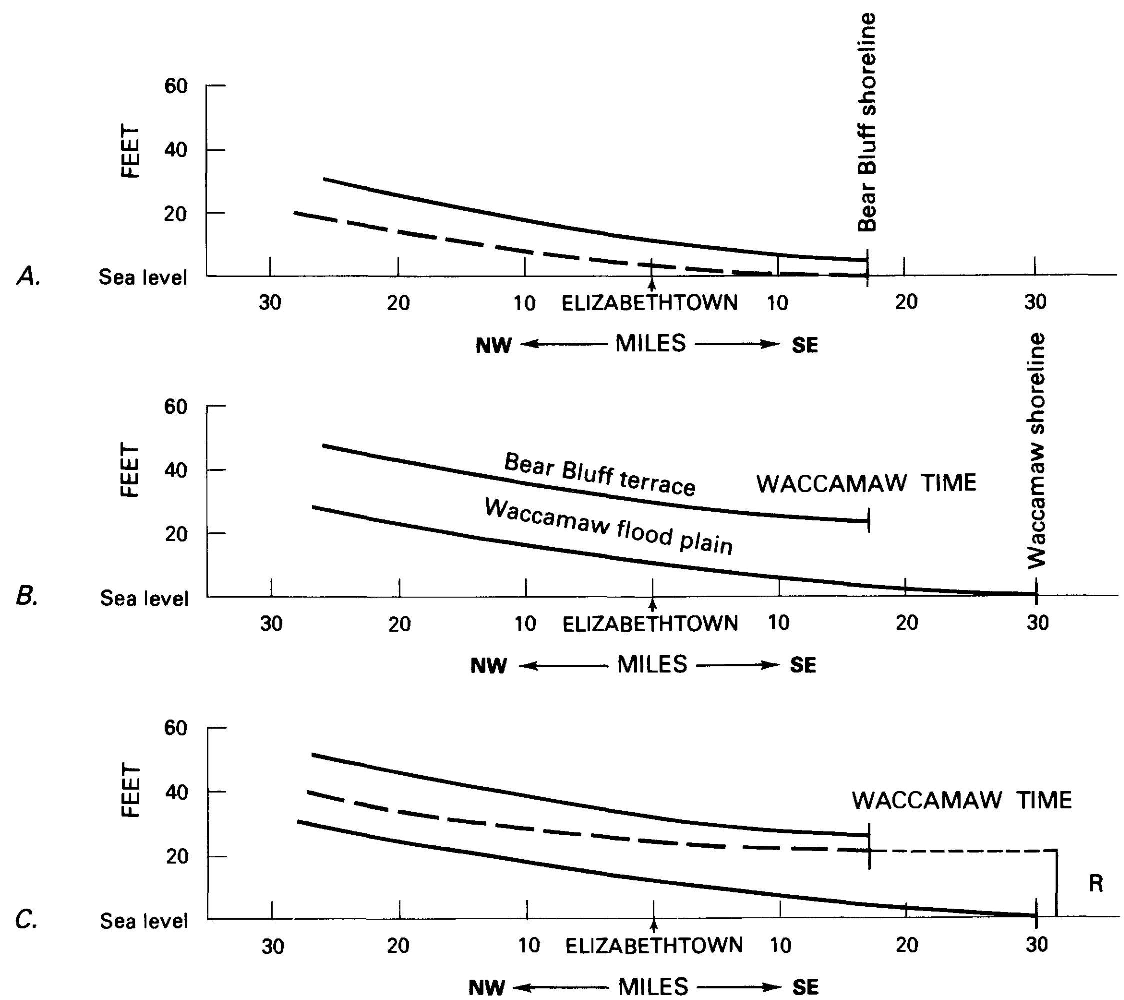


FIGURE 25.—Profile reconstruction used to assess the accuracy of the best fit procedure for profiling, using the configuration of the terraces existing during Waccamaw time as an example. In *A*, the Bear Bluff flood plain (dashed line) is uplifted by the amount determined in figure 24 and values are projected toward the shoreline. Solid line shows position of uplifted Bear Bluff surface. In *B*, the Waccamaw shoreline is located in the correct relationship to the uplifted Bear Bluff shoreline and the Waccamaw flood plain is drawn. The difference in elevation between the Bear Bluff and Waccamaw surfaces should agree with that measured in figure 21; if it does not, the best fit of the data is reassessed. In *C*, which is a composite of *A* and *B*, the difference in elevation between the Waccamaw and original Bear Bluff shorelines ("R") is a measure of the regional uplift, on which the localized warping of the terraces is superimposed.

profiles are shown in figure 26. Since these uplift values are verified by profile reconstruction, and values for the regional uplift or sea level drop have been derived, local uplift rates along three transects and a regional rate were calculated (table 4). To compute rates, age of deposits had to be generalized (see table 1). These computed rates are quite low relative to rates from tectonically active island arcs (Bloom, 1980), and are slightly less than data for South Carolina as reported by Cronin (1981). However, Cronin was not able to differentiate local and regional forces, and computed rates for the entire interval since a unit was deposited (e.g., Bear Bluff time (late Pliocene) to Holocene).

The data gathered for the Cape Fear River valley indicate that the locus and intensity of local tectonism have varied with time. These rates, and the geologic map and migration rate data, lend support to the following conclusions. From at least Bear Bluff time gentle uplift to the north and (or) northeast of the river valley caused a southwestward (lateral) river migration. The uplift became relatively less intense at the northwestern end of the valley after Bear Bluff time. An increase in the rate of uplift at the northwestern end,

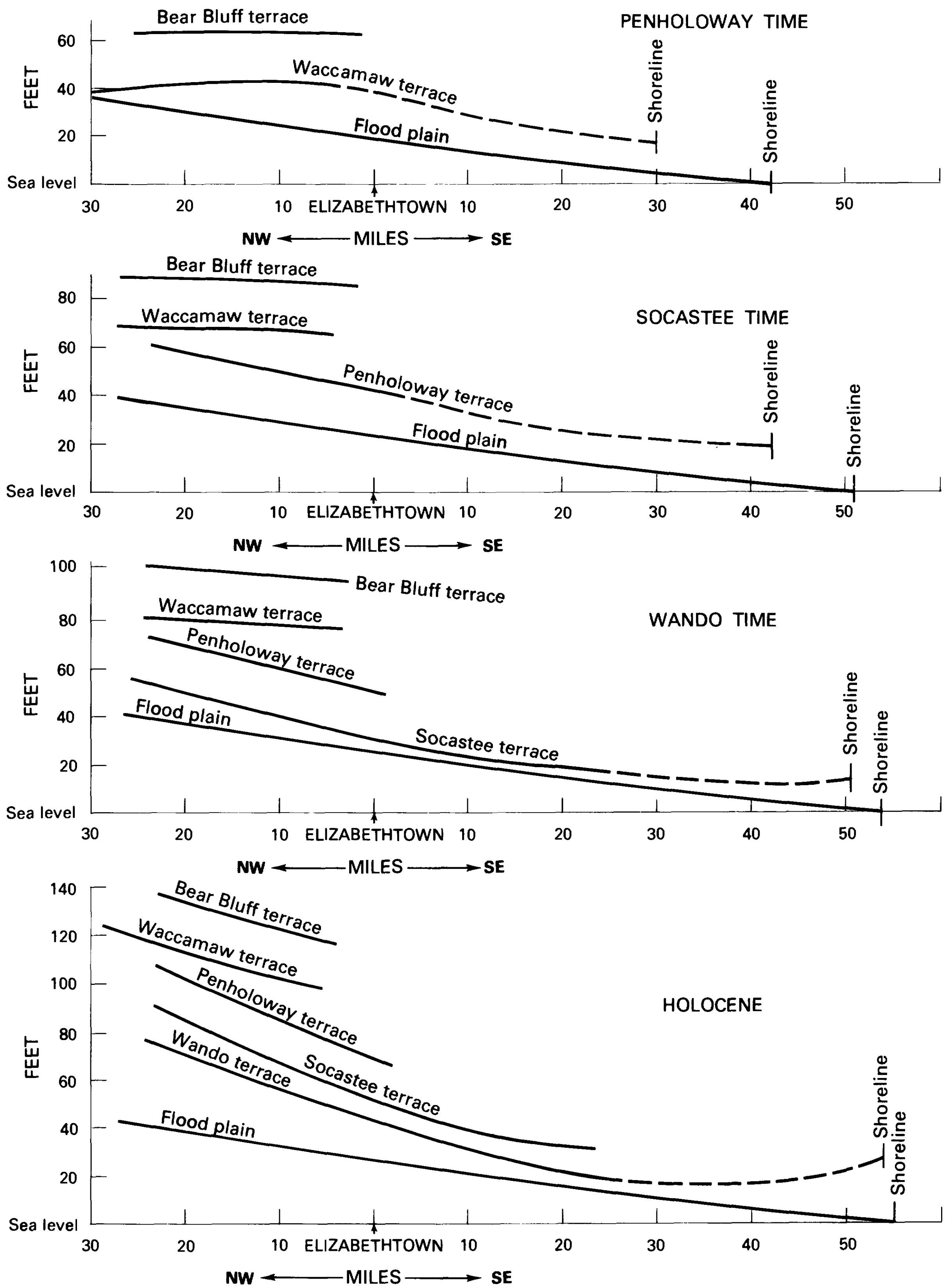


FIGURE 26.—Reconstructed profiles for all intervals of deposition since Waccamaw time (early Pleistocene). Dashed lines show projected profiles of youngest terraces. The close agreement of the profiles for the Holocene with figure 21 indicates that the best fit procedure and the values for uplift or sea level variation are reasonable.

TABLE 4.—*Uplift rates in the Cape Fear River valley*[1]
[Feet per 1,000 years]

Time interval[3]	Location of calculation[2]			Regional
	20 mi NW	Elizabethtown	20 mi SE	
Bear Bluff-Waccamaw	0.011	0.008	[4]0.005	0.019
Waccamaw-Penholoway	.002	.01	[4].003	.015
Penholoway-Socastee	.025	.018	[4].025	.022
Socastee-Wando	.09	.03	.03	.07
Wando-Holocene	.21	.05	-.04	.12

[1]Uplift rates were calculated from a best fit analysis of longitudinal profiles.
[2]Calculations were made along lines 20 miles upvalley (to the northwest) from Elizabethtown, N.C., at Elizabethtown, and 20 miles downvalley (southeast). A regional uplift was also calculated.
[3]The time interval is, for example, the interval between deposition of the Bear Bluff and Waccamaw formations.
[4]Projected value.

near the Piedmont, once again is detected in the Socastee-age terrace. The rate of uplift at the northwestern end of the valley increased progressively since Penholoway time and overshadowed the lesser, but persistent, uplift from the north or northeast in post-Socastee time. While uplift rates were increasing in the upper, northwestern end of the valley, rates were decreasing in the lower, southeastern end; maximum uplift in the northwest occurred in post-Wando time concurrent with probable subsidence in the lower valley. It would seem that these two areas of different uplift histories are related; a simple tilting of the Coastal Plain along the valley length, up from the direction of the Piedmont, could account for these data.

In summary, two localized tectonic events have shaped the valley: (1) a persistently low rate of uplift from the north or northeast, transverse to the valley length and largely responsible for the succession of unpaired terraces in the central section of the valley, and (2) uplift from the direction of the Piedmont (parallel to the valley length) that has been intermittently active and most recently tilted up to the northwest and down to the southeast, causing deep entrenchment of the Cape Fear River into the Wando terrace in the upper valley and burial of terraces beneath the flood plain in the lower valley.

These uplift values were derived from a series of measurements which involve the cumulative effects of several variables, and from necessary assumptions that include age of deposition and depositional style. The geologic map, however, supports the data interpretation, which is therefore thought to be a reasonable approximation. Some consideration of the source of the localized tectonism and the regional factor is therefore warranted.

The local uplift described above requires either a monocline or series of flexures, or a series of faults. Broad, gentle flexures such as the Cape Fear arch are substantially supported by field evidence. Evidence to support a faulting mechanism to explain the surficial geology is less well established on the outer Coastal Plain; in the thin Coastal Plain sediments adjacent to the Piedmont in northeastern Virginia, however, Neogene faulting has been observed (Mixon and Newell, 1977, 1982). On the outer Coastal Plain of North Carolina, faulting has been inferred by several authors. LeGrand (1955) noted anomalously high salt water concentrations of an artesian seepage in Bladen County, N.C., in the lower Cape Fear River valley. He attributed this phenomenon to artesian movement of brines upward, along a fault plane. This observation and preliminary geophysical data (MacCarthy, 1936) prompted Ferenczi (1959), in a review article, to propose a fault zone extending from Conway, S.C., to northeastern North Carolina. This feature has been named the "Carolina Fault"(Baum and others, 1978), apparently on the basis of LeGrand's observations. Harris and others (1979) assumed the existence of the fault, and modified the position of its trace. The existence of a deep-seated, northeast-trending basement fault is neither supported nor rejected by the present study. However, the brine seepage reported by LeGrand (1955) could be explained by another mechanism, without invoking faulting. Erosion and truncation of tilted Cretaceous strata by the pre-Holocene Cape Fear River may have exposed a pressured aquifer, which is currently releasing brine through terrace sediments to the surface. The "Carolina Fault"may exist, but field evidence to support this idea is lacking.

A small-scale flexure, related to larger scale tectonism on the Cape Fear arch, seems to explain the terrace pattern in the Cape Fear River valley. The upper surface of crystalline basement rocks as shown in figure 4*B* depicts the broad outline of the Cape Fear arch beneath the Coastal Plain in North and South Carolina. To the north and east of the Cape Fear River valley, a bulge in the basement structure contours indicates a localized zone of more intense uplift, parallel to and superimposed on the Cape Fear arch. This localized uplift lies in the correct position relative to the Cape Fear River valley to account for the uplift history of the valley, and is therefore considered the source of uplift that shaped the valley. Most of the Cape Fear River valley lies over the local bulge; the lower part of the valley, where relative subsidence has occurred, lies on the southern margin of this structural high where uplift is relatively less than on the structural axis of this feature. This basement feature seems to account for the gradual downvalley change in the nature of the river valley, from the pattern of strath and depositional terraces that occur where the river traversed the localized uplift, to relative subsidence in the lower valley, which does not overlie the uplifted area.

CAUSE OF THE ELEVATED SHORELINES

The cause of the regional elevation of the Coastal Plain deposits has been the subject of debate since the late 1800's, when the "marine terraces"were recognized. There are merits to both the isostatic uplift mechanism proposed by Kerr (1875) and Shattuck (1906) and the glacioeustatic sea level oscillations invoked by Cooke (1930). Although it is established that the sediments beneath the terraces were laid down during warm, interglacial highstands of the sea (Cronin, 1981), a purely eustatic model is incorrect because worldwide, or even regional, correlation of units based solely on elevation have been found invalid. In a study of Holocene datums, leveling surveys, and older Holocene tidal marsh surfaces, Newman and others (1980) concluded that the east coast of the United States has been differentially warped in the past 12,000 years. In a supplementary study using newly published radiocarbon-dated marsh peats, Cinquemani and others (1981) refined the earlier conclusions and suggested that the Cape Fear arch is an inflection zone for the Holocene, with the coast to the north subsiding relative to the southern coast. Winker and Howard (1977), in a correlation study of Pliocene and Pleistocene shorelines on the southern Atlantic Coastal Plain, found the shorelines to be warped, not horizontal. While that study relied on geomorphic criteria, the general trends are valid and follow major, persistent structural trends (the Cape Fear and Ocala arches and the southeast Georgia and south Florida basins). The warping of shorelines was also detected for the early Pleistocene, using biostratigraphy to correlate deposits (Cronin, 1981; Cronin and others, 1984). The data from these studies support the contention that major structural features of the Atlantic margin have remained active throughout the Pleistocene, warping the land surface and preventing regional correlation of deposits based solely on elevation.

The map pattern of elevated shoreline deposits on the Coastal Plain likely reflects a combination of eustatic and tectonic forces. An estimation of the magnitude of glacioeustatic sea level fluctuations during the Pleistocene can be derived from oxygen isotope data (Shackleton and Opdyke, 1973). As shown in figure 27, from Szabo (1985) as modified from Shackleton and Opdyke (1973), decreases in the oxygen isotope fractionation ratio correspond to ocean highstands and periods of deposition. The fractionation ratios are most extreme for stage 5e (Wando Formation), which suggests that previous Pleistocene glacioeustatic sea level highstands were perhaps lower than in Wando time. More important, however, is the lack of a trend that would suggest a decrease in magnitude of sea level oscillation with time; therefore, while glacioeustatic sea level fluctuation is the mechanism that governs whether a landscape is being aggraded or degraded, the pattern of elevated shorelines must have been the result predominantly of crustal uplift, not of a decreasing magnitude of sea level fluctuations. Although interglacial sea level almost certainly fluctuated within some range, the data from this study do not permit the resolution of paleo-sea levels from the greater contribution of crustal uplift, especially in the older deposits where the cumulative effects of this uplift are most pronounced. In the following discussion it is initially assumed that past interglacial sea levels have been at approximately the present level; minor adjustments to this assumption were necessary for only a portion of the late Pleistocene.

Cronin (1981) summarized the potential mechanisms that could have produced the elevated shorelines along the Atlantic coast. It is not within the scope of the present study to comment on these various theories; the purpose, instead, is to investigate the uplift history along the Cape Fear arch. As mentioned previously, J.P. Owens (USGS, personal commun., 1985) has noted

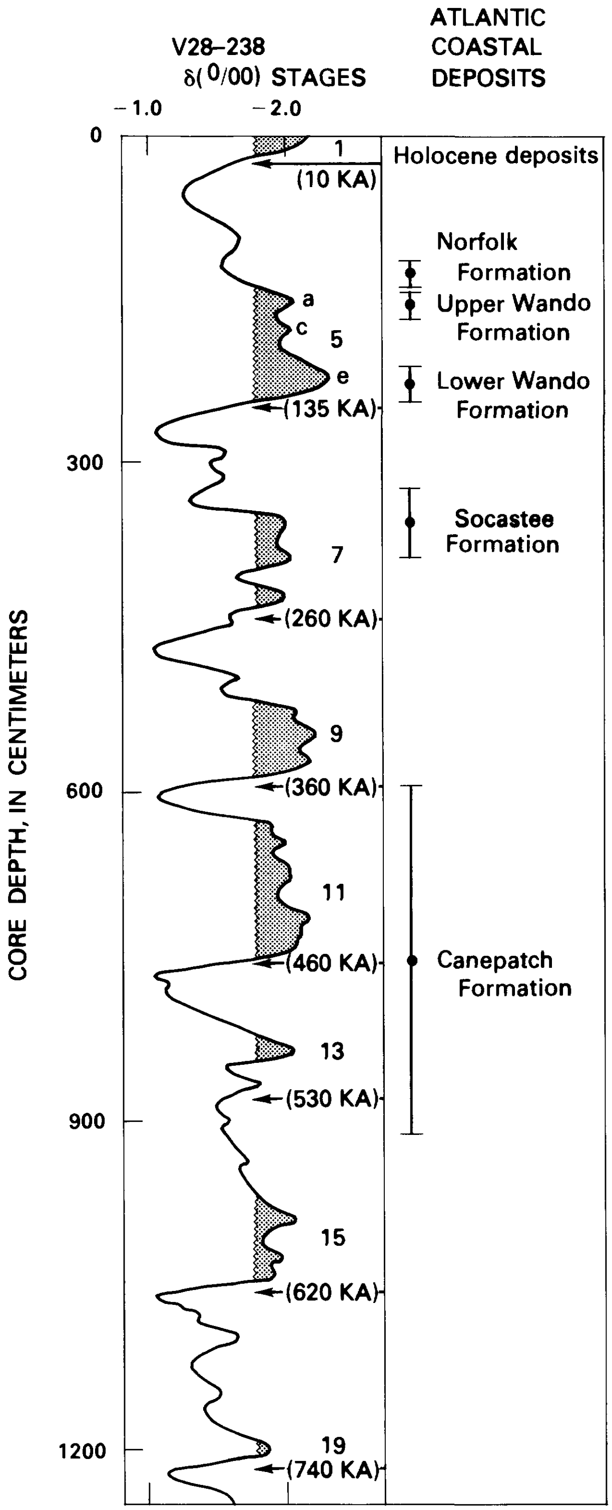


FIGURE 27.—Comparison of southeastern Atlantic Coastal Plain deposits with major interglacial oxygen isotope stages. Note that the maximum negative ratio (i.e., sea level highstand during interglacial) is attained at stage 5e (Wando Formation). Modified from Szabo (1985), figure 6.

the tectonic complexities of the Cape Fear arch; evidence suggests not only that the arch has migrated with time (see fig. 4*A*), but also that it has not always been a positive tectonic feature. Given the complex nature of the arch, only the most general comments on the cause of elevated shorelines are in order.

In southeastern North Carolina, the Coastal Plain was elevated at a fairly constant rate of 0.015 to 0.022 feet per 1,000 years between deposition of the Bear Bluff and Socastee Formations (see table 4). This slow, steady rate of elevation most likely reflects upwarping of the Cape Fear arch, or a larger area of the Coastal Plain. Superimposed on this uplift were the localized flexures on the Cape Fear arch that preserved the unpaired terraces in the Cape Fear River valley.

The rate of elevation since Socastee time far exceeds the earlier rates. Assuming that the regional uplift discussed in the preceeding paragraph was still operative after deposition of the Socastee Formation, and allowing for variability in the rates, a value consistent with the earlier rates (their mean plus 2 sigma) was subtracted from the post-Socastee rates. The resultant rate indicates a change in sea level elevation between the Socastee, Wando, and Holocene depositional events which cannot be due to the slower, more persistent tectonic mechanism discussed above; the elevational changes must therefore be caused by variations in sea level highstands. Although the resultant rates are high relative to pre-Socastee rates, the actual change in paleo-sea levels is low: sea level during Socastee time is calculated to have been roughly 4.5 feet higher than during Wando time, which in turn was about 9.5 feet higher than during the Holocene. These small differences can be readily accounted for by glacioeustatic fluctuations; estimates of eustatic sea levels in late Pleistocene time derived from a comparison of coral reefs in Barbados, oxygen isotope data from deep sea cores, and climatic modeling indicate interglacial highstands of 20 feet or less above present sea level during Socastee and Wando time (Cronin, 1981).

Although glacioeustatic fluctuations appear, on the basis of this series of assumptions and calculations, to contribute to the appearance of elevated shorelines in post-Penholoway time, their detection in the older record is uncertain because the cumulative effects of uplift tend to obscure the eustatic contribution to the present elevations of these paleo shorelines. Despite the assumptions concerning past sea levels, it is in fact uncertain whether sea level during pre-Socastee deposition was equal to the present level. Oxygen isotope data (Shackleton and Opdyke, 1973) and figure 27 suggest that past interglacial sea levels were at or below the Wando level, but the limitations of these studies and the data within this report, specifically the data derived from longitudinal profiling of the older terraces, must be stressed. It is indeed possible that variations in pre-Socastee interglacial sea levels were of a magnitude similar to those during Socastee and Wando time, but the cumulative effects of uplift on paleo-shoreline elevation far exceed the probable variation in sea level highstand elevations and therefore prevent the detection of eustatic variations. In the absence of other evidence, and given the consistent rates of shoreline elevation during pre-Socastee time, crustal uplift is assumed to be the dominant mechanism.

COMPARISON OF THE CAPE FEAR AND PEE DEE RIVER VALLEYS

The distribution of fluvial terraces within the Cape Fear River valley is the product of tectonic warping of the Coastal Plain during the late Tertiary and Quaternary. The regional or local extent of this tectonism can be assessed by a geomorphic and lithologic comparison of the Cape Fear River valley with the next major Piedmont-draining river the the south, the Pee Dee.

Thom (1967) mapped the river terraces in the lower part of the Pee Dee River valley below Florence, S.C., and the lower reaches of the Little Pee Dee-Lumber Rivers and Waccamaw River and correlated these fluvial deposits with the series of old barrier beaches paralleling the coastline in northeastern South Carolina, near the city of Myrtle Beach. This detailed study, as well as geomorphic analyses of the Pee Dee River during the present study and recent regional mapping (Owens, in press), provides a unique opportunity for comparison of two major river systems and their responses to tectonism. A general location map including the two rivers is shown in figure 28.

Both rivers drain the Piedmont, and their length on the outer Coastal Plain is approximately 100 miles. The inner boundary of the outer Coastal Plain in the Carolinas is the Orangeburg scarp. These rivers are entrenched as they exit the higher upland northwest of the Orangeburg scarp; extensive terraces and flood plains have developed from the reduced gradients in both rivers on the outer Coastal Plain. The rivers are eroding the same units between the Orangeburg scarp and the ocean; therefore, differences in valley morphology are attributable to some influence other than lithology of the outer Coastal Plain sediments, in this case, presumably tectonism.

Obvious and striking differences in valley morphology, terrace sediments, geomorphic expression, and river hydraulics exist between the two river valleys. In the Cape Fear River valley, terraces are unpaired and decrease in age and elevation southwestward toward

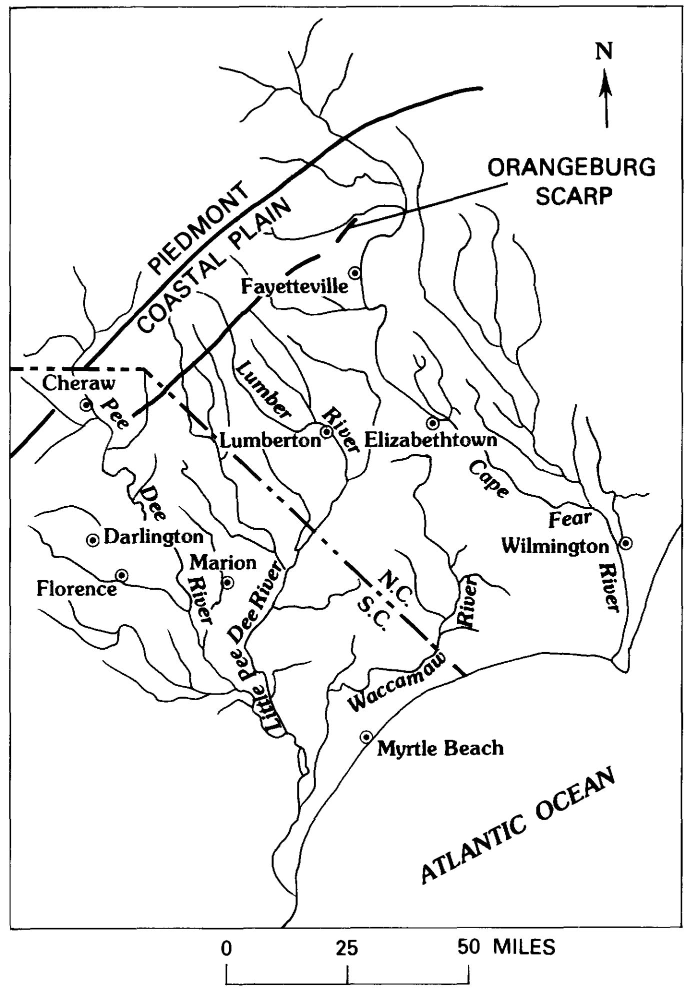


FIGURE 28.—Generalized location map showing major drainage and cities between the Pee Dee and Cape Fear Rivers, South Carolina and North Carolina.

the river; this pattern has been shown elsewhere in this paper to be the result of uplift generally from the north or northeast. The Cape Fear River flows strongly, is entrenched in its upper reaches on the outer Coastal Plain, is eroding its southwest valley wall, and has a low sinuosity value, between 1.09 and 1.18 for that portion of the profile currently being uplifted. The Pee Dee River crosses the Orangeburg scarp below Cheraw, S.C., and flows in a sluggish, meandering course to the ocean. The sinuosity is 1.56 for the first 50 miles below Cheraw; a similarly high value is approached in the Cape Fear River only as the river exits the wide portion of the valley, below LB-316.

Sediments in the Pee Dee River valley are largely very fine sand, silt, and clay (Thom, 1967), while medium and coarse sand are common in the Cape Fear River valley. This contrast in sediment load must have persisted for a great length of time, as the oldest terrace is late Pleistocene, about 2.75 million years old. Because sand was abundant in the Cape Fear River channel, dune fields and Carolina bays could develop in the valley when conditions permitted. These features are ubiquitous and well preserved in the valley; the Carolina bays are among the best examples on the Coastal Plain and have been carefully studied (Johnson, 1942; Prouty, 1952). The fine sediments of the Pee Dee River valley have not provided a source of sand adequate for the development of dunes or Carolina bays; the terrace surfaces therefore bear little resemblance to those in the Cape Fear River valley.

The terraces in the Pee Dee River valley are fewer and narrower than those in the Cape Fear River valley and are quite limited to the north of Darlington, S.C., where the river appears to migrate freely between the valley walls or flows in the center of the valley. However, downriver from Darlington, river terraces widen and are upaired, decreasing in age and elevation southwestward toward the river, as in the Cape Fear River valley.

Thom (1967) described three terraces in the lower Pee Dee River valley and its major tributaries between Florence, S.C., and the coast; this distance, 58 miles, is roughly equivalent in the Cape Fear River valley to the distance from Elizabethtown to the coast. Recent mapping by Owens (in press) subdivides the middle terrace. These four terraces are correlated with terraces in the Cape Fear River valley by their shared backbarrier-beach deposits, and rarely by geomorphic characteristics.

In the Pee Dee River valley, Thom mapped a low terrace clearly marked with large meander scars and designated this Terrace I. The fill is sandy, and dune fields occur on the surface. In the Little Pee Dee and Waccamaw river valleys, Thom noted that the gradients of Terrace I in the lower valleys becomes convex and the Terrace I surface drops below the level of the flood plain. Thom correlated Terrace I with a Wisconsinan sea level rise near, but below, present sea level. In the Pee Dee, convexity downvalley was not observed. Terrace I is correlated with the Wando-age terrace in the Cape Fear River valley, because meander scars are so prominent and unique to this terrace and because this is the youngest unit preserved in either valley. A coastal equivalent of this fluvial deposit does not exist in the Myrtle Beach area.

On the basis of regional mapping by Owens (in press), Thom's Terrace II was subdivided into a Socastee and a Penholoway-age terrace. The lower of these two terraces (Socastee Formation) is the fluvial extension of the backbarrier-barrier beach complex adjacent to the coast and underlying Myrtle Beach, at an elevation of approximately 20 feet above sea level. The upper terrace is correlated with the Penholoway Formation, whose surface parallels the shoreline inland from the Socastee Formation at approximately 30 to 40 feet above sea level. Thom referred to the nonfluvial part of the Penholoway surface as the "Conway backbarrier flat" and correlated it with Terrace III. Thom's Terrace III is mapped by Owens as the fluvial facies of the Waccamaw Formation, which lie to the southwest of the Pee Dee River outside the area mapped by Thom.

The three pre-Wando terraces in the Pee Dee River valley are lithologically different from those in the Cape Fear River valley. In the Pee Dee River valley, sediment beneath the three terraces is dominated by very fine sand, silt, and clay, Carolina bays are absent, and, while the Socastee and Penholoway terraces are broad, flat, and only slightly weathered, the Waccamaw terrace is only a narrow, weathered, and dissected remnant. In contrast, the three terraces in the Cape Fear River valley are widespread features capped by dune fields and Carolina bays and the fluvial sediment is mostly medium to pebbly coarse sand. The Waccamaw terrace in the Little Pee Dee River valley is more similar to the Waccamaw Formation in the Cape Fear River valley than that in the Pee Dee River valley; sand dominates in the Little Pee Dee River valley, Carolina bays and dunes are common, and the terrace is broad and flat. The discrepancy in weathering and dissection between Waccamaw terraces in the Pee Dee and the Little Pee Dee river valleys may be due to the narrow outcrop pattern of this terrace in the Pee Dee River valley; erosion would proceed more quickly there than on a broad, flat surface, and increased dissection improves water circulation and increases the rate of weathering. The similarity between the Waccamaw terrace in the Cape Fear and Little Pee Dee River valleys exists strictly because sand was available for deposition. The rivers, and the reason for a sand supply, are quite different: the Cape Fear River is a fast-flowing Piedmont-draining river that did not deposit its fine sediment beneath the terrace, while the Little Pee Dee River is a smaller, Coastal Plain-draining tributary that erodes sandy sediment and redeposits it nearby.

UPLAND DRAINAGE PATTERNS AND REGIONAL TECTONICS

There are two patterns of drainage on the uplands between the two river valleys, one dominantly down to the southeast and the other down to the southwest. On the upper portion of the outer Coastal Plain, northwest of a line approximately through Marion, S.C., and Lumberton, N.C., and crossing the Cape Fear River valley at the large entrenched meander south of

Fayetteville, N.C., the streams and major rivers flow generally to the southeast (see fig. 28). On the southeast side of this roughly placed line, drainage has a less preferred orientation but seems to be generally to the southwest, especially in the Little Pee Dee-Lumber River and the Waccamaw River. The course of the Pee Dee River below this line also appears to reflect some impetus to flow in a southwesterly direction; it is deflected to the southwest valley wall and assumes a more southerly course as it erodes into the upland sediments. This area of dominantly southwest drainage extends to the northwest across the Cape Fear River valley in the region where a sustained tilt, up from the north or northeast, has been demonstrated. Although the Cape Fear River flows southeast downriver from this line, it is migrating southwestward owing to the tilting.

The relative ages of these two drainage patterns can be inferred, and if the streams are not wholly controlled by the preexisting topography, the age and nature of the tectonic forces that shaped the current drainage patterns can also be assessed, as was done for the Cape Fear River valley in another section of this paper. Topographic control of drainage between the Pee Dee and Cape Fear Rivers is quite possible, however, and the southward projection of tectonic uplift measured in the Cape Fear River valley is conjectural. The similarity between drainage patterns around the Cape Fear River and to the south is worth some conjecture since uplift has been established for part of this area.

Several terraces are preserved on the northeast side of the Pee Dee River valley south of Darlington, as mentioned earlier, and in the Little Pee Dee-Lumber River valley. This suggests that the rivers have been migrating south and southwest during the interval between formation of the oldest terrace and the present. This conclusion is supported by this study of the Cape Fear River valley, where uplift from the northeast has been sustained for a period long enough to preserve five terraces. In contrast, river valleys in the area of predominantly southeast drainage are filled with Socastee-age to Holocene sediments; older fluvial sediments do not occur except as narrow benches along the flanks of the Pee Dee River valley. This drainage is therefore considered younger than the southwest pattern. If this southeast drainage has resulted from tectonism, the requisite north to northwest uplift, from the direction of the Piedmont, postdates the tectonism postulated for the southwest drainage pattern. This conclusion is also supported by the tectonic history of the Cape Fear River valley.

Some of the tectonic forces that have shaped the Cape Fear River valley can be extrapolated southward on the basis of gross geomorphic evidence, as outlined above, and noting the probable topographic control of tributary drainage. The magnitude of these forces appears to be subdued to the south, as would be expected farther from the source of uplift. The Cape Fear River flows on the axis or southern limb of an uplifted area; the Pee Dee River is much farther removed from the area of maximum uplift and has responded less dramatically than the Cape Fear River owing to the slower rate of uplift. The differing uplift rate and sediment loads in the two rivers have been responsible for the sharply differing nature of the two valleys; dunes and Carolina bays dominate the landscape of the Cape Fear River valley and the surface is well drained, while the Pee Dee River valley is less well drained and generally is lacking in dunes and Carolina bays.

SUMMARY

The purpose of this study was to investigate the geologic history of a major river valley on the Atlantic Coastal Plain. The Piedmont-draining rivers have deposited large quantities of sediment; they are geologically significant features, and worthy of study. The geologic record preserved in these river valleys has been largely unknown; therefore, many of the conclusions reached in this study cannot be compared with other river systems but instead await integration into, or comparison with, future research.

While the geomorphology of the Cape Fear River valley revealed a wealth of data on the configuration of the terraces, and on the magnitude, location, and age of uplift in the area, the importance of mineralogic analysis of the sediments cannot be emphasized too greatly. In the current study, geomorphic inspection was used to identify a series of sloping surfaces near the Cape Fear River, which by inference may be considered river terraces. In several cases, these terraces could be correlated with nearby strandline deposits on the basis of elevation, which, as demonstrated in this paper, can be a tenuous assumption in a region with a tectonic history. Because of the limitation of geomorphic correlations, the mineralogy and lithology of the deposits were studied carefully to provide evidence for the relative ages of the various units and the depositional environment. Mineralogy has been shown to be a useful correlation tool, and the mineralogy of the marine and strandline deposits is known in several places on the Coastal Plain. Using a combination of lithologic, mineralogic, and geomorphic data, the Cape Fear River valley was studied and the following conclusions were reached.

1. Five river terraces are present between Fayetteville and Wilmington, N.C. Several late Tertiary

and Pleistocene depositional events have been previously identified and dated in the Carolinas; the terraces were with these age-dated formations. Fluvial deposition occurred roughly at 2.75 Ma (Bear Bluff age), 1.75 million Ma (Waccamaw age), 750 ka or more (Penholoway age), 200 ka (Socastee age), 100 ka (Wando age), and during the Holocene.

2. Based on the age of the oldest fluvial terrace, an ancestral Cape Fear River began draining the region at least 2.75 Ma (late Pliocene). The maximum age of this river can be inferred; no fluvial terrace is preserved that can be correlated with the next older unit, which is about 3.25 Ma. However, it is possible that the Cape Fear River is much older, and that prior to 2.75 Ma the region was tectonically stable; an absence of cross-valley tilt would cause erosion of the 3.25-Ma fluvial deposit during the next depositional episode.

3. The terraces are unpaired, and all lie northeast of the modern Cape Fear River, which is eroding into the uplands on the southwest wall of the valley. The youngest terrace lies adjacent to the river, with the age and elevation of the terraces increasing away from the river, toward the northeast. This map pattern suggests migration of the Cape Fear River, to the southwest, since the late Pliocene. This is supported by the pattern of tributaries; they are numerous and often large on the northeast (valley) side of the river, while, owing to stream capture, tributaries on the southwest margin of the valley are quite few and limited in development.

4. The valley is astride the Cape Fear arch, and has been tectonically warped by movement along the arch throughout the latest Tertiary and Pleistocene. The response of the Cape Fear River to tectonism was modeled by geomorphic analysis of the fluvial surfaces. The tectonic forces can be defined as follows. A gentle uplift centered to the north or northeast of the valley that persisted from at least the late Pliocene to late Pleistocene (from more than 2.75 to less than 100 ka) has caused a southwestward migration of the Cape Fear River and consequent preservation of fluvial terraces to the northeast of the river. Normal to this flexure is another flexure, along the valley (and arch) axis. Relative uplift and subsidence have occurred along this trend. Since at least Penholoway time (at least 750 ka), the valley northwest of Elizabethtown, N.C., near the Piedmont, has been uplifted while relative subsidence is likely for the river's lower course. Prior to this episode, the upper valley did not undergo uplift in this direction and seems to have subsided relative to the south. These local flexures, inferred by geomorphic data, correlate with a localized zone of relatively intense uplift; this local phenomenon is superimposed on a more gentle, widespread uplift of the arch that elevated the deposits along the Coastal Plain and prevented their complete erosion by subsequent transgressions.

5. Deposition on the Coastal Plain occurred during interglacial highstands of the sea, which were of roughly equal magnitude. Geomorphic analysis of fluvial terraces seems to support this assumption and indicates that a gentle, persistent crustal uplift has occurred at least since deposition of the oldest terrace.

6. The series of flexures identified in the Cape Fear River valley also seems to explain the nature of the Pee Dee River valley. The Pee Dee River has deposited finer sediments than the Cape Fear River, is not entrenched on the outer Coastal Plain, and has produced a terrace pattern similar to the Cape Fear River valley in only one stretch of the river. The Pee Dee River flows far down the southern limb of the Cape Fear arch, and the effects of arch uplift are minimal here. Strictly based on geomorphic reconnaissance, the effects of flexures on the arch seem to be manifested in the drainage pattern of tributaries on the uplands to the south of the Cape Fear River and in the Pee Dee River valley, and in the lack of Carolina bays in the Pee Dee River valley due to the predominance of finer grained sediment.

7. Sediments under terraces in the Cape Fear River valley generally consist of tan to pale gray, pebbly, coarse-grained quartzose river channel sands that fine upward to silty fine sand or clayey overbank deposits. Roughly 22 to 30 feet in thickness, the fluvial sediments are usually capped by dune sand of variable thickness. Sediments under the flood plain are finer overall than under the terraces. The flood plain is generally narrow and is nearly absent north of Elizabethtown. Wood sampled at a depth of 35 feet, just above the base of the flood plain at Elizabethtown, N.C., is ~3,540 years old. Apparently, the majority of the flood-plain sediments were rapidly deposited as the river backfilled during the Holocene sea level rise.

8. The abundance of hornblende, epidote, and feldspar in fluvial samples is much greater than in upland samples. The uplands are characterized by backbarrier and barrier beach facies whose sediments have been reworked extensively. The upland sediments are also generally older than the fluvial deposits. The mineralogic differences between upland and fluvial sediments are therefore considered to be due to age and reworking.

9. Hornblende, epidote, and feldspar are less abundant beneath older terraces than younger ones, and less abundant farther downvalley. In the upland deposits, the abundance of labile minerals decreases upward in the section; in the valley, hornblende, epidote, and feldspar are less abundant on older terraces than on younger ones, but their abundance does not decrease toward the surface in most drill holes. In contrast to the

vertical percolation of water that normally leaches surficial units (e.g., upland sediments) from the top down, the fluvial deposits are generally capped with silt or clay and floored by fine-grained Cretaceous sediments, and behave somewhat as confined aquifers. Water movement is subhorizontal, and intrastratal solution of minerals occurs throughout the fluvial interval, not just in the upper portion. The older terraces have undergone intrastratal solution for a longer period of time than the younger terraces and are more depleted in labile minerals. The downvalley decrease in these minerals is due to the effects of dilution, as Coastal Plain-draining tributaries contributed resistant minerals to the Cape Fear River channel.

10. Clay mineralogy of fluvial deposits is uniform and therefore is not useful in distinguishing the various terraces. The suite contains kaolinite and mixed-layer expandable (with ethylene glycol treatment) clay, with minor vermiculite and gibbsite.

11. The clay-sized fraction of upland and fluvial sediments is mineralogically distinct; the clay fraction in upland sediments either has more vermiculite and less mixed- layer clay than the fluvial sediments, plus kaolinite, gibbsite, and goethite, or contains only kaolinite. In those samples of upland deposits containing only kaolinite in the clay fraction, the upper portion of the profile has probably been stripped off, leaving a clay mineral assemblage in the truncated profile that is unusually immature for the age of the deposit. The more mature, intact profiles contain gibbsite, goethite, and vermiculite and indicate moderate to intense desilication.

12. Dune sands, derived from upland sediments and river terraces, are readily distinguished from the underlying fluvial deposits on the basis of lithology, and sand and clay mineralogy. Labile sand-sized minerals are less abundant in the dunes than in the fluvial sediments. In addition, the abundance of labile minerals is lower in dunes farther from the Cape Fear River. More than one generation of dunes are therefore likely on the terraces. Well-crystallized vermiculite, gibbsite, and quartz are present in the clay-sized fraction of dune sands, but not in fluvial sediments. These clay-sized minerals have been eroded and transported from older, weathered sediments in the uplands.

13. The organic acids in humate-rich sediments have leached the labile sand-sized minerals and all clay-sized minerals except quartz. The chemical environment in these humate-rich sediments is much different from that in peat-rich sediments. Leaching of minerals in peat samples is no more intense than in adjacent, organic-poor samples.

14. At least two generations of dunes are present in the valley. The younger dune event has been dated at less than 7,700 yBP from a peat beneath the dune sheet; these dunes are confined to the southwest edges of the Wando and Socastee terraces and contain an immature sand mineral assemblage. These dunes were a short-lived phenomenon, as sand was available from the incised Cape Fear River flood plain only until approximately 3,540 yBP, when backfilling of the channel began. The older dune sheet(s) cover the Socastee and older terraces in thicknesses of 3 to 22 feet. Peats from the bases of these dunes and Carolina bay rims are older than 35,000 years, and the sand mineralogy is mature. It was not possible to subdivide these dunes into two or more generations, but the possibility of multiple ages remains.

15. If all Carolina bays were formed by a single, unique climatic event, then the bays formed after deposition of the Socastee terrace and prior to deposition of the Wando terrace (between roughly 100 and 200 ka) and are equivalent in age to the older dunes in the valley. If there are several generations of Carolina bays, then the above age is still valid for those bays on the Socastee terrace but is a minimum age for all other bays.

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APPENDIX: DRILL-SITE LOCATIONS

Donoho Creek Landing	Bolton 15' quad. Cutbank on southwestern side of Cape Fear River, just downriver from confluence with Donoho Creek.
LB-273	Elizabethtown 15' quad. Follow rte. 701 north from Elizabethtown approx. 1 mile to Suttons Corner. Hole was on left before intersection.
LB-274	Elizabethtown 15' quad. Hole was at junction of rtes. 701 and 53, approx. 1 mile west of White Lake on the northwest rim of Wam Squam Bay.
LB-275	White Lake 15' quad. Follow rte. 41 east from White Lake to dirt road just north of Causeway Bay. Turn right. Hole was less than 0.1 mile before intersection on right side.
LB-276	White Lake 15' quad. From LB-275, follow rte. 41 east to Smith Crossroads. Hole was just before intersection on right side.
LB-277	White Lake 15' quad. From Smith Crossroads, follow rte. 41 east for 0.5 mile. Turn left onto dirt road. Hole was on right side just after the turn.
LB-278	White Lake 15' quad. Follow rte. 41 less than 0.2 mile east of Tomahawk. Hole was on left, just past paved road to the left.
LB-279	Garland 15' quad. South of Clinton, turn left off rte. 701 approx. 1.5 miles south of junction with rte. 421. Take first road to the left, approx. 0.8 mile. Hole was 0.1 mile ahead on right side of the road.
LB-289	Bolton 15' quad. From intersection with rte. 74-76, go north on rte. 141 for 0.6 mile to a road on the left. Hole was on the left (west) side of rte. 141, 0.1 mile past the road.
LB-290	Bolton 15' quad. From intersection with rte. 87, go south on rte. 141 approx. 1.5 miles, past a paved road to the left, to a dirt road on the left. Hole was on left side of rte. 141 just past dirt road.
LB-291	Bolton 15' quad. From intersection with rte. 87, go north on rte. 141, across Cape Fear River, approx. 3 miles. Hole was on right side of rte. 141, approx. 0.1 mile past a dirt road on the left.
LB-292	Acme 15' quad. From intersection with rte. 210, go south on rte. 141 approx. 0.8 mile. Hole was on left side of the road just past paved road.
LB-293	Acme 15' quad. From intersection with rte. 210, go north on rte. 141 approx. 1 mile. Hole was on right side of road approx. 0.1 mile past dirt road.
LB-294	Acme 15' quad. From intersection of rtes. 53 and 141 at Long View, go north on rte. 53 approx. 1.6 miles to a dirt road and sand pit on left, at top of rise. Hole was on left side of road just past dirt road.
LB-295	Elizabethtown 15' quad. From White Oak, follow rte. 53 southeast approx. 1.6 miles to dirt road. Hole was 0.1 mile past dirt road on left side of rte. 53.
LB-296	Roseboro 15' quad. From White Oak, follow major road trending northeast, parallel to Ellis Creek, for approx. 4 miles. Hole was on left side of road, on southeast rim of Bushy Bay.
LB-297	Roseboro 15' quad. Follow rte. 242 north from Ammon, approx. 3.3 miles. Hole was on left side of road, on southeast rim of Mill Bay.
LB-298	Roseboro 15' quad. South of Roseboro. From intersection with rte. 210, go north on rte. 242 approx. 3.5 miles, up the rise to a dirt road on right. Hole was at intersection.
LB-309	Elizabethtown 15' quad. From Elizabethtown, go north on rte. 701 and turn left just before the Cape Fear River bridge. Hole was on the flood plain, at end of road next to the Cape Fear River.
LB-310	Elizabethtown 15' quad. From Elizabethtown, go north on rte. 242 to Jones Lake. Turn left on a secondary road approx. 1.5 miles past Jones Lake. Hole was approx. 0.9 mile from intersection, northeast of Salters Lake.
LB-311	Bladenboro 15' quad. From Clark Chapel, go northeast on first major road past Tarheel Landing and across the Cape Fear River. Turn left onto dirt road and double back toward river. Hole was in a clearing on right side, approx. 0.1 mile from the Cape Fear River.
LB-312	Elizabethtown 15' quad. From Shiloh Church (west of McNeil Bay), go southwest on main road approx. 1.2 miles to intersection with main road and turn left. Hole was on right approx. 0.4 mile from intersection.

LB-316 Bolton 15' quad. Follow rte. 53 northwest from Kelly approx. 0.4 mile, then turn left. Go 0.8 mile to a T-stop. Hole was on right, at intersection.

LB-317 White Lake 15' quad. From intersection of rte. 33 and Whitehall Road, go northeast on Whitehall Road approx. 1.8 miles to south rim of Tedder Bay. Hole was on left side of road.

LB-318 Acme 15' quad. From intersection with rte. 210 near Still Bluff, follow Caintuck Loop Road south approx. 0.7 mile to dirt road on left. Follow road approx. 2 miles after sharp turn to right (due south) and go approx. 0.2 mile to dirt road on left. Hole was at intersection.

LB-319 Acme 15' quad. From Currie, Follow main road north (not rte. 210) approx. 0.8 mile to intersection. Go northeast on main road approx. 1.5 miles, past Black Swamp (bay) to intersection. Stay on main road. Hole was approx. 0.1 mile past intersection on right side.

LB-321 Acme 15' quad. Follow rte. 421 south from intersection with rte. 210 approx. 4 miles, past Goose pond, to a dirt road on right. Hole was approx. 0.1 mile onto dirt road near bed of old Atlantic Coast Line.

LB-322 Wilmington 71/2' quad. Go south on rte. 421 to exit for U.S. North Carolina Battleship Memorial, go toward battleship, turn right onto paved service road that parallels the Cape Fear River, and proceed approx. 1.5 miles. Hole was located on left side of road, between Eagle Island landfill and industrial development to the north. Area was undisturbed by excavations.

LB-380 Saint Pauls 15' quad. From Cumberland Church (west of Big Alligator Swamp), go west on road toward the Cape Fear River approx. 0.7 mile. Turn left onto road. Hole was approx. 0.1 mile from intersection, near a cemetery.

LB-381 Saint Pauls 15' quad. From Cumberland Church, go east on Johnson Road, toward Big Alligator Swamp. After approx. 1.2 miles, the road bends to the right. Follow for 0.2 mile to dirt road on left. Hole was on Johnson Road, approx. 0.1 mile past dirt road on right.

LB-382 Roseboro 15' quad. From the intersection of Johnson Road and rte. 53, go north on rte. 53 approx. 0.5 mile, around the northwest side of Black Ground Bay, to a dirt road on the right. Turn right. Hole was approx. 0.1 mile down dirt road, on the northern rim of Black Ground Bay.

LB-383 Roseboro 15' quad. From the intersection of Cedar Creek-Stedman Road and rte. 210, go southwest on Cedar Creek-Stedman Road approx. 0.4 mile to a right bend in the road, between Buckhorn and Harrison Creek bays. Hole was on left side of road, just after the bend.

LB-384 Roseboro 15' quad. From Autryville, go southeast on rte. 24 approx. 1 mile to road across from Piney Pocosin. Turn right. Hole was on left side less than 0.1 mile from intersection, just across the railroad track.

LB-385 Garland 15' quad. From Garland, go north on rte. 411 approx. 4 miles and turn onto major road on left to Parkersburg. Hole was less than 0.1 mile from intersection, on right side.

LB-386 Saint Pauls 15' quad. From Newlight School, go north on major road, paralleling the Cape Fear River lying to the west, for approx. 3 miles to intersection (the road to the right heads toward Bakers Lake and Thoroughfare Bay). Proceed on major road approx. 0.5 mile. Hole was on left side of road.