REPORT ON THE PROPOSED "NEUSE FAULT"

Prepared for
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1.0 INTRODUCTION

Ferenczi (1959) was first to suggest the existence of a fault in the region of the Neuse River. He referred to it as the "Cape Lookout-Neuse River Fault Zone". His evidence was based on the following arguments:

1) The Castle Hayne Formation (of Eocene age) was not deposited north of the Neuse River. However, since then Castle Hayne outcrops have been mapped north of the Neuse River (Brown and others, 1972; Baum, 1981 and Otte, 1981).

2) The occurrence of silicified zones along the strike of the "fault zone". More recently, however, Otte (1981) has shown that the silicified zones are not restricted to the alignment of the postulated fault. Silicification resulted from the diagenetic alteration of siliceous sponges (Otte, 1981).

3) The abrupt change in depth to basement as recorded from two wells 24 kms apart (at Havelock and Moorehead City). More detailed work on depth to basement maps in this region shows that the change in depth is not oriented perpendicular to the trend of the postulated fault and bears no relation to it (Brown and others, 1972).

Independently, and without reference to Ferenczi (1959), Gibson (1967 and 1970) suggested the existence of a "northwest-southeast positive element" parallel to the Neuse River. Evidence for this positive element was based on isopachous mapping and structural contouring of the Yorktown Formation (of Miocene age). More recent work by Baum

(1981) indicates that if a "Neuse Fault" exists, it had no discernible effect on the deposition of the Trent Formation (of Oligocene age). The Trent Formation is equivalent to the River Bend Formation of Ward and others (1978) and both are older than the Yorktown Formation (Miocene).

Baum and others (1978) citing the work of Ferenczi (1959) and Gibson (1967 and 1970) espoused the existence of a "Neuse Fault" (equivalent to the Cape Lookout-Neuse River fault zone). In later publications Harris and others (1979a and b) proposed a change in the location and orientation of the Cape Lookout-Neuse River fault zone (the Neuse fault). In so doing, they invalidated the third argument of Ferenczi (1959) for the existence of a fault. The new position resulted in having the two deep wells to basement on the same (north side) of the fault (Figure 1).

To date, the literature does not advance any evidence of observed faulting, displacement, recognizable surface expression or associated seismicity that is directly or indirectly attributable to movement on a "Neuse fault". The postulated evidence for faulting so far was either disproved or is presently disputed by experts in Coastal Plain geology of North Carolina (Brown and others, 1972; Otte, 1981; Jones, 1982; Berggen and Aubry, preprint on file; Hazel and others, preprint on file). Therefore, in Ebasco's opinion, the indirect arguments that have been presented so far do not in any manner support the existence of a "Neuse Fault" in the Coastal Plain of North Carolina.

The evidence for and against the existence of a "Neuse Fault" is summarized below and discussed in greater detail in the remainder of this report.

1.1 POSTULATED EVIDENCE FOR THE EXISTENCE OF A NEUSE FAULT

1. Abrupt change in depth to basement between Havelock and Morehead City, N. C. (Ferenczi, 1959).

2. The restricted spatial distribution of the Castle Hayne Formation (Eocene) to the south side of the Neuse River (Ferenczi, 1959).

3. The occurrence of silicified zones along the alignment of the Neuse Fault (Ferenczi, 1959).

4. The spatial distribution, petrology, and correlation of units of Eocene age (Baum and others, 1978 and Harris and others, 1979b).

5. The thickening of Cretaceous units north of the Neuse Fault (Harris and others, 1979b).

6. The present tilted attitudes of the Dupin Plain of early Pliocene age (9 m in 60 km), the Wacamaw-Canepatch Plain of Plio-Pleistocene age (2 m in 60 km), and the Socastee Plain, approximately 32,000 years old (4.6 m in 12 km) between the New River and Wilmington, N.C. (Zullo and Harris, 1979).

1.2 EVIDENCE AGAINST THE EXISTENCE OF A NEUSE FAULT

1. The change in depth to basement between Havelock and Morehead City, in light of additional well data, is neither coincident nor associated with the Neuse Fault (Brown and others 1972).

2. The Castle Hayne limestone has been shown to occur north of the limits known to Ferenczi (Brown and others, 1972 and Otte, 1981).

3. Silicified zones are not restricted to the alignment of the postulated zone and are the result of diagenetic alteration of silicious sponges (Otte, 1981).

4. Many stratigraphic experts dispute the stratigraphic subdivisions of the Castle Hayne and the Upper Eocene age assigned to its upper units by Baum, Harris, and Zullo (1978, 1979b) and Harris and Zullo, (1980). Brown and others (1972), Ward and others (1978), Jones (1982), Berggen and Aubry (preprint on file), Hazel and others (preprint on file) consider the entire formation to be Middle Eocene. Facies changes and age relationships are not well enough agreed upon to define the specific depositional basins which Baum, Harris and Zullo infer were created by movement along the Neuse Fault. Even if Baum, Harris and Zullo's stratigraphic and age interpretations are accepted the spatial disposition of their units do not require movement along the postulated Neuse Fault to explain the prevalent depositional environment.

5. Structural contours on top of basement and on top of Cretaceous units (Brown and others, 1972) do not show any evidence of movement along the postulated Neuse Fault.

6. The original slopes of the Plio-Pleistocene Coastal Plain terraces are not shown to have been initially horizontal. Given the extremely low tilts that are invoked the original attitudes of the plains must be demonstrated first in order to validate the conclusions that are reached.

7. The region of the postulated Neuse Fault is aseismic. (Figure 2.5.2-1 -2, SHNPP FSAR).

2.0 DISCUSSION OF THE USE OF THE TERM NEUSE FAULT

The first person to propose a northwest trending fault parallel to the Neuse River in the Coastal Plain was Ferenczi (1959). Ferenczi called the feature the Cape Lookout-Neuse Fault and gave three lines of evidence to support his conclusions:

1. A difference in depth to basement across the fault based on 2 wells 24 km apart, one at Havelock and the other near Morehead City. Brown and others (1972) using additional well data generated a top of basement contour map which show the maximum slope change perpendicular to a north-south axis and not perpendicular to the proposed trend of the Neuse Fault. He interpreted the change in basement surface elevation to a steepening of slope away from the Cape Fear Arch.

2. Ferenczi also thought that the Castle Hayne Formation was not deposited north of the Neuse River and that the Neuse Fault provided a structural boundary limiting the basin of deposition, Brown and others (1972) and Otte (1981) both refer to outliers of the Castle Hayne beyond this boundary.

3. Finally, Ferenczi interpreted the occurrence of silicified Eocene outcrops aligned along his fault zone as evidence of faulting. Otte (1981) showed that the Wayne County outcrops are silicified because of the presence of silicious sponges which provided a ready source of silica. He also observed that the silicified sediments are more widespread than Ferenczi realized and are not restricted to his fault alginment.

The next published reference to the "Neuse Fault" occurs in Baum and others (1978). This is primarily a biostratigraphic paper. However, the authors by referring to the work of Gibson (1967), who identified a positive element trending parallel to the Neuse River, and the work of Ferenczi (1959), consider this sufficient evidence to use the term Neuse Fault without providing any additional supporting evidence. Baum and others (1978) use the postulated Neuse Fault as part of a model to explain the distribution of the Eocene to Miocene strata and facies changes in the Coastal Plain of North Carolina.

It is important to distinguish here the difference between postulating a fault to create a model which will explain the deposition of the strata and proving the existence of a fault based upon stratigraphic evidence. At no point have any strata been shown to be offset by the Neuse Fault. The problems with the Ferenczi's work are discussed above and Gibson (1967, 1970) does not invoke faulting as an explanation for his "positive element". Thus the use of the term fault by Baum, Harris and Zullo (1978), Harris and others (1979a and b) and Harris (1982) should not be regarded as proof of its existence but merely as one convenient explanation of the distribution of Tertiary sediments in the Coastal Plain of North Carolina. At this point it should also be noted that there are alternate models which explain the distribution of Eocene and Miocene formations in the Coastal Plain of North Carolina, without recourse to faulting along the Neuse Fault. Brown and others (1972) propose that the stratigraphic framework and spacial distribution of the Atlantic Coastal Plain is controlled by northeast and north/south trending hinge zones. Gibson (1967, 1970) postulates a positive element north of the Neuse Fault, during the deposition of Miocene strata, however he does not attribute this positive element to a northwest trending fault. Otte (1979, 1981) attributes the facies distribution and thickness of the exposed Eocene Castle Hayne Formation to structural control by the Cape Fear Arch and pre-existing topography.

The U.S. Army Corps of Engineers (USCOE) in its Phase I Report on Earthquake Design Analysis of Philpott Dam (1982) refers to papers on the subject published in the Field Trip Guidebook of the Carolina Geological Society and Atlantic Coastal Plain Geological Association

(Baum, Harris, and Zullo, 1979, editors). The USCOE reports and adopts, without discussion, the position espoused by Ferenczi (1959), Baum and others (1978), Harris and others (1979) and Baum and others (1979).

Within the guidebook are two papers which discuss postulated tectonic movements along the Neuse Fault. These papers are: 1) Harris and others (1979b), Tectonic effects on Cretaceous, Paleogene, and early Neogene sedimentation, North Carolina, and 2) Zullo and Harris (1979), Plio-Pleistocene Crustal Warping in the Outer Coastal Plain of North Carolina. These are the papers which label all the lines on the map (Figure 1) as "faults". The second paper by Zullo and Harris (1979) proposes that the Neuse Fault moved in the Quaternary. As will be discussed below, the conclusions of these papers are in conflict with those of other workers and their evidence for movement along the proposed Neuse Fault is not sufficient to substantiate faulting.

2.1 Discussion of paper by Harris, Zullo and Baum (1979b).

This paper (Harris and others, 1979b) is controversial with respect to the Eocene, Oligocene and Miocene stratigraphy. Ward and others (1978) and Brown and others (1972) are a few of the workers who had previously published their versions of the stratigraphic correlation between the same rock units. Since the publication of the guidebook, the controversy has continued with publications by Baum (1981), Harris and Zullo (1980, 1982) and Harris (1982) on one side and Jones (1982), Berggen and Aubry (preprint, on file) and Hazel and others (preprint, on file).

Harris and others (1979b) major evidence for movement of a "Neuse Fault" in the Paleogene is the distribution of the New Bern Formation (as defined by Baum and others, 1978), which is restricted to the area north of the Neuse Fault. Harris and others (1979b) consider the New Bern Formation to be latest Eocene (Jacksonian in age) and younger than the Castle Hayne Limestone as they define it. Harris and others (1979b) also state that these strata represent "a major lithologic change from a carbonate dominated regime to a clastic dominated regime", a change they interpret as caused by faulting of the Late Eocene Castle Hayne during the latest Eocene and the deposition of the latest Eocene New Bern Formation in the resulting structural low north of the "Neuse Fault". Ward and others (1978) do not recognize the New Bern Formation as being a separate formation from the Castle Hayne and call it the Spring Garden Member (Middle Eocene) of the (Middle Eocene) Castle Hayne Formation. Although Harris and others (1979b) use the Neuse Fault as an explanation for the restricted distribution of the rock they call the New Bern (upper Eocene), the distribution of the New Bern Formation itself is not primary evidence of faulting and such a conclusion is especially tenuous if the age relationships (middle or late Eocene) are in question. Cook and Macneil (1952), Brown and others (1972), Ward and others (1978), Jones (1982), Berggen and Aubry (preprint on file) and Hazel and others (preprint on file) consider the Castle Hayne, which includes the "New Bern" of Baum and others (1978), to be middle Eocene in age. If this interpretation is accepted the shallow water facies of the Castle Hayne, north of the postulated fault, is only a facies of the Castle Hayne limestone south of the

postulated fault. As a result, no intervening fault needs to be evoked to explain what is a normal stratigraphic transition.

Harris and others (1979b) also state that the restriction of the middle Miocene Pungo River Formation to the area north of the fault indicates that the "Neuse Fault" was active in the middle Miocene. The distribution of the Pungo River Formation is limited not only to the north of the proposed Neuse Fault, but the western boundary of the formation strikes north-south and is entirely east of the proposed Neuse Fault (Gibson, 1967) (Miller, 1982). Although Gibson (1967) proposes a positive feature north of the "Neuse Fault" as being responsible for the restricted deposition of the Pungo River Formation, he does not call it a fault. Miller (1982) attributes the restricted deposition of the Pungo River Formation to the north-south hinge line of Brown and others (1972) which is parallel to the strike of the formation and coincident with its western boundary. In light of the detailed work done by Miller (1982) the conclusions of Harris and others (1979b) and Harris (1982) cannot be considered evidence of movement along the "Neuse Fault" in the middle Miocene. The Oligocene Trent, Silverdale and Belgrade Formations (River Bend and Belgrade Formations of Ward and others 1978) are older than the Pungo River Formation and closer to the "Neuse Fault" than the Pungo River Formation. The Oligocene Formations do not appear to be related to tectonic activity according to Harris and others (1979b).

Elsewhere in their paper Harris and others (1979b) discuss Cretaceous movement of the "Neuse Fault". Their conclusions are based upon a structural contour map of the top of Cretaceous unit F and on an isopach map of the same unit F which immediately overlies the basement in North Carolina. Both maps were generated by Brown and others, (1972). Harris and others (1979b) propose that since unit F thickens considerably north of the Neuse Fault and south of the Cape Fear Arch, which they also call a fault, the area between the two features was positive between these two proposed faults during deposition of unit F, resulting in thicker deposits in the basins to the north and south. They believe that the movement was syn-depositional because the structural contour map of the top of unit F does not exhibit any structural relief in the vicinity of the Neuse Fault. The absence of structural relief on the top of unit F is not only good evidence that there was no movement immediately after deposition of unit F in the region of the proposed Neuse Fault but that there was no movement ever along the proposed Neuse Fault after the deposition of unit F.

Brown and others (1972) also prepared a structural contour map of the top of basement rocks which immediately underlie unit F. Although this map is not mentioned by Harris and others (1979b), it does not show any structural relief along the alignment of the Neuse Fault either, implying that there has been no movement along the Neuse Fault since the Cretaceous.

In comparing the figures from both papers, Harris and others (1979b) have apparently mislabeled a contour line on the isopach map of unit F. A contour line north of the "Neuse Fault", which should be labeled 500 m, is labeled 1000 m on the figure of Harris and others (1979b).

Although Cretaceous unit F does thicken north of the proposed Neuse Fault, the Neuse Fault does not coincide with the greatest change in thickness of the unit.

Most workers consider the Cape Fear Arch to have exerted a major structural control over Cretaceous and younger deposition in the Carolina Coastal Plain. The theory that a proposed Neuse Fault was also active in the Cretaceous and Tertiary appears to be both unsubstantiated and unnecessary.

2.2 Discussion of paper by Zullo and Harris (1979)

Zullo and Harris (1979) submit that the proposed Neuse Fault was active throughout the Tertiary and Pleistocene. The arguments of the authors are based upon the identification of Plio-Pleistocene marine scarps and terraces in the area between the New River and Wilmington, N.C. (see Figure 1) and the measurement of the elevation of the marine terraces at points that are distant from each others. They conclude that because the terraces are not at present, uniformly horizontal plains, but rather, are slightly tilted along a northeast-southwest axis, perpendicular to the proposed Neuse Fault, they were tilted by tectonic activity, specifically by movement along the proposed Neuse Fault in the last 32,000 years.

In order to discuss the ramifications of Zullo and Harris' paper a brief digression on Coastal Plain scarps and terraces is presented below:

Numerous workers (for example: Flint, 1940 and 1941; Cooke, 1941; Daniels and others, 1966 and Oakes and Dubar, 1974) have described erosional marine scarps and associated shoreline features which record former higher sea level stands on the North Carolina Coastal Plain. A series of at least three marine scarps are found between the modern coast and the edge of the Coastal Plain, up to elevations of about 90 m. They can be most easily identified on topographic maps and areal photographs. The terraces or plains (with slopes that are less than one meter per kilometer) are interconnected by scarp faces (with slopes on the order of 15 m per kilometer). These scarps are difficult to recognize in the field, but are fairly obvious when compared to the average slope of the North Carolina Coastal Plain (with slopes that are less than one meter per kilometer) (Daniels and others, 1966). Evidence for a marine erosional origin of the scarps by wave action, during relatively stable sea level stands, includes their arcuate nature, the persistence of the scarps over tens and even hundreds of kms, the consistency of the scarp toe elevations over these distances, and the deposition of marine units seaward of these scarps. Although the toe elevations of the scarps are generally remarkably uniform, the height of the scarps may not be (Wheeler and others, 1979).

The terraces or plains between the scarps are commonly formed by either erosional or depositional processes and thus may be underlain by deposits laid down during the occupation or retreat of the sea level stand which cut the scarp. Also, they may be underlain by older sediments which were modified by the transgressing or regressing sea. In either case, both shoreline features such as dunes, bars and channels and subaerial/fluvial processes such as stream erosion may modify the surfaces. Some of these features are beautifully shown on the aerial photographs in Mixon and Pilkey (1976) and on Landsat imagery.

The underlying assumption of Zullo and Harris' paper is that the "Duplin" plain, the "Waccamaw-Canepatch" plain and the "Socastee" plain were formed as horizontal surfaces. They conclude that the presently observable slopes and slope directions on the plains indicate that episodic and differential uplift have occurred in the region. However, the assumption of original horizontality of the plains is not substantiated. Topography along the present day Atlantic margin slopes offshore and is modified by bars and channels, only the actual contact of the shoreline and the sea may represent a near horizontal surface (toe of the scarp). Furthermore, Zullo and Harris (1979) do not define how they measured the average slope of their plains. A cursory examination of 7 1/2 minute topographic maps of the area confirms the existence of fairly uniform and slightly sloping plains, but does not indicate that the north-south elevation changes are of sufficient magnitude to represent the top of initially level horizontal planes which have been tilted by tectonic activity (see Sections 1.1 and 1.2). The tilt of the plains can also be explained by primary depositional slopes of an offshore marine area, and/or subaerial or subaqueous post depositional modifications, since the area is incised by tributaries of the New and Cape Fear Rivers.

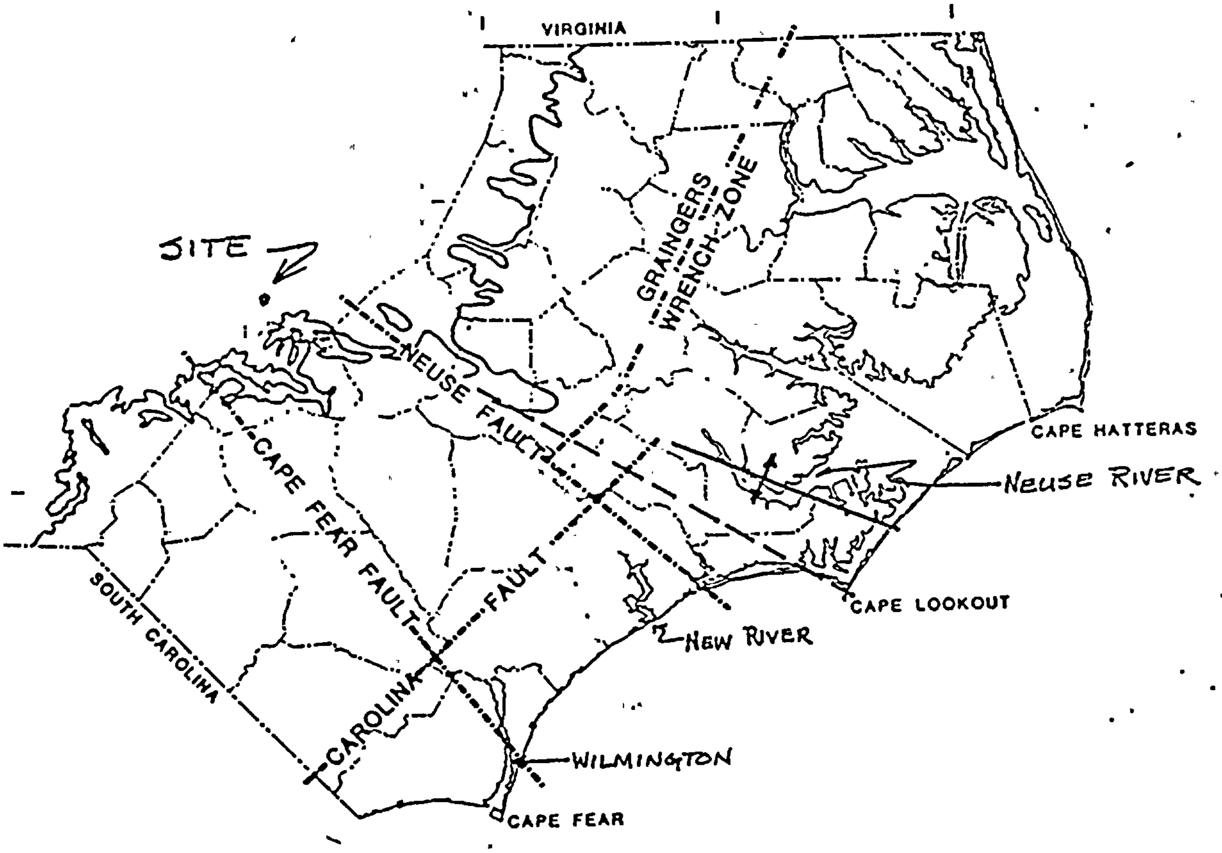
In addition to the issues discussed above, other assumptions and conclusions of the authors remain unsubstantiated. For example: The authors attribute the tilt of the plains to movement of a block bounded by the Cape Fear Arch (or proposed Fault) on the south and by the proposed Neuse Fault on the north. Yet the data presented in their paper is restricted to the area between central New Hanover County and the south side of the New River. The New River is south of the trace of the proposed Neuse Fault. No explanation is given as to why the proposed Neuse Fault was chosen as the northern boundary of the block as no data on either the area between the New River and the "Neuse Fault" or the area north of the "Neuse Fault" is provided to show that there is a structural boundary there. Even if the concept of tilting is adopted, it represents a regional tilt; it does not constitute proof of sharp displacement across a fault boundary.

Zullo and Harris (1979) state that the Waccamaw Sea transgressed only as far inland as the Hanover Scarp (top of scarp elevation less than 10 m, or 35 ft) in the study area while the same sea occupied the Surry Scarp (toe elevation 30 m, or 94 ft) north of the New River and south of the Cape Fear River (Harris and others p. 38). This they say is evidence that the study area was structurally higher relative to the adjacent areas in the early Pleistocene than at present. However, their Figure 4 p. 36 shows Waccamaw-Canepatch equivalents as having been deposited inland from the Hanover Scarp, illustrating that the Hanover Scarp was not the landward limit of the Waccamaw Sea, and that there is no evidence to indicate that the study area was uplifted relative to adjacent areas at this time. Furthermore, the Surry Scarp appears to cross the proposed trace of the "Neuse Fault" without disturbance (Daniels and others, 1966).

In a recently published paper (Harris, 1982), Harris again refers to the "Neuse Fault" and reiterates his previous conclusions presented in Harris and others (1979a and 1979b) and Harris and Zullo (1979), stating that the Neuse Fault was active in the latest Cretaceous and intermittently throughout the Tertiary as well as in the Quaternary. However, Harris (1982) does not present any new data as evidence for the "Neuse Fault" or movement along it.

2.3 Conclusion

In conclusion, no evidence has been presented that proves either the existence of the proposed Neuse Fault or that it has moved in the last 32,000 years. In addition, the seismicity of the area around the proposed Neuse Fault is discussed in Section 2.5.2 of the SHNPP FSAR. As shown on Figure 2.5.2-1 of the FSAR there is no seismicity associated with the alignment of the proposed Neuse Fault, and no seismic evidence suggest that the proposed Neuse Fault exists.



---- Faults proposed by Harris and others (1979a,b)

— — Neuse River-Cape Lookout Fault proposed by Ferenczi (1959)

↕ Positive Element of Gibson (1967)

Figure 1

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PREPRINT

Rb-Sr Glauconite Isochron of the Eocene Castle Hayne Limestone, North Carolina – Further Discussion

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INTRODUCTION

The 11-meter thick lectostratotype of the Castle Hayne Limestone Formation selected by Baum et al. (1978) in New Hanover County, North Carolina (Fig. 1; also see Ward et al., 1978), has become the subject of considerable interest in recent times because of a reported Rb-Sr glauconite isochron date of 34.8 Ma±1 Ma (Harris, 1979; Harris and Zullo, 1980; Fullagar et al., 1980) at a stratigraphic level (Fig. 2) interpreted as belonging to calcareous nannoplankton zones NP19 and NP20 (= late Eocene, Priabonian Stage) (Turco et al., 1979; Worsley and Turco, 1979). Different age interpretations based upon other biostratigraphic evidence were reviewed by Harris and Zullo (1980) but the authors opted for a late

Eocene age based on the calcareous nannoplankton evidence. In a critique of the above studies Jones (1982) has presented evidence from planktonic foraminifera suggesting that the Castle Hayne Formation is of Zone P11-12 age (middle Eocene, Lutetian Stage), while, in a reply, Harris and Zullo (1982) defend and retain their late Eocene age interpretation.

Accurate radiometric dates are important both as calibration points and consistency checks in the formulation of geological time-scales. Of paramount importance is precise biostratigraphic control on radiometrically dated levels so that they may serve as internal consistency checks upon each other as additional data are compiled over the years. In recent years there have developed two "schools" of thought regarding the age of the Eocene/Oligocene boundary, a so-called orthodox school (Berggren, 1972; Hardenbol and Berggren, 1978) who believe that the boundary has an age of about 37 Ma; a vocal minority (Odin, 1978; Odin <u>et al.</u>, 1978; Glass and Zwart, 1977; Harris and Zullo, 1980, 1982) believe the boundary is considerably younger, <u>ca</u>. 33-34 Ma. A third group has taken an intermediate position with age estimates in the 34-35 Ma range.

Because of the controversy surrounding Paleogene chronology in general, and the Eocene/Oligocene boundary in particular, we have decided to make a comment on this particular study and, what we view, as some anomalous results. In order to treat the problem in its proper perspective it is necessary to bring in data from a variety of fields and to range over a spectrum of Paleogene stratigraphy. However, we shall try, to the extent possible, to confine the discussion, as much as possible, to middle Eocene and upper Eocene stratigraphy. A comprehensive review of Paleogene bio- and chronostratigraphy, and magneto- and radiochronology, and a thoroughly revised Paleogene time scale is being prepared by W. A. Berggren, Dennis Kent and John T. Flynn. In this paper we shall demonstrate that the Castle Hayne Formation:

1) is of late Middle Eocene (late Lutetian to early Bartonian) age,
2) is no older than planktonic foraminiferal zone P12, nor younger than P14, and is most likely correlative with upper Zone P12 to Zone P13,
3) belongs to calcareous nannoplankton Zones NP16 (upper part) to NP17 (lower part),
4) as a maximum spans the interval represented by magnetic polarities 20 to 18 (= 46-42 Ma, LaBrecque _et al._, 1977; = 45-41 Ma, Ness _et al._, 1980), as a minimum spans the interval bracketing the base of anomaly 18 (= 43-42.5; LaBrecque _et al._, 1977; = 42-41.5 Ma, Ness _et al._, 1980).

Further we shall show that available data now support an age of 36.5-37 Ma for the Eocene/Oligocene boundary.

CUSSION

We shall address ourselves to various points raised in the pers by Harris and Zullo (1980, 1982) and the critique by Jones 982) and present our own interpretations and evaluations of blished data as well as our own investigations on material upplied to us of the Castle Hayne Formation.

Biostratigraphy

Calcareous Nannoplankton

The Castle Hayne Formation has been assigned to calcareous nannoplankton zones NP19 and 20 by Turco *et al.* (1979) and Worsley and Turco (1979) based primarily on the basis of the presence of *Zygolithus* (*vel Neococcolithites*) *dubius*, *Chiasmolithus grandis* and *Sphenolithus pseudoradians*, among other taxa. They mention the possibility that "the Castle Hayne extends down into the Middle Eocene" if one includes an outlier of Eocene chalk on State Route 701, which belongs to Zone NP18 (considered of late Eocene age) on the basis of the presence of *S. pseudoradians*, with a form "intermediate between *Z. dubius* and *Isthmolithus recurvus*" (Worsley and Turco, 1979: 72).

Jones (1982: 180) observed that all the calcareous nannoplankton taxa mentioned by Worsley and Turco (1979: 71) from the lectostratotype Castle Hayne Formation "have world wide stratigraphic ranges that extend down into the middle Eocene". In this he is correct. Harris and Zullo (1982: 182) reply that there

are three taxa listed by Worsley and Turco (1979) which "unequivocally have ranges beginning above the middle Eocene [Chiasmolithus oamaruensis, Sphenolithus pseudoradians, and Helicosphaera reticulata (T.R. Worsley, personal commun.)"]. It should be borne in mind here that these species names are derived from a list of taxa identified in North Carolina Coastal Plain wells (Worsley and Turco, 1979: 70), two of which penetrated strata assigned to the Castle Hayne Formation. Let us look closer at these three taxa:

a. Chiasmolithus oamaruensis is listed only as a "?" in a single sample (230' below the surface) in the Evans #1 well.

b. Sphenolithus pseudoradians is recorded in the Castle Hayne limestone in both wells (Evans #1 and 1-0 core) and in the upper part of the outcrop lectostratotype. This taxon has been recorded in several tropical sites in Zone NP16 (Muller, 1976: 612), and Martini (1976: 383) has indicated that this taxon has its initial appearance in the Equatorial Pacific much earlier than in high latitudes. This species has been observed in Zone NP16 in several sites from the Atlantic, Pacific and Indian oceans (Aubry, work in progress).

c. Helicosphaera reticulata is not listed in either of the two wells that penetrated the Castle Hayne Formation from which Worsley and Turco (1979: 70) listed taxa, but it appears on a chart of composite ranges of Paleogene calcareous nannoplankton taxa to be restricted to Zones NP 19 and 20 (Worsley and Turco, 1979: 69). We have no way of evaluating the stratigraphic distribution of this taxa relative to the Castle Hayne Formation.

More pertinent to the problem of the ages of the Castle Hayne Formation is the general nature of the calcareous nannoflora listed from this formation. *Neococcolithites dubius* and *Chiasmolithus grandis* became extinct in the latest middle Eocene, within or at the top of, Zone NP17. The latter taxon has its LAD close to the top of Zone NP17, approximately coincident with the FAD of *Chiasmolithus oamaruensis* and, indeed, the LAD of *C. grandis* is often used to denote the NP17/18 boundary in instances where *C. oamaruensis* is [illegible]e or absent. *Micrantholithus procerus* has been suggested to be a useful marker form for distinguishing middle and upper Eocene strata (Bukry and Bramlette, 1969), and Bybell and Gartner (1972) recorded it from the upper middle Eocene of the Gulf Coast, France, Mexico, Brazil, the Indian Ocean, and JOIDES Hole 3 from the Blake Plateau.

In figure 3 we have listed the known global ranges of the various calcareous nannoplankton taxa mentioned by Worsley and Turco (1979) from the Castle Hayne Formation and by Jones (1982) from a supposed (outcrop) equivalent of the Castle Hayne Formation. In addition we have examined several samples (R2204B-E collected by the U.S. Geological Survey and CHMM-2, from approximately the same stratigraphic level as R2204-E, collected by Gary Jones, Union Oil Co. of California) from the Comfort Member of the Castle Hayne Formation (as described by Ward *et al.*, 1978) at the lectostratotype locality of Baum *et al.* (1978) in the Martin Marietta Company rry, New Hanover County, North Carolina (see Fig. 2). In dition, Edwards *et al.* (in press) list the nannoflora and noflagellates from the Comfort Member at this locality. The Rb-Sr auconite isochron date of 34.8±1 Ma was obtained from a tratigraphic level between samples R2204C and D (Harris, 1979; ullagar and others, 1980; Harris and Zullo, 1980). Samples R2204B (New Hanover Member of Ward *et al.*, 1978), and C from the lower Comfort Member are virtually barren; however, samples R2204D and E, and CHMM-2 contain a numerically scarce but rather diversified, moderately well preserved calcareous nannoflora.

Twenty-six taxa have been identified in samples R2204D and E and CHMM-2 from the Castle Hayne Formation (see Table 1). Nine of these have their FAD in the early middle Eocene (NP14, NP15) or earlier and range into the late Eocene or younger (*Discoaster barbadiensis*, *D. saipanensis*, *Zyghablithus bijugatus*, *Ericsonia formosa*, *Cyclococcolithus luminus*, *Chiasmolithus titus*, *Coccolithus pelagicus*, *Micrantholithus vesper* and *Lanternithus minutus*). A further nine have their FAD in the late middle Eocene (Zone NP16) and extend into the late Eocene or younger: *Reticulofenestra bisecta*, *Helicosphaera compacta*, *Cyclococcolithus floridanus*, *Reticulofenestra hesslandii*, *Cyclococcolithina protoannula*, *Coccolithus eopelagicus*, *Reticulofenestra reticulata*, *R. samodurovi*, *Sphenolithus spiniger*. Six taxa are restricted to the middle Eocene (Lutetian-Bartonian): *Micrantholithus crenulatus*, *Cruciplacolithus delus*, *Wiseorhabdus inversus*, *Cyclococcolithus pseudogammation*,

tosphaera sigmoidalis and Rhabdosphaera spinula.
iculofenestra reticulata first appears in Zone NP16 and
iciplacolithus delus became extinct within Zone NP17. Although
orly known the last occurrence of Cyclococcolithus pseudogammation
within the upper part of Zone NP16 or lower part of Zone NP17.
onsequently, it appears that the Castle Hayne Formation can be
ssigned either to the upper part of Zone NP16 or to the lower part
of Zone NP17. On the basis of the absence of the species which
characterize epicontinental sediments belonging to Zone NP17
(Clathrolithus spinonus, Corannulus germanicus, Sphenolithus celsus
among others, Aubry, work in progress), we prefer an assignment of
the Castle Hayne Formation to Zone NP16. The lack of the zonal
markers has no significance since Chiasmolithus solitus, as well as
Chiasmolithus grandis are absent or very rare in shallow-water
sediments. On the other hand, if the Castle Hayne Formation had
been of late Eocene age (NP19-20), one could have expected the
occurence of Ismolithus recurvus, a form known to occur commonly in
shallow epicontinental environments.

It is clear that the stratigraphic overlap of taxa shown in Fig. 3 occurs in the interval of the upper part of Zone NP16 and the lower part of Zone NP17. The calcareous nannoplankton evidence suggests that the Castle Hayne belongs to the interval of Zone NP16 which is of late Lutetian to Bartonian Age (late Middle Eocene; Cavelier and Pomerol, 1976; Hardenbol and Berggren; 1978).

Planktonic Foraminifera

Jones (1982) has drawn attention to the fact that Harris and Zullo (1980) indicate that planktonic foraminiferal evidence suggests that the Castle Hayne Formation is of middle Eocene (Claibornian) age but do not cite the evidence. He then cites data from his own detailed Ph.D. studies (Jones, 1981) which clearly indicate a middle Eocene age for the Castle Hayne Formation. In their reply Harris and Zullo (1982: 182) dismiss Jones' (1982) evidence with the statement that "a list of species which are not figured does not..." as Jones (1982: 181) states..."prove the middle Eocene age of the Castle Hayne Formation..." (The same can be said for the calcareous nannoplankton lists provided by Worsley and Turco, 1979 upon which Harris and Zullo relied so heavily for their paleontological calibration but the authors conveniently overlook this point.) Harris and Zullo (1982: 182) dismiss Jones' evidence on the basis that evidence presented based on data from a Ph.D. dissertation in progress "is a preconceived conclusion made prior to completion of and in critical review of the work". However, they could have availed themselves, in preparing their reply (Harris and Zullo, 1982) to Jones' (1982) critique, of Jones' (1981) Ph.D. thesis and the evidence contained therein. This we have done, in addition to examining material from samples from the Castle Hayne Formation.

Jones (1981, 1982) has documented a taxonomically varied, if numerically poor in some cases, planktonic foraminiferal fauna (which he assigns to upper Zones P11 and P12) in the lectostratotype(s) and other outcrops of the Castle Hayne Formation and core samples from nine counties in North Carolina. The taxonomic composition (low-conical morozovellids, non-carinate acarininids and Truncorotaloides i.al.) is typical of the middle Eocene. We have examined the same samples mentioned above (under the Calcareous Nannoplankton). Foraminifera are present in all samples and we have verified essentially the same planktonic foraminiferal fauna as that cited by Jones (1981, 1982) although not all the taxa he mentions have been observed owing to small sample sizes. Nevertheless the presence of Acarinina bullbrooki, Truncorotaloides collactea, T. rohri, Planorotalites renzi, Morozovella spinulosa-coronata group represent a typical middle Eocene fauna similar to that reported by Jones (1981, 1982).

In Fig. 4 we have plotted the stratigraphic ranges (Berggren, 1977; Blow, 1979) of some of the stratigraphically diagnostic taxa documented by Jones (1981, 1982) from the Castle Hayne Formation. The presence of Acarinina, Truncorotaloides and Morozovella precludes an age assignment of the Castle Hayne Formation younger than Zone P14 or basal P15 (Blow, 1979; figs. 50, 53, 58-61; p. 290-292). The overlap of diagnostic taxa occurs within the interval of Zones P12 and P13. However, there are several indications that this can be narrowed down to the interval of Zone P13, namely the presence of Planorotalites renzi, Morozovella lehneri, M. coronata, Globigerapsis kugleri, Acarinina bullbrooki - all of which have their LAD's in Zone P13 (Blow, 1979) and of Hantkenina longispina which has its FAD in this Zone (Blow, 1979).

Harris and Zullo (1982: 182) note that Huddlestun (in a personal communication, 1981) attributes the Castle Hayne Formation to Zone P13 based on an examination of numerous foraminiferal samples from this unit. At the same time they observe that Huddlestun has some misgivings about some of the species or their ranges (Jones, 1982) questions the absence in Jones' (1982) list of such taxa as Globorotalia bullbrooki Bolli, G. crassata, G. crassula, G. densa, G. rotundimarginata and G. spinuloinflata, which "are common to abundant in middle Eocene deposits. They then conclude that this "indicates a problem in the planktic foraminiferal data". Does it really? We hardly think so, if one is familiar with the taxonomy of planktonic foraminifera.

a. Acarinina bullbrooki is a senior synonym of Acarinina densa (the holotype of the latter taxon having been lost (Berggren, 1977: 260, 261; Blow, 1979: 915-917). Jones (1981) describes it from the Castle Hayne Formation and provides an excellent illustration (pl. 7, figs. 15-17) from core CR-C2-79, 57 ft. 8 in., Craven County.

Globorotalia crassata is best considered *nomen non conservandum* (Blow, 1979: 1013) because of the loss of the lectotype selected by Bandy (1964) for Cushman's (1925) taxon. The remaining syntypic series of specimens do not appear to be synonymous with Bandy's (1964) lectotype. On the other hand, observations in 1967 by Berggren and Blow (see Berggren, 1977: 247) have shown that *crassata* Cushman, 1925 = *spinulosa* Cushman, 1927. This is confirmed by subsequent studies in 1970 on the type material of these taxa (Blow, 1977: 1012-1013) except that the holotype of *crassata* had been lost in the interim. Blow (1979: 1012-1013) suggests substitution of the name *spinulosa* for those forms previously attributed to *crassata* as well as forms subsequently identified as *spinulosa* in the literature. A new subspecies *spinulosa coronata* (Blow, 1979: 1016-1017) was described for forms with a "more widely open (not closed) umbilicus which is surrounded by a coronet of muricae borne on the ventral extremities of the umbilical shoulders of the chambers of the last convolution of the test". Its range is from Zone P10-P13 (Blow, 1979: 1017). Again, *Morozovella spinulosa* is listed and illustrated by Jones (1981: pl. 8, figs. 10-12) from the Castle Hayne Formation, well CR-A40-62, and from Neuse River, Stop 1, Craven County. Its morphology is typical of *M. coronata* (and it is listed as such in Fig. 3).

c. *Acarinina rotundimarginata* Subbotina, 1953 is conspecific with *Globorotalia spinuloinflata* Bolli, 1957 (*non* Bandy, 1949) and both are synonymous with *Globorotalia* (*vel Truncorotaloides*) *collactea* Finlay, 1939 (Jenkins, 1971: 134; Berggren, 1977: 261-262; Blow, 1979: 919). *Acarinina* (*vel Truncorotaloides*) *rotundimarginata* = *T. collactea* is present in Castle Hayne samples we have examined from the Martin Marietta Quarry and appears to have been identified as *Acarinina pentacamerata* (Subbotina) by Jones (1981: pl. 7, figs. 12-14) from the Ideal Cement Company Quarry, New Hanover County, North Carolina and well BEA-T-38, Beaufort County, North Carolina.

d. *Globorotalia crassula* Cushman and Stewart is a mid-Pliocene-Pleistocene taxon and its presence in the Castle Hayne Formation would be cause for considerable alarm.

In summary, the biostratigraphic evidence of calcareous nannoplankton (NP16-17) and planktonic foraminifera (P12-P13) are in close agreement in assigning the Castle Hayne to the late middle Eocene (late Lutetian to early Bartonian Age).

Dinoflagellates

Dinoflagellate biostratigraphy of the Comfort Member of the Castle Hayne Formation lectostratotype is treated in greater detail by Lucy Edwards in ~~Bybell~~ *et al.* (this volume). Suffice to observe here that the microflora indicate correlation with the upper part of the *Coleothrypta kisselovia* Zone (Costa and Downie, 1976) which, in turn, suggests an age assignment no older than the *Areosphaeridium arcuatus* (B-4) Assemblage Zone nor younger than the *Cyclonephelium intricatum* (B-5) Assemblage Zone (Eaton, 1976; Bujak *et al.*, 1980) of the upper Bracklesham Beds of the Isle of Wight. These latter two zones are only slightly lower (older) than the basal Bartonian *Heteraulacacysta porosa* (Bar-1) Assemblage Zone (Bujak *et al.*, 1980) which is equivalent to the lower part of the *Rhombodinium draco* Zone (Costa and Downie, 1976). The Bracklesham Beds correspond predominantly to the Lutetian Stage of the Paris Basin: the uppermost part corresponds to the basal part of the Auversian (Chateauneuf and Gruas-Cavagnetto, 1978: 72,76; Chateauneuf, 1980: fig. 45). The upper part of the stratotype Lutetian belongs to Zone NP16 (Aubry, in prep.), and it and the lower part of the Auversian beds are placed in the *Wetzeliella* aff. *articulata* Zone by Chateauneuf and Gruas-Cavagnetto (1978, 1980). The overlying Auversian beds (upper NP16, Aubry, in prep.; and equivalent to the lower Bartonian) are placed in the *Rhobodinium draco* Zone (*loc. cit.*). Thus the dinoflagellate stratigraphy suggests a latest Lutetian or earliest Bartonian age assignment for the Comfort Member of the Castle Hayne Formation.

Magnetobiochronology

Recent correlations between calcareous plankton biostratigraphy and magnetostratigraphy in the Contessa section(s), Gubbio, Italy (Lowrie, Alvarez *et al.*, 1982) provide additional constraints on the chronology of the Castle Hayne Formation:

1. The LAD of *Chiasmolithus grandis* is associated with upper magnetic anomaly 18 (Lowrie, Alvarez *et al.*, 1982).
2. The FAD of *Morozovella lehneri* (Zone = P11/12 boundary) is associated with mid-anomaly 20, the LAD of *Truncorotaloides rohri* and the FAD of *Globigerapsis semiinvoluta* (= Zone P14/15 boundary) is associated with the top of anomaly 18, and the extremely brief Zone P13 is shown to bracket the base of anomaly 18 (Lowrie, Alvarez *et al.*, 1982).

The biostratigraphic data reviewed above suggest that the Castle Hayne Formation at an outside maximum could span, or be located within the interval between, anomalies 20 and 18. At a minimum, it is correlative with an interval bracketing the base of anomaly 18. We suggest that the most probable correlation is within an interval bracketed by the top of anomaly 19 to the top of anomaly 20. The chronology derived from a purely magnetic stratigraphy (LaBrecque *et al.*, 1977) or an integrated biostratigraphically calibrated magnetostratigraphy (Ness *et al.*, 1980) are quite similar, the former having been based on lateral or downward (older) linear extrapolation based on the assumption of constant rates of sea-floor spreading from radiometric calibration points between 0 time and the late Neogene, the latter scale having been prepared by interpolation between the same late Neogene calibration points and a paleontologically controlled biochronologic age estimate near anomaly 24 and the Paleocene/Eocene boundary. The values of these two scales (in Ma) are shown below in Table 2.

TABLE 2.

Anomaly Number	LaBrecque *et al.* (1977)	Ness *et al.* (1980)
Anomaly 18	42.44	41.40
	42.88	41.82
Anomaly 20	44.85	43.69
	46.40	45.18

Table 2. Estimated magnetic chronology of anomalies 18 and 20 (top and bottom values shown in proper vertical order) in Ma.

Table 2 shows that the Castle Hayne Formation has an age range of = 45 - 41 Ma (maximum), but if, as the biostratigraphic evidence presented here suggests, it is essentially correlative with an interval bracketed by anomalies 19 and 20, its age should be more properly in the 43 - 45 Ma range. This value should be compared with the Rb-Sr isochron date of 34.8$\pm$1 Ma obtained on the Castle Hayne Formation (Fullager, 1979; Harris and Zullo, 1980).

AGE OF THE EOCENE/OLIGOCENE BOUNDARY

Harris and Zullo (1980: 591) indicate that the "volcanic ages of Evernden and others (1964), the glauconite ages of Ghosh (1972) and of Odin and others (1978), and the mikrotektite ages of Glass and others (1973) and Glass and Zwart (1977) indicate a much younger age [than the 37-37.5 Ma suggested by Funnell, 1964; Berggren, 1972; and Hardenbol and Berggren, 1978] for the boundary, between 33-35 Ma." They cite in support of their viewpoint the fact that "Odin et al. (1978) determined glauconite ages of marine sequences in England (type Barton Beds) [apparently unaware that the Bartonian is of late middle Eocene age, biostratigraphically approximately equivalent to Zones P13-P14 and NP16-NP17: Hardenbol and Berggren, 1978] and in Germany and suggested that the age of the boundary was about 33 m.y."

It is impossible to enter into a detailed analysis of the problems associated with the various age estimates made on the Eocene/Oligocene boundary. This is currently being done by Berggren, Kent and Flynn and will be presented elsewhere. Suffice at this point to make several observations.

1. The (revised) glauconite determinations made by Odin et al. (1978: 487) on the type Barton Beds (ca. 39-40 Ma) are viewed as anomalous in the light of other evidence discussed below.
2. A K-Ar (glauconite) date of 37.5±0.5 Ma has been obtained (Gramann et al., 1975) on the Siberberg Beds at Helmstedt, NW Germany with a calcareous nannoflora assigned to Zone NP21 (Martini, 1971; Haq, 1972) which brackets the Eocene/Oligocene boundary.
3. A number of K-Ar (glauconite) dates have been obtained from the underlying Gehlberg Beds at Helmstedt ranging in age from 37.4 - 39.6±0.7 Ma. The biostratigraphic position of these beds is difficult to determine, but they are certainly late Eocene in age and post Zones NP15-16.
4. K-Ar (glauconite) dates of 36.4±0.7 Ma and of 39.4±0.9 Ma and 39.6±0.6 Ma, have been obtained (Gramann et al., 1975) on the upper, and lower, part, respectively of the *Ostrea queteleti* Beds at Lehrte, east of Hannover, with a similar NP21 flora (Martini, 1971; Haq, 1972).

5. Ghosh (1972) has obtained similar K-Ar (glauconite) dates of 37.6 Ma on the Pachuta Member (Jackson Formation), 37.9 Ma on the Shubuta Member (Jackson Formation), 38.2 Ma on the Moodys Branch Formation, and 39 Ma and 39.4 Ma on the Yazoo Formation - all of which are of late Eocene (Priabonian) age. The Pachuta and Shubuta Members of the Jackson Formation contain a latest Eocene P16-P17 fauna and a NP19/20 and NP21 flora, respectively (Bybell, 1982). The dates of Ghosh (1972) support those obtained in NW Germany and the age estimate of 37 Ma made for the Eocene/Oligocene boundary by Hardenbol and Berggren (1978). In fact it was primarily on the basis of Ghosh's (1972) determinations that Hardenbol and Berggren (1978: 228, fig. 6) chose the value of 37.0 in estimating the age of this boundary. The statement by Harris and Zullo (1980: 591) that the ages of Ghosh (1972) support a younger age estimate is surprising in this context.

6. The volcanic ages of Evernden et al. (1964) do not, as Harris and Zullo contend (1980: 591), support a significantly younger (ca. 33-35 Ma) age for the Eocene/Oligocene boundary. Combined studies on magnetostratigraphy and mammalian biostratigraphy on continental sections of Chadronian land-mammal "age" in two key sections at Flagstaff Rim, Natroma County and Toadstool Park, Sioux County, Nebraska and the integration of four high temperature K-Ar dates (Evernden et al., 1964) on ash-beds in the Flagstaff Rim section have recently provided important and

much needed calibration points for, and constraints upon, mid-Tertiary magnetogeochronological scales (Prothero and Denham, 1981; Prothero et al., in press a,b).

The radio- and magnetochronologic relationships are as follows (Prothero et al., in press a,b). The top of anomaly 12 is dated at 32.4 Ma, the top of anomaly 13 at 34.6 Ma, a level in the reversed polarity interval between anomalies 12 and 13 at 33.5 Ma, and the base of anomaly 13 at 36.1 Ma. Recalibration of the radiometric dates (37.4 and 37.7 Ma) on the Bracks Rhyolite which lies stratigraphically below the Ash Spring and Airstrip local faunas (= Chadronian land mammal "age") of the Capote Mountain Formation, Vieja Group, Southwest Texas and reinterpretation of the magnetic polarity stratigraphy of Testarmata and Gose (1979) which suggests that the Bracks Rhyolite may be associated with the anomaly 15-16 interval provides limiting dates on a late Eocene level. Thus the anomaly 12 (top) to 15-16 interval is bracketed by high temperature K-Ar dates of ca. 32.4 - 37.7 Ma.

Where, in this sequence, does the Eocene/Oligocene boundary lie? Integrated calcareous planktonic biostratigraphic studies in the Mediterranean (Lowrie, Alvarez et al., 1982) and on hydraulic piston cores taken by the Glomar Challenger for the Deep Sea Drilling Project in the South Atlantic (Poore, personal communication, 1982) have shown that the Eocene/Oligocene boundary, determined by the LAD's of Turborotalia cocoaensis-cerroazulensis group, Hantkenina, and the rosette shaped discoasters D. barbadiensis and D. saipanensis, occurs approximately midway within the interval of reversed polarity between anomalies 13 and 15. In terms of the radiometrically calibrated magnetostratigraphy cited above the Eocene/Oligocene boundary would be constrained by the limiting values of ca. 36.1 (near the base of anomaly 13) and 37.4 and 37.1 Ma (within the anomaly 15-16 interval). A numerical value of about 37 Ma is suggested by the radiometric data, which the K-Ar (glauconite) dates cited above appear to support. Alternatively a magnetochronologic age estimate can be made based upon which time-scale is used. That of LaBrecque et al. (1977) yields an age estimate closer to 36.5 Ma, that of Ness et al. (1980) an estimate of about 35.7 Ma, clearly too young.

We disagree with the conclusion expressed by Fullager and others (1980, p. 430) that the Claiborne/Jackson boundary is between 35-37 Ma and that the Eocene/Oligocene boundary is less than 34 Ma based on their K-Ar (glauconite) date of 34.8±1 Ma on the Castle Hayne Limestone, 36.7±0.6 Ma on the Santee Limestone of South Carolina (= Cubitostrea lisbonensis and C. sellaeformis assemblage zones = Zone NP16-17 = Zone P12-13) and 34.1±1.5 Ma also on the Santee Limestone of South Carolina. The Santee Limestone is essentially correlative with the Castle Hayne Formation, and is of late Middle Eocene (Lutetian-Bartonian) age. The dates cited by Fullagar and others (1980) and Harris and Zullo (1980) are from upper middle Eocene strata and do not provide age estimates of late Eocene chronology.

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	R 2204E CHDM2 R	R 2204D R	R 2204C RR	R 2204B R
Chiasmolithus titus	X	X		
Coccolithus eopelagicus				
Coccolithus pelagicus	X	X	X	
Cruciplacolithus delus				
Cyclococcolithina protoannula	X	X		
Cyclococcolithus floridanus	X	X		
Cyclococcolithus luminis	X			
Cyclococcolithus pseudogammation	X	X		
Discoaster barbadiensis	X			
Discoaster saipanensis	X			
Ericsonia formosa	X	X		
Ericsonia cf. subdisticha	X	X		
Helicosphaera compacta	X			
Lanternithus minutus	X			
Micrantholithus crenulatus	X			
Micrantholithus vesper		X		
Neococcolithites minutus	X			
Pontosphaera sigmoidalis		X		
Reticulofenestra bisecta	X	X		
Reticulofenestra hesslandii	X			
Reticulofenestra reticulata	X	X	X	
Reticulofenestra samodurovi	X			
Rhabdosphaera spinula		X		
Sphenolithus spiniger	X	X		
Wiseorhabdus inversus	X	X		
Zygrhablithus bijugatus	X		X	

Table 1. Stratigraphic distribution of calcareous nannoplankton taxa in the Comfort Member, Castle Hayne Formation, Martin Marietta Quarry, New Hanover County, North Carolina.

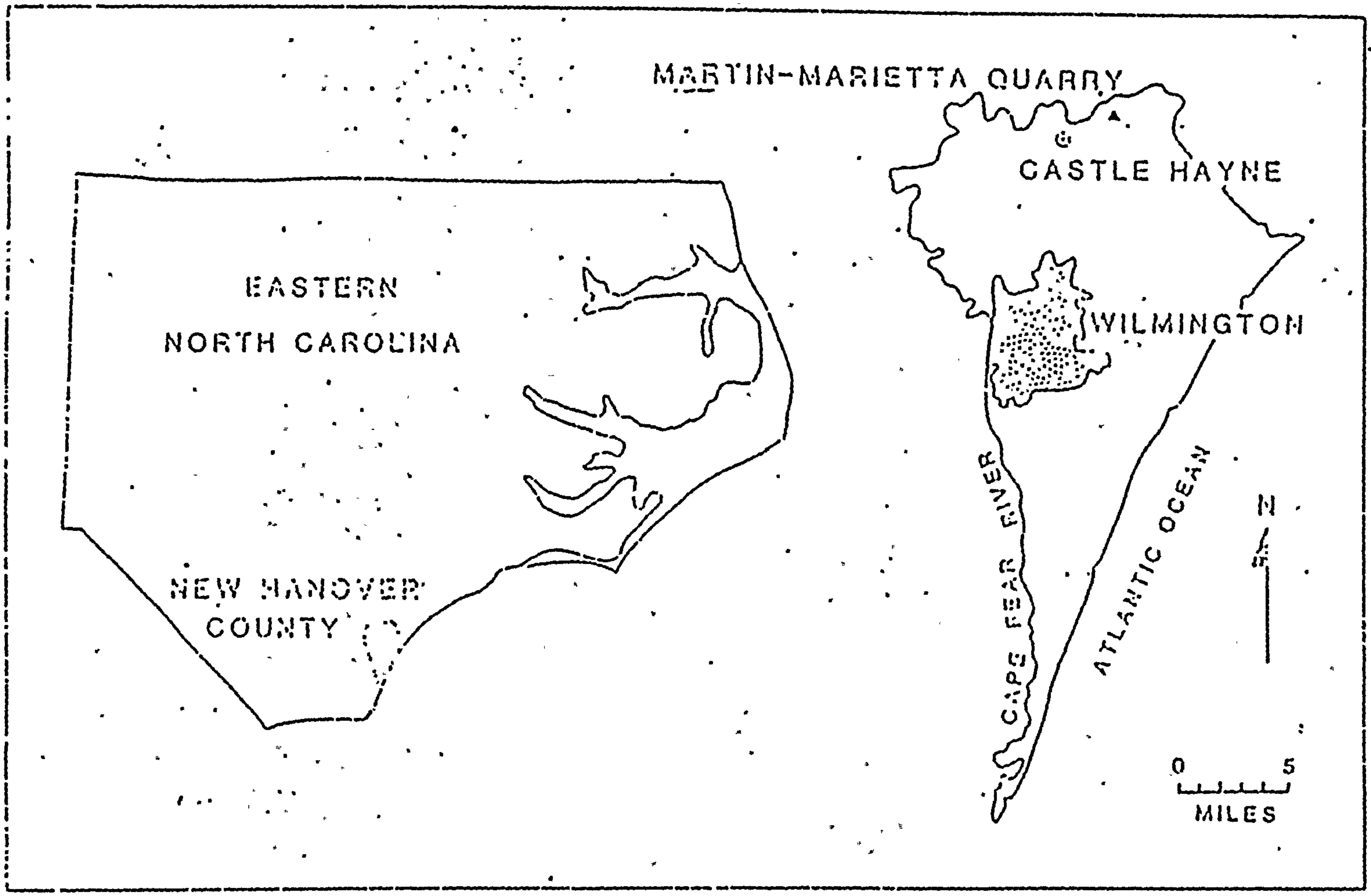
TEXT FIGURES

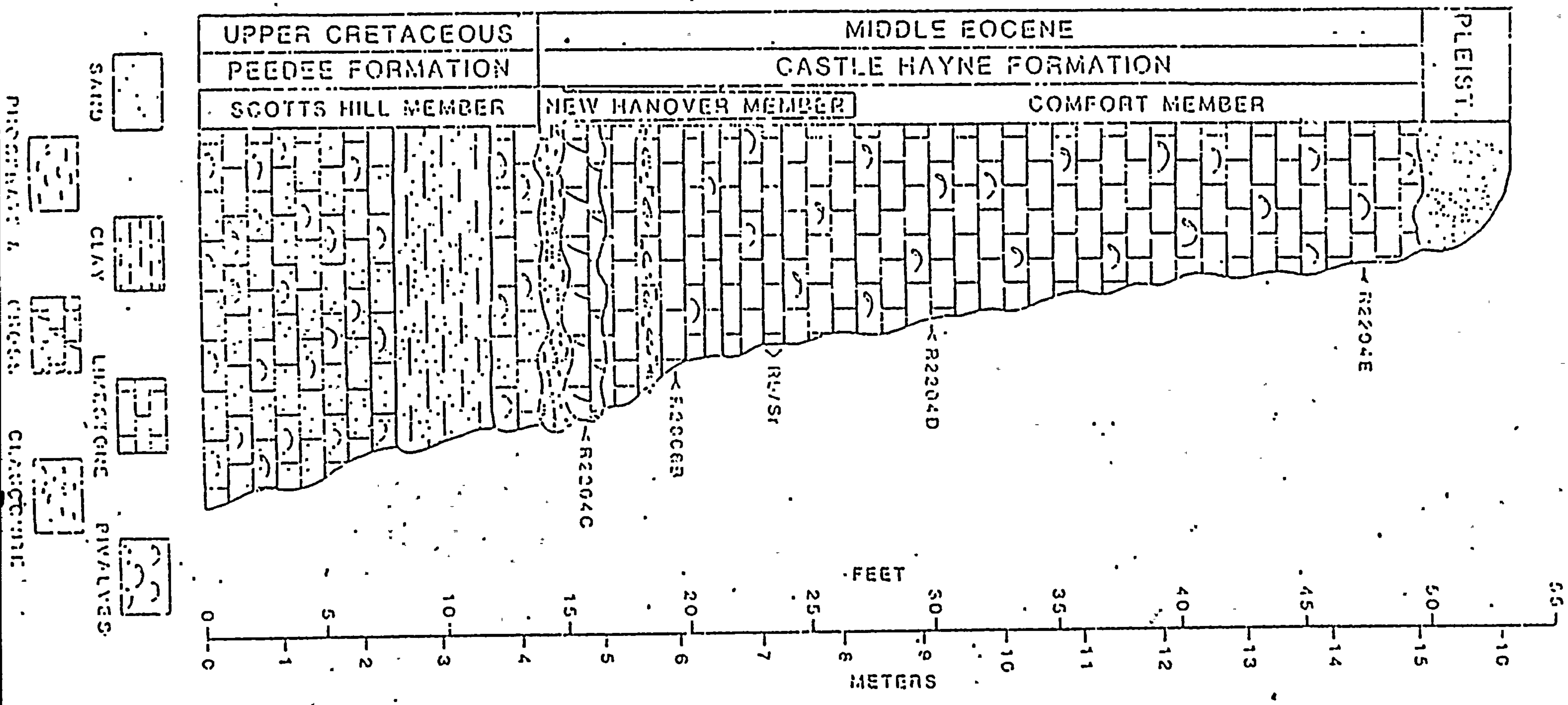
1. Location of maps.

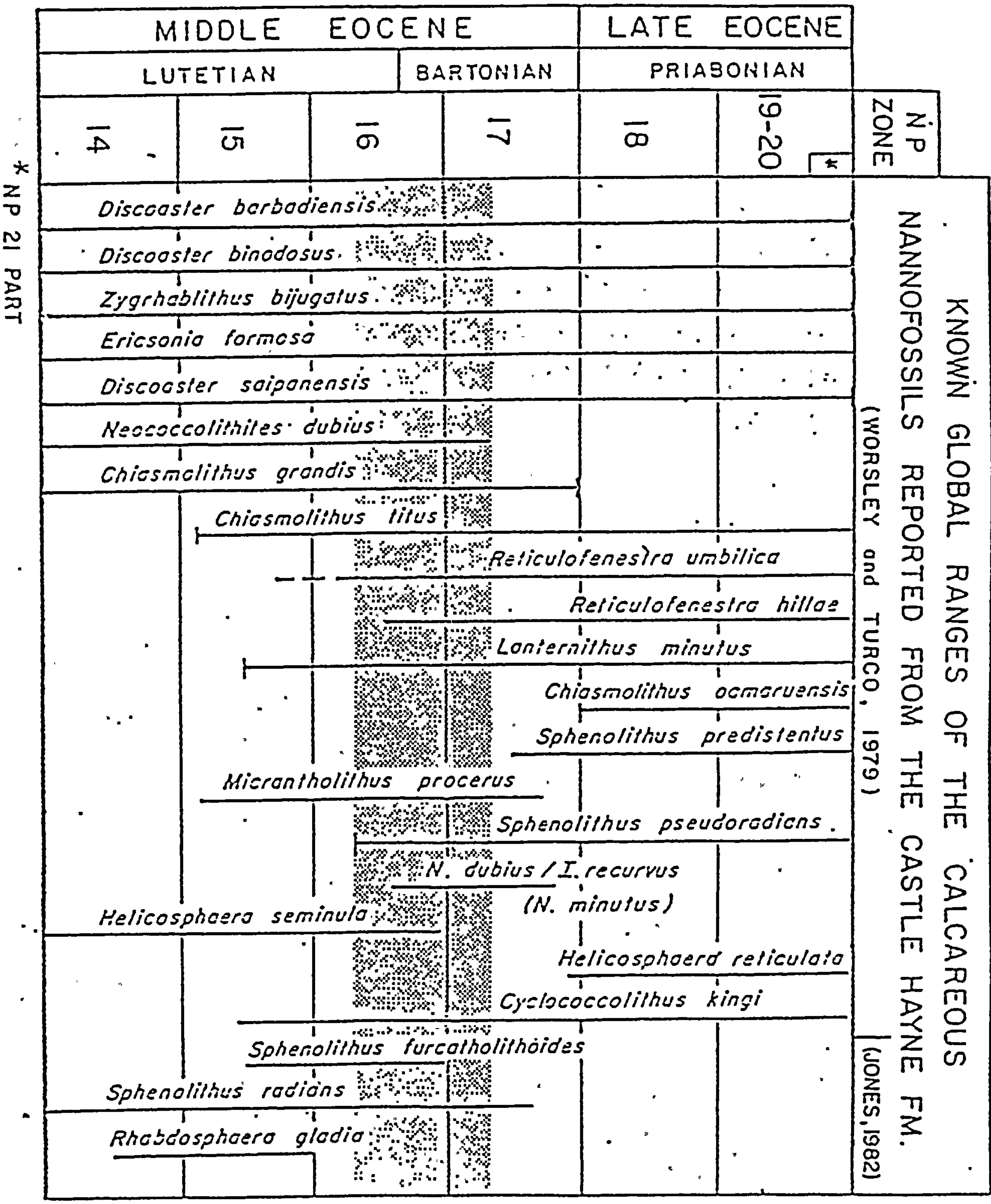
2. Lithologic section exposed at lectostratotype of Castle Hayne Formation, Martin Marietta Quarry, New Hanover County, North Carolina.

3. Known global ranges of the calcareous nannofossils reported from the Castle Hayne Formation.

4. Stratigraphic distribution of some biostratigraphically important taxa of planktonic foraminifera.


MARTIN-MARIETTA QUARRY
CASTLE HAYNE
EASTERN
NORTH CAROLINA
WILMINGTON
NEW HANOVER
COUNTY
CAPE FEAR RIVER
ATLANTIC OCEAN
N
0
5
MILES


UPPER CRETACEOUS
PEEDEE FORMATION
SCOTTS HILL MEMBER
MIDDLE EOCENE
CASTLE HAYNE FORMATION
NEW HANOVER MEMBER
COMFORT MEMBER
PLEIST.
SAND
CLAY
BIVALVES
R2204C
R2204D
R2204E
FEET
0 5 10 15 20 25 30 35 40 45 50 55
METERS
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16



KNOWN GLOBAL RANGES OF THE CALCAREOUS NANNOFOSSILS REPORTED FROM THE CASTLE HAYNE FM.

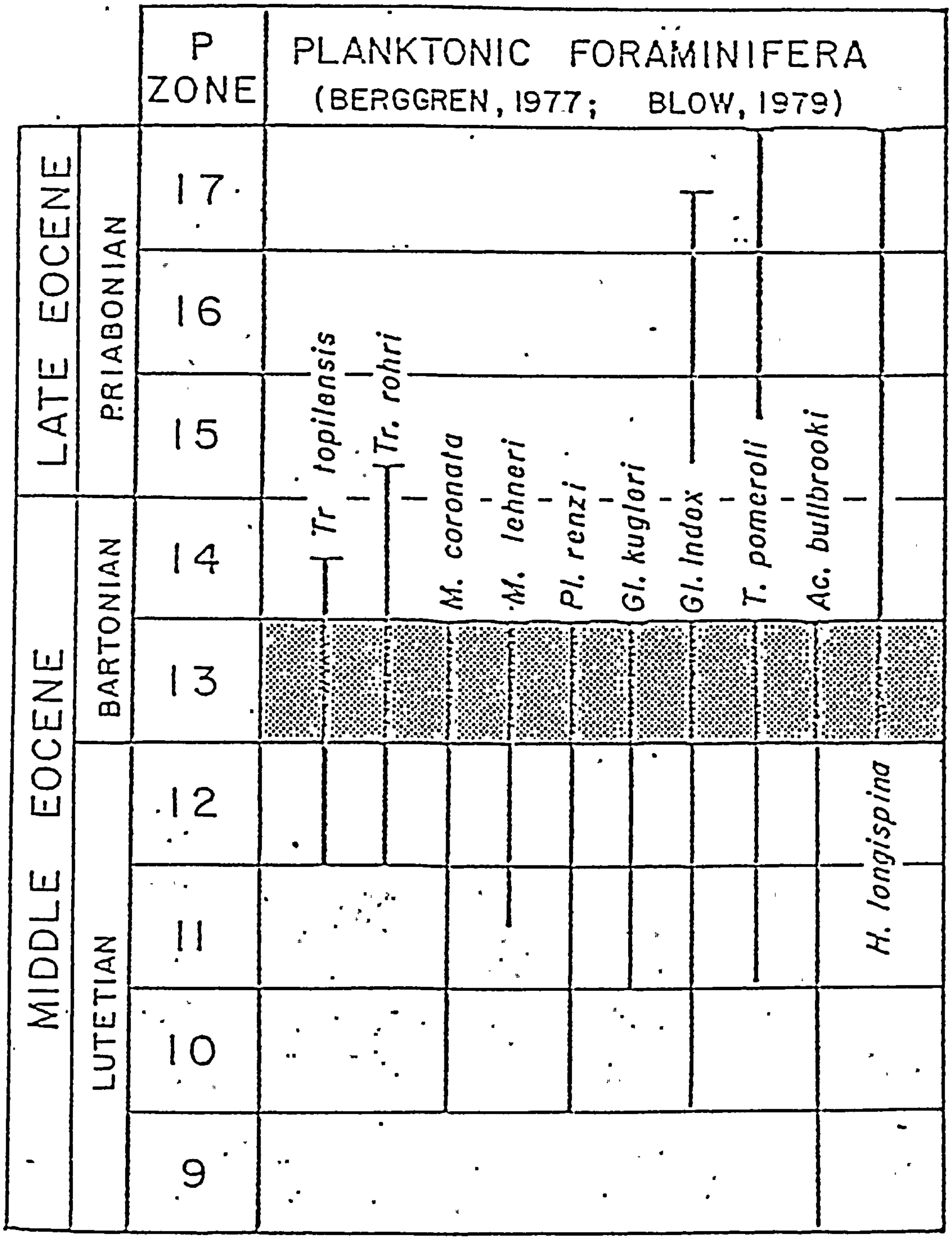

P ZONE
PLANKTONIC FORAMINIFERA
(BERGGREN, 1977; BLOW, 1979)
LATE EOCENE
MIDDLE EOCENE
PRIABONIAN
BARTONIAN
LUTETIAN
17
16
15
14
13
12
11
10
9
Tr. topilensis
Tr. rohri
M. coronata
M. lehneri
Pl. renzi
Gl. kuglori
Gl. Indox
T. pomeroli
Ac. bullbrooki
H. longispina

Figure 3. Correlation chart showing the position of Atlantic and Gulf Coastal Province lithostratigraphic units in a biostratigraphic, chronostratigraphic, chronometric, and magnetostratigraphic model (after Hazel and others, in press). The maximum limit, based on the fauna and flora, of the Comfort Member at the Castle Hayne Formation is indicated by the shaded band. Time is expressed in megaannums (Ma) before present. The "unit" column contains the composite unit scale values derived from Graphic Correlation modelling of numerous measured sections (see Shaw, 1964; Miller, 1977; Murphy and Edwards, 1977). The calcareous nannofossil zones are based on those of Martini (1971) or Bukry (1978); the asterisk in the lower block of the nannofossil column stands for the *Zygodiscus sigmoides* Zone. The dinoflagellate zonation is from Costa and Downie (1976). The foraminifer zonation is after Stainforth and others (1975) except for the middle Eocene which follows Toumarkine and Bolli (1977), and for the definition of the *Planorotalites pseudomenardii* Zone, which follows Blow (1979). The "cycle" column indicates our estimate of the position of the coastal onlap cycles of Vail and Mitchum (1979). The calibration of the magnetic anomaly sequence to the zonations is based on Graphic Correlation modelling of fossiliferous measured sections with magnetics presented in Lowrie and others (1982) and Poore and others (in press). In the "series" column, the numbers 1 through 4 indicate the possible positions of the Paleocene-Eocene boundary (see Pomerol, 1977).

Table 1. Planktic foraminifers from the Comfort Member of the Castle Hayne Formation at the Martin-Marietta quarry.

Table 2. Calcareous nannofossils from the Comfort Member of the Castle Hayne Formation at the Martin-Marietta quarry.

Table 3. Dinocysts from the Comfort Member of the Castle Hayne Formation at the Martin-Marietta quarry.

	CHMM-3	CHMM-glauc.	CHMM-4	CHMM-1	CHMM-2
Acarinina cf. A. bullbrooki (Bolli, 1957)					X
A. aff A. pentacamerata (Subbotina, 1947)	X			X	X
A. sp. A	X		X	X	X
Catapsydrax perus (Todd, 1957)	X	X	X	X	X
Chiloguembelina cubensis (Palmer, 1934)		X			
C. martini (Pijpers, 1933)	X	X		X	X
Globigerinatheka kugleri (Bolli, Loeblich, & Tappan, 1957)				X	
G. mexicana mexicana (Cushman, 1925)	X	X		X	X
Globorotalia cf. T. frontosa (Subbotina, 1953)	X				
Morozovella spinulosa coronata (Blow, 1979)	X	X	X	X	X
Planorotalites renzi (Bolli, 1957)	X	X	X	X	X
Pseudohastigerina micra (Cole, 1927)					
P. sharkriverensis Berggren and Olsson, 1967	X	X		X	
P. cf. sharkriverensis Berggren and Olson, 1967			X		
P. wilcoxensis (Cushman & Ponton, 1932) s.l.	X	X	X	X	X
Subbotina eocaena (Gumbel, 1868)	X	X	X	X	X
S. linaperta (Finlay, 1939)	X	X	X	X	X
Testacarinata inconspicua (Howe, 1939)	X			X	X
Truncorotaloides rohri Bronnimann & Bermudez, 1963 T. topilensis (Cushman, 1925)	X	X X	X X	X	X X

> I prefer the genus Turborotalia for this species.

TABLE 2

Species Name	R2866	R2204D	R2204E
Blackites sp.	X	X	X
Braarudosphaera bigelowi (Gran & Braarud, 1935) Deflandre, 1947	X		
Campylosphaera dela (Bramlette & Sullivan, 1961) Hay & Mohler 1967	X		X
Chiasmolithus grandis (Bramlette & Riedel, 1954) Hay, Mohler & Wade, 1966	X		
C. solitus (Bramlette & Sullivan, 1961) Hay, Mohler, Wade, 1966	X		
C. titus Gartner, 1970	X		X
Coccolithus eopelagicus (Bramlette & Riedel, 1954) Bramlette & Sullivan, 1961	X		
C. pelagicus (Wallich, 1877) Schiller, 1930	X	X	X
Cyclococcolithus formosus Kamptner, 1963	X	X	X
C. protoannulus (Gartner, 1971) Haq & Lohmann, 1975	X	X	X
Dictyococcites bisectus (Hay, Mohler, & Wade, 1966) Bukry & Percival, 1971	X	X	X
D. scrippsae Bukry & Percival, 1971	X	X	X
Discoaster barbadiensis Tan Sin Hok, 1927	X	X	X
D. saipanensis Bramlette & Riedel, 1954	X		X
Helicosphaera compacta Bramlette & Wilcoxon, 1967	X	X	X
H. lophota (Bramlette & Wilcoxon, 1967) Locker, 1973	X		

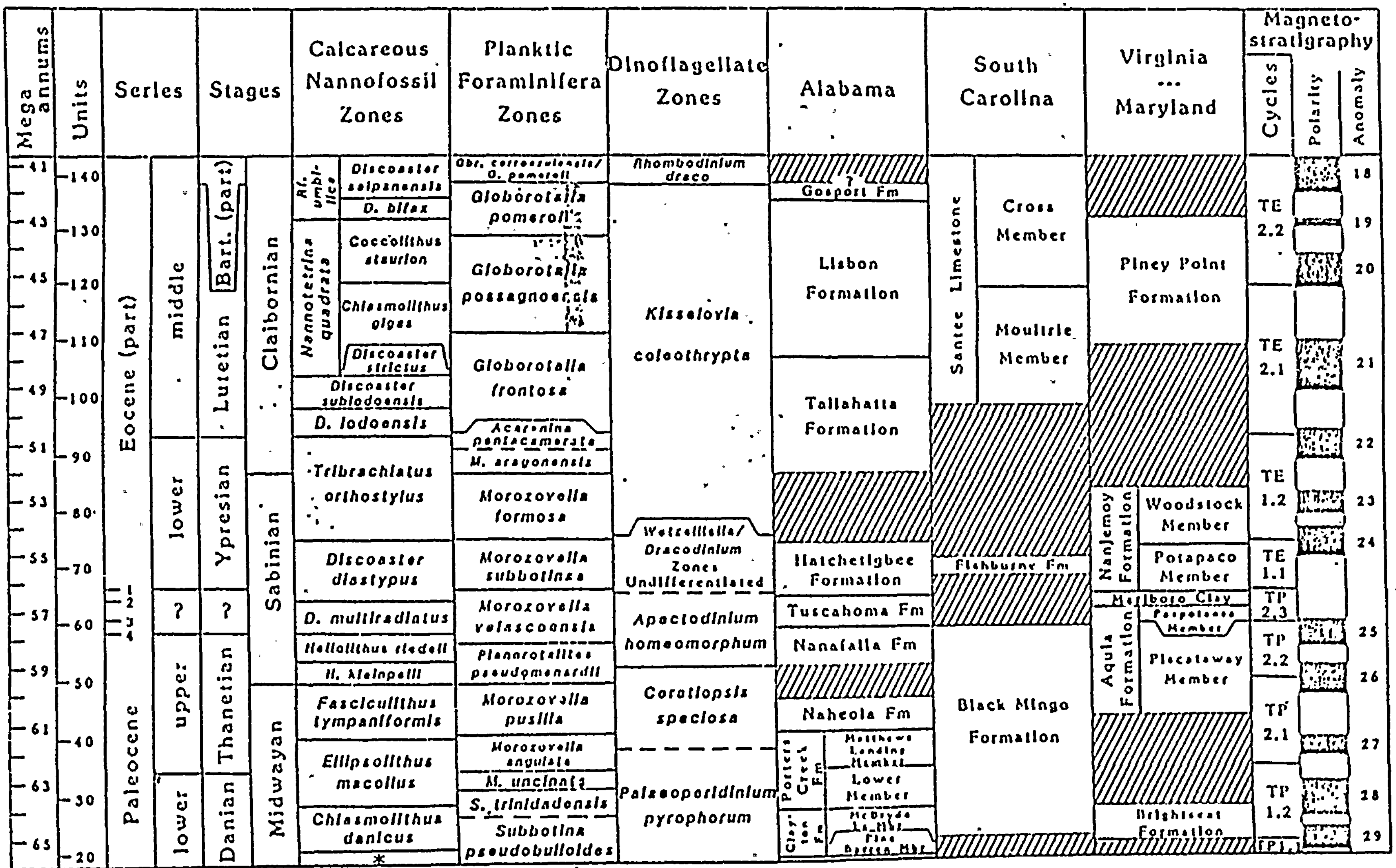
H. reticulata Bramlette & Wilcoxon, 1967	X		
Markalius inversus (Deflandre, 1954) Bramlette & Martini 1964	X		X
Micrantholithus sp. aff. M. crenulatus Bramlettte & Sullivan, 1961	X		
M. procerus Bukry & Bramlette, 1969	X		
Neochiastozygus sp.	X		
Pedinocyclus larvalis (Bukry & Bramlette, 1969) Bukry & Bramlette 1971	X		
Reticulofenestra floridana (Roth & Hay, 1967) Bybell, 1982	X	X	X
R. hillae Bukry & Percival, 1971	X	X	
R. reticulata (Gartner & Smith, 1967) Roth & Thierstein, 1972	X	X	X
R. umbilica (Levin, 1965) Martini & Ritzkowski, 1968	X		X
Sphenolithus moriformis (Broennimann & Stradner, 1960) Bramlette & Wilcoxon, 1967	X	X	X
S. pseudoradians Bramlette & Wilcoxon, 1967	X		X
Transversopontis pulcheroides (Sullivan, 1964) Perch-Nielsen, 1971	X	X	X
Zygrhablithus bijugatus (Deflandre, 1954) Bramlette & Sullivan, 1961	X	X	X

TABLE 3

Species	R 2866 B	R 2204 D
Areoligera coronata (Wetzel, 1933) Lejeune-Carpentier, 1938	X	
Areosphaeridium dictyostylum (Menedez, 1965) Sarjeant, 1981	X	
Cordosphaeridium gracile (Eisenack, 9154) Davey & Williams, 1966b	X	X
Diphyes colligerum (Deflandre & Cookson, 1955) Cookson, 1965	X	
Dinopterygium cladoides sensu Morgenroth, 1966	X	
Homotryblium tenuispinosum Davey & Williams, 1966b	X	X
Hystrichokolpoma rigaudiae Deflandre & Cookson, 1955	X	
Impagidinium n. sp.	X	
Lingulodinium machaerophorum (Deflandre & Cookson, 1955) Wall, 1967	X	X
Meiourogononyaulax sp. I of Manum 1976	X	
Melitasphaeridium pseudorecurvatum (Morgenroth, 1966) Bujak, 1980)	X	
Millioudodinium cf. M. giuseppei (Morgenroth, 1966) Stover & Evitt, 1978	X	
Pentadinium goniferum Edwards, 1982	X	X
Pentadinium laticinctum Gerlach, 1961, subspecies granulatum Gocht, 1969	X	
Pentadinium polypodum Edwards, 1982	X	X
Rhombodinium glabrum (Cookson,, 1956) Vozzhennikova, 1967	X	
Samlandia chlamydophora Eisenack, 1954	X	
Samlandia reticulifera Cookson & Eisenack, 1965	X	

Spiniferites pseudofurcatus (Klumpp, 1953) Sarjeant, 1970	X	X
Spiniferites ramosus (Ehrenberg, 1838) Loeblich & Loeblich, 1966, subspecies gracilis Davey & William	X	
Systematophora placacantha (Deflandre & Cookson, 1955) Davey, Downie, Sarjeant, & Williams, 1969	X	X
Thalassiphora pelagica (Eisenack, 1954) Eisenack & Gocht, 1960	X	
Wetzeliella? sp.	X	
Glaphrocysta intricata (Eaton, 1971) Stover & Evitt, 1978		X
Glaphrocysta undulata (Eaton, 1976) Stover & Evitt, 1978		X
Hystrichokolpoma salacium Eaton, 1976		X
Hystrichostrogylon membraniphorum Agelopoulos, 1964		X
Polysphaeridium zoharyi (Rossignol, 1962) Bujak, Downie, Eaton, & Williams, 1980		X

where is the shaded band — the Comfort Member?
(I put in the blue color)



perhaps NP and P zones should also be listed here?

large number of sediment samples does not prove wide distribution in the source area, as the montmorillonite occurrence in the Ni Basin indicates. Neither is an evaluation of provenance on safe grounds from a consideration of the relative intensity of basal reflections in the sediments. Though montmorillonite has only 17% frequency in the Garnetts Creek source area it has a higher average intensity than kaolinite, which, by volume, is far more prevalent than montmorillonite. Relative intensity of the minerals one to another seems to carry over from the source area to streams except for montmorillonite and mixed-layer structures.

Near absence of mixed-layer structures in the stream sediments of both basins is perplexing for such small drainage nets. Most authorities do not believe a fresh water stream capable of significant alteration of clay mineral structures. Certainly for such small streams alteration would not be expected. The data reported here do not permit a conclusive statement in explanation of this relationship. Chemical data on the waters could prove beneficial in the solution of this problem, but it is felt that physical phenomena are perhaps the operating mechanisms.

Two processes could explain the situation in these basins. Differential transportation and deposition is one alternative and indeed may be operative with certain minerals such as montmorillonite. However, on the basis of unpublished data (Brown, 1958) collected during a study of the entire York River tributary basin, it is felt that the apparent "unmixing" of mixed-layer structures may be a real physical unmixing.

An explanation of the reduced intensities given by sediment clays relative to source samples is not readily explained. Possible explanations may be an increase in percentages of amorphous materials or finer sized sediment particles, the latter being compatible with the physical unmixing hypothesis.

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Koss

(20)

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ABSTRACT

The North Carolina Coastal Plain is not a simple homoclinal structure. The Great Carolina Ridge is an area of uplift and the Hatteras Axis is one subsidence; both are transverse to the Appalachian trend. Midway between those two features is the Cape Lookout-Neuse Fault Zone, also transverse to the Appalachians. Several data in the literature suggest a fourth structural feature, a to date unnamed fault zone with a trend parallel to the Appalachians. As a possible fifth feature, a "zone of subterranean disturbances", suggested by Shaler, 1871, but not proved to date, is mentioned. In conclusion it is suggested that the capes along the present shoreline have been controlled by these structural features.

The basement rock beneath the sedimentary cover has the character of peneplained block mountain rather than that of a folded mountain chain.

INTRODUCTION

This paper attempts to collect and evaluate opinions about the structural conditions of the Atlantic Coastal Plain, especially in North Carolina. Consideration of this problem developed during studies in the University of Virginia, Charlottesville, Va., 1954-55, in connection with preparation of annotated bibliographies for the Hydrographic Office, U. S. Navy, on harbor approaches along the Atlantic Coast. My curiosity was wakened by the peculiar surface features of the subsea prolongation of Cape Hatteras, Cape Lookout, and Cape Fear, by coastal arcs connecting them, by the fairly equal distances between them, and by what relationship they have to structural features of the Coastal Plain. During my following three years with the North Carolina State College, Raleigh, N. C., I became more familiar with the geology of the Coastal Plain of North Carolina.

ACKNOWLEDGMENT

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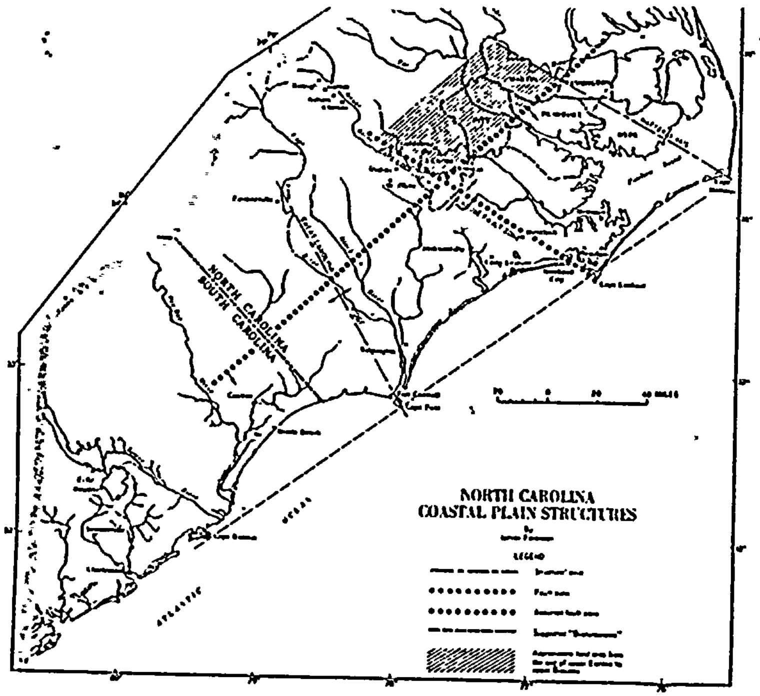

Figure 1

THE GREAT CAROLINA RIDGE (CAPE FEAR ARCH)

Dall in 1892 described a structural feature under the name "Great Carolina Ridge" as "an elevated ridge of perhaps very ancient origin, whose extension may be seen in the contours of the sea bottom far off the coast" (p. 182).

In 1926 Stephenson dealt with it again as "a broad upwarp, having its axis near the boundary between North Carolina and South Carolina", and so indicated it on an accompanying sketch map (p. 468, and pl. 1), although well records described by him in 1912 (p. 163-167, and p. 169-171) suggested an axis farther northeast.

In 1927 Mansfield showed the existence of this elevated ridge by comparing surfaces of the basement rocks as determined in the Havelock, N.C., Wilmington, N.C., Fort Caswell, N.C., and Summerville, S.C., wells. He concluded that this "seems to verify the opinion of Stephenson that the course of the Cape Fear River across the Coastal Plain approxi-

sediments, and the uplifted position has been maintained without [illegible] subsidence until the present" (p. 11). This location of the axis of the Great Carolina Ridge was accepted by Stephenson in a sketch map in his 1928 paper in which he described it as a "broad upwarp in the Cape Fear region in Eocene time" which "raised Upper Cretaceous beds to the surface near the coast" (p. 892 and p. 889, fig. 1).

MacCarthy and his coauthors in a short abstract in 1933 described evidence that this area, especially referring to the southwest flank of the Great Carolina Ridge, reflects differences in the dip of the basement rock surface, a relatively steeper dip toward the coast line than inland. They interpreted this as two erosional surfaces with their intersection about 17 miles west of Conway, S. C. (p. 21).

Prouty used the name "Cape Fear Arch" instead of the former name Great Carolina Ridge, marking it as "an anticlinal fold (arch) through Wilmington running parallel with the Cape Fear River basin toward the north-west." He indicated it also on his sketch map adapted from Stephenson, as well as on his diagram (Prouty, 1936, p. 485, p. 486, fig. 1, and p. 487, fig. 2).

In 1936 Cooke dealt with the area between the Santee River in South Carolina and the Cape Fear River in North Carolina, where "the present land for a considerable distance inland from the present coast both north and south of that area was submerged. This old land area, the Great Carolina Ridge of Dall, may have projected for many miles into the Atlantic as a peninsula, separating an enlarged Chesapeake embayment from an enlarged Gulf of Mexico, Florida being at the time submerged" (p. 99). Jackson (Eocene) time began "with a crustal movement that raised the region between the Cape Fear River in North Carolina and the Santee River in South Carolina, thus producing the Great Carolina Ridge and depressed the regions on both sides of it" (p. 156). Describing the structural conditions of the South Carolina Coastal Plain, he stated that "only deposits of the Upper Cretaceous and Eocene formations are in South Carolina conspicuously deformed on the west limb of the Great Carolina Ridge, whose crest or axis lies not far from the North Carolina-South Carolina state line and nearly parallel to it and whose northeast limb is in North Carolina" (Cooke, 1938, p. 158). "Upon the beveled surface lie thin patches of nearly horizontal marine Miocene formations (remnants separated by erosion)" (p. 159).

As a result of further magnetometer investigations, MacCarthy mentioned that "evidence supporting Stevenson's suggestion of a northwest-southeast uplift near Wilmington has been obtained" whereas roughly parallel to the coast a "magnetically disturbed zone . . . consisting of a series of subparallel highs and lows, has been found", which has been traced from Myrtle Beach, S. C., to the vicinity of Wilmington, N. C., "with further evidence suggesting that it may continue through Burgaw toward extreme northeastern North Carolina", representing "a folded and perhaps fractured zone" (MacCarthy, 1936, p. 405). In 1937 MacCarthy and Straley gave a more detailed picture of these magnetic disturbances referring to the "Wilmington anticline". They stated that "magnetic evidence for or against the existence of this uplift might be expected but because of the nature of the country, observations have not been made" (p. 363). A short abstract by MacCarthy and Straley in 1938 gave as "results to date: (1) a magnetically disturbed area in the neighborhood of the Wilmington, N. C., arch, . . . (3) a series of low magnetic highs extending in an interrupted irregular line from the latitude of that of Beaufort" (p. 1953). Johnson's remarks on magnetic disturbances in northeastern North Carolina are published only in a short abstract (p. 1951).

Richards in 1945 dealing with well records of North Carolina Coastal Plain wrote about a "conspicuous high . . . noted in the vicinity of Cape Fear, North Carolina", which "has been recognized for a long time and is known as the Great Carolina Ridge", indicated on three cross sections (p. 953 and p. 941-943, figs. 20-21). In 1947 Richards wrote: "In any case the basement and all formations rise sharply near Cape Fear. This is one of the most conspicuous structural features of the East Coast and is called the Great Carolina Ridge or the Cape Fear Arch. At Wilmington, the basement rises to a depth of only 1,109 feet and then dips again toward South Carolina" (p. 47). The Ridge is shown on a generalized cross section from Fort Monroe, Va., to Hilliard, Fla., (p. 46). A third paper in 1948 again reflects the elevated position of the Great Carolina Ridge in a cross section from Fort Monroe, Va., to Paris Island, S.C., (Richards, 1948, p. 55, fig. 2).

Straley and Richards in 1950 gave the same cross section, and with reference to the Ridge stated that the "basement rises at Wilmington to within 425 meters" (correctly 338 meters = 1,109 feet) "of the surface, and extends north-westward toward the Piedmont at an equal or greater elevation" (p. 88, fig. 2).

Berry in 1951 described the "Carolina Ridge", as "one of the most prominent features of the basement" oriented "roughly parallel with the valley of Cape Fear River"(1951, p. 414). He also noted seaward change on the basement slope (1948, p. 87, fig. 1, and 1951, p. 412-413, fig. 116).

Likewise Eardley described the "Cape Fear Arch" as "the most conspicuous feature of the Coastal Plain" (p. 131), indicating it on the index map (p. 70, fig. 22) as a broad bulge of the Cretaceous formations. However, he remarked that "this structure is not truly an arch" as such a structural feature was defined by him in chapter 2 of his book. He concluded that "the unconformities around the Cape Fear Arch indicate the principal times of uplift and erosion to have been at the close of the Cretaceous and again at the close of the Early Miocene".

LeGrand in 1955 referring to the Carolina Ridge stated that "the assumed single homoclinal structure of the Atlantic Coastal Plain becomes complex" in its vicinity. Besides changes in the extension of various Cretaceous formations covering the area, he mentioned a fault line with northeastward trend between Cape Fear River and Black River a few miles from their confluence, and a broad dome-like area, based on presence of brackish ground-water, west of Wilmington, N.C. Although it had "received scant geological attention in the past, the Great Carolina Ridge contains complex structures" (p. 2036-2037).

After this review of opinions, it may be stated that below the area of the Great Carolina Ridge there is a large block of pre-Cretaceous basement rocks, which moved up or down either as a unit, or as smaller blocks independent of adjoining areas of the Atlantic Coastal Plain. This large block of the basement rock extends on its northeast side to Havelock, N.C., and on its southwest side to the neighborhood of Summerville, S.C. At both places the surface of the basement rocks was found at relatively great depths, 2,318 feet at Havelock, and 2,450 feet at Summerville. The crest line is in the vicinity of Wilmington, N.C., where this surface is at its least depth, 1,109 feet, and extends northwestward, approximately parallel to the course of the Cape Fear River, toward Fayetteville, where the block joins the Piedmont. Within this large block are smaller units separated by faults that run at right angle to the northwest-southeast direction of the Great Carolina Ridge, i.e., parallel to the main trend of Appalachian structure. Structural elements of this type were proved by the magnetic investigations of MacCarthy and his associates, and more recently by the observations of LeGrand concerning brackish water areas in the sedimentary cover of the Great Carolina Ridge (LeGrand, 1955, p. 2036).

The separate movement of blocks in the Great Carolina Ridge is "very ancient", as was thought by its first describer, (Dall, 1892, p. 182), but it was proven by LeGrand that movements occurred also within Cretaceous time. The absence of the Tuscaloosa Formation in four deep wells between Conway, S.C., and Jacksonville, N.C., implies a land barrier within the area of the Great Carolina Ridge during Tuscaloosa time. Likewise the apparent absence of the basal strata of the Black CreekFormation in the Wilmington, N.C., well indicated this barrier was above the sea until the latter part of Black Creek time (LeGrand, 1955, p. 2036).

The area was submerged in late Black Creek and Peedee time, but this submergence was followed by an uplift in Paleocene time, since such sediments have not been reported in the area. Submergence during Eocene time only lowered the northeastern flank of the Great Carolina Ridge below the sea, as indicated by surface patches and well data of Upper Eocene limestone. The patches of Middle Eocene (?) sediments near Fayetteville and Raleigh, N.C., also are confined to this flank of the Ridge. On the southwest flank the Black Mingo Formation and overlying younger members of the Eocene series appear only at much greater distances from the crestline of the Ridge.

The submergence during Eocene time was followed by an uplift of greater extent. Along the length of the Great Carolina Ridge the presence of Oligocene sediments has been suggested only by McLean with a questionable reference by Richards (1948, p. 62), from the shallow well at Camp Lejeune, Onslow Co., N.C. On the southwest flank of the Great Carolina Ridge no sediments have been definitely determined as of Oligocene age. The nearest area in South Carolina where such sediments (Flint River Formation) occur lies far distant from the Great Carolina Ridge, near the Savannah River. Also in case if the Cooper Marl of South Carolina repeatedly "shifted back and forth between the Eocene and the Oligocene" by subsequent authors, should be definitely verified as of Oligocene age, as Cooke and MacNeil wrote, the area covered by it lies on the southwesternmost flank of the Great Carolina Ridge (1952, p. 27).

The total absence of Lower and Middle Miocene sediments in the area of the Great Carolina Ridge, as shown by Brown's recent study of well logs from the Coastal Plain of North Carolina (1958, figs. 7-9), is good proof that the entire length of the Great Carolina Ridge during the Early Miocene and Middle Miocene was still above sea level. A new submergence in the Late Miocene resulted in the southeastern part of the north flank of the Ridge being covered by the transgression of the Yorktown sea, while the northwestern portion of this flank remained uncovered. Only during the youngest phase of Upper Miocene transgression, the time of the deposition of the Duplin Formation, was the whole area perhaps below sea level, except for an area on the south bank of the Neuse River near Mt. Olive, N.C., which remained as a peninsula.

Evidence is lacking concerning movements in post-Miocene time.

Perhaps a southeastern strip along the shore line was covered by Pliocene and Pleistocene seas.

THE HATTERAS AXIS

The first author who suggested that "the projection of Cape Hatteras is due to subterranean disturbances" was Shaler in 1871 (p. 112), when he considered the causes "which have led to the production of Cape Hatteras". Although he did not specify the direction of these disturbances, the fact that he linked them with a ridge between Richmond, Va., and Weldon, N.C., clearly reveals a northeast-southwest direction parallel with the Appalachian trend.

In 1891, McGee twice referred to the "Hatteras Axis" - in neither case specifying any direction - "as an axis of interruption or change in epeirogenetic movement during every geologic period since the Cretaceous" (p. 403), and as "an axis of minimum subsidence and minimum uplift" (p. 503).

In 1894, Hayes and Campbell mentioned the Hatteras Axis, and gave its direction as northwest-southeast, a transverse line to the Appalachian trend. This may be deduced from their statement that if the direction of the Hatteras Axis is continued "across the Ohio River its direction will be found to coincide with that of the main or northwestward branch of the Cincinnati Arch" (p. 81), whereas the "Charleston-Memphis axis", passing Atlanta, Ga., forms "a tangent to the great northwestward bend of the Tennessee River" (p. 82). Since then the Hatteras Axis has always been considered as a structural feature transverse to the Appalachian trend.

In 1899, Glenn discussed the Hatteras axis pointing to its role in sedimentation during the Triassic period and also in the Middle Miocene, and referred to it being not "a narrow belt with a close approach to the idea of a line but rather a broad belt or region" (p. 379).

In 1926, Stephenson, referring to major features in geology of the Atlantic and Gulf Coastal Plain, indicated it on his sketch map as an axis, in which two downwarped basement surfaces, - one dipping to the southwest, the other to the northeast, - cross each other (pl. 1). In the text, however, he only states: "North of Cape Hatteras the downwarping in late Tertiary and in Quaternary times affected the Coastal Plain more completely than it did south of this point" (p. 472). In a second paper he shows another line more northward, crossing the shore line somewhere near the Virginia-North Carolina boundary (Stephenson, 1928, p. 889, fig. 1). In his text he referred to "a downwarp affecting the North Atlantic Coastal Plain from Maryland to northern North Carolina" which "resulted in the transgression of the Upper Miocene sea inland to the inner edge of the Coastal Plain in North Carolina and Virginia" (p. 891).

Prouty, in general adopting the data from Stepehenson's 1928 sketch map, does not refer to the "Hatteras axis", but replaces it with a "synclinal fold (trough)" in the area of Norfolk, Virginia(p. 485-486, fig. 1). Later Gardner mentioned it as a zone of transition, where northern faunal elements of the Upper Miocene Yorktown formation were replaced by southern types (p. 70, p. 131, etc.).

Richards in 1945 published two cross sections showing subsurface conditions; both show a low in the basement surface at the well at Havelock, N.C., (p. 941-942, figs. 20-21). In 1947 he stated that "the basement drops decidedly between Fort Monroe (2,246 feet) and Hatteras (9,878 feet). However, the north-south slope is not as great as might be indicated since Hatteras is well out to sea. . . If we were to contrast Fort Monroe (2,246 feet), with Havelock, N.C. (2,318 feet) or Morehead City, N.C. (4,036 feet), the slope should not be as great" (p. 47). His generalized cross section in this case indicated the "Hatteras Low" in the line of the Morehead City well (p. 46). Similarly in a 1948 paper he indicated a low in the basement surface in the line of the Morehead City well (Richards, 1948, p. 55, fig. 2). On the other hand, he stated in the same paper that "a study of samples from the deep well at Cape Hatteras shows a thickening of most formations. Also several formations have been recognized in the well that do not crop out in North Carolina" (p. 73).

A cross section in a 1950 paper by Straley and Richards is s milar (p. 88, fig. 2). However, they emphasized the "notable feature . . . the basin between the Dismal Swamp and the Carolina Ridge at Cape Fear" (p. 88).

The last cross section found was published by Spangler in 1950; he again indicated the lowest point on the basement surface as at the Hatteras well (25, p. 120-121, fig. 7).

After this review of opinions, it may be stated that the Hatteras Axis represents a line where all formations are at their greatest depth. The line trends northwestward from the Hatteras well.

The southwest limit of the Hatteras Axis area and the northeast limit of the Great Carolina Ridge block is marked by the Cape Lookout-Neuse Fault Zone (a third transverse structural feature to be discussed later in this article). From this fault zone northeastward well records show the thickening of formations toward the Hatteras Axis. Likewise, on the northeast side of the Hatteras Axis formations thicken southwestward toward the Axis, as already referred by several authors, e.g., by Berry (1951, p. 414).

The Lower Cretaceous series, for example, shows this thickening. Although such sediments were distinguished in the Merrimon and Morehead City, N.C., wells, they are not known in surface outcrops nor in well records in the entire area of the Great Carolina Ridge. Upper Cretaceous formations thicken from both directions toward the Hatteras Axis.

Paleocene sediments are limited mostly to the northeast flank of the Great Carolina Ridge, and are not known to occur south and west of Pitt County, as stated by Brown in his Correlation Chart (1958, table 1). The gradually progressing Late Eocene transgression deposited sediments in the area of the Hatteras Axis; such sediments are missing in surface outcrops, and from the subsurface in an area north of the Neuse River. If the thin unit questionably indicated in Brown's cross section as "unnamed Oligocene" (1950, fig. 4), is proved to be Oligocene, then this unit is likewise restricted to the Hatteras Axis area. It is known only in the records of the Hatteras well and in the Pamlico Sound well, as described by Richards (1948, p. 61), and not in surface outcrops.

While the Great Carolina Ridge remained during the Early and Middle Miocene above sea level, probably continuous sedimentation occurred in the area of the Hatteras Axis. In his Correlation Chart Brown does not show proved sediments of Lower Miocene age, but indicates a thickness of nearly 400 feet in the Hatteras well as "unnamed Lower Miocene (?) unit" (Brown, 1958, table 1 and fig. 4). Brown indicates Middle Miocene sediments also by a question mark, and in his Correlation Chart (table 1) states that these phosphate sand sediments are "not known to occur in outcropping sections", but that their "subsurface distribution" is "localized in Beaufort, Washington, Gates and Hyde Counties", i.e., the area of the Hatteras Axis.

The Late Miocene transgression of the Yorktown sea covered the entire area of the Hatteras Axis and deposits accumulated to considerable thicknesses, such as 325 feet at Edenton, N.C., and more than 500 feet in the Hatteras well, as shown in Brown's cross section (fig. 4). Moreover, faunal evidences prove, according to Gardner (1944, p. 70, pl. 131, etc.), that the sediments of the Yorktown formation north of the Neuse River were deposited in an embayment that was removed from the influence of warmer oceanic waters. This embayment was protected by the peninsula which remained during the Late Miocene time above sea level in the area of Mount Olive, N.C., on the south side of the Neuse River valley.

THE CAPE LOOKOUT - NEUSE FAULT ZONE

Besides the two main structural features which have just been discussed, two others are indicated. The Cape Lookout-Neuse Fault Zone - a third northwest-southeast directed feature transverse to the Appalachian trend - is midway between the Great Carolina Ridge and the Hatteras Axis. Its existence is indicated by the difference in the depth of the basement rock surface, 2,318 feet on the southwest side of the fault zone in the Havelock well and 4,000 feet on the northeast side in the Merrimon test wells, as well as in the Morehead City well. Nearer the Piedmont, in the area of Goldsboro, N.C., the presence of such a fault zone is suggested in that the Upper Eocene Castle Hayne Limestone, which on the right bank of the Neuse River overlies the eroded surface of the Black Creek and Tuscaloosa Formations, is missing both in surface outcrops and in well records from the left bank area north of the Neuse River. Moreover, the well data and cross sections of Brown (1958, figs. 2-9) indicate that north of the Neuse River there is an area, bounded approximately on the south by the Neuse and on the east by a line drawn along the eastern boundaries of Martin, Pitt, and Lenoir counties, where the Miocene Yorktown sediments directly overlie the Cretaceous formations, without intervening Middle or Upper Eocene sediments. This elevated block must have been above sea level until the end of Middle Miocene time, but sank with the oncoming transgression of the Late Miocene Yorktown sea independently of the adjoining areasouth of Neuse River, which remained above sea level during Late Miocene Yorktown time.

The "Cape Lookout-Neuse Fault Zone" is suggested also by a line along which older sediments became silicified during emergence of this area between Late Eocene and Late Miocene times. The occurrences of silicified older sediments in the Piedmont area, such Eocene deposits in the railway cut at Garner, N.C., and at the boundary of Wake and Johnston Counties on old Highway 70 between Clayton and Auburn, N.C., (Richards, 1951, p. 14), the Eocene outcrop with silicified Bryozoan stocks southwest of Dudley, Wayne Co.,[1] finally the silicified sandstone southeast of Kinston, N.C., (Stuckey, 1928, p. 22-23), lie in an approximate northwest-southeast line, coinciding with the Cape Lookout-Neuse Fault Zone. This fault zone limits the block of older sediments on the southwest side, which was not covered by the Eocene Castle Hayne Limestone, but became inundated by the Yorktown sea in the Upper Miocene.

STRUCTURAL FEATURES PARALLEL TO APPALACHIANS

In addition to the three structural features discussed in the foregoing paragraphs, two structural features may be mentioned which run parallel to the Appalachian trend. The first of these is the line of "subterranean disturbances", as suggested by Shaler (1872, p. 112). This features, however, so far is only suggested by the parallelism of the present coast line southwest of Cape Hatteras to the main trend of the Appalachians.

The second feature, indicated on the sketch map as "Unnamed Fault Zone" is more evident. Its southwest-northeast trend is indicated at the area about 17 miles west of Conway, S.C., where the seaward slope of the basement surface becomes steeper (MacCarthy, 1936, p. 399, fig. 1), the northwestern limit of the magnetically disturbed-zone west of Wilmington, N.C., (MacCarthy, 1936, p. 399, fig. 1), the location of the fault near the confluence of the Cape Fear and the Black River (LeGrand, 1955, p. 2036), and the line along the eastern boundary of Martin, Pitt, and Lenoir counties (mentioned on page 12 as a line where the Upper Miocene Yorktown sediments overlie the Cretaceous formations without intervening Eocene sediments). This data indicate a zone of movements, which in its continuation, is perhaps reflected in the magnetic anomalies observed by Johnson in northeastern North Carolina (1938, p. 1951). This is also the zone where the slope of the basement surface steepens in the North Carolina Coastal Plain area, as illustrated by the cross sections of Berry (1951, p. 413, fig. 116).

MORPHOLOGICAL REFLECTIONS OF STRUCTURE

The morphology of the North Carolina Coastal Plain appears to be connected with the structural features. The drainage areas of the Neuse River and the Cape Fear River within the Coastal Plain have a peculiar asymmetry. The left bank tributaries of the Neuse River and the Cape Fear River are longer, and the slopes on the north banks are steeper, in part, almost escarpment-like. The course of the Roanoke River follows, at least in part, the direction of the Hatteras Axis. The sharp northeast turn of the Neuse River near Kinston, N.C., relates to the unnamed fault zone.

It would seem to be not an accidental coincidence that the peculiar configuration of the capes along the present shore line of both Carolinas has developed relating to these structural features: Cape Hatteras to the Hatteras Axis, Cape Lookout to the Cape Lookout-Neuse Fault Zone, Cape Fear to the Great Carolina Ridge, and perhaps, Cape Romain in South Carolina at the southwest boundary of the Great Carolina Ridge. Such a relation between Cape Canaveral, Fla., and structural lines was recently determined by White (1958, p. 1718-1719). How these structural features, although differing in character, have led to formation of the individual capes needs more detailed studies. The coincidence, in any case, is noteworthy. It should be even more interesting if the northeast-southwest "subterranean disturbances" suggested by Shaler (but not

[1] Information from Richard D. Pusey, U.S. Geol. Survey, Ground Water Branch, Raleigh, N.C., and also personal observation.

proved to date), as a cause leading to the "projection of Hatteras", could be proven as a further structural feature of the Coastal Plain of the Carolinas. In this case the crossing of the transverse structures with this northeast-southwest structure would provide another basis to the idea that the capes have not been formed accidentally at their locations, but through the influence of structural control.

CONCLUSIONS

The North Carolina Coastal Plain is not a simple homoclinal structure but is more complex. The transverse structural features, the Great Carolina Ridge and the Hatteras Axis, influenced the transgression and regression of the seas in different geological times. The middle feature, the Cape Lookout-Neuse Fault Zone had a similar role, but the movements along this zone affected smaller areas of deposition. Besides the parallelism of the assumed northeast-southwest line of Shaler to the main trend of Appalachian structure, an unnamed zone of structural disturbances is suggested. Movement along these features also influenced the morphology of the North Carolina Coastal Plain, and such an influence may be suggested for the whole extent of the Atlantic Coastal Plain. The basement rock beneath the sedimentary cover has the character of a peneplained block mountain rather than that of a folded mountain chain. Former folds, if they were once present, have been obliterated by fault systems developed since the Appalachian Revolution. Structural conditions of the Atlantic Coastal Plain, and gravity and other anomalies indicated more recently by Skeels (1950, plates I-IV, figs. 1-2), can perhaps be more easily interpreted by referring them to blocks in the basement rock mass differing in position.

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Bendeleben, and Windy Creek plutons; the Hunter Creek and Granite Mountain plutons; and the Selawik Hills pluton.

120. Computer Mapping of Surficial Geology near Halloran Springs (California) using 12 Channel Multispectral Data. JEFFREY L. EHRENZELLER (Indiana State University), ROBERT C. HOWE (Indiana State University), STEVEN A. STANLEY (Indiana State University).

A recent trend in geological remote sensing has been to map surficial materials using multispectral data collected by satellite and aircraft scanners. Research by the authors at the Indiana State University Remote Sensing Laboratory (ISURSL) involved the production of a surficial geology map of the Halloran Springs area from analysis of 12 channels of aircraft multispectral data collected by the Environmental Research Institute of Michigan (ERIM) with the Michigan M-7 optical mechanical scanner system. To produce the surficial geology map, a correction was made for albedo. Then a ratioing technique, in which the revised spectral values in a thermal band (channel 11) were divided by the values in a visible band (channel 6), was used to produce an alphanumeric printout displaying relative differences in thermal inertia. This map was similar to a large scale surficial geology map prepared by conventional field mapping methods. Two rock types, quartz monzonite and basalt, which were not separable in this area using conventional remote sensing methods, were separated on the thermal inertia map. These techniques, although still in the developmental stage, are encouraging.

121. Recording Time Frames and the Sun-Earth Linkage. MALCOLM MORROW, (U.S. Gov. Ret.)

Solar energy input (SEI) varies daily. Data for daylight and the night following (before the next SEI) make a natural 24 hour unit. Time frames for local time and U.T. are not congruent with these limits; both mix observations from two SEI's, distorting averages, etc. The corpus of recorded data is contained in these time frames. Selection and regrouping produce congruent data. These, with an improved curve comparison technique, yield important findings; primarily, proof of a fine-structured sun-earth linkage. Sunspot (SS) variations produce a basic terrestrial chart pattern of maximum daily temperatures (T max) that is inversely correlated with the SS curve. Temperature curves above the tropopause are correlated inversely with the T max curve, and directly with the SS curve. (Polar areas omitted for lack of data.) Local station curves are variants of the basic curve, modulated primarily by latitude and longitude. Typical local T max curves move as a unit, the configuration intact but for systematic modulations, from W to E about 15° daily. Examples of T max correlations: (direct) sunshine, cal. cm^2; ultra-violet radiation; water temperature in evaporation pan; (inverse) atmospheric ozone, esp. with data from potassium papers; air pressure; (complex) geomagnetism. Finally, there is a solar-based 27 day recurrence tendency in the conformation of the T max curve.

122. Heaving Shale- Evidence for the Role of Bacteria in its Development. EMMY BOOY (Colorado School of Mines).

Shales in the areas around the cities of Pittsburgh, Pennsylvania, Cleveland, Ohio, and Ottawa, Ontario have expanded up to several inches subsequent to buildings being built upon bedrock. This expansion is caused by the growth of gypsum and, frequently, jarosite between layers of pyritiferous calcitic shale. Damage to the overlying structures has been extensive, sometimes forcing their abandonment.

Similar shale in the White Pine Mine in Michigan has shown no evidence of such expansion in the mine openings. The major difference between heaving and non-heaving shale appears to be the presence or absence of Thiobacillus ferrooxidans whose metabolism generates sulfuric acid which reacts with the calcite present in the rock to form gypsum.

Where conditions favorable for population explosions of the bacteria occur (pH 2, 35° C), severe expansion of the shales have occurred. In conditions, as at White Pine, which the bacteria find hostile, mineralogically similar shales show no evidence for expansion due to the growth of secondary minerals.

123. The Ice-Sheet-Tectonics Synthesis. RALPH FRANKLIN WALWORTH, Earth Science Consultant.

Earth's topography is studied to discern the forces which shaped the surface. Sixty percent of Earth's surface consists of two vast, weather-planed "terraces." The Upper Terrace, comprising 30% of the total area, ranges about present sea level. The Lower Terrace (sometimes "rise") lies c. 5000 meters below present sea level on the ocean floors. River-cut canyons and valleys run down to deltas and alluvial fans on the Lower Terrace.* The Synthesis relates these features and proposes that all Arcuate Elevations (island arcs, mountain ranges) on the Upper Terrace were formed by ice-sheet tectonics; and that ice-sheet dehydration of continental seas precipitates salts, metals, calcites, silicates, and sometimes petrifies life forms in the process. Petroleum, coal and "phosphate rock" are interred simultaneously. Below the Lower Terrace oceans are then dehydrated by evaporation, resulting in precipitation of the same minerals on ocean floors, but without organics. The formation locations of minerals and organic deposits may be deduced from these factors.

* WALWORTH, Ralph Franklin with SJOSTROM, Geoffrey Walworth: SUBDUE THE EARTH, Delacorte, 1977.

124.

Structural Control of Mesozoic-Cenozoic Deposition, North and South Carolina Coastal Plain. HARRIS, W. BURLEIGH, ZULLO, VICTOR A. (Univ. North Carolina at Wilmington, North Carolina 28403) and BAUM, GERALD R. (College of Charleston, Charleston, South Carolina 29401).

Abrupt changes in distribution, facies, thickness and attitude of Cretaceous and Cenozoic sediments in the Carolina Coastal Plain reflect episodic and differential movements along three fault zones. "Santee fault", Cape Fear arch (fault) and Neuse fault are subparallel, trend NW-SE, and extend from the inner Coastal Plain to the coast between Georgia and Hatteras embayments. Pre-late Cretaceous movement along Cape Fear and Neuse faults is indicated by offset and changes in thickness of lower Cretaceous beds that are not reflected in overlying units. Paleogene activity along Neuse fault is indicated by a shift in structural and depositional strike in post-Paleocene units and abrupt thickening of upper Eocene through lower Miocene units north of the fault. Sporadic Paleogene uplift of the south side of Cape Fear fault is suggested by the angular unconformity between Cretaceous and overlying units and by intraformational disconformities and Dorag dolomitization in the middle Eocene Castle Hayne Limestone. Post-medial Eocene activity along "Santee fault" is shown by rapid thickening of upper Eocene beds southwest of the fault and their onlap of older units to the northwest. Plio-Pleistocene activity along Cape Fear and Neuse faults is indicated by tilting of coastal terraces, distribution of marine sediments and derangement of drainage patterns.

125. Hydrogeology and Development of Spring Cave, Colorado. R. MARK MASLYN (Consulting Geologist), JAMES A. PISAROWICZ (University of Denver).

Spring Cave is developed wholly within the early Mississippian Leadville Limestone. In this area the Leadville is approximately 62 m thick and consists of two main divisions, an upper massive cliff forming lime packstone and a lower section of thinner limestone and dolomite beds. The cave entrance is located at the head of a small drainage tributary to the South Fork of the White River at an elevation of 2380 m. This

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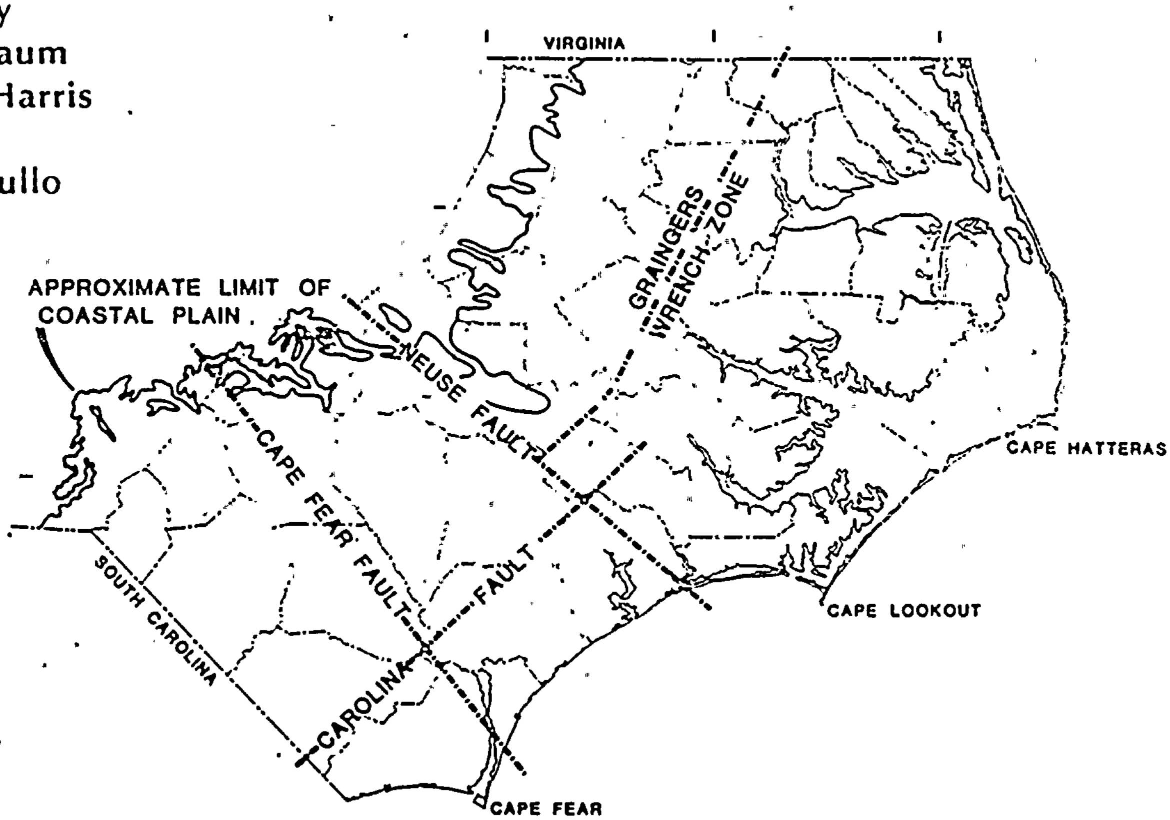
STRUCTURAL AND STRATIGRAPHIC FRAMEWORK FOR THE COASTAL PLAIN OF NORTH CAROLINA

Edited By
Gerald R. Baum
W. Burleigh Harris
And
Victor A. Zullo



Carolina Geological Society
And
Atlantic Coastal Plain Geological Association

Field Trip Guidebook
October 19-21, 1979
Wrightsville Beach, North Carolina

TECTONIC EFFECTS ON CRETACEOUS, PALEOGENE, AND EARLY NEOGENE SEDIMENTATION, NORTH CAROLINA

by

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INTRODUCTION

The Atlantic Coastal Plain Province is an oceanward thickening wedge of SE dipping Mesozoic-Cenozoic sediments and sedimentary rocks that unconformably overlie an oceanward dipping pre-Cretaceous basement. Three major structural features modify the general oceanward slope of the basement: Cape Fear fault in North Carolina, Ft. Monroe uplift (Norfolk arch) in Virginia, and Normandy arch in New Jersey.

Traditionally, the Atlantic Coastal Plain has been considered the stable western limb of an offshore geosyncline that has experienced little or no fault activity, only gravity induced subsidence and concomittant uplift (Murray, 1961). Consequently, most geologic interpretations of Coastal Plain geology have been governed by this tradition, with most workers not considering that tectonic activity may have affected Mesozoic-Cenozoic sediment deposition. Therefore, in many cases, the lack of recognition and consideration of the effects of tectonic activity have lead to a general misunderstanding and misinterpretation of Coastal Plain geology.

Hobbs (1904) recognized major lineaments along the Atlantic border region and suggested that the lineaments were the result of a crustal fracture field. Brown et al. (1972) in a study based on subsurface data established a regional tectonic framework for the Atlantic Coastal Plain and found that many of their structural axes coincided with those of Hobbs. Recently other workers have suggested Cretaceous and/or Tertiary deformation in the Coastal Plain of Maryland (Jacobeen, 1972), Virginia (Mixon and Newell, 1977; Dischinger, 1979), North Carolina (Brown et al., 1977; Baum et al., 1978); South Carolina (Inden and Zupan, 1975; Zupan and Abbott, 1975; Higgins et al., 1978; Rankin et al., 1978; Zoback et al., 1978; Baum and Powell, 1979), and Georgia (Prowell et al., 1975; Cramer and Arden, 1978; Cramer, 1979).

The main purpose of this study is to refine and detail the basement-rooted tectonic framework introduced by Brown et al. (1972) for the Atlantic Coastal Plain and to show its sequential effect on Cretaceous, Paleogene, and early Neogene sedimentation in North Carolina. Tectonic activity also has affected Plio-Pleistocene sedimentation, drainage and geomorphology, and is discussed by Zullo and Harris in the following paper.

GEOLOGIC SETTING

The emerged North Carolina Coastal Plain is underlain by Lower Cretaceous to Quaternary sediments and sedimentary rocks that extend from a feather-edge along the Fall Line to a maximum thickness greater than 3 km at Cape Hatteras. The area represents a typical belted Coastal Plain with younger beds progressively cropping out closer to the coast. Structurally, four major features rooted in the pre-Coastal Plain basement have periodically affected Mesozoic-Cenozoic sedimentation: Cape Fear fault, Neuse fault, Carolina fault, and Graingers wrench zone (Fig. 1). Interpretations of the times of tectonic activity are discussed later in this paper.

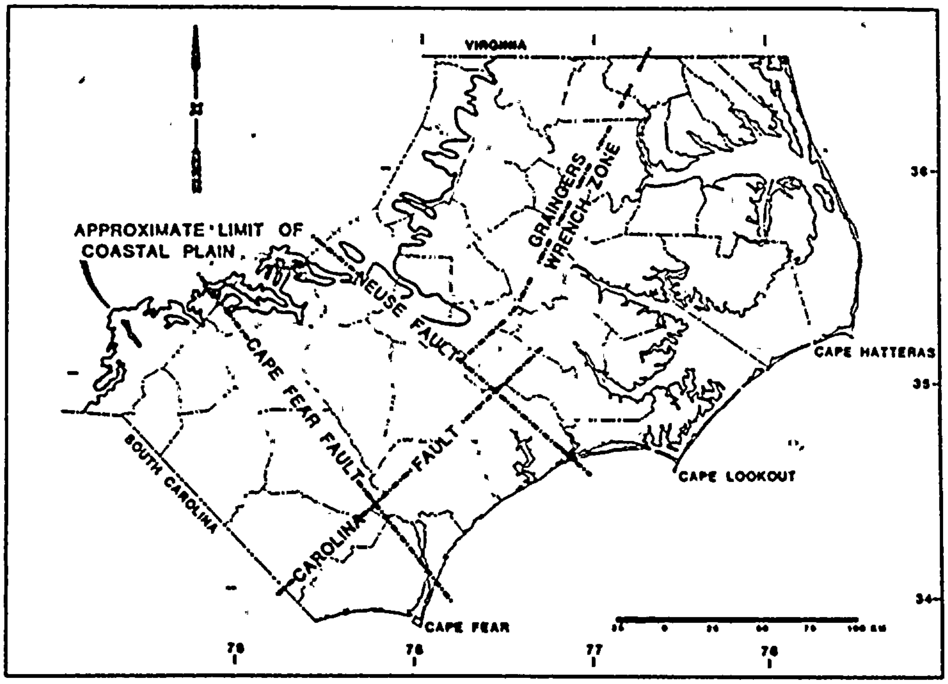


Figure 1. Major structural features of the North Carolina Coastal Plain.

Cape Fear Fault

Dall and Harris (1892) originally recognized a major positive feature (Cape Fear arch) along the Cape Fear River; however, Stephenson (1923) is usually given credit for first delineating the structure. Since then many workers have documented the presence of a structure along the Cape Fear River that has undergone periodic movement (MacCarthy, 1936; Mansfield, 1937; Richards, 1945; Straley and Richards, 1950; Baum et al., 1977; Harris et al., 1977). Harris et al. (1979) suggested that the Cape Fear arch represents a basement fault that has experienced episodic and differential movement from Lower Cretaceous through the Quaternary.

Cape Fear fault trends NW-SE and can be traced from about Fayetteville, Cumberland County, to Carolina Beach, New Hanover County. The approximate location of the fault is NE of the line separating the Peedee drainage basin from the Cape Fear drainage basin. The direction of relative movements along Cape Fear fault has periodically reversed.

Neuse Fault

Ferenczi (1959) postulated that a fault occurred along the Neuse River and called the feature the Cape Lookout-Neuse River fault zone. Baum et al. (1978) also recognized the feature and shortened the name to the Neuse fault. Subsequently, Harris et al. (1979) changed the trend of Neuse fault. Neuse fault trends NW-SE parallel to Cape Fear fault and can be traced from about Smithfield, Johnston County, to Bogue Inlet at the mouth of the White Oak River, Onslow-Carteret County line. The fault is probably part of a series of basement faults that occur between the Neuse and New Rivers that have a sense of relative movement with the north side down. Movement along Neuse fault has occurred periodically from Lower Cretaceous through the Quaternary.

Carolina Fault

LeGrand (1955) and Ferenczi (1959) postulated a fault zone trending NE-SW, parallel to the coast, that could be traced through the vicinity of Kinston, Lenoir County. The unnamed fault was suggested by the occurrence of saltwater incursion near the confluence of the Cape Fear and Black Rivers. Baum et al. (1978) named the feature Carolina fault and showed that the fault can be traced from the confluence of the Cape Fear and Black Rivers, Pender County, to Kinston, Lenoir County. Recent work suggests that the trace of the fault passes through Cove City, Craven County.

Graingers Wrench Zone

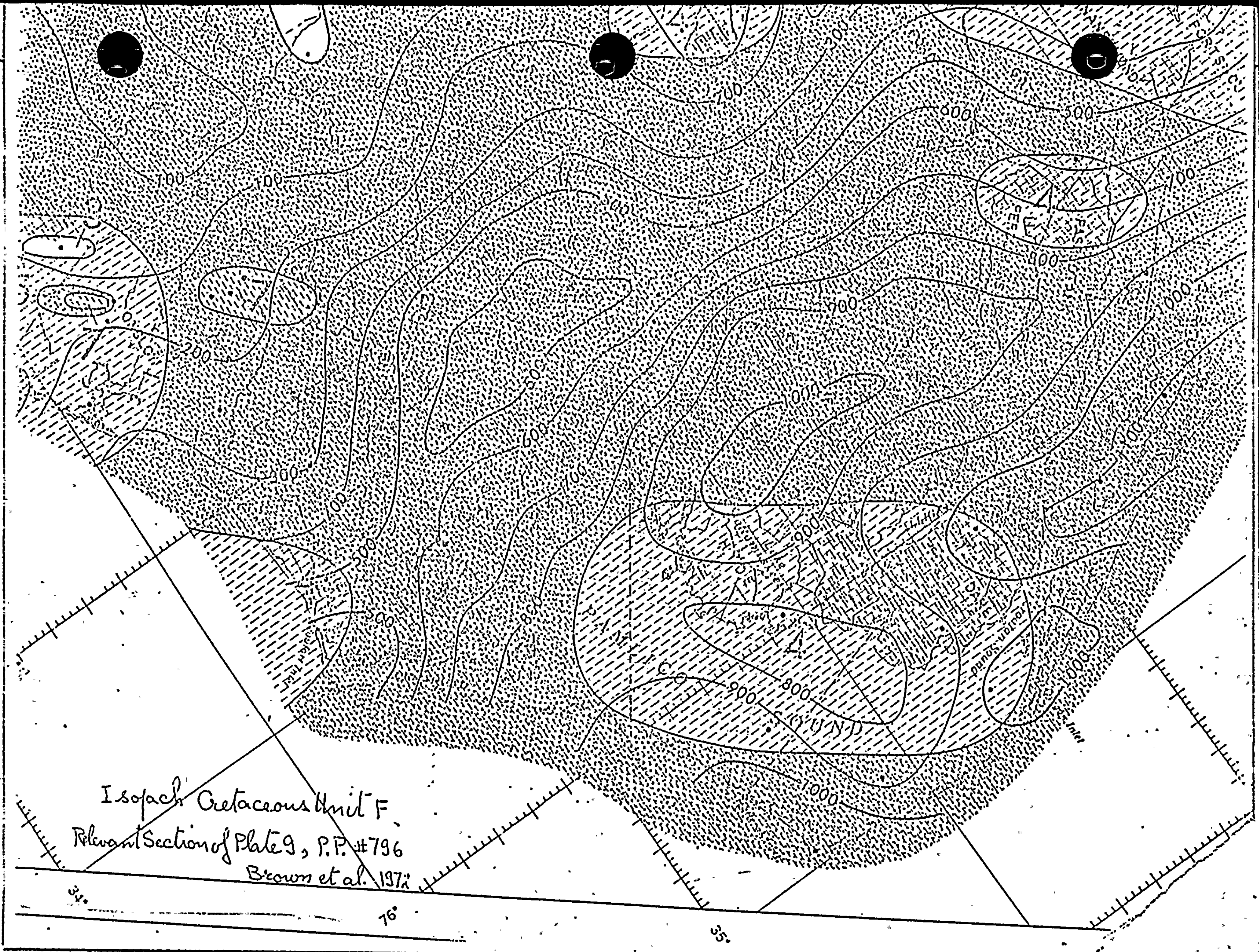
Graingers wrench zone was proposed by Brown et al. (1977) to explain surface topography and anomalous exposures of the Paleocene Beaufort Formation in the Kinston area, Lenoir County. The wrench zone trends NE-SW (parallel to the Carolina fault) and can be traced through the town of Graingers, Lenoir County. Because the projected trace of Graingers wrench zone corresponds to gravity anomalies identified by Johnson (1975), and to geomorphic and stratigraphic features in southeast Virginia, Graingers wrench zone may extend for 250 km. Brown et al. (1977) interpret that the most recent movement along the fault zone has resulted from wrenching along a pre-Coastal Plain basement fault.

Graingers wrench zone consists of a series of en echelon faults that extend north from Neuse fault. Although the sense of relative movement on each individual fault varies within the zone, there is an overall sense of downward movement progressively toward the east. Won et al. (1979) suggest that the Graingers wrench zone coincides with a Triassic Basin border fault and have identified the width and length of the basin from gravity data. The 20 km wide basin occupies the areas bounded by the Graingers wrench zone and Carolina fault. The Graingers fault was active as early as the Triassic (pre-Coastal Plain sedimentation), but wrench movement probably occurred during the Paleocene and maybe as recently as the Quaternary.

DISCUSSION

Cretaceous

Clastic sediments of the Fredericksburg and Washita Stages (Cretaceous Unit F of Brown et al., 1972) represent the earliest widespread deposition of Mesozoic sediments in North Carolina. Unit F only crops out south of the Neuse fault, along the Fall Line, but is widespread throughout the Coastal Plain (Fig. 2). The distribution, thickness, and attitude of Cretaceous Unit F suggests that syn-depositional tectonic activity affected Fredericksburg and Washita deposition.



Isopach Cretaceous Unit F
Relevant Section of Plate 9, P.P. #796
Brown et al. 1972

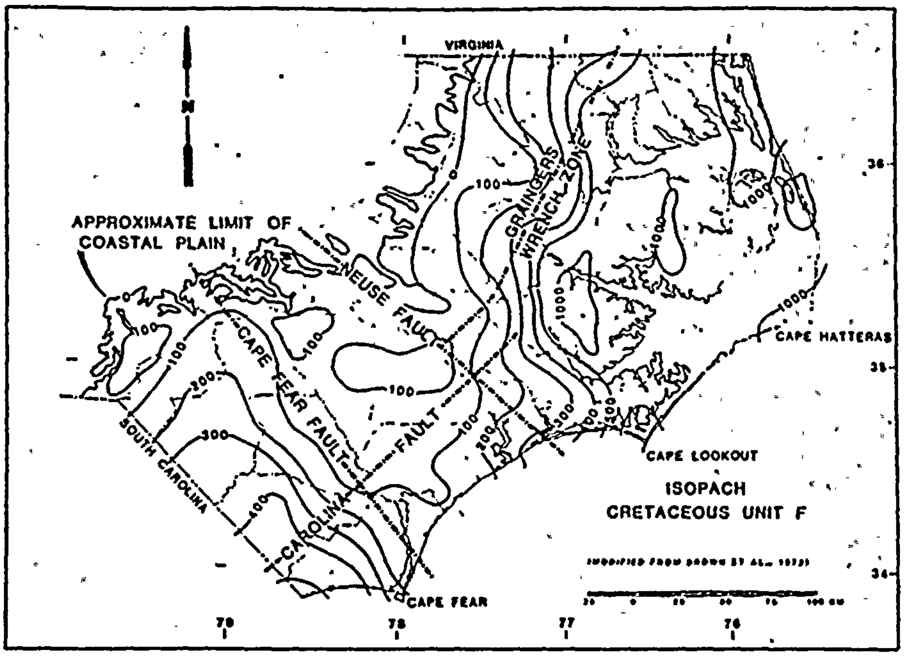


Figure 2. Isopachous map of Cretaceous Unit F (modified from Brown *et al.*, 1972).

Isopachous mapping of Cretaceous Unit F (Fig. 2) reveals that the unit attains a thickness of about 100' (30 m) between the traces of Cape Fear and Neuse faults. South and north of the faults, respectively, Cretaceous Unit F obtains a thickness of about 500' (150 m). Because isopachous relationships are related to basin configurations, as well as tectonism, three possible interpretations can explain the isopach map of Cretaceous Unit F:

1) pre-depositional subsidence north of Neuse fault and south of Cape Fear fault,
2) syn-depositional subsidence north of Neuse fault and south of Cape Fear fault, with sediment deposition equaling subsidence,
3) post-depositional uplift of the area between Neuse and Cape Fear faults.

Comparison of structure contours on top of Cretaceous Unit F (see Brown *et al.*, 1972, Plate 9) with the isopach map of the unit favors interpretation 2.

If pre-depositional uplift elevated the block between Cape Fear and Neuse faults, consequently controlling sedimentation, structure contours on top of Cretaceous Unit F should indicate a structural nose or positive area between the faults that mimics the thinning of the unit illustrated by the isopachous map. Because no high or structural positive is present, pre-depositional uplift probably was not important.

By the same line of reasoning, if post-depositional uplift elevated the block between Cape Fear and Neuse faults, structure contours on top of Cretaceous Unit F should also indicate a positive area between the faults. In addition, if the assumption is made that post-depositional uplift occurred prior to deposition of overlying Cretaceous Unit E, then an isopach of Unit E should mimic the isopach of Unit F by indicating thick areas north and south of Neuse and Cape Fear faults, respectively. Also, lithofacies distributions of Cretaceous Unit E would indicate that the uplifted area had served as a source area during deposition. Because the isopachs of Cretaceous Unit E and Unit F are dissimilar in pattern and because available evidence suggests that Unit E did not serve as a source area, post-depositional uplift of the area between Cape Fear and Neuse faults probably is not responsible for the distribution and thickness of Cretaceous Unit F.

We suggest then that isopachous mapping and structure contours on top of Cretaceous Unit F support syn-depositional subsidence south and north of basement-rooted Cape Fear and Neuse faults, respectively, with sediment deposition balancing subsidence. Regardless of whether syn-depositional subsidence occurred independent of pre- or post-depositional uplift, isopachous mapping of Cretaceous Unit F documents that faulting was active in controlling deposition of the unit. Differences in the amount of dip on Cretaceous Unit F north and south of Neuse fault and the position and outcrop pattern of the unit along the Fall Line suggests some post-depositional shifting or readjustment of the block north of Neuse fault. Available data suggests that Carolina and Graingers faults were not active during the Lower Cretaceous.

There is no evidence of movement along Cape Fear and Neuse faults and Graingers and Carolina faults during the Upper Cretaceous.

Paleogene

Paleocene. The Paleocene Beaufort Formation crops out in Lenoir and Craven Counties and contains Danian (Brown et al., 1977) and Thanetian equivalents (Harris and Baum, 1977). Danian beds are referred to as the Jericho Run Member and are locally present as a silicified mudstone assigned to the P1 planktic foraminifera zone (Brown et al., 1977). Thanetian beds are unnamed and disconformably overlie the Jericho Run Member of the Cretaceous Peedee Formation. These beds consist of consolidated sandy, glauconitic foraminiferal biomicrosparite and unconsolidated sandy, foraminiferal biomicrite. They correlate with the P4 planktic foraminiferal zone of Berggren (1971). Authigenic glauconites from the Thanetian beds have been dated by Harris and Baum (1977) at 55.7 and 57.8 m.y.

Outcrops of the Beaufort Formation occur near the intersection of Neuse fault and Graingers wrench zone and are related to a structural mosaic of horst, graben, and half grabens with the faults trending NE-SW (Brown et al., 1977) (Fig. 3). These en echelon faults overlie a buried Triassic Basin (Won et al., 1979). Variations in thickness and sudden lateral terminations of the Jericho Run Member and Thanetian

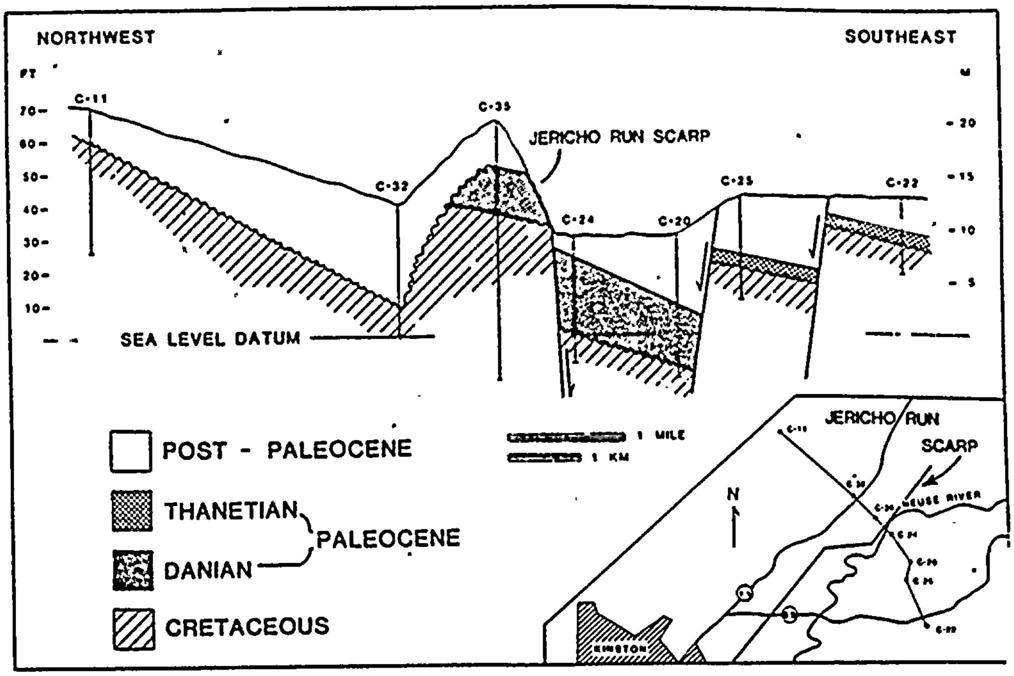


Figure 3. Northwest-southwest section across Graingers Wrench Zone.

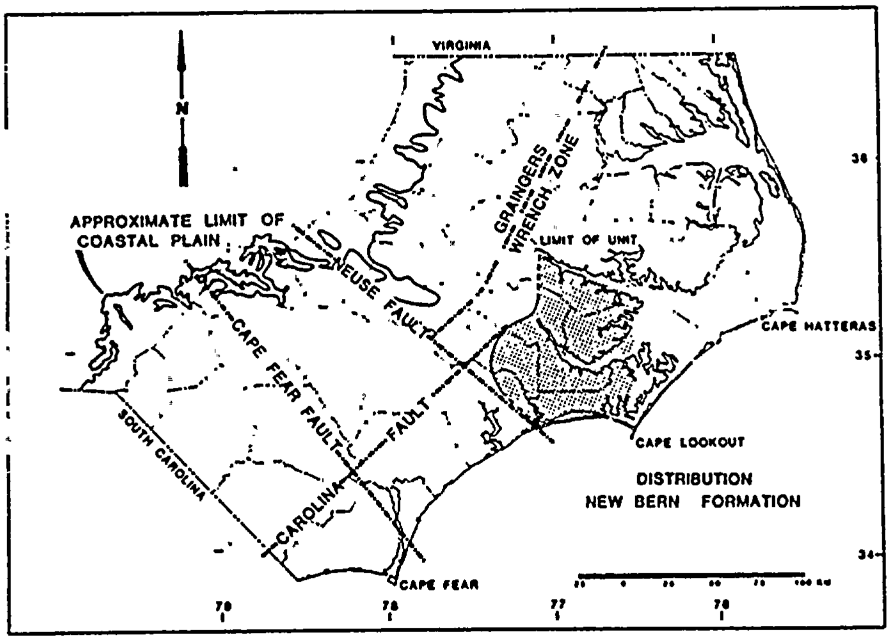


Figure 4. Distribution of the upper Eocene New Bern Formation.

sediments (Fig. 3), suggest that these faults experienced episodic movement during the Paleocene(?) and the post-Paleocene. Except for a minor reentrant of Paleocene beds along the Pender-Onslow County line (see Brown et al., 1972, Plate 15), the Beaufort Formation is restricted to the area north of Neuse fault. The lack of a regionally recognizable marker horizon overlying the Beaufort Formation circumvents establishing the time of post-Paleocene movement. However, offset of the Eocene Castle Hayne Limestone suggests post-Eocene deformation. Brown et al. (1978) recognize the following features that are associated with a NE-SW trending scarp that borders Jericho Run: 1) an uplifted stratigraphic marker horizon, 2) triangular faceting of the scarp, 3) extensive parallel ravinement normal to the scarp, and 4) the presence of breccias along the toe of the scarp. The excellent preservation of these features in a humid environment suggests some Quaternary movement along the Graingers wrench zone.

Eocene. One of the most extensive transgressions of the Cenozoic in North Carolina occurred during the middle to upper Eocene. Eocene seas transgressed most of the Coastal Plain reaching the Fall Line, depositing tropical marine carbonates atypical of other Cenozoic sedimentary units in North Carolina.

The middle to upper Eocene Castle Hayne Limestone consists of three prominent facies: lower phosphate-pebble conglomerate, middle bryozoan biosparrudite, and upper bryozoan-sponge biomicrudite. Bryozoan biosparrudite and bryozoan biomicrudite are the two dominant facies of the Castle Hayne Limestone. Numerous diastems and Dorag dolomitization in the bryozoan biomicrudite in the lower Cape Fear area (Brunswick and New Hanover Counties), suggests movement of Cape Fear fault during middle and upper Eocene.

The upper Eocene New Bern Formation consists of sandy, pelecypod-mold biomicrosparrudite and represents the youngest outcropping Eocene strata in North Carolina (see Baum et al., this volume). "Outcrops of the New Bern Formation are confined to an area lying between the Neuse and Trent Rivers..." (Baum et al., 1978). The New Bern Formation is restricted to the area north of the Neuse fault and east of Carolina fault (Fig. 4). Because of this restriction, and because the New Bern Formation represents a major lithologic change from a carbonate dominated regime (Castle Hayne Limestone) to a clastic dominated regime (New Bern Formation), the area north of Neuse fault was downdropped during latest Eocene. Movement on Neuse fault appears to coincide with movement along "Santee" fault, in the Charleston area of South Carolina (Harris et al., 1979; Baum and Powell, 1979; Baum et al., this volume).

Oligocene. The Oligocene Trent Formation is restricted to the area north of New River, Onslow County, east of Carolina fault. However, the distribution, thickness, and lithofaciés of the Trent Formation do not suggest Oligocene movement of Neuse and Carolina faults.

Neogene

Miocene. The lower Miocene Belgrade and Silverdale Formations (and the Crassostrea beds) are restricted to the area east of the Trent Formation and do not appear to be related to tectonic activity. Depositional strike of these units is N-S; consequently, because of the orientation of the North Carolina coast, they do not crop out south of New River. Fossils assignable to the lower Miocene have been found on Onslow and Topsail beaches, suggesting that these units are exposed on the continental shelf south of New River.

The middle Miocene Pungo River Formation is restricted to the area north of Neuse fault and east(?) of Graingers and Carolina faults (Fig. 5). Miller (1971) suggested that deposition of this unit was controlled by NE-SW trending faults. Deep-water deposits (100-200 m) of phosphate, diatomite, and carbonate suggest that the rate of subsidence exceeded the slow supply of terrigenous sediments (Gibson, 1967).

SUMMARY

1. Mesozoic and Cenozoic deposition in the North Carolina Coastal Plain was affected by four basement-rooted structural elements: Cape Fear fault, Neuse fault, Carolina fault, and Graingers wrench zone.
2. During the lower Cretaceous (Fredericksburg and Washita stages), syn-depositional tectonism along Cape Fear and Neuse faults resulted in elevation of the area between the faults. Consequently, isopachous mapping of Fredericksburg and Washita sediments reflect thick areas south and north of Cape Fear and Neuse faults, respectively, with an intervening thin area. Structure contours on top of Fredericksburg and Washita sediments do not reflect this uplift, therefore, sediment supply and deposition kept pace with the rate of uplift.
3. The Paleocene Beaufort Formation is restricted to the area north of Neuse fault, and appears to be related to reactivated Triassic faults. Graingers wrench zone and Carolina fault bound and limit Paleocene deposits and reflect movement during the Paleocene. The distribution, thickness and lithofacies of Danian and Thanetian beds support Paleocene movement. The excellent surface preservation of a surface scarp coincident with Graingers wrench zone suggests Quaternary movement.
4. Middle to upper Eocene sediments (Castle Hayne Limestone) support Eocene tectonism in the Coastal Plain. Numerous diastems and Dorag dolomitization in the upper biomicrudite in the lower Cape Fear region suggests late Eocene movement along Cape Fear fault. The restricted occurrence of the upper Eocene New Bern Formation to the east of Carolina fault and north of Neuse fault suggests latest Eocene activity along Carolina and Neuse faults.
5. The distribution, thickness, and lithofacies of Oligocene sediments (Trent Formation) suggests no tectonic activity during that epoch.
6. The distribution of Belgrade and Silverdale Formations and the Crassostrea beds do not suggest tectonism during the lower Miocene. The restriction of the middle Miocene Pungo River Formation to the area north of Neuse fault suggests that Neuse fault was active with the north side down.

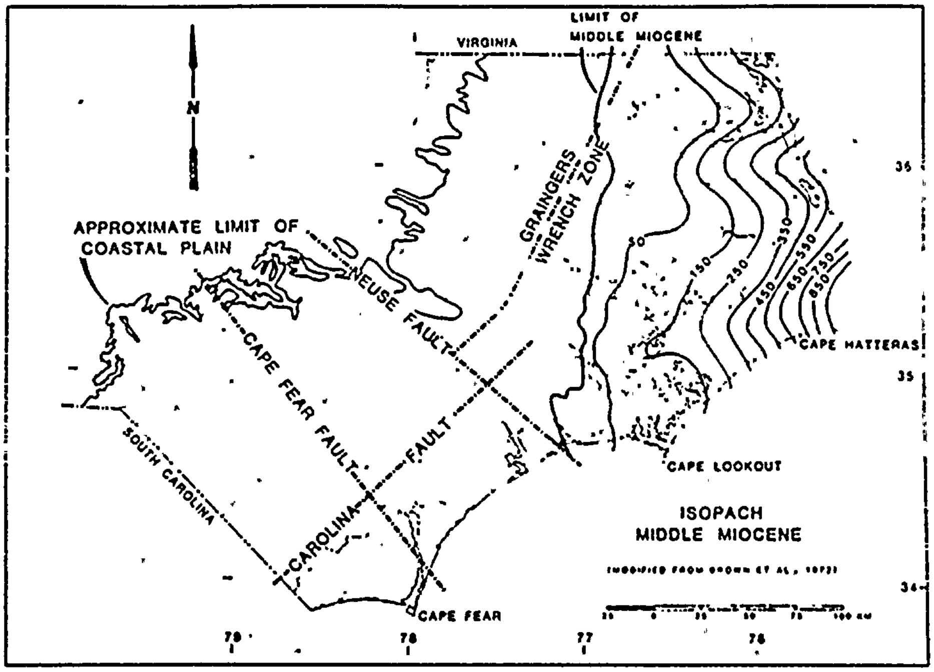


Figure 5. Distribution and thickness of the middle Miocene Pungo River Formation (modified from Brown et al., 1972).

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Rb-Sr glauconite isochron of the Eocene Castle Hayne Limestone, North Carolina

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ABSTRACT

The 11-m-thick lectostratotype of the Castle Hayne Limestone in New Hanover County, North Carolina, consists of lower phosphate pebble biomicrudite; middle bryozoan biosparrudite; and upper bryozoan-sponge biomicrudite. The relative age of the Castle Hayne Limestone is equivocal. The planktic foraminiferal fauna and part of the molluscan fauna suggest that the entire formation should be correlated with the Gulf Coast Claibornian Stage (middle Eocene), whereas calcareous nannofossils, bryozoans, barnacles, and some mollusks indicate that the upper bryozoan-sponge biomicrudite is a Gulf Coast Jacksonian Stage (upper Eocene) equivalent. Because of problems correlating the Castle Hayne Limestone to equivalent Gulf Coast stages, the lectostratotype was dated by application of the Rb-Sr glauconite isochron.

Five hand-picked glauconite concentrates analyzed for Rb, Sr, and Sr-isotopic composition yielded an isochron age of 34.8 ± 1 m.y. ($\lambda Rb^{87} = 1.42 \times 10^{-11} yr^{-1}$) with an initial $(Sr^{87}/Sr^{86})_0$ ratio of 0.7083 ± 0.0004. The determined initial $(Sr^{87}/Sr^{86})_0$ ratio is in good agreement with previous estimates of the Sr-isotopic composition of sea water during the Eocene. Although the age is younger than the value of 37 m.y. earlier proposed for the Eocene/Oligocene boundary, it agrees with fission-track and K-Ar ages of tektites and microtektites, and K-Ar ages of bentonites and glauconites in upper Eocene marine and nonmarine units throughout the world.

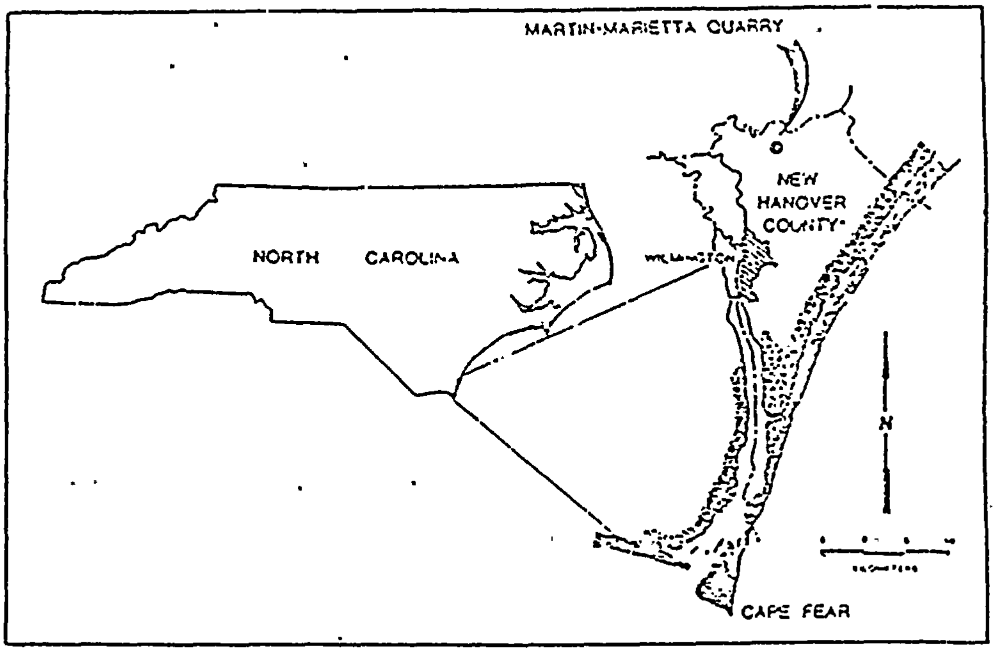

Figure 1. Location of Martin-Marietta quarry, New Hanover County, North Carolina. Sample of Castle Hayne Limestone was collected at this quarry.

INTRODUCTION

Recent work in the United States by Ghosh (1972), Owens and Sohl (1973), Harris and Bottino (1974), Harris (1976), and Harris and Baum (1977), and in Europe by Priem and others (1975), Odin (1978), and Odin and others (1978) has demonstrated that glauconite ages can have direct application to conversion of the standard geologic column to a radiometric time scale. In addition, the accuracy of glauconite ages has also demonstrated that they can aid in the resolution of problems in correlation where faunal data differ.

As a result of these recent successful applications of Rb-Sr and K-Ar dating methods to glauconite, the Eocene Castle Hayne Limestone of the North Carolina Coastal Plain was selected for radiometric and faunal study. The Castle Hayne Limestone has been correlated with the Jacksonian Stage (Clark, 1909; 1912; Canu and Bassler, 1920; Kellum, 1925, 1926; Cheetham, 1961; Copeland, 1964). Brown (1958) and Baum and others (1978) correlated the unit with both the Jackson and Claiborne Stages; however, Brown and others (1972) and Ward and others (1978) correlated the unit with the Claiborne Stage. Therefore, because of problems in correlating the Castle Hayne Limestone with equivalent stages in the Gulf Coastal Plain or in Europe, the lectostratotype was examined for diagnostic fauna and was radiometrically dated by application of the Rb-Sr isochron method to glauconites.

GEOLOGIC SETTING

The Castle Hayne Limestone occurs throughout eastern North Carolina; how-

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ever, the unit crops out only between the Neuse and Cape Fear Rivers. Miller (1912) named the unit for exposures in the vicinity of Castle Hayne, New Hanover County, North Carolina. Because Miller did not designate a type section of the Castle Hayne Limestone, Baum and others (1978) designated the Martin-Marietta quarry, 4.5 km northeast of Castle Hayne, the lectostratotype (Fig. 1).

The Castle Hayne Limestone consists of three units: a lower phosphate pebble biomicrudite, a middle bryozoan biosparrudite, and an upper bryozoan-sponge biomicrudite (Baum and others, 1978). As defined by Baum and others (1978), the Castle Hayne Limestone does not include the overlying Spring Garden Member of Ward and others (1978).The phosphate pebble biomicrudite (New Hanover Member of Ward and others, 1978) forms a discontinuous conglomerate at the base of the Castle Hayne Limestone that does not exceed 1.5 m in thickness. It is present along the outcrop belt and is thickest where it overlies the Rocky Point Member of the Peedee Formation of Late Cretaceous age.

The bryozoan biosparrudite unit disconformably overlies the basal pebble biomicrudite of the Castle Hayne Limestone. It occurs as isolated patches in the vicinity of the Cape Fear fault and thickens to the northeast to a maximum of 12.2 m, where it interfingers with the overlying bryozoan-sponge biomicrudite. Bryozoan-sponge biomicrudite occurs throughout the area between the Cape Fear and Neuse Rivers and is the dominant unit exposed in outcrop. In the area of the Cape Fear fault, it contains numerous diastems and is locally dolomitized (Baum and others, 1978). The bryozoan biosparrudite and bryozoan-sponge biomicrudite lithofacies are the Comfort Member of the Castle Hayne Limestone of Ward and others (1978).

At the lectostratotype, the Castle Hayne Limestone is 11 m thick; it disconformably overlies the Cretaceous Rocky Point Member of the Peedee Formation, and disconformably underlies post-Eocene sand and gravel or Pliocene(?) sediments (Fig. 2). The lower contact of the Castle Hayne is the Cretaceous-Tertiary boundary and is a regional disconformity characterized by solution pits, phosphate, and glauconite. All three units of the Castle Hayne occur at the lectostratotype; however, the bryozoan-sponge biomicrudite forms the dominant part of the section. It consists of loose, unconsolidated carbonate sediment which contains a 1-m-thick dolomitized zone about 1.5 m above the disconformity that separates the bryzoan biosparrudite lithofacies from the overlying bryozoan-sponge biomicrudite lithofacies. The glauconite sample that was used for radiometric dating in this study was collected from a 25-cm-thick glauconite-rich zone immediately below the dolomitized zone in the bryzoan-sponge biomicrudite facies (Fig. 2).

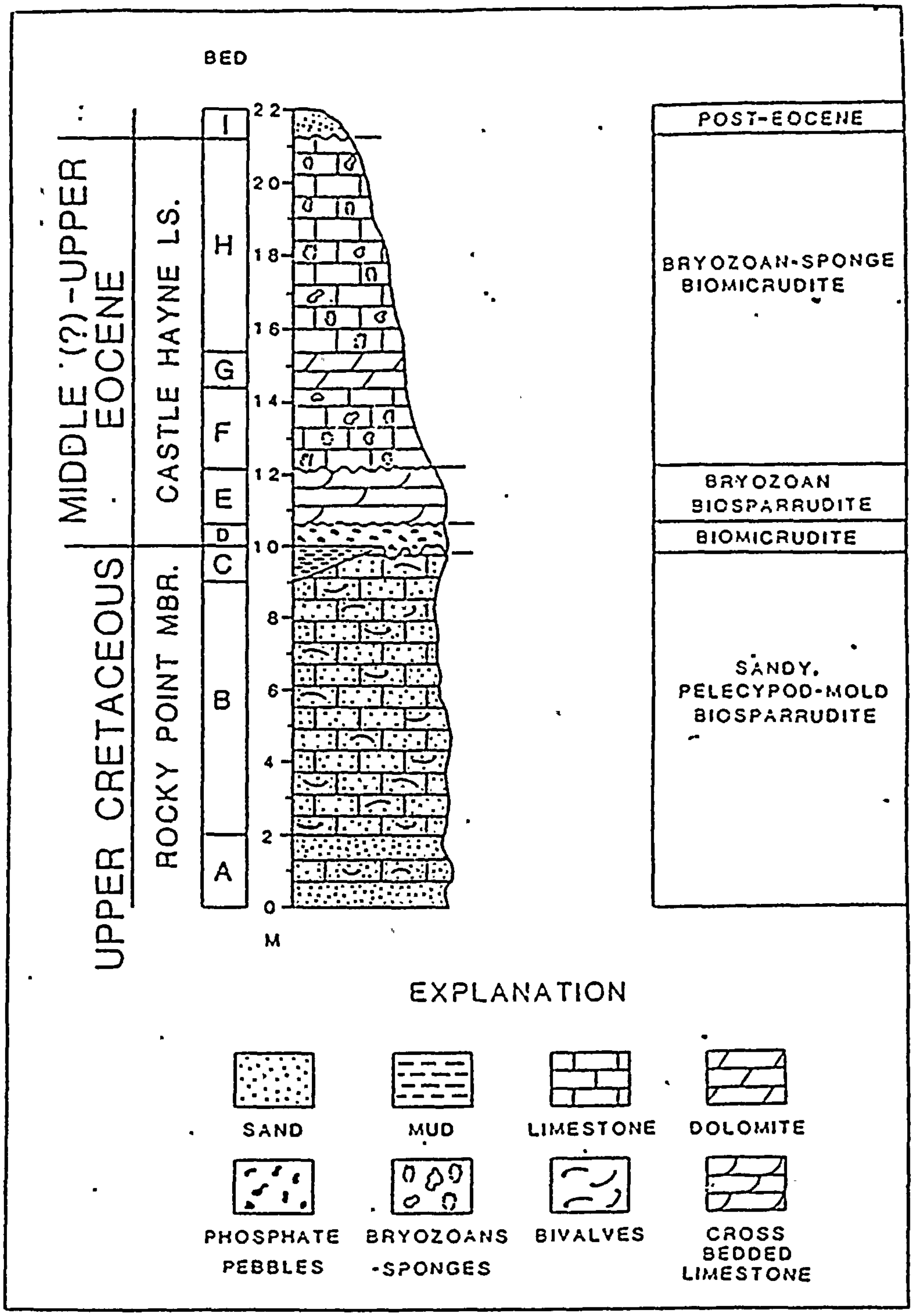

Figure 2. Columnar section of the lectostratotype of the Castle Hayne Limestone. Sample dated in this study was collected from the lower part of the bryozoan-sponge biomicrudite. Bed D is the New Hanover Member.

PALEONTOLOGIC ANALYSES AND RESULTS

The fauna of the Castle Hayne Limestone was considered equivalent to Jacksonian Stage (late Eocene) faunas of the Gulf Coast until the publication of Cooke and MacNeil's (1952) revision of South Carolina

Tertiary stratigraphy. In that paper, Cooke and MacNeil concluded that the lower part of the Castle Hayne Limestone (the basal phosphate pebble biomicrudite and the overlying biosparrudite facies) in the type area was equivalent to the Santee Limestone of South Carolina and the middle Claibornian Stage (middle Eocene) of the Gulf Coast. The upper part of the formation (the bryozoan-sponge biomicrudite) was correlated with newly discovered strata overlying the Santee Limestone in South Carolina. Fossils from these beds were correlated with the fauna of the Gosport Sand that is considered uppermost Claibornian in Alabama. Cooke and MacNeil (1952) cited the following fossils in the Castle Hayne as indicative of Claibornian age: late Claibornian: *Crassatella alta;* middle Claibornian: *Eurhodia rawneli (= E. rugosa), Hemipatagus subrostratus, and Ostrea sellaeformis.*

Species previously considered as Jacksonian indicators were discounted because they were thought to have been misidentified, or were found only at localities far removed from the type area of the Castle Hayne Limestone, or were known to occur as well in Gulf Coast Claibornian units.

LeGrand and Brown (1955) recognized both Claibornian and Jacksonian foraminiferal and ostracod assemblages from presumed Castle Hayne Limestone localities between the Cape Fear and Neuse Rivers. The single Claibornian fauna listed is from the vicinity of Fort Barnwell, Craven County. Microfaunal assemblages described from localities in the type area were considered of Jacksonian age. LeGrand and Brown concluded that the Castle Hayne Limestone was a time-transgressive unit in which deposition began in Claibornian time and lasted through Jacksonian time. Brown (1958), on the basis of ostracod assemblages from wells in the North Carolina Coastal Plain, recognized Claibornian and questionable Jacksonian strata in presumed subsurface equivalents of the Castle Hayne Limestone. In the southeastern counties of North Carolina, in the vicinity of the type area, only Jacksonian(?) strata were encountered. In the central counties, between the New and Neuse Rivers and in the region where the New Bern Formation of Baum and others (1978) overlies the Castle Hayne Limestone, both Jacksonian(?) and Claibornian microfossil assemblages were recognized; in the northeast, only Claibornian strata were encountered.

Brown and others (1972), again primarily on the basis of ostracod zonation, but also utilizing foraminiferal evidence, did not recognize any unit of Jacksonian age in the subsurface in North Carolina. All subsurface sediments associated with the Castle Hayne Limestone or the overlying New Bern Formation were considered Claibornian equivalents. These subsurface data were not related to previously described outcrops of the Castle Hayne Limestone, nor were previous determinations of subsurface Jacksonian microfossil assemblages (for example, Brown, 1958; Copeland, 1964) discussed.

Baum and others (1978) and Zullo and Baum (1979) also considered that most of the Castle Hayne Limestone was Claibornian but suggested that the uppermost unit, the bryozoan-sponge biomicrudite, might extend into the Jacksonian Stage. The overlying New Bern Formation was considered Jacksonian. Ward and others (1978) regarded the Castle Hayne Limestone and the overlying New Bern Formation as Claibornian equivalents. They cited the presence of *Cubitostrea sellaeformis* in the basal phosphate pebble biomicrudite (their New Hanover Member), of *Crassatella alta, Pecten clarkeanus,* and *P. membranosus* in the overlying biosparrudite and biomicrudite lithofacies (their Comfort Member), and of *Crassatella alta, Macrocallista neusensis* (Harris), and *Bathytormus protextus* (Conrad) in the New Bern Formation as evidence of Claibornian age.

Cheetham (1961) argued for a Jacksonian age for the Castle Hayne fauna. From a biostratigraphic analysis of 155 cheilostome bryozoan species described by Canu and Bassler (1920) from the type area of the Castle Hayne Limestone, Cheetham concluded that a late Jacksonian age was indicated. He also suggested that such previously determined Claibornian indicators, such as *Crassatella alta* and *Cubitostrea sellaeformis* were misidentified, as these identifications were based on molds, casts, or juvenile forms. Zullo (1979), in an analysis of the barnacle fauna from the bryozoan biomicrudite facies, concluded that the majority of species, including *Arcoscalpellum jacksonense, Euscalpellum* n. sp., and *Solidobalanus* n. sp. A, were indicative of Jacksonian age. The remaining species were undiagnostic. Studies on calcareous nannofossils from the bryozoan-sponge biomicrudite unit of the lectostratotype by Turco and others (1979) and by Worsley and Turco (1979) indicated that this unit is assignable to zones NP-19 and NP-20, or Jacksonian. Worsley and Turco also noted the presence of zone NP-18 nannofossils from an isolated outcrop near Newton Grove, Sampson County; the NP-18 zone is considered basal Jacksonian (Bybell, 1975).

As noted by both Cheetham (1961) and Brown (1963), and as evidenced by the paleontological discussion, the relative age of the Castle Hayne Limestone is as much disputed now as it has always been. The lack of conformity of opinion is a result of a complex of factors. The Castle Hayne fauna is highly endemic, although it has been suggesid that some so-called endemics may be conspecific with Gulf Coast species (for example, Ward and others, 1978). The value of some species that do appear to afford an opportunity for interregional correlation is lessened because of doubts concerning their identification and stratigraphic range both in the Atlantic and Gulf Coastal Plains, and because of the lack of updated systematic treatments of the genera or species groups to which they are assigned. Another major factor contributing to the dispute is the overwhelming tendency to include the Santee Limestone (in the broadest sense) of South Carolina in any discussion of the age of the Castle Hayne Limestone.

Although depositional environments represented by Paleogene sediments in South Carolina are similar to those in North Carolina, it is not correct to presume that similar sediment types in the two regions are contemporaneous. It has long been recognized that Cretaceous and Tertiary deposition in the Carolinas has been influenced by episodic movement along the Cape Fear fault (for example, Stephenson, 1912; Richards, 1950; Baum and others, 1978). More recently, it has been demonstrated that additional structural elements ("Santee fault," Neuse fault, Graingers wrench zone, Carolina fault) have affected Cretaceous and Cenozoic intrabasinal sedimentation in the Carolinas (Brown and others, 1972; Baum and others, 1978; Harris and others, 1979; Zullo and Harris, 1979). The net result of these discoveries is to emphasize the fact that the stratigraphic column cannot be interpreted merely in terms of eustatic transgressive-regressive cycles on a passive foreland. Rather, it is clear that the effects of eustatic sea-level change were specifically modified by tectonism.

Lithologic similarities between the Castle Hayne and Santee Limestones reflect regional paleogeography. The absence of clastics and the prevalence of calcareous bank deposits suggest a broad, low-lying foreland over which the sea transgressed rapidly, and an adjacent hinterland of low relief whose sluggish streams transported little sediment to the sea. Individual deposi-

TABLE 1. Rb-Sr ANALYTICAL DATA FOR THE EOCENE CASTLE HAYNE LIMESTONE, LECTOSTRATOTYPE, NEW HANOVER COUNTY, NORTH CAROLINA

Sample	Rb (ppm)	Sr (ppm)	Rb^{87}/Sr^{86}	$(Sr^{87}/Sr^{86})_N$
MM[illegible]T	202.08	13.39	43.77	0.7301
MM1[illegible]HM	195.91	26.85	21.14	0.7182
MM1-100HF	199.80	29.66	19.52	0.7188
MM1-70HT	189.78	50.25	10.94	0.7135
MM1-70HF	196.96	19.48	29.31	0.7223

ple 70a, K-feldspar, the one-standard-deviation experimental errors are ± 0.0005 for the Sr^{87}/Sr^{86} and 1.0% for the Rb^{87}/Sr^{86} ratios.

The Sr^{87}/Sr^{86} values in Table 1 have been normalized to Sr^{86}/Sr^{88} = 0.1194. The value obtained from the Massachusetts Institute of Technology standard Eimer and Amend carbonate sample during the period of analyses was $(Sr^{87}/Sr^{86})_N$ = 0.7090. The isochron age was calculated using the recently proposed decay constant of λRb^{87} = $1.42 \times 10^{-11} yr^{-1}$ (Steiger and Jager, 1978).

The Rb-Sr mass spectrometry was performed with a single-focusing, 12-in., triple-filament mass spectrometer. Data were collected and analyzed with a Nuclide DA/CS-III automation and data-reduction computer system.

The [illegible]lts on the five glauconite samples have [illegible]calculated as an isochron age using [illegible]ast-squares regression method of York (1966). The isochron plot for the five glauconite samples indicates an age of 34.8 ± 1 m.y. for the Eocene Castle Hayne Limestone with an initial $(Sr^{87}/Sr^{86})_0$ = 0.7083 ± 0.0004 (Fig. 4).

DISCUSSION AND CONCLUSIONS

Funnell (1964), Berggren (1972), and Hardenbol and Berggren (1978) placed the Eocene-Oligocene boundary between 37.5 and 37 m.y. on the basis of a compilation of various age types. However, the volcanic ages of Evernden and others (1964), the glauconite ages of Ghosh (1972) and of Odin and others (1978), and the microtektite ages of Glass and others (1973) and Glass and Zwart (1977) indicate a much younger age for the boundary, between 33 and 35 m.y. Odin and others (1978) determined glauconite ages of marine sequences in England (type Barton beds) and in Germany and suggested that the age of the Eocene-Oligocene boundary was about 33 m.y. In marine sequences in North America, Glass and others (1973) and Glass and Zwart (1977) considered the Eocene-Oligocene boundary less than 34.2 to 34.6 m.y. on the basis of microtektite ages; this conclusion is supported by the glauconite and bentonite ages of Ghosh (1972) from marine exposures in Mississippi and Alabama. Data from nonmarine sediments in North America and East Africa place the Eocene-Oligocene boundary between 33.9 and 37.5 m.y. (Evernden and others, 1964). In addition, Tarling and Mitchell (1976) used isotopic age determinations of sediments overlying oceanic magnetic anomalies to suggest that the "probable stratigraphic age ..." for the Eocene-Oligocene boundary is close to 35 m.y.

Several conclusions may be drawn from this study. An abundance of published radiometric ages of glauconite, tektites and microtektites, and volcanics indicates that the Eocene-Oligocene boundary is closer to 33 than to 37 m.y.; this age is supported by the 34.8 m.y. isochron age of the Castle Hayne Limestone. Secondly, the glauconite isochron method can provide accurate ages for conversion of the standard geologic column to a radiometric column. Although many Rb-Sr glauconite ages may be young because of the preferential loss of radiogenic Sr relative to Rb^{87} (Thompson and Hower, 1973), the agreement of the Rb-Sr isochron age of the Castle Hayne Limestone with published ages from Europe, Africa, and North America indicates that this is not a problem in this study.

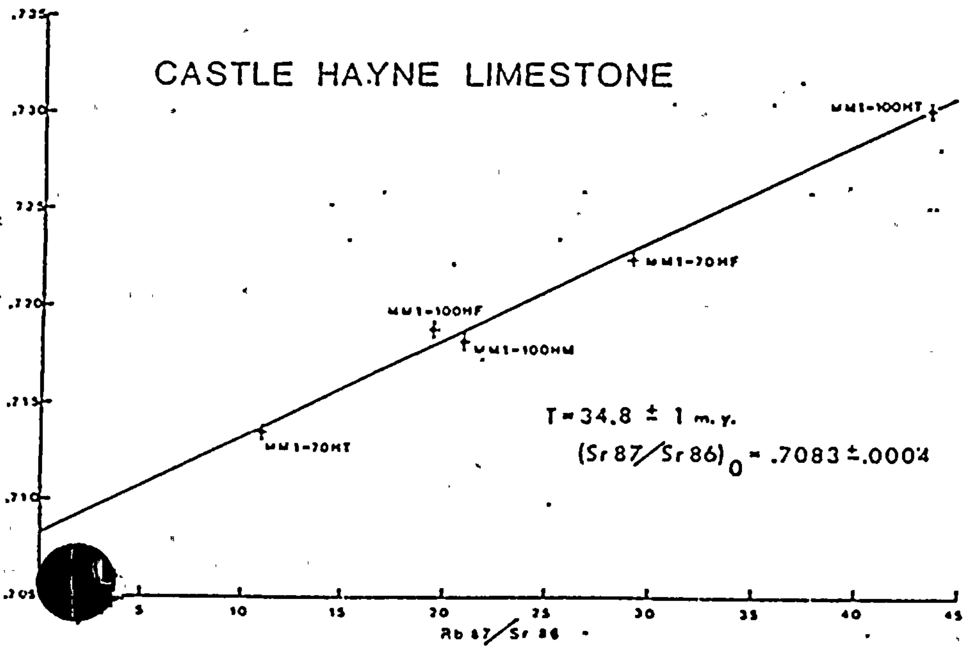

Figure 4. Plot of $(Sr^{87}/Sr^{86})_N$ versus Rb^{87}/Sr^{86} for glauconites from the Castle Hayne Limestone, New Hanover County, North Carolina.

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Rb-Sr glauconite isochron of the Eocene Castle Hayne Limestone, North Carolina: Discussion and reply

Discussion

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Harris and Zullo's (1980) recent paper is an important step in the collection of evidence needed for correct positioning of the Castle Hayne Limestone within the Cenozoic time scale. The authors mention in the abstract that the "planktic foraminiferal fauna . . . suggest(s) that the entire formation should be correlated with the Gulf Coast Claibornian Stage (middle Eocene)." Nowhere in the text do Harris and Zullo cite either the planktic foraminiferal evidence or a reference to such evidence. This discussion is an effort [illegible]nd discuss the planktic foraminiferal evidence based on data [illegible] Ph.D. dissertation (Jones, 1981) on the lower Claibornian rocks of the North Carolina Coastal Plain. In addition, the calcareous nannoplankton ages cited by Worsley and Turco (1979) for the Castle Hayne Limestone as defined by Baum and others (1978) are discussed.

Samples of the bryozoan biosparrudite and bryozoan-sponge biomicrudite facies of the Castle Hayne Limestone (=Comfort Member of Ward and others, 1978) were collected from the lectostratotype of Baum and others (1978) at the Martin Marietta Quarry, Castle Hayne, North Carolina (Figs. 1, 2). In addition, samples of the same two facies were collected from the lectostratotype of the Castle Hayne Limestone of Ward and others (1978) at the Ideal Cement Company Quarry (Figs. 1, 2). All samples from both lectostratotypes yielded diverse populations of planktic foraminifera. Collectively, the species identified include: *Truncorotaloides topilensis* (Cushman, 1925); *T. rohri* Bronnimann and Bermudez, 1953; *Globigerinatheka mexicana mexicana* (Cushman, 1925); *G. mexicana kugleri* (Bolli, Loeblich, and Tappan, 1957); *G. mexicana barri* (Bronnimann, 1952); *Morozovella spinulosa* (Cushman, 1927); *M. lehneri* (Cushman and Jarvis, 1929); *Turborotalia cerroazulensis frontosa* (Subbotina, 1953); *T. cerroazulensis pomeroli* (Toumarkine and Bolli, 1970); *Subbotina linaperta* (Finlay, 1939); *S. eocaena* (Gumbel, 1868) s.l.; *Pseudohastigerina micra* (Cole, 1927); *P. sharkriverensis* Berggren and Olsson, 1967; *Acarinina pentacamerata* (Subbotina, 1947); and *Planorotalites renzi* (Bolli, 1957). The overlapping ranges of these species provide the evidence for placing the entire Castle Hayne section above the phosphate pebble biomicrudite (=New Hanover Member of Ward and others, 1978) at both lectostratotype localities within the upper *Globigerinatheka subconglobata* Zone, P 11, and the *Morozovella lehneri* Zone, P 12 (Stainforth and others, 1975; Hardenbol and Berggren, 1978). In addition, a split of the sample used in Harris and Zullo's Rb-Sr analysis was kindly provided by the authors (Fig. 1). It yielded the same upper P 11 and P 12 zone determination. Furthermore, outcrop and core samples of the bryozoan-sponge biomicrudite and bryozoan biosparrudite facies collected from nine counties in North Carolina have all yielded middle Eocene ages that include the upper P 11 and P 12 zones (Fig. 2). A few samples with low numbers of planktic foraminiferal species yielded age determinations consisting of all or most of the zones in the middle Eocene (P 10 through P 14). Most samples, however, have diverse planktic assemblages and yielded upper P 11 and P 12 zonal determinations, thus correlative with the lower part of the Claibornian Stage of the Gulf Coast which is equivalent to zones P 11, 12, 13, 14 (Huddleston and others, 1974).

As discussed by Harris and Zullo, Worsley and Turco (1979) analyzed lower Tertiary calcareous nannofossils from various locations in the North Carolina Coastal Plain. Rare *Zygolithus dubius* and *Chiasmolithus grandis* from the lower half of the lectostratotype of the Castle Hayne Limestone of Baum and others (1978) suggested correlation with zone NP 20. A nannoflora similar to that from the lectostratotype was identified from Natural Well, Duplin County, and suggested correlation with the upper NP 19 and lower NP 20 zones. In the Evans no. 1 well, Onslow County, Worsley and Turco (1979) reported the consecutive, local extinctions of *C. grandis*, *Discoaster saipanensis*, *Cyclococcolithina formosa*, and *Reticulofenestra umbilica* as suggestive of a continuous sequence of zones NP 19-23, although the presence of *Sphenolithus pseudoradians* suggested the base of the section may be as young as zone NP 20. An outlier of Castle Hayne Limestone in Sampson County yielded *S. pseudoradians* and a form intermediate between *Z. dubius* and *Isthmolithus recurvus*, which suggested correlation with zone NP 18. The local extinctions of these marker species could be due to

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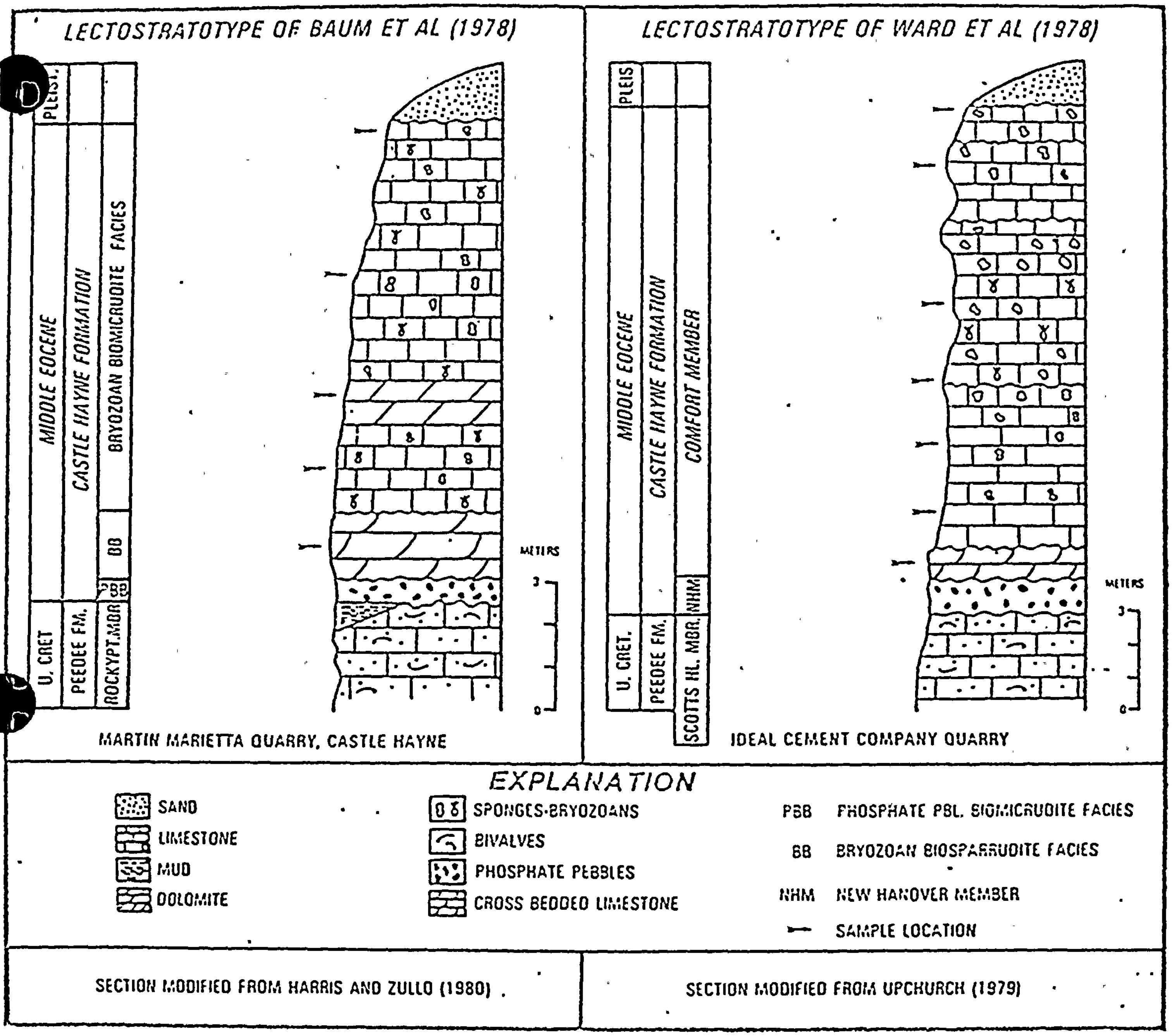

Figure 1. Simplified composite sections of the lectostratotypes of the Castle Hayne Limestone as proposed by Baum and others (1978) and Ward and others (1978). Arrows indicate sample locations of this study.

changing environments or dissolution rather than a result of evolutionary events. Furthermore, all of the calcareous nannofossil species listed above have world-wide stratigraphic ranges that extend down into the middle Eocene (Martini, 1971; Haq, 1978; T. R. Worsley, 1981, personal commun.). Although some of the calcareous nannofossil evidence suggests, to Worsely and Turco, an upper Eocene age for some localities of the Castle Hayne Limestone, their data do not include species whose world-wide stratigraphic ranges begin above the middle Eocene.

A sample of bryozoan–sponge biomicrudite facies (PC-3, Fig. 2) collected by me but not studied by Worsley and Turco (1979) yielded a diverse nannoflora, including *S. radians, Rhabdosphaera gladius,* and *S. furcatolithoides,* suggestive of zone NP 15, middle Eocene (T. R. Worsley, 1981, personal commun.). This facies may not be contiguous with the lectostratotypes in New Hanover County.

As stated above, an upper P 11 and P 12 zonal determination has been obtained for widely spaced samples of the Castle Hayne Limestone from both outcrop and subsurface sections, throughout the North Carolina Coastal Plain. The zonal determination is based partly on species whose world-wide stratigraphic ranges are restricted to the middle Eocene. Planktic foraminifera with stratigraphic ranges beginning above the middle Eocene have not been identified from the Castle Hayne Limestone (Jones, 1981). Thus,

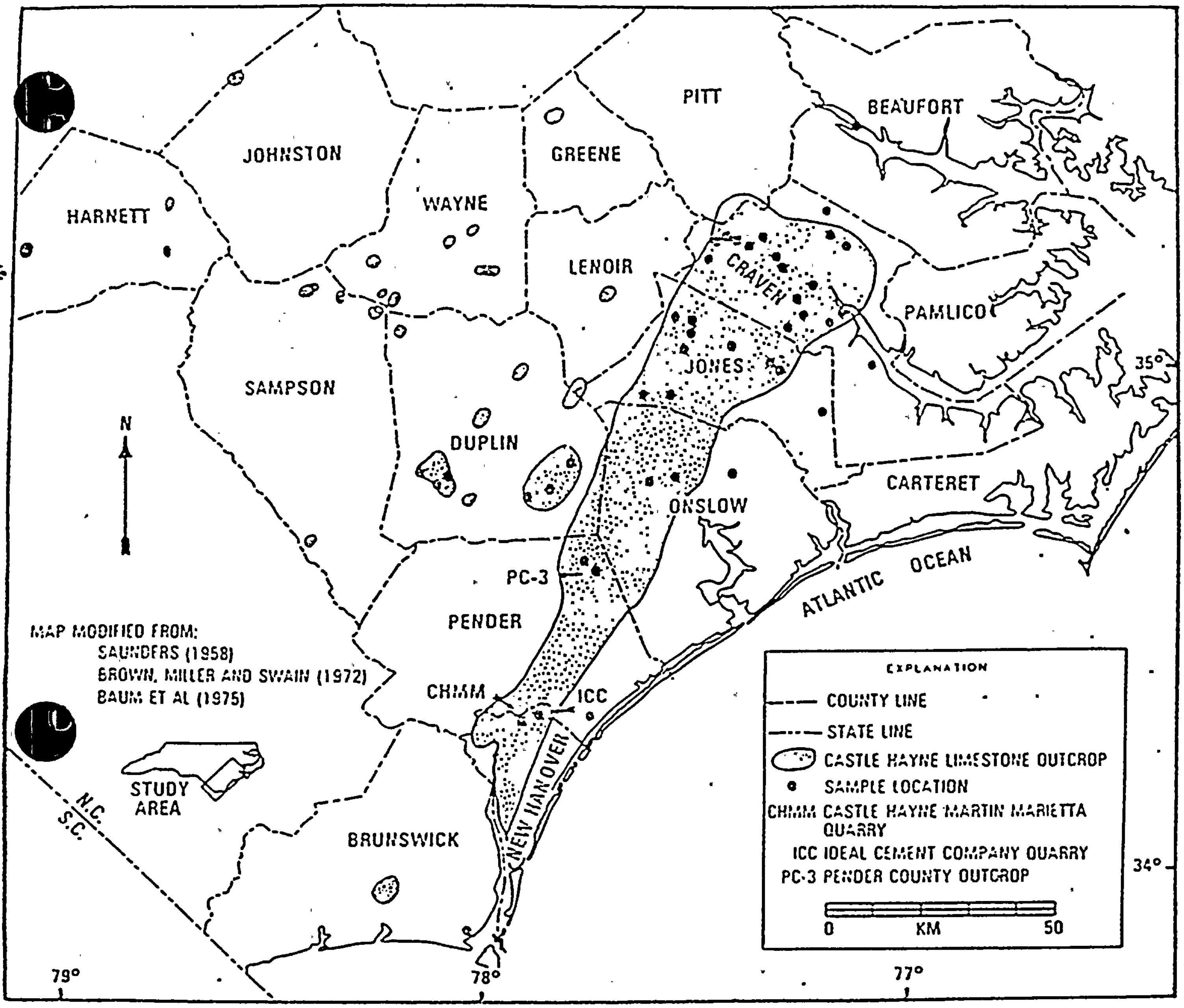

Figure 2. Areal distribution of outcrops and near-surface outcrops of the Castle Hayne Limestone and sample locations of this study.

the planktic foraminifera prove the middle Eocene age of the Castle Hayne Limestone, and the calcareous nannofossil data can be interpreted as being consistent with such an age.

Harris and Zullo's (1980) Rb-Sr age for the Castle Hayne Limestone, therefore, appears to be too young and does not support their conclusion that the glauconite isochron method can provide accurate ages for conversion of the standard geologic column. Rather, their data support Thompson and Hower (1973) who presented evidence indicating that Rb-Sr glauconite ages may be young because of preferential loss of radiogenic Sr relative to Rb^{87}.

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Reply

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We thank Garry D. Jones for his recent discussion of our paper (Harris and Zullo, 1980), and we are pleased that it further emphasizes one of our major statements—from biostratigraphic data, the relative age of the Castle Hayne Limestone is equivocal. The major purpose of our paper was to provide an alternative method for relative correlation of the Castle Hayne Limestone to the standard Gulf Coast Eocene sections, and to determine the feasibility of the Rb-Sr glauconite isochron in solving correlation problems where faunal data are in conflict. We have achieved this purpose as further indicated by the Discussion of Jones.

Jones suggested that we should have fully discussed and developed the planktic foraminiferal evidence which indicates that the Castle Hayne Limestone correlates with the Gulf Coast Claibornian stage. As this information has not been published and was only made aware to the authors through personal communications with Jones, Paul Huddleston, and other workers, it was not our purpose to discuss unpublished biostratigraphic data. Rather, interested readers were made aware of the published foraminiferal evidence through reference to LeGrand and Brown (1955) and Brown and others (1972).

Jones provided a collective list of the planktic foraminifera that he has identified from the Castle Hayne Limestone; however, a list of species which are not figured does not "prove the middle Eocene age of the Castle Hayne Limestone." In addition, a discus[illegible] of "the planktic foraminiferal evidence based on data from a [illegible]. dissertation in progress . . . on the lower Claibornian rocks of the North Carolina Coastal Plain" is a preconceived conclusion made prior to completion of and critical review of the work. Paul Huddlestun (1981, personal commun.) has examined numerous planktic foraminiferal sediments from the Castle Hayne Limestone and has placed the unit approximately in Blow's P13 zone (upper Claibornian). He has further suggested that some of the species or their ranges presented by Jones (see his Discussion above) as indicative of zones P11-P12, are problematical. For example, Huddlestun suggested that *Turborotalia cerroazulensis frontosa* (Subbotina, 1953) and *Globigerinatheka mexicana mexicana* (Cushman, 1925) do not occur together in the standard Gulf Coast Claibornian stratotypes. Rather, *T. cerroazulensis frontosa* occurs in the Tallahatta Formation (P10-P11?), and *G. mexicana mexicana* occurs in the Lisbon Formation (P13= *Cubitostrea sellaeformis* zone). As Huddlestun has never recognized *T. cerroazulensis frontosa* or *Acarinina pentacamerata* (Subbotina, 1947) in the Castle Hayne Limestone, he also questions their occurrence there. He further suggests that the absence in Jones' (1981) species list of *Globorotalia bullbrooki* Bolli, *G. crassata, G. crassula, G. densa, G. rotundimarginata*, and *G. spinuloinflata* (Stainforth and others, 1975) which are common to abundant in middle Eocene deposits, indicates a problem in the planktic foraminiferal data. This further emphasizes the conflicting biostratigraphy of the Castle Hayne Limestone described in our paper.

The argument that Worsley's and Turco's (1979) nannofossil study of the Castle Hayne Limestone does "not include species whose world-wide stratigraphic ranges begin above the middle Eocene" is incorrect. Worsley and Turco (1979) reported 19 selected lower Tertiary nannofossils from the Castle Hayne Limestone. Three of their listed species are world-wide stratigraphic indicators which unequivocally have ranges beginning above the middle Eocene [*Chiasmolithus oamaruensis, Sphenolithus pseudoradians,* and *Helicopontosphaera reticulata,* (T. R. Worsley, 1981, personal commun.)]. Of the seven species discussed by Jones, all but one

(*Sphenolithus pseudoradians*) do have ranges beginning below the upper Eocene; however, when only selected species are discussed and not entire assemblages, any conclusion about age can be drawn. In addition, Jones suggests that because many Eocene nannofossil [illegible] are only locally recognizable, selective dissolution or local environmental conditions have negated the usefulness of nannofossil biostratigraphy in the North Carolina Coastal Plain. It appears that Jones believes that planktic foraminifera are the only reliable biostratigraphic tool, whereas it is possible to advance the same arguments to explain the discrepancies in the planktic foraminiferal evidence. According to T. R. Worlsey (1981, personal commun.), the nannofossil assemblages are preserved consistently in their proper evolutionary sequence, therefore, Jones' argument that the paucity of these marker species is related to local extinctions or selective dissolution is not supported by the data. Calcareous nannofossil data support an upper Eocene age for the Castle Hayne Limestone at the lectostratotype and are not consistent with a middle Eocene age.

Suggesting that the Rb-Sr isochron age of the Castle Hayne Limestone is too young because of preferential loss of radiogenic Sr87 suggests unfamiliarity with the literature, particularly in light of the numerous age determinations on units from other parts of the world which support the age (see discussion in Harris and Zullo, 1980, p. 591). In addition, recent Rb-Sr glauconite isochron ages of Eocene strata from South Carolina support the age of the Castle Hayne Limestone. Fullagar and others (1980) reported ages from the upper Santee Limestone of Baum and others (1980) (= *Cubitostrea sellaeformis* zone) of 36.7 ± 0.6 m.y. and the Cross Formation of 34.1 ± 1.5 m.y. The restricted Santee Limestone of Baum and others (1980) is considered to represent calcareous nannofossil [illegible] NP16 and NP17 (Hazel and others, 1977), and the Cross [illegible]ion of Baum and others (1980) nannofossil zones NP18, NP19, and NP20 (L. M. Bybell, 1978, personal commun.). Therefore, recent Eocene age determinations from other parts of the southeastern Atlantic Coastal Plain support the age of the Castle Hayne Limestone reported by Harris and Zullo (1980) and provide further evidence that the Rb-Sr glauconite isochron method can provide accurate ages for conversion of the standard geologic column.

Numerous lithofacies that transgress time have been included in the Castle Hayne Limestone, each with their own distinctive faunal and floral assemblages. Until detailed lithostratigraphic relationships of all facies assigned to the Castle Hayne Limestone as well as the Eocene have been determined, problems will exist in the biostratigraphic data. If the complexity of Castle Hayne facies is not recognized, then any conclusions about correlation of sediments considered to be equivalent to it at localities other than the lectostratotype, are premature. For example, Jones (1981) apparently does not recognize that the lectostratotype of the Castle Hayne Limestone proposed by Baum and others (1978) contains different lithofacies than the lectostratotype proposed by Ward and others (1978). Also, with no detailed information on the exact locality or horizon in which Jones collected samples for study, collective lists of fauna from different localities are useless.

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AGE OF THE COMFORT MEMBER OF THE CASTLE HAYNE FORMATION (EOCENE) OF NORTH CAROLINA

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ABSTRACT

The biostratigraphic and chronostratigraphic position of the Comfort Member of the Castle Hayne Formation has been the subject of much debate. At the Martin-Marietta quarry at Castle Hayne, New Hanover County, North Carolina, the planktic foraminifers indicate an age no older than the Globorotalia possagnoensis Zone of Toumarkine and Bolli (1970) (which approximates the upper part of the Globogerinatheka subconglobata and Globorotalia lehneri Zones as used by Stainforth and others (1975)) and no younger than the Globorotalia pomeroli Zone (which approximates the Orbulinoides beckmanni Zone). The calcareous nannofossils indicate an age no older than the Coccolithus staurion Subzone of the Nannotetrina quadrata Zone of Bukry (1978) and no younger than Bukry's Discoaster bifax Subzone of the Reticulofenestra umbilica Zone. The dinocyst data indicate placement in the upper part of the Kisselovia coleothrypta Zone of Costa and Downie (1976). All of these zonal units are considered to be within the middle Eocene and, based on the time scale used in the present model, indicate placement in the time interval between 42.1 and 45.3 megaannums (Ma.). The samples in this study bracket the bed from which Harris and Zullo (1980) obtained a Rb/Sr isochron age of 34.8 ± 1.0 Ma. This date is clearly in error and cannot be used to date the Comfort Member.

INTRODUCTION

Miller (1912) named the Castle Hayne Formation for the limestone exposures near Castle Hayne in New Hanover County, North Carolina, but did not designate a type locality. Recently, in nearly simultaneous publications, Ward and others (1978) and Baum and others (1978) sought to correct this deficiency. Ward and others (1978) designated the exposure at the Ideal Cement Co. quarry at Castle Hayne as the lectostratotype and divided the formation into three members: the New Hanover Member, a phosphatic lithocalcirudite; the Comfort Member, a bryozoan-echinoid calcirudite; and the Spring Garden Member, a molluscan-mold biocalcirudite. Baum and others (1978) chose the exposure at the Martin-Marietta quarry near Castle Hayne (hereafter referred to as the Martin-Marietta quarry) as the lectostratotype and divided it into three informal units, a biomicrudite (the New Hanover Member of Ward and others, 1978), a bryozoan biosparrudite, and a bryozoan-sponge biomicrudite. The latter two lithologies constitute the Comfort Member of Ward and others (1978). Baum and others (1978) considered the Spring Garden Member of Ward and others (1978) to be a separate formation, the New Bern Formation. Figures 1 and 2 give the location and section at the Martin-Marietta quarry; the nomenclature follows Ward and others (1978).

The biostratigraphic and chronostratigraphic position of what is now referred to as the Comfort Member of the Castle Hayne Formation has been the subject of much debate. Traditionally, it has been considered correlative with Gulf Coast units assigned to the provincial Jacksonian Stage, which is generally equated with the late Eocene (Clark, 1909; 1912; Canu and Bassler, 1920; Kellum, 1925; 1926; Cheetham, 1961; Copeland, 1964). However, Cooke and MacNeil (1952) and later, Brown and others (1972) concluded that beds now included in the Comfort Member of the Castle Hayne are entirely of Claibornian age, which is generally equated with the middle Eocene. Recently, Baum and others (1978) and Zullo and Baum (1979) suggested that the uppermost bryozoan-sponge biomicrudite of the Comfort Member may extend into the onian Stage. Ward and others (1978) considered the entire Castle Hayne Formation, ding their uppermost Spring Garden Member, to be Claibornian. Turco and others) and Worsley and Turco (1979) presented evidence from calcareous nannofossils to ort an age determination of late Eocene.

A new dimension was added to the controversy by Harris and Zullo (1980). They ained glauconite samples from the Comfort Member of the Castle Hayne at the tin-Marietta quarry, New Hanover County, N.C. and dated these using the Rb-Sr chron technique. An isochron of 34.8 $\pm$1.0 Ma (megaannums) was obtained. In view of conflicting paleontological results, they used this isochron to suggest a late Eocene e for the Comfort Member and also advance the hypothesis that the Eocene-Oligocene undary is close to 33 Ma.

In a discussion of the Harris and Zullo (1980) paper, Jones (1982) stated that the omfort Member contains a planktic foraminiferal assemblage inconsistent with a late ocene age. Because Harris and Zullo (1980) themselves stated that many Rb-Sr glauconite ages may be too young because of the preferential loss of radiogenic Sr relative to 87Rb, Jones concluded this was the more likely alternative. He also stated that the nannofossil data of Worsley and Turco (1979) (could be consistent) with a middle Eocene age. necessary parentheses?

The present study is based on the examination of microfossils (planktic foraminifers, calcareous nannofossils, and dinoflagellates) from the Comfort Member of the Castle Hayne at the Martin-Marietta quarry, both above and below the horizon of Harris and Zullo's (1980) glauconite material (Fig. 2). The purpose of this paper is to determine the age and correlation of the Comfort Member at this locality. The data show that the Comfort there is of middle Eocene age; therefore the Rb-Sr isochron age of 34.8 Ma (Harris and Zullo, 1980) is about 10 Ma too young and has no direct bearing on providing an age estimate for the Eocene-Oligocene boundary.

Planktic Foraminifera

Table 1 shows the distribution of planktic foraminfers in five samples from the Comfort Member at the Martin-Marietta quarry. The lowest foraminiferal sample (CHMM-3 at 5.5 m) contains 14 planktic species. The presence of *Globigerinatheka mexicana mexicana* (Cushman, 1925) indicates a chronostratigraphic placement no older than the *Globorotalia possagnoensis* Zone of Toumarkine and Bolli (1970) and the presence of *Globorotalia frontosa* (Subbotina, 1953), which defines the top of the *G. possagnoensis* Zone, places the sample in that zone.

Sample CHMM-Glau. was taken from the bed that Harris and Zullo (1980) obtained the glauconite for their analysis. The presence of *Planorotalites renzi* (Bolli, 1957), *Truncorotalites topilensis* (Cushman, 1925), *T. rohri* Bronnimann and Bermurdez, 1953, and *Morozovella spinulosa coronata* (Blow, 1979) indicates a middle Eocene age. According to Blow (1979, p. 1017) *M. spinulosta coronata* does not occur as high as the youngest middle Eocene. The assemblage at 11 m (CHMM-1) is virtually the same; the presence of ~~*Acarinina*~~ *Testacarinata* *inconspicua* (Howe, 1939) is further evidence of a middle Eocene age. The highest foraminiferal sample, CHMM-2 at 15.2 m contains virtually the same assemblage as CHMM-1.

The foraminiferal data indicate that the Comfort Member at this quarry is of middle Eocene age. It is no older than the *Globorotalia possagnoensis* Zone of Toumarkine and Bolli (1970), which approximates the upper *Globigerinatheka subconglobata* and *Globorotalia lehneri* Zones as used by Stainforth and others (1975). The lower sample examined represents the *Globorotalia possagnoensis* Zone, the remainder of the Comfort could represent this zone or the younger *Globorotalia pomeroli* Zone. The presence of *Morozovella spinulosa coronata* precludes an age assignment younger than the *Globorotalia pomeroli* Zone, which approximates the *Orbulinoides beckmanni* Zone as used by Stainforth and others (1975) (see Toumarkine and Bolli, 1970; Blow, 1979).

Calcareous Nannofossils

Table 2 is a list of the calcareous nannofossils found in the Comfort Member. The lower sample contains Chiasmolithus solitus (Bramlette and Sullivan, 1961) Locker, 1968, whose last appearance defines the boundary between the Discoaster saipanensis and Discoaster bifax Subzones of the Reticulofenestra umbilica Zone of Bukry (1978). Also present is Campylosphaera dela (Bramlette and Sullivan, 1961) Hay and Mohler (1967), which last appears shortly before the last Chiasmolithus solitus. Dictyococcites scrippsae Bukry and Percival, 1971, also occurs in this sample; this species has its first appearance datum at or very close to the last appearance datum of Chiasmolithus gigas (Bramlette and Sullivan, 1961) Hay, Mohler, Wode, 1966, which defines the top of the Chiasmolithus gigas Subzone of Bukry (1978). Thus, the lower Comfort represents the middle part of the middle Eocene and the Coccolithus staurion Subzone of the Nannotetrina quadrata Zone and/or the Discoaster bifax Subzone of the Reticulofenestra umbilica Zone of Bukry (1978). Bukry (1978, p. 56) considers these two subzones as correlative with the upper part of the Nannotetrina alata Zone (NP15) through Discoaster tani nodifer Zone (NP16) of Martini (1971).

Calcareous nannofossils are less diverse and more poorly preserved in the upper part of the Comfort. However, the presence of Campylosphaera dela in R2204E, indicates an age no younger than middle Eocene, probably no younger than the Discoaster bifax Subzone of Bukry (1978).

The nannoflora in the Comfort at the study site provides evidence that the unit is of middle Eocene age and can be assigned to the chronozone of the Coccolithus staurion and/or Discoaster bifax Subzones of the middle part of the middle Eocene.

Dinoflagellates

Table 3 lists the occurrence of dinoflagellate cysts in the Comfort Member at the study cality. The flora indicates biostratigraphic placement in the upper (middle Eocene) ortion of Costa and Downie's (1976) Kisselovia coleothrypta Zone. The first appearance of Rhombodinium draco Gocht, 1955, defines the base of Costa and Downie's overlying R. draco Zone; this species is notably absent in the Comfort.

On the basis of the dinocyst evidence, the Comfort correlates with the upper Bracklesham in England. The presence of Areosphaeridium dictyostilum (Menendez, 1965) Sarjeant, 1981, senior synonym of A. arcuatum Eaton, 1971, indicates that the Comfort is no older than the upper Bracklesham B-4 assemblage of Eaton (1976) and Bujak and others (1980). The Comfort does not appear to be as young as the basal Barton, in which the base of the R. draco is found. The joint occurrence of Pentadinium goniferum Edwards, 1982 and Pentadinium polypodum Edwards, 1982 suggest correlation with the upper part of the Lisbon Formation or the Gosport Sand of the Gulf Coast. (It is interesting to note that the coleothrypta/draco zone boundary occurs in the upper part of the Gosport Formation of Alabama.) The dinoflora compares favorably with the flora cited by Chateauneuf (1980) from the upper Lutetian/lower Auversian of the Paris Basin.

Biostratigraphic Conclusion

The combined microfossil data lead to the conclusion that the Comfort Member of the Castle Hayne Formation at the Martin-Marietta quarry, at a maximum, is no older than the Coccolithus staurion Subzone of the Nannotetrina quadrata Zone or the upper Globorotalia possagnoensis Zone. Further, the unit in not younger than the Discoaster bifax Subzone of the Reticulofenestra umbilica Zone or the Globorotalia pomeroli Zone.

Figure 3 (modified from Hazel and others, in press) is a chart showing the correlation of Midwayan, Sabinian, and Claibornian strata from three areas in the Coastal Province. The chart is based on a biostratigraphic, magnetostratigraphic, and chronometric model developed by us in which first appearance and last appearance datums for calcareous nannofossils, planktic foraminifers, and dinoflagellates are calibrated to each other, to Paleogene magnetic anomalies, and to time. Insufficient data are available at the present time to present an entire column for southeastern North Carolina; however, the maximum biostratigraphic and magnetostratigraphic position of the Comfort Member is indicated by the shaded band.

The model indicates that the Coccolithus staurion Subzone to Discoaster bifax Subzone interval correlates with the reversed interval between magnetic anomalies 18 and 19 to the lower part of anomaly 20. Ness and others (1980) give an age for the base of anomaly 18 of 41.82 Ma and an age of 45.18 for the base of anomaly 20. On the time scale developed for the USGS model this interval is between 42.1 and 45.3 Ma.

The combined dinoflagellate (tab. 3) and molluscan assemblages (which includes Crassatella texanus Heilprin, 1890 Pholadomya claibornensis Aldrich, 1886 and Pecten membranosus Morton 1834) of the Comfort at the Martin-Marietta quarry strongly suggest correlation of the Comfort with the uppermost part of the Lisbon Formation of Alabama. This in turn suggests that the Comfort falls in the chronozone of the lower part of the Discoaster bifax Subzone and the lower part of the Globorotalia pomeroli Zone. This level is calibrated to time at about 43.0 Ma.

The 34.8 ±1.0 Ma isochron age for the Comfort Member of the Castle Hayne Formation (Harris and Zullo, 1980) is clearly in error. It has no bearing even on the age of the middle Eocene-upper Eocene boundary (39.6 to 40.4 Ma depending on paleontologic definition), much less the Eocene-Oligocene boundary.

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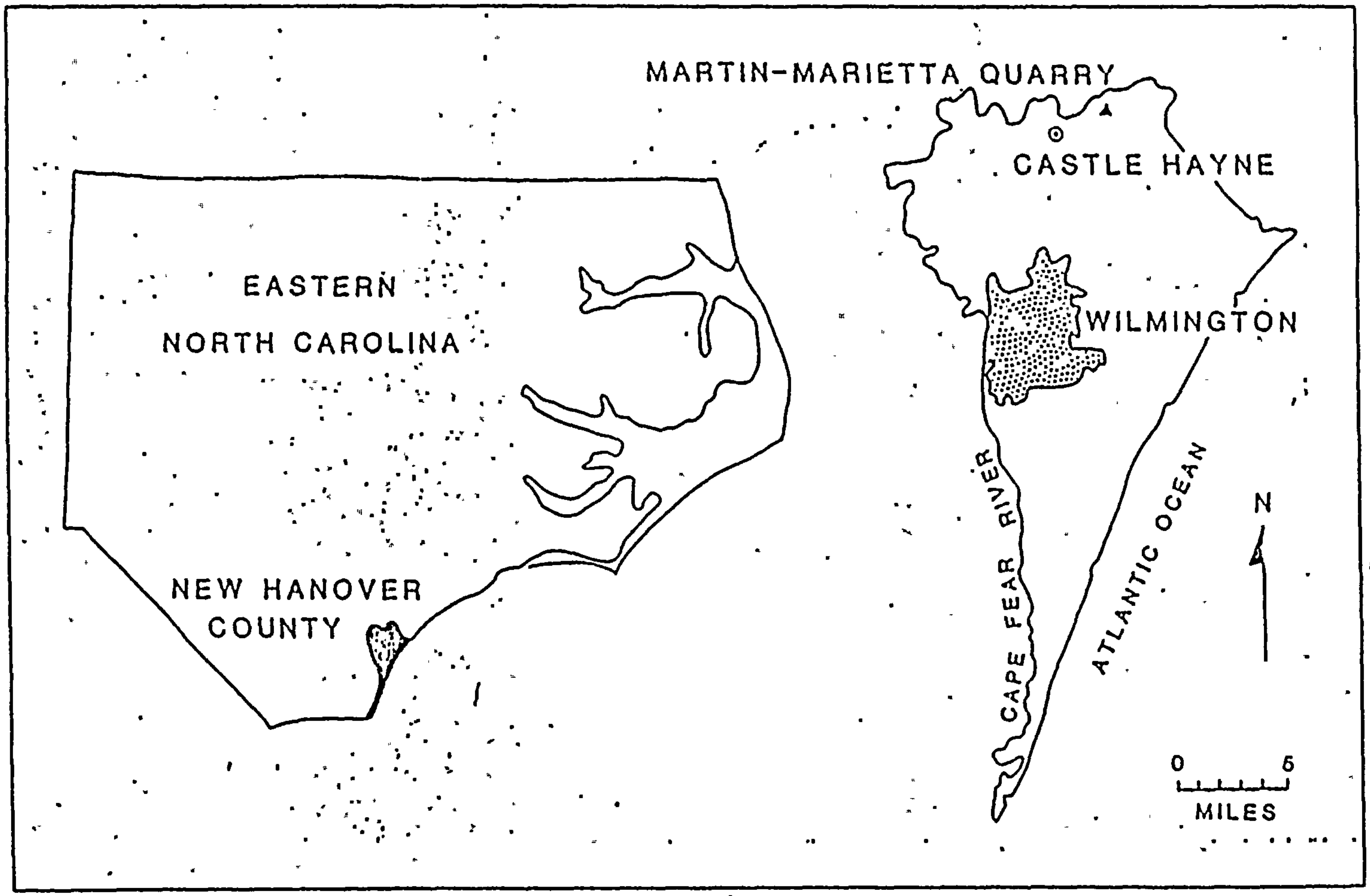
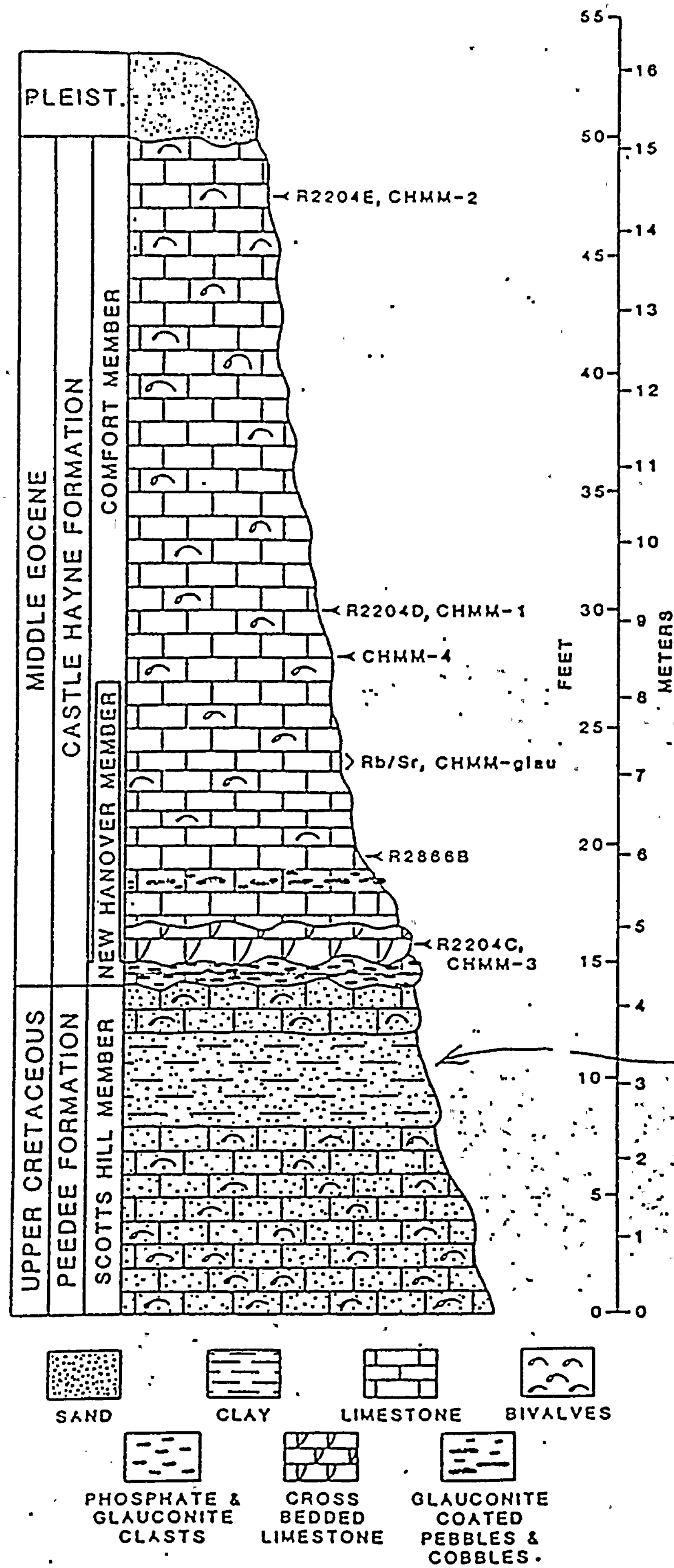
Figure 1. Locality map of eastern North Carolina showing the location of the Martin-Marietta quarry near Castle Hayne, New Hanover, County.



MARTIN-MARIETTA QUARRY
CASTLE HAYNE
EASTERN
NORTH CAROLINA
WILMINGTON
CAPE FEAR RIVER
ATLANTIC OCEAN
N
NEW HANOVER
COUNTY
0 5
MILES

New Hanover County.

gure 2. Section at the Martin-Marietta quarry, New Hanover County. Lithostratigraphic nomenclature is after Ward and others (1978). Location of samples used in this study is indicated by arrows. R coded samples were examined for calcareous nannofossils and dinoflagellates. CHMM samples were examined for foraminifers.



Might want to show this clay pinching-out as it is not found everywhere in the quarry

LEE JAMES OTTE. Petrology of the exposed Eocene Castle Hayne Limestone of North Carolina (under the direction of DANIEL A. TEXTORIS)

ABSTRACT

The Eocene Castle Hayne Limestone crops out on the southern half of the North Carolina Coastal Plain. The major outcrop belt, marking the updip limit of continuous rock, is 160 km long and up to 24 km wide, striking southwestward through southern Pitt, western and central Craven, western Jones, central Onslow, central Pender, northwestern New Hanover, and possibly southern Brunswick Counties. Outliers, erosional remnants preserved in preCastle Hayne stream valleys, are found in Duplin, Lenoir, Sampson, and Wayne Counties. Outliers of molluscan-rich, clastic-dominated sediments in Harnett, Hoke, Johnston, Moore, and Wake Counties are tentatively dated as Eocene. If these sediments correlate with the Castle Hayne Limestone, the marine transgression that initiated deposition of this formation extended to the eastern edge of the Piedmont Province of the Appalachian Highlands, a full 200 km west of the present coast.

The Castle Hayne fauna contains a mixed Middle and Upper Eocene assemblage. Correlation with regional tectonics and depositional history of the North Atlantic suggests a Middle Eocene age for the Castle Hayne Limestone.

Sixty-three outcrops of the Castle Hayne have been reported in the literature during the past 145 years. These are divided into five major lithofacies: (1) a shallow-water, low-energy, phosphate-pebble

biomicrudite; (2) a shallow-water, high-energy, bryozoan biosparrudite; (3) an intermediate-depth, low-energy, sandy, sponge-spicule-bearing biomicrite; (4) an intermediate-depth, low-energy, sandy, foraminiferal, echinoderm biomicrite; and (5) a deep water, low-energy, bryozoan biomicrudite. Three minor, locally occurring lithofacies are also recognized: (1) a dolomitized biomicrite; (2) a molluscan-mold, bryozoan biomicrudite; and (3) a bryozoan, foraminiferal biomicrite.

Major diagenesis of the Castle Hayne Limestone occurred in four environments: (1) shallow marine, represented by authigenic glauconite; (2) mixed shallow marine and fresh-water vadose, represented by phosphate precipitation and development of calcitic drusy rim cement, in association with diastemic surfaces; (3) fresh-water vadose, represented by dissolution of aragonite and leaching of Mg^{++} from high-Mg calcite bioclasts; and (4) fresh-water phreatic, represented by development of syntaxial overgrowths on echinoderm fragments and recrystallization of micrite. Additional diagenetic alterations include scattered occurrences of silica in the form of chalcedony, iron oxides, and iron sulfides. A local zone of dolomitized biomicrite in New Hanover County is attributed to the Dorag model of diagenesis.

Distribution of sponges, foraminifera, and bryozoans, plus the lack of benthonic algae, suggests water depths of more than 100 m during deposition of the bryozoan biomicrudite facies and 30 to 45 m for the sandy, sponge spicule-bearing biomicrite facies. The sandy, foraminiferal, echinoderm biomicrite facies, with its high micrite content and occasional zones of winnowed sediment, suggests deposition in a water environment that experienced occasional periods of high energy. Reverse cross-bedding and bimodal grain alignment in the bryozoan biosparrudite facies indicate tidal influence, and deposition above wave base. The above four facies were deposited in a transgressive episode. The basal phosphate pebble biomicrudite facies, separated from the other lithofacies by a prominent erosion surface that developed during a significant marine regression, was deposited during an earlier transgression. Thick sections of Castle Hayne-like sediments, found beneath typical Castle Hayne Limestone in numerous outliers in Duplin, Sampson, and Wayne Counties, may also belong to this earlier episode.

Except for the bryozoan biosparrudite facies, typical shallow water to supratidal carbonates are lacking in the Castle Hayne Limestone. The environments in which these sediments formed were possibly situated on the Cape Fear Arch, a structural high extending along the southwestern edge of the Castle Hayne Embayment. The lack of terrigenous detritus in outcrops closest to the Arch suggests that during Castle Hayne time the Arch was under water and perhaps covered with a veneer of carbonate sediments.

The overall distributional pattern of sediment types and faunal components suggests deposition in a coastal embayment, open to the ocean on the southeast side, but surrounded by either land or very shallow water on the other three sides.

Paleogeography, Paleocirculation, and Paleoclimate

The North American continent has drifted westward since its separation from Africa and Europe (Smith and Briden, 1977; Fig. 7) and has wandered both northward and southward during that time. Sixty million years ago the North Carolina coast was situated at about 33°N, but by the beginning of the late Eocene (40 million years ago) it had shifted to about 29°N. By 20 million years ago, however, the North Carolina coast had reverted to about 32°N. The portion of the North Carolina Coastal Plain on which the Castle Hayne Limestone is now located lies between 34°N and 35°N. During time of deposition the Eocene coastline was positioned at approximately 30°N, the latitude in which north Florida is presently located.

During the Late Cretaceous, a widespread marine regression drastically reduced the size of the North Atlantic epicontinental seas and altered the major circulation pattern in the North Atlantic (Berggren, 1978). Within the Gulf of Mexico, a major wind-driven current with a clockwise circulation pattern similar to the current Gulf Stream circulation developed. The outflowing water from this current contributed to the generation of the Gulf Stream and the North Atlantic Drift (Berggren, 1978). This surface circulation of warmer waters from low latitudes to high latitudes has continued since that time. Consequently, during Eocene time, a major northward-flowing current somewhere off the North Carolina coast brought warm water and possibly some tropical faunal elements into the area.

Through the Cenozoic Era, from about 63 million years ago up to present, the earth's surface has changed from largely tropical

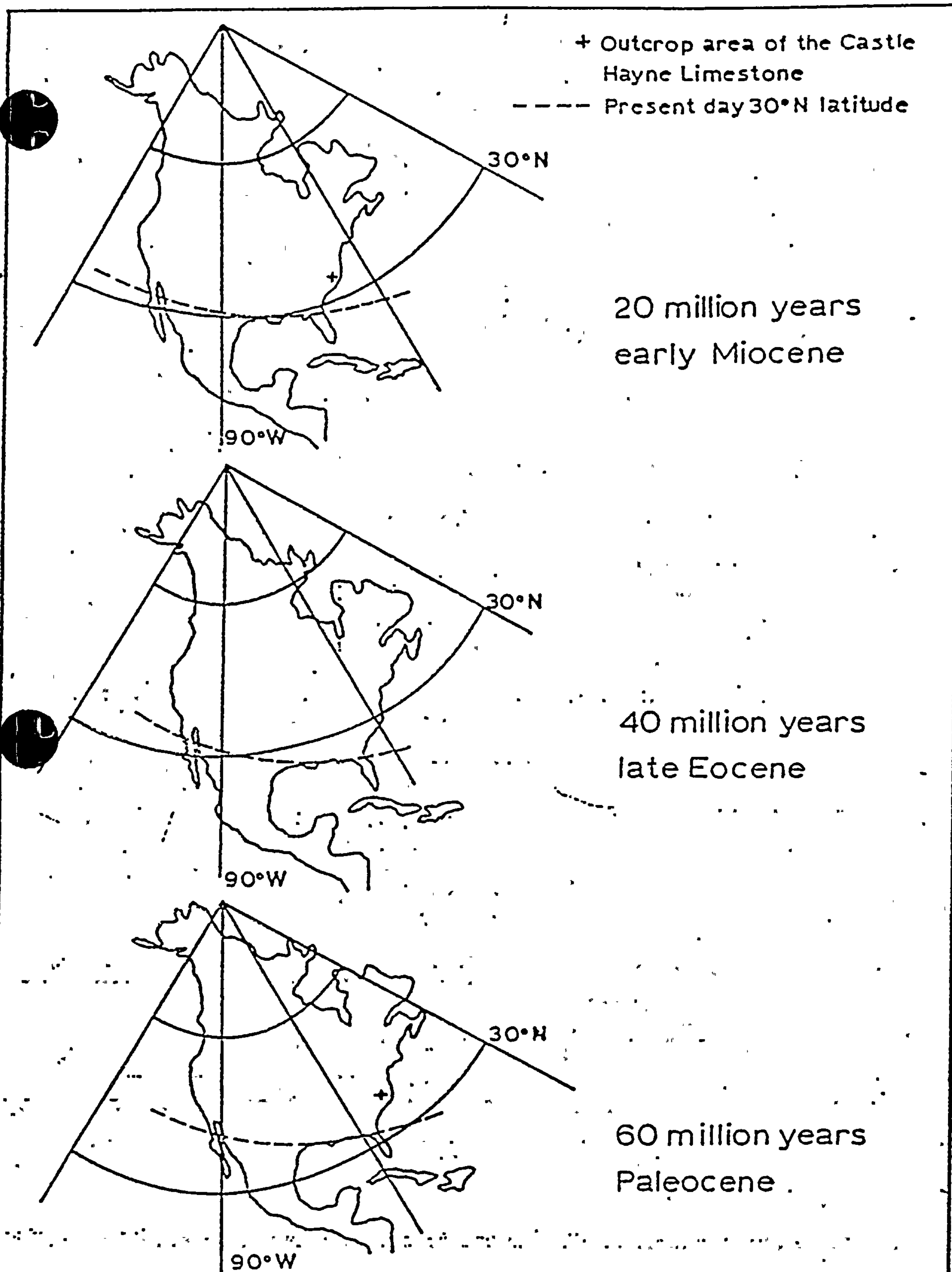


Figure 7. Change in paleolatitude and paleolongitude of North America during the Cenozoic Era, based on Smith and Briden (1977).

and subtropical to temperate (Berggren, 1978), though this trend has not been uniform. Major climatic changes progressed at different rates in different parts of the world. For instance, in the North Atlantic, a warm trend is recorded for the Paleocene and the Lower Eocene, with climatic zones expanding poleward (Berggren, 1978). Two events took place during the Eocene, however, that drastically reversed this trend for the Atlantic Ocean. During the late Early Eocene (about 50 mya) the final opening of the northern perimeter of the Atlantic Ocean occurred when Greenland and Europe separated (Berggren, 1978). The development of a deep ocean basin in the northern latitudes allowed colder Arctic bottom waters to flow into the North Atlantic and sink, thereby generating a deep cold-water circulation pattern. Cool surface waters also began to push down from the Arctic. This influx of cooler water started a slow but steady cooling of the North Atlantic, aided in the gradual development of a boreal faunal realm, and increased the differentiation of temperate and tropical realms.

By the middle Late Eocene (about 40 mya) Antarctica, which had earlier split off from South America, moved into the high southern latitudes over the south pole and caused fundamental changes in oceanic circulation and world climate (Berggren, 1978). Direct sedimentological evidence exists for Antarctic glaciation already having occurred in the Middle Eocene, though it was closer to the Eocene-Oligocene border (about 38 mya) that the glaciers became [illegible] enough to cover large portions of the continent & to develop deep sea masses (Berggren, 1978). This glaciation provided a mechanism to cool southern waters and also a method to lower global sea level during the Late Eocene.

Late in the Cretaceous, sea level reached a maximum of more than 300 m higher than present-day sea level (Vail and Hardenbol, 1979). The Tertiary was marked by a great number of fluctuations in sea level, although the overall trend had been a gradual lowering of sea level since the Cretaceous high. The regressions appear to have been very rapid, having occurred over a period of a million years or so, and many of them seem to have been a magnitude of more than 100 m (Vail and Hardenbol, 1979). During the Eocene two major sea level falls are recorded, one at about 49.5 million years ago and one at 40 million years ago (Vail and Hardenbol, 1979), in rough correlation with the opening of the North Atlantic and the beginning of major Antarctic glaciation, respectively. Paleontologic evidence supports a major drop in sea level of up to 300 to 400 m from 50 to 49 million years ago (the Upper Cretaceous-Middle Eocene break on the North Carolina Coastal Plain) (Vail and Hardenbol, 1979). Pitman (1979) reported a major Paleocene to Middle Eocene regression from 65 to 45 million years ago that moved the western Atlantic shoreline rapidly seaward. This event could explain the lack of Paleocene and Lower Eocene sediments from large portions of the Atlantic Coastal Plain.

A major transgression in the Middle Eocene, during which the Eocene seas reached their maximum extent (Hallam, 1963), is found on most continental shelves. One of the few places where a regression is seen at this time is on the Gulf Coast of North America (Hallam, 1963). The transgression is well documented for the western Atlantic Ocean and the Atlantic Coastal Plain, and is the result of a combination of factors. During the Middle Eocene the rate of sea level fall decreased well below the Cenozoic average for the western Atlantic, to about 0.37 cm/1000 years (Rona, 1973). The post-Cretaceous subsidence rate at the shelf edge east of Cape Hatteras, North Carolina, has averaged about 2.5 cm/1000 years (Rona, 1973). When rate of subsidence exceeded the rate of sea level fall, a transgression occurred (Rona, 1973; Pitman, 1979). Based on microfossil assemblages of Eocene sediments in New Jersey, Olsson (1978) also recognized a major Eocene trasgression in which increase in water depth on the shelf was significant.

The relative rise in sea level during the Middle Eocene, accompanied by a cessation of terrigenous sediment influx from the continent (Gibson, 1970), set the stage for the development of carbonate units along the Atlantic Coastal Plain. Sheridan et al. (1978) found so little terrigenous sediment reaching the Atlantic shelf during Eocene time that subsidence exceeded accumulation and waters on the outer shelf deepened to bathyal depths. Deep-sea Eocene sediments in the western North Atlantic consist of calcareous and siliceous oozes with almost no detrital component (Gibson, 1970). The shallow-water Eocene strata on the Coastal Plains of Virginia, Delaware, and New Jersey primarily consist of glauconitic units several hundred feet thick (Gibson, 1970). Eocene strata in North Carolina and to the south consist mostly of carbonate sediments.

Sea level fell once more at the end of the Middle Eocene, then rose again during Late Eocene (Vail and Hardenbol, 1979), though the Late Eocene transgression was not nearly as extensive as the one during the Middle Eocene (Hallam, 1963). An additional large-scale fall in sea level at the Eocene-Oligocene boundary (Vail and Hardenbol,

1979) caused the erosion of a substantial proportion of the Eocene sediments.

Middle Eocene time apparently represents a period of major transgression on the Atlantic Coastal Plain, accompanied by minimal influx of terrigenous sediments from the Appalachian Highlands. The North Atlantic Ocean was gradually cooling because of inflow of surface and deep, cold-water currents, yet, like today, a Gulf Stream-type current flowed northward, possibly at or near the edge of the continental shelf, providing an insulating effect on water temperatures.

Age Assignment

Numerous geologists have dated the Castle Hayne Limestone as Middle and/or Upper Eocene. Recently, Brown et al. (1972) recognized no Upper Eocene (Jackson) sediment in North Carolina, and, except for a small amount of Lower Eocene beneath the outer counties in northeastern North Carolina, concluded that all the remaining Eocene is Claibornian (Middle). Baum et al. (1978), however, named the Eocene New Bern Formation and placed it in the Upper Eocene, with this new formation resting disconformably on their Middle Eocene Castle Hayne. They cited the occurrence of several typical Upper Eocene fossils in the New Bern Formation and used this to defend the Jackson age. However, they also stated that typical Claibornian fossils were found in the New Bern Formation and, likewise, that typical Jacksonian fossils were found in their Claibornian Castle Hayne Limestone. Ward et al. (1978), on the other hand, call the New Bern Formation of Baum et al. the Spring Garden Member of the Castle Hayne Formation, making it the uppermost facies of their Middle Eocene Castle Hayne Limestone.

Upchurch (1973). Textoris presented a general survey of the Castle Hayne Limestone, while Cunliffe and Upchurch concentrated their works on individual quarries, Cunliffe the Martin Marietta quarry and Upchurch the Ideal Cement quarry, both near the town of Castle Hayne in New Hanover County. Through these investigations an unexpected variety of microfacies was found within the Castle Hayne Limestone. The recognition of these and the correlation with modern carbonate environments provided a very important beginning in deciphering the Castle Hayne. Baum (1977) and Baum et al. (1978) continued this study on a larger, though more generalized scale throughout the major outcrop belt of the Castle Hayne Limestone.

Structure

Brown et al. (1972) recognized three levels of tectonic control along the Atlantic Coast of North America that help form the geologic features seen currently on the North Carolina Coastal Plain. Their research " . . . indicates that the coastal margin is a margin where the principal mobility takes the form of block faulting or flexing, accompanied by a rotational realignment of the axes of positive and negative structures in the region" (Brown et al. 1972).

Phase one of Brown et al.'s deformation is aligned northeast-southwest and is composed of parallel series of positive structural features (fault-block anticlines) and of adjacent negative structural features (half-grabens). This phase is associated with the major structural alignment of the Appalachian Highlands and appears to be the controlling factor in the overall positioning of the Atlantic

Continental Shelf. The axes of their phase two deformations are aligned northwest-southeast and are composed of parallel positive structural features (compressional anticlines) and of adjacent negative features (compressional synclines). In phase three positive features (compressional anticlines) and adjacent negative features (grabens) diverge with the axes of the positive features aligned northwest-southeast and the axes of the adjacent negative features variously aligned either north-south or north to northwest-south to southeast.

Along the North Carolina segment of the Atlantic Coastal Plain four structural features have been proposed to aid in control over the deposition and erosion of Cenozoic and Mesozoic strata (Fig. 8): (1) the Cape Fear Arch, corresponding to the phase two deformation of Brown et al. (1972), (2) the Neuse Fault, which may correspond to the phase three deformation of Brown et al. (1972), (3) the Carolina Fault, and (4) the Graingers Wrench Zone, which does correspond to the phase three deformation.

These structural features, active at various times during the Cenozoic Era and at least the Cretaceous Period of the Mesozoic Era, produced different topographic highs and lows, resulting in changing depositional basins and centers of erosion on the North Carolina Coastal Plain. Several geologists, most noticeably Ferenczi (1959), Baum et al. (1978), and Harris et al. (1979) have attempted to show that these features controlled the deposition of the Castle Hayne Limestone. The Cape Fear Arch definitely controls the depositional and erosional history of the Castle Hayne Limestone. The Graingers Wrench Zone may not have been active until post-Eocene time, and, as will be

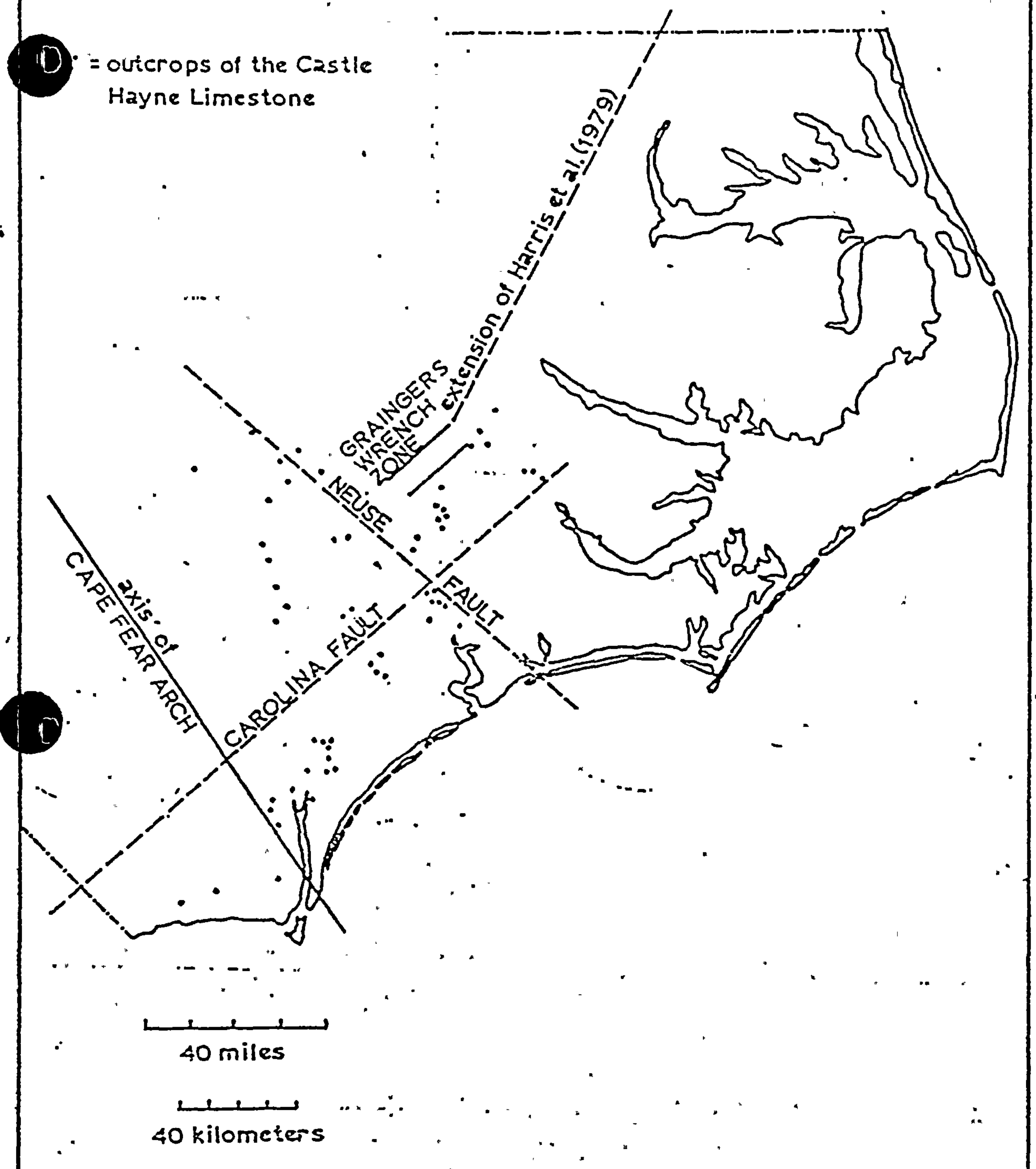


Figure 8. Major structural features on the southern portion of the North Carolina Coastal Plain, after Harris et al. (1979).

discussed, no evidence is available that actually proves the existence of the Neuse and Carolina Faults, much less their control over the Castle Hayne Limestone.

Cape Fear Arch. This positive, northwest-southeast trending arch; part of the phase two deformation of Brown et al. (1972) and first recognized by Dall and Harris (1892), has been shown by numerous geologists to have periodically controlled deposition and erosion of Mesozoic and Cenozoic sediments found adjacent to and on the Arch. The arch represents a basement high with crystalline rock rising to within 335 m of ground surface. Within North Carolina, between Wilmington in New Hanover County and Cape Hatteras in Dare County, the basement surface drops from 338 m to 3012 m below ground surface into the Chesapeake-Delaware Embayment, a difference of 2763 m in 250 km. In North Carolina most sedimentary units either become thin over the rch and grade into shallower water or nonmarine facies or are completely absent due to nondeposition or erosion.

Neuse Fault. This fault, originally described as the Cape Lookout-Neuse Fault Zone (Ferenczi, 1959), later shortened to the Neuse Fault (Baum et al., 1978) corresponds to phase three of Brown et al. (1972). Ferenczi (1959) recognized the fault on three lines of evidence. Further study, however, indicated that his three criteria were not proof of a fault zone.

Ferenczi (1959) suggested that a difference in depth to basement between a well at Havelock (707 m) and a well in Morehead City (1220 m), both in Carteret County and separated by about 24 km, was caused by a basement fault. Brown et al. (1972), however, with the aid of additional lls to the basement, interpreted change in basement surface as a steepening of slope from the Cape Fear Arch toward the Chesapeake-Delaware Embayment.

West of the above mentioned area, Ferenczi used the presence of the Castle Hayne Limestone on the south bank of the Neuse River and its absence in both outcrop and wells north of the Neuse River, in the vicinity of Goldsboro in Wayne County, as evidence for the existence of the fault in this area. In Wayne County, however, the Castle Hayne occurs only as erosional remnants. It is impossible to determine the original geographical extent of the Castle Hayne in this area. Ferenczi's third line of evidence consisted of a number of Eocene outcrops aligned along his fault zone which had undergone silicification. This diagenetic alteration, he argued, resulted from emergence between late Eocene and late Miocene times, due to the movement along the fault. However, several of the Wayne County outcrops are silicified because the presence of siliceous sponge spicules provided a ready source of silica. Silicified sediments along the fall-line segment of Ferenczi's fault are much more widespread than he originally thought and are not just along the fault.

<u>Carolina Fault</u>. This fault zone strikes northeast-southwest (Fig. 8) parallel to the present day coast and does not appear to correspond to any of the deformational phases of Brown et al. (1972). Ferenczi (1959) originally suggested the presence of this fault. Baum et al. (1978) named this feature the Carolina Fault and believed they could trace it from the confluence of the Cape Fear and Black Rivers in Pender County to Kinston in Lenoir County. Several geologic features (Ferenczi, 1959) suggest the existence of this fault: (1) the northwest limit of a magnetically disturbed zone west of Wilmington (MacCarthy,

1936); (2) the location of a zone of subsurface brackish water at the confluence of the Cape Fear and Black Rivers (LeGrand, 1955); (3) a line along the eastern boundary of Martin, Pitt, and Lenoir Counties, where the Upper Miocene Yorktown sediments overlie Cretaceous sediments without intervening Eocene sediments; and (4) a zone where the slope of the basement surface steepens under the North Carolina Coastal Plain, as illustrated by Berry (1951).

Again, as with the Neuse Fault, all of these criteria provide evidence for possible faulting, but none proves conclusively that a fault exists. MacCarthy (1936) reported a parallel series of magnetic highs running from South Carolina across the Cape Fear Arch into North Carolina. However, all his work is south of the Cape Fear River, whereas the Carolina Fault is extended well north of the Arch (Fig. 6). MacCarthy only postulated that the magnetic highs continue northeastward beyond the Cape Fear River. He suggested several possible causes for the highs, but not a fault zone. The salt-water incursion, documented by LeGrand (1955), is good evidence for some type of subsurface structural disturbance but this phenomenon is found only in New Hanover County and small portions of Pender and Brunswick Counties. Interestingly Ferenczi (1959) indirectly suggested that the northwest boundary of the Castle Hayne was controlled by the Carolina Fault; yet the western limit of continuous Castle Hayne in Duplin County is 16 to 32 km east of the proposed fault and north of this area, in Jones and Craven Counties, the western limit is 16 to 32 km west of the proposed fault. These limits are based on the western boundary of the Castle Hayne Limestone as illustrated by Brown et al. (1972) and by the position of known outcrops. The position of the fault zone is based on Ferenczi

(1959) and Harris, Zullo, and Baum (1979).

<u>Graingers Wrench Zone</u>. This fault zone (Fig. 8) is better documented than any other on the North Carolina Coastal Plain. Brown et al. (1977) have found in Craven and Lenoir Counties "tilted and partially exposed blocks of Navarroan, Midwayan, and Claibornian sedimentary rocks" that comprise a "structural mosaic of horst, graben, and half-graben that are arranged in a right-handed, en echelon pattern." The zone characterized by a set of relative left-lateral displacements. Brown et al. (1977) figured that this zone coincided with the phase three deformation of Brown et al. (1972).

The zone is about 24 km wide, but its length is presently unknown. The axis of the zone strikes about N25° to 30°E. Harris et al. (1979), based on gravity anomalies in Virginia (Johnson, 1975) and on geomorphic and structural features in southeast Virginia, extended the wrench zone more than 250 km through the northern part of the North Carolina Coastal Plain and into Virginia (Fig. 8).

overgrowths are most abundant in the high-energy sediments and are nearly absent in the low-energy, micrite-rich sediments.

In the high porosity sediments the overgrowths have frequently grown large enough to contact adjacent allochems, thus helping to hold the sediment together. In micrite-poor sediment, especially in the bryozoan biosparrudite facies, this type of cement is the major sediment binding agent.

Dolomite

Dolomite was found only in the Martin Marietta quarry (NH-4) in New Hanover County, where an extensive lense of highly dolomitized biomicrite is located. The dolomitization was confined to a micrite-dominated layer, overlain and underlain by a less micritic bryozoan biomicrudite. Dolomite reaches a maximum of 100% replacement in the thicker parts of the altered zone (Fig. 24A), whereas a maximum of 56% occurs in thinner portions (Fig. 24B) closer to the outer edges of the lense. Throughout the lense maximum alteration lies close to the vertical center of the lense and decreases both up- and down-section though decrease is less rapid upsection (Fig. 15).

Baum (1977) and Baum et al. (1978a; 1978b) proposed the Dorag model of dolomitization (Badiozamani, 1973) as the most likely way to develop this lense. They envisioned positive movement on the Cape Fear Arch to initiate a relative, local drop in sea level in the vicinity of the quarry as a way to develop the fresh water lense needed for the generation of a fresh water-salt water interface along which maximum dolomitization occurs.

Magnesium necessary for the formation of dolomite ($Mg,Ca(CO_3)_2$) could have come from within the Castle Hayne Limestone. The limestone contains large amounts of echinoderms, the skeletons of which are composed of high-Mg calcite, and even larger quantities of bryozoans, some of which are also high-Mg calcite. In a subaerial vadose environment, magnesium is "preferentially removed from high-magnesium calcite by leaching" or the grains are changed to low-magnesium calcite by solutioning and redeposition on a microscale (Friedman, 1964). This is one of the first products of diagenesis in the vadose environment and could have produced the magnesium needed for dolomitization of underlying sediments in a mixed fresh-sea water phreatic environment.

Silica

Diagenetic silica, in the form of chalcedony, is found in the Castle Hayne Limestone. The nature of its presence demonstrates that an internal source of silica was available. This source is most evident at site W-1, where an average of 10% of the sediment consists of molds of what once were opaline sponge spicules (Fig. 24C). The fine texture and dominance of micrite at this site preserve the small molds (generally less than one millemeter long). Within the W-1 section are thin lenses of chalcedony (Fig. 24D), up to several meters long and a few centimeters thick that mark local diagenetic concentrations of mobilized silica.

Within the main body of the Castle Hayne, chalcedony is found as isolated pockets of silicified limestone. When found, it most commonly occurs in association with sponges; probably because these organisms initially contain opaline spicules.

DEPOSITIONAL ENVIRONMENTS

Based on all the information gathered during the course of this investigation, a geologically reasonable depositional model for the Castle Hayne Limestone can be developed. This model makes use of distinctive geographical and stratigraphical patterns in lithofacies distribution and in patterns of sediment component and sediment texture variation throughout the formation. The only word of caution concerning this model is the fact that most of the information was gathered from surface outcrops. Very little subsurface control is available. The small amount of subsurface data that are available, including thicknesses and general lithologies, supports the model proposed below. It is believed that, with the overall abundance of surface data used to define the model, the use of additional subsurface data would result in a refinement and not a major change in the model.

Embayment Model

The Castle Hayne Limestone, as defined in this report (p. 22) was deposited in an embayment on the southeastern portion of the North Carolina Coastal Plain (Fig. 35). The remaining, uneroded portion of this embayment has a maximum northeast-southwest width of 160 km and is at least 120 km deep from the present coast northwestward. Based on regional stratigraphy and the depositional history of the western North Atlantic Ocean the Eocene terrigenous clastic-dominated outliers

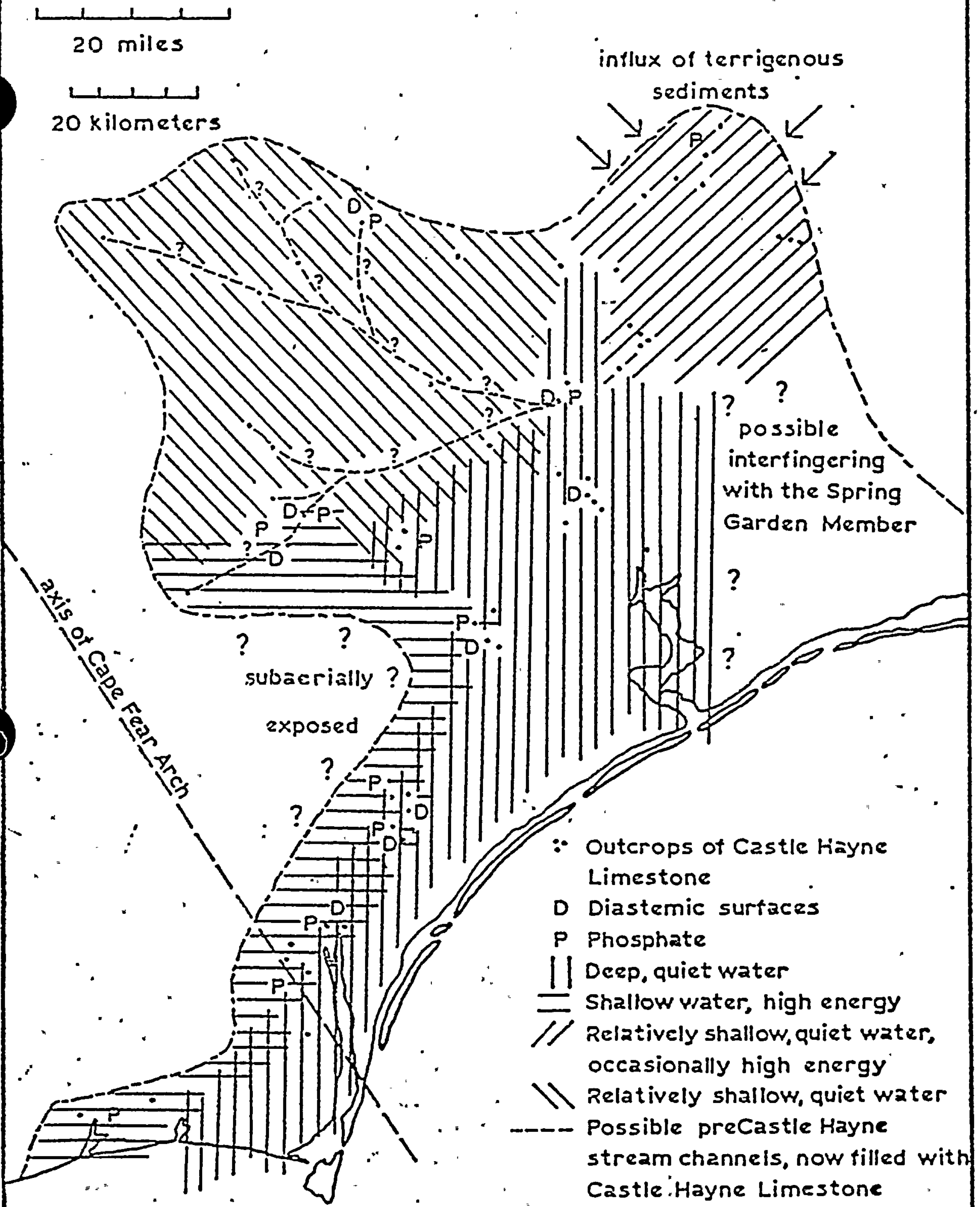


Figure 35. Depositional environments of the Eocene Castle Hayne Embayment on the lower southeastern North Carolina Coastal Plain.

in Hoke, Harnett, Johnston, Moore, and Wake Counties should be Castle Hayne in age. If so, the embayment extended an additional 90 km to the west.

The outcrops in New Hanover, Pender, Onslow, Jones, and Craven Counties mark the present updip limit of continuous Castle Hayne sediment. The outcrops in the western half of the embayment, in Duplin, Lenoir, Sampson, and Wayne Counties, are isolated outliers of what was probably a continuous sheet of carbonate sediment.

The upper slopes of the northeast flank of the northwest-southeast trending Cape Fear Arch serves as the southwestern limit of the embayment, though it is possible that during Castle Hayne time the arch was submerged and covered with a thin veneer of carbonate sediment (long since eroded away). In the New Hanover and Brunswick County area (Fig. 35), farther out on the oceanward end of the arch, the Castle Hayne Limestone actually wraps around the crest of the arch. Sediments along the western and northern sides of the embayment become more enriched in terrigenous clastics, suggesting the original shoreline in these areas was not too distant. To the southeast the embayment opens to the ocean and the continental shelf, where Eocene sediments are dominated by deep water foraminiferal biomicrites and diatomites. No barrier of any type is known to isolate the embayment from the open ocean. The abundant marine fauna found throughout the eastern half of the embayment implies a normal marine environment.

The limestone in the center of the embayment is relatively thick. For a northeast-southwest distance of 100 to 130 km through northern Pender, Onslow, and southern Jones Counties up to 70 m of limestone can be found (Fig. 28), with thickness increasing oceanward. To both the north (Craven and northern Jones Counties) and the south (western New Hanover and southern Pender Counties) the limestone rapidly thins to less than 15 m and extends thus for 30 to 50 km around the edge of the embayment (Fig. 28). In the western half of the embayment, the limestone is found only as scattered outliers where the rock occupies low spots in the Pre-Castle Hayne topography (Figs. 19, 20, and 28).

During Castle Hayne time the area occupied by the embayment was topographically, the lowest part of the coastal plain. Numerous geologists (p. 35) have postulated structural control of the deposition of the Castle Hayne Limestone, suggesting that down-dropped blocks were the primary method of producing an area below sea level in which the deposition of carbonate sediment occurred. The data gathered for this report suggests a geomorphic control rather than a structural control. As discussed on pages 24 through 29, prior to the deposition of the Castle Hayne Limestone, the Atlantic Coastal Plain experienced a major marine regression, with sea level perhaps as low as several hundred meters below present sea level. The absence of Paleocene and Lower Eocene sediments beneath the Castle Hayne Embayment (most of the Castle Hayne lies disconformably on Cretaceous sediment) confirms a major erosional event in this area. Three dimensional reconstruction of the D-5 outlier (Figs. 19 and 20) shows deposition of limestone in an elongate, branching depression. Many of the other outliers are restricted in size but are relatively deep (many are greater than 30 m). It is here proposed that these outliers occupy remnants of a major Pre-Castle Hayne drainage system (Fig. 35). This drainage system produced a series of relatively deep stream valleys and also lowered the overall elevation in the embayment area. This stream system possibly fed from the outlier area (Duplin, Sampson, and Wayne Counties) into the deeper part of the eastern half of the embayment (southern Jones, Onslow, and northern Pender Counties) through southern Lenoir County and southwestern Jones County. Site J-6, in southwestern Jones County contains at least 30 m of limestone. As seen in photographs of southern Lenoir County, an area up to 10 km wide, running eastward into Jones County, contains a high concentration of sinkholes, signifying a well developed unit of underlying limestone. Just east of this area (Onslow County), based on well logs, is the deepest portion of the embayment.

Based on the distribution of premanent diastems exposed around the present-day geographical edge of the formation (Fig. 35) especially in New Hanover, Duplin, Wayne, and Sampson Counties, at least two episodes of transgression and limestone deposition are recognizable. These two episodes are separated by a prominent erosion surface, marking a significant marine regression. In the eastern section of the outcrop belt the lower depositional cycle is represented by the phosphate pebble biomicrudite facies. In the outlier belt, at sites D-5 and D-8 the two depositional cycles are separated by a prominent erosion surface that displays local relief of up to 12 meters. Samples collected from the lower cycle show an increase in quartz sands upsection, suggesting the regression allowed terrigenous clastics to feed into the area of these particular sites before development of the erosion surface.

Sediments accumulated during the second episode of deposition dominate the formation. Within this episode the terrigenous sediment component decreases upsection (Fig. 37) and laterally turned the center of the embayment (Fig. 36). Lithofacies around the edge of the basin average from 10 to 20 percent quartzsand, while the bryozoan biomicrudite

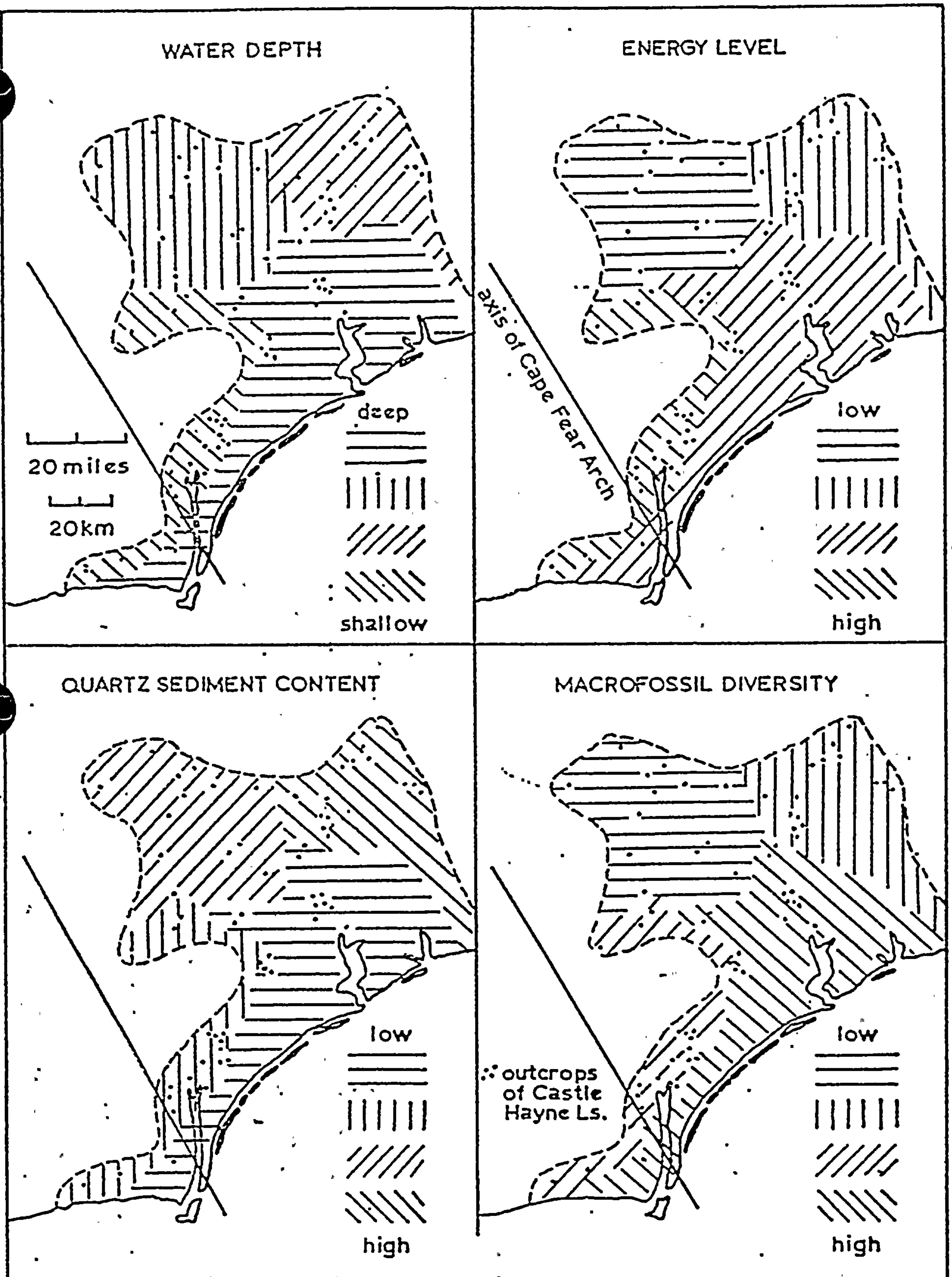


Figure 36. Geographical distribution of selected environmental features within the Castle Hayne Embayment.

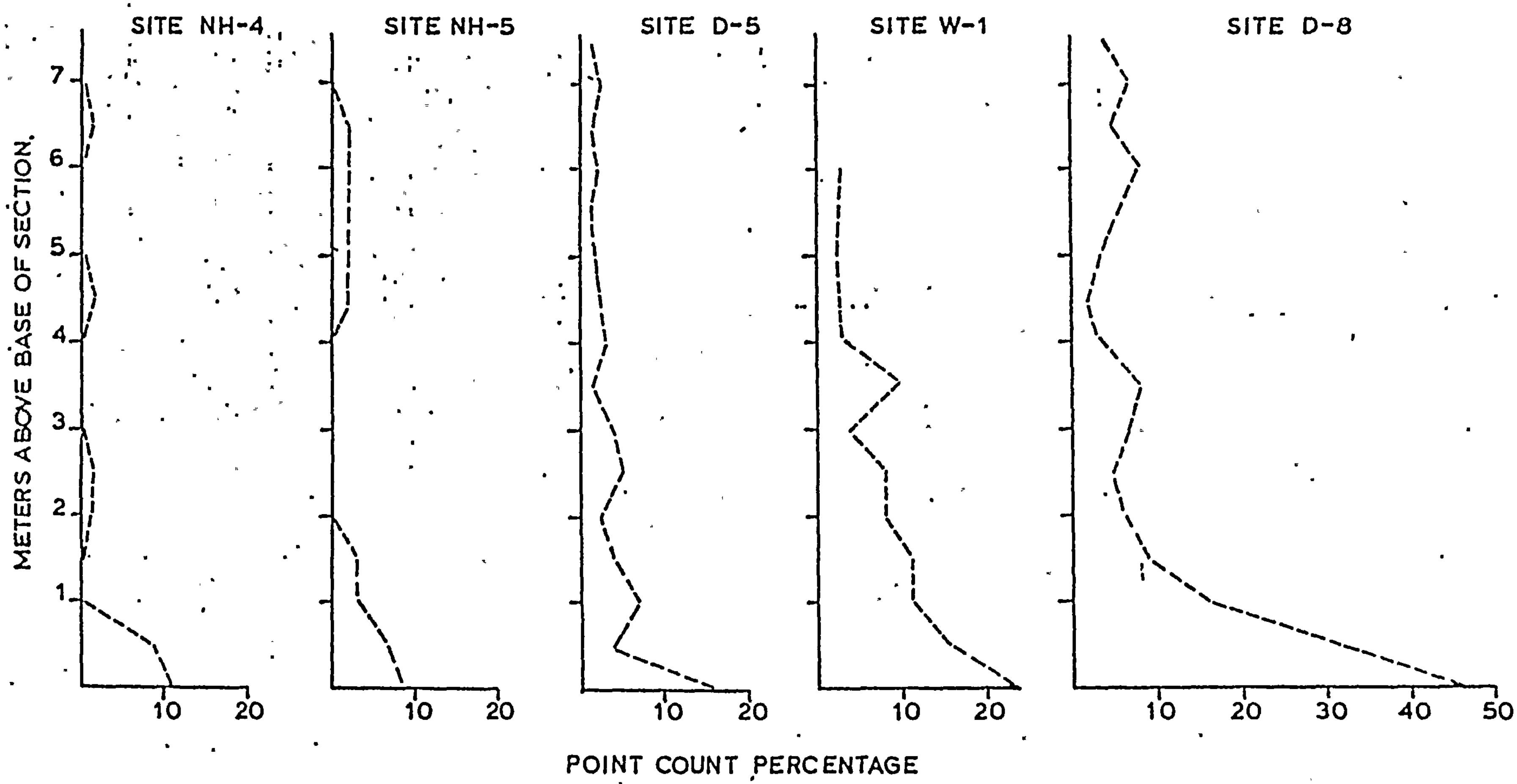


Figure 37. Decrease in quartz sand and silt content upsection at five selected sites. Site NH-4 and NH-5 are in the major outcrop belt and sites D-5, D-8, and W-1 are outliers.

facies in the center of the embayment averages about two percent quartz sand. In comparison with the terrigenous strata above and below the Castle Hayne even the 20% terrigenous content of the peripheral facies is minor. The relatively low, overall terrigenous content, the presence of reworked clasts composed of rock from the underlying Paleocene and Cretaceous formations, the lack of unstable terrigenous grains (feldspars, micas, clays, and other common minerals), and the dominance of quartz grains favor a local reworking of underlying and surrounding noncarbonate sediments rather than an influx of sediment from the crystalline Appalachians to the west as a source of the terrigenous component of the Castle Hayne sediments.

The various components of the different lithofacies provide ample information for deciphering the depositional environments within the embayment. Generally, a shallow-water, high-energy bryozoan biosparrudite facies extends along most of the western edge of the embayment (Fig. 31 and 35). The northern and northwestern portion of the embayment is filled with micrite dominated sediments deposited in a relatively shallow, low-energy environment, including the sandy, sponge spicule-bearing biomicrite facies to the northwest and the sandy, foraminiferal, echinoderm biomicrite to the north. The center and oceanward portion of the embayment is dominated by the deep, relatively quiet water bryozoan biomicrudite facies.

The bryozoan biomicrudite facies contains an abundant normal marine fauna and is the most fossiliferous unit in the embayment (Fig. 36). The growth forms of the bryozoans (p. 141) and the types of siliceous sponges (p. 139) in this facies suggests water depths in excess of 100 m. The high energy bryozoan biosparrudite facies also contains an abundant marine fauna, though it does appear as though a large portion of the bioclastic debris may have been washed in from deeper waters. The overall faunal content of the sandy, foraminiferal, echinoderm biomicrite and the sandy, sponge spicule-bearing biomicrite facies is much more restricted (Fig. 36). It is possible that the high micrite content in these two facies may have kept a large number of the filter-feeding organisms out of this part of the embayment.

The distribution of high versus low energy environments (Fig. 36) and of macrofossil diversity (Fig. 36) suggests that the center of the embayment and the southwestern edge of the embayment were more open to the ocean that the northern half of the embayment. It is possible that normal wind driven marine currents flowed into the more open end of the embayment and washed against the eastern edge of the Cape Fear Arch, where the high energy bryozoan biosparrudite facies is found. The concentration of phosphate mineralization in this area relative to the northern half of the embayment (Fig. 26) also suggests deeper water currents flowing into the lower part of the embayment. The northern half of the embayment appears to have been more restricted, perhaps a function of the shape of the embayment and of the direction of current movement within the embayment.

The faunal variety and faunal abundance in the Castle Hayne Limestone imply normal marine environments, but not necessarily tropical environments. Ample documentation now shows that carbonate sediments accumulate in all climatic regions. Organisms that produce carbonate skeletons inhabit continental shelves at every latitude (Chave, 1967). Skeletal debris can accumulate to form carbonate-dominated sediments wherever the influx of terrigenous clastics is insufficient to dilute the carbonate fraction. Rivers, the major suppliers of terrigenous debris to the continental shelves, and not climate, apparently control the distribution of carbonate sediments (Chave, 1967).

Based on extensive sampling of shelf sediments carbonates are found to dominate hundreds and even thousands of square kilometers of temperate shelves (Lees and Buller, 1972). Table 6 presents data summarized in extensive surveys of mid-latitude, temperate water carbonates and low latitude, tropical carbonates (Lees and Buller, 1972; Nelson, 1978). When compared with data characterizing the Castle Hayne Limestone, it is seen that the Castle Hayne shares characteristics of both tropical and temperate carbonates (Table 6). This mixing suggests that the limestone was deposited in a subtropical to warm temperate climate.

The benthic fauna of the Castle Hayne Limestone displays a high degree of endemism (Canu and Bassler, 1920; Kellum, 1926; Cheetham, 1961; Fallow, 1962; Kier, 1980; Rigby, 1980-per. com.) The fauna is most similar to the fauna of the Eocene of South Carolina, but is very different from the Eocene of the remainder of the Atlantic Coastal Plain or the Gulf Coastal Plain. This endemism suggests some form of barrier between the North Carolina embayment and the more southern shelf environments. Several features could have produced this isolation. The Cape Fear Arch, whether subaerially exposed or submerged under shallow marine waters during Castle Hayne time, could have acted as an effective migratory barrier. The strong, north-flowing offshore Gulf Stream, already active during the Eocene (p. 24) could have prevented any southward drift of benthonic species that possessed planktonic or nektonic larvae. Southerly species, however, could have

become introduced by way of the Gulf Stream into the North Carolina region. Once introduce, however, they would have had to successfully compete with an already well established fauna. If successful they were incorporated into the Castle Hayne ecosystem. It is evident, though, that few species were able to do so.

A combination of the Cape Fear Arch, a north-flowing offshore current, the deep water conditions of the Eocene North American Atlantic shelf (p. 27) and the fact that the limestone was deposited in an embayment served to isolate the Castle Hayne and produce a highly endemic fauna. It is probably this endemism and a resultant difficulty in correlating with other Eocene faunas that has caused problems with biostratigraphically dating the exact age of the Castle Hayne Limestone.

CONCLUSIONS

The Eocene Castle Hayne Limestone, the northernmost essentially pure carbonate formation on the Atlantic Coastal Plain, crops out on the southern half of the North Carolina Coastal Plain. The major outcrop belt, marking the updip limit of continuous rock, is 160 km long and up to 24 km wide, and strikes southwestward through southern Pitt, western and central Craven, western Jones, central Onslow, central Pender, northwestern New Hanover, and possibly southern Brunswick Counties. Outliers, erosional remnants preserved within Pre-Castle Hayne stream valleys, are found in Duplin, Lenoir, Sampson, and Wayne Counties. Outliers of molluscan-rich, clastic dominated sediments in Harnett, Hoke, Johnston, Moore, and Wake Counties are dated as Eocene in age. These sediments appear to be Castle Hayne equivalents, and thus extends the marine transgression that initiated deposition of this formation to the eastern edge of the Piedmont Province of the Appalachian Highlands, a full 200 km west of the present coast.

Sixty-three outcrops of the Castle Hayne have been reported in the literature during the past 145 years. These are divided into five major lithofacies that were deposited in a coastal embayment: (1) a shallow-water, low-energy, phosphate-pebble biomicrudite; (2) a shallow-water, high-energy, bryozoan biosparrudite; (3) an intermediate-depth, low-energy, sandy, sponge-spicule-bearing biomicrite; (4) an intermediate-depth, low-energy, sandy, foraminiferal, echinoderm biomicrite; and (5) a deep water, low-energy, bryozoan biomicrudite. Three minor lithofacies, representing localized changes in depositional environment, are also recognized: (1) a dolomitized biomicrite; (2) a molluscan-mold, bryozoan biomicrudite; and (3) a bryozoan, foraminiferal biomicrite.

Distribution of sponges, Foraminifera, and bryozoans, plus the lack of calcareous algae, suggest water depths of more than 100 m (328 ft) during deposition of the bryozoan biomicrudite facies, the easternmost facies in the outcrop area, and 30 to 45 m for the sandy, sponge-spicule-bearing biomicrite facies, the westernmost facies in the outcrop area. The sandy, foraminiferal, echinoderm facies, bearing characteristics of both of the above-mentioned lithotypes, is considered to be intermediate in depth. The bryozoan biosparrudite facies, bearing evidence of tidal influence, was deposited above wave base. The above four facies were deposited in a transgressive environment. The basal phosphate-pebble biomicrudite facies, separated from the other lithologies by a prominent erosion surface that developed during a significant marine regression, was deposited during an earlier transgressive episode. Thick sections of Castle Hayne-like sediments, found beneath typical Castle Hayne Limestone in numerous outliers in Duplin, Sampson, and Wayne Counties, may also belong to this earlier transgressive episode.

Except for the Echinoidea, the Foraminifera, and the Bryozoa, the Castle Hayne fauna is still poorly understood, both taxonomically and paleoecologically. A large portion of the known Castle Hayne fauna is endemic, thus posing a problem when trying to biostratigraphically correlate the formation with other Eocene strata. This endemism is possibly the reason why paleontologists disagree on the age of the limestone, some placing it in the Middle Eocene, others in the Upper Eocene, and still others in both the Middle and Upper Eocene. The history of the North Atlantic Ocean, accompanied by the pattern of sea level fluctuation and marine deposition of deep sea sediments suggests that the Castle Hayne is Middle Eocene, a time of major transgression in the north Atlantic coastal regions, and abundant evidence for such a transgression is available for the Atlantic North American continental shelf.

Diagenesis of the Castle Hayne environment occurred in four major environments: (1) shallow marine, represented by glauconite formation; (2) mixed shallow marine and fresh water vadose, associated with hardgrounds and including phosphate precipitation, development of drusy rim cements, and microkarstic features, (3) fresh-water vadose, represented by dissolution of aragonite and leaching of Mg^{++} from high-Mg calcite bioclasts; and (4) fresh-water phreatic, represented by development of syntaxial overgrowths on echinoderm fragments. One additional minor diagenetic alteration is the dolomitization of a biomicrite lense, exposed in the Martin Marietta quarry in New Hanover County. This diagenetic process occurred along a zone of marine and fresh-water mixing in a near-shore subaerial environment associated with the Cape Fear Arch.

The depressions in which the limestone is preserved in Duplin, Lenoir, Sampson, and Wayne Counties are interpreted as preCastle Hayne stream valleys. The exact dimensions of these outliers are not known. It is possible that some of these outliers, although separated by many kilometers on the surface, may be continuous in the shallow subsurface.

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STRUCTURAL AND STRATIGRAPHIC FRAMEWORK FOR THE COASTAL PLAIN OF NORTH CAROLINA

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PLIO-PLEISTOCENE CRUSTAL WARPING IN THE OUTER COASTAL PLAIN OF NORTH CAROLINA

by

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INTRODUCTION

Many landforms and associated sedimentary deposits in the southeastern Atlantic Coastal Plain preserve a record of repeated inundations and withdrawals of the sea during the Pliocene and Pleistocene. Traditional interpretations of the origins of these features presume that the Coastal Plain was a stable crustal region during and after their formation. Periodic glacio-eustatic transgressions and regressions of the sea are correlated with interglacial and glacial stages, respectively, of the Pleistocene (e.g., Oaks and DuBar, 1974), or with earlier displacements of ocean basin waters onto the land during episodes of increased sea floor spreading (Le Pichon, 1968).

Reliance on a stable crust model for the explanation of Plio-Pleistocene events is in marked contrast to the conclusions derived from studies of older Tertiary and Cretaceous Coastal Plain sediments, whose distribution and character are known to have been influenced by episodic activity along major structural features (e.g., Baum et al., 1978; Brown et al., 1972, 1977; Ferenczi, 1959; Harris et al., 1979; Richards, 1950). Few studies have suggested that tectonic activity in the Coastal Plain might have played a role in modifying the effects of eustatic sea level change during the Pliocene and Pleistocene. Doering (1960) concluded that upwarping of the Cape Fear arch (fault) in southeastern North Carolina, together with regional uplift of the Appalachian Highlands and inner Piedmont, preceded Pleistocene glacio-eustatic oscillations. Winker and Howard (1977), based on a re-interpretation of relict shoreline sequences in the Atlantic Coastal Plain south of the Cape Fear River, North Carolina, arrived at similar conclusions, and provided tentative evidence for Pleistocene uplift along the Cape Fear fault.

Direct evidence of Pleistocene tectonic activity in the Coastal Plain is difficult to obtain. Faulting of units in subsurface is obscured because of the minor amounts of displacement involved, and because of the thinness, lithologic similarity, and discontinuity of Pleistocene sediments. Surface fault scarps are rapidly obliterated by fluvial erosion of the unconsolidated surficial sediments. Instead, reliance must be placed upon recognition of the secondary effects of tectonic activity on regional geology and geomorphology.

CAPE-FEAR - NEW RIVER COASTAL PLAIN, NORTH CAROLINA

The geology of Plio-Pleistocene deposits in the outer Coastal Plain between the Cape Fear and New Rivers, North Carolina (Fig. 1) has not been studied in detail. The region is a structural and geomorphic entity bounded to the southwest by the Cape Fear fault, and to the northeast by the Neuse fault. The Cape Fear fault, whose axis is approximated by the course of the Cape Fear River, has been active periodically since Aptian-Albian time and has had a profound influence on the distribution and thickness of Cretaceous and Tertiary units on either side of its axis. Initial movement along the Neuse fault, which can be traced from the vicinity of Smithfield, North Carolina southeast to the coast between New and Neuse Rivers, also occurred during Aptian-Albian time. Changes in structural and depositional strike and thickness of

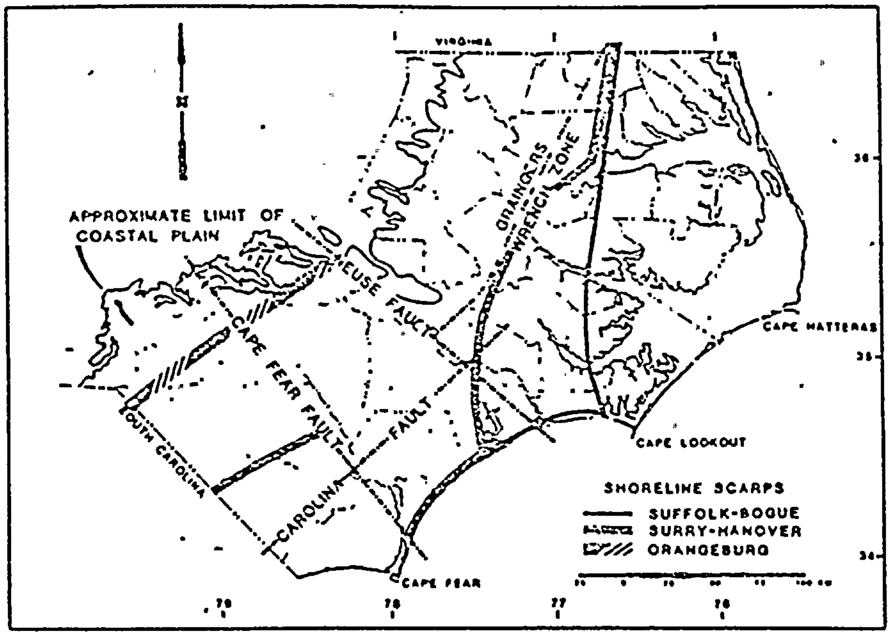
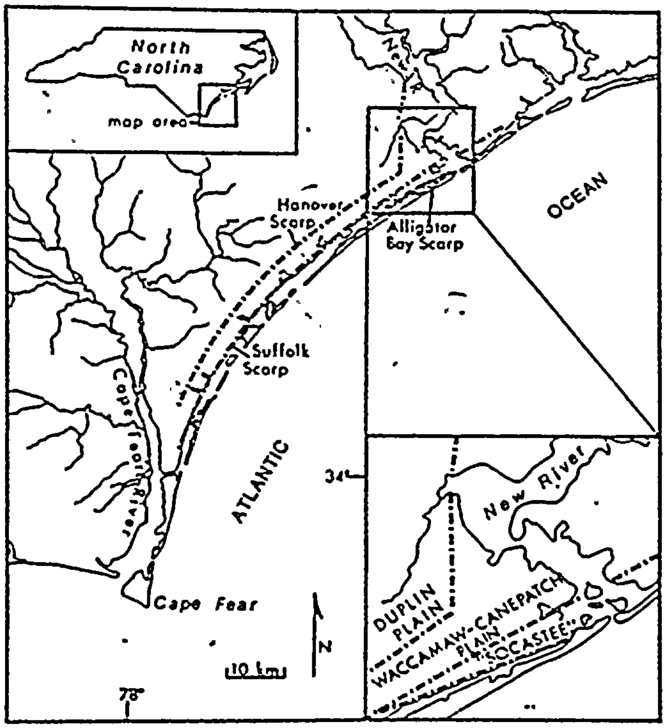
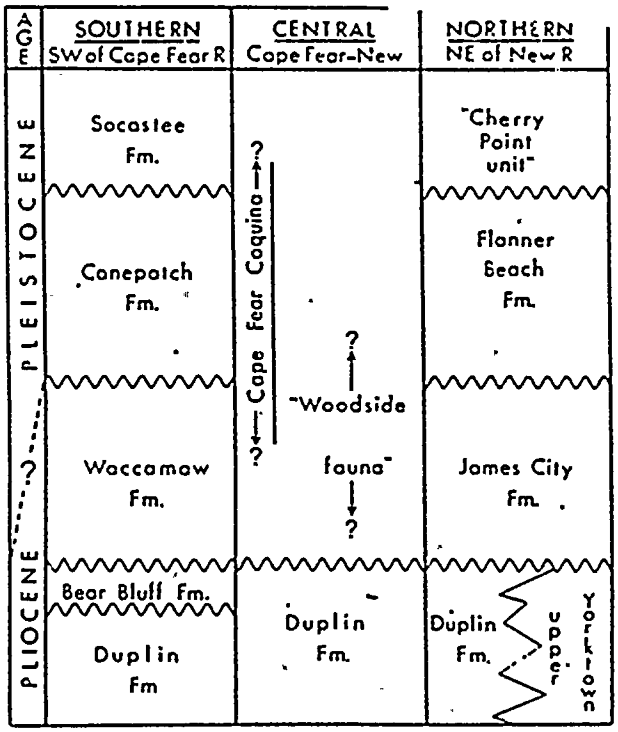


Figure 1. Relation of scarps to major structural Features, North Carolina Coastal Plain.





)
e 2. Location of scarps and plains, outer Coastal Plain, North Carolina.

(right)
Figure 3. Stratigraphic relationships of upper Cenozoic units, Carolina Coastal Plain.

Tertiary units crossing the fault indicate periodic movement in the Paleogene and early Neogene (see Harris et al., this volume).

Unlike adjacent sections, this part of the Coastal Plain is characterized by a dearth of either large scale relict shoreline features or Pleistocene marine deposits. In addition, drainage development and direction of flow differ markedly from those seen in adjacent Coastal Plain sections. Detailed geomorphic analysis, utilizing recently completed 7.5' topographic quadrangles with 5-foot (1.5 m) contours, coupled with field mapping and analysis of subsurface data have been used to delimit relict shorelines and associated marine deposits in the region. The data obtained from these varying lines of evidence indicate that the unusual geologic and geomorphic features of the Cape Fear-New River region are the result of episodic tectonic activity during the Pliocene and Pleistocene.

SHORELINE SCARPS AND ASSOCIATED MARINE DEPOSITS

An erosional scarp, here designated Hanover Scarp, with an average relief of 5 m can be traced from central New Hanover County northeastward to the west side of New River, Onslow County (Fig. 2). At this point, Hanover Scarp turns abruptly north and is traced for 20 km along the west side of New River. A second scarp, located seaward of Hanover Scarp and essentially delimiting the modern mainland coastline, parallels Hanover Scarp between central New Hanover County and New River. This scarp, as predicted by Mixon and Pilkey (1976), is the southwesterly continuation of their Bogue Scarp. Bogue Scarp as mapped by Mixon and Pilkey (1976) continues northeastward past New River and parallels the shoreline into central Carteret County where it connects with elements of the north-trending Suffolk Scarp of southeastern Virginia (Mixon and Pilkey, 1976; Oaks and DuBar, 1974). A third scarp, here designated Alligator Bay Scarp, occurs seaward of Suffolk Scarp between Spicer and Alligator Bays west of New River. This minor scarp is submerged southwest of Spicer Bay, and has not been mapped northeast of New River, although Mixon and Pilkey (1976) indicate the presence of what is presumed to be this scarp along part of the Bogue Sound shoreline.

These three scarps form the seaward borders of tilted plains. These plains are immediately underlain by a veneer of nonmarine deposits that overlie fossiliferous marine sediments, and are regarded as subaerially modified sea floors of former marine transgressions. The surface landward of Hanover Scarp is an extensive, moderately dissected, west to northwest sloping plain. This plain is continuous inland to the seaward edge of the inner Coastal Plain (Orangeburg Scarp), and is characterized by the development of Carolina Bays. In the Early Pliocene the outer and middle southeastern Atlantic Coastal Plain was inundated by the Duplin sea to Orangeburg Scarp. The youngest fossiliferous marine deposits underlying the plain between Orangeburg and Hanover Scarps are outliers of the Duplin Formation (Figs. 3 and 4). The dissected plain between Hanover and Suffolk Scarps is underlain by younger marine deposits including equivalents of the Waccamaw and overlying Canepatch formations of South Carolina (Figs. 3 and 4).

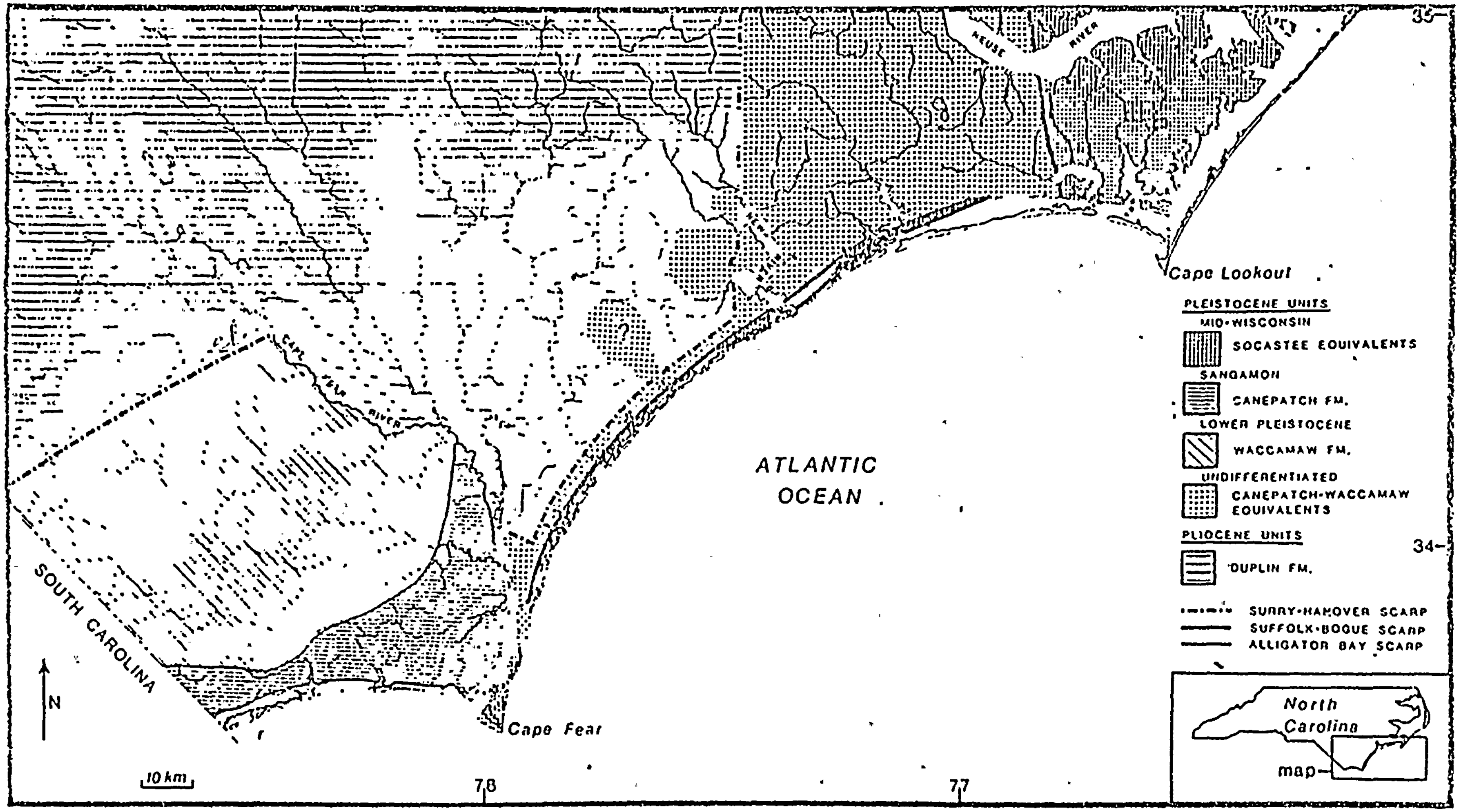


Figure 4. Geology of Plio-pleistocene deposits, outer Coastal Plain, North Carolina (modified from DuBar et al., 1974; Mixon and Pilkey, 1976).

The age of these units is not established conclusively, but the marine transgressions responsible for their deposition appear to have occurred between the beginning of the Pleistocene (Calabrian) and the end of the Sangamon Interglacial (Campbell et al., 1975; DuBar et al., 1974).

Subsurface stratigraphy of the plain between Alligator Bay and Suffolk Scarps is not known. Richards (1950) reports fossils from this area that may be related to the Socastee fauna. The Socastee Formation overlies the Canepatch Formation in South Carolina and is considered to represent a minor transgression during the mid-Wisconsin Interstadial (Figs. 3 and 4). Socastee equivalents are known east of Suffolk Scarp farther to the north in the Neuse River region, North Carolina (Mixon and Pilkey, 1976; Oaks and Dubar, 1974).

DISCUSSION

Assuming that the Duplin plain, the Waccamaw-Canepatch plain, and the "Socastee" plain were formed as nearly horizontal surfaces, the presently observable slopes and slope directions on the plains indicate that episodic and differential uplift have occurred in the region (Fig. 5C). The Duplin plain is at an elevation of 12.2 m in central New Hanover County, but to the northeast, over a distance of 60 km, its elevation gradually increases by nearly 9 m to 21 m on the west side of New River. The Waccamaw-Canepatch plain rises less than 2 m between central New Hanover County and New River, and the "Socastee" plain, although only traceable for about 12 km southwest of New River, rises from sea level to 4.6 m at New River. Although all three plains presently dip west or southwest from an axis along New River, the observed differences in slopes of these plains are indicative of at least three periods of tectonic activity between the time of withdrawal of the Duplin sea and the present.

The divergence of slopes of the Duplin and Waccamaw-Canepatch plains towards New River indicates uplift along the Neuse fault after withdrawal of the Duplin sea and prior to Canepatch transgression (between three million and 75,000 years ago). The divergence of slopes of the Waccamaw-Canepatch and "Socastee" plains toward the Cape Fear fault indicates uplift of the fault after Canepatch sea withdrawal and prior to transgression of the Socastee sea (between 75,000 and 32,000 years ago). The divergence of slopes of the "Socastee" plain and the modern sea level plain towards New River indicates uplift along the Neuse fault in the past 30,000 years that resulted in the present attitude of the plains (Figs. 5A-C).

Post-Waccamaw-Canepatch uplift along the Cape Fear fault also is recorded by regional anomalies in distribution and elevations of these formations. The Cape Fear River, whose course approximates the trend of the Cape Fear fault, forms the boundary between two distinct geologic regions. Northeast of the river the Waccamaw and Canepatch formations are restricted to the narrow coastal strip seaward of Hanover Scarp, and are not found above +5 m elevation. Southwest of the river the Waccamaw Formation extends about 80 km inland to the seaward edge of the middle Coastal Plain (Surry Scarp), and its base is found at elevations up to +28 m (DuBar et al., 1974; Howard, 1974). The Canepatch Formation extends about 20 km inland, and its base occurs at elevations up to +13.7 m (DuBar et al., 1974).

The widespread distribution and higher elevations of Pleistocene formations southwest of the Cape Fear River indicate that: (1) prior to Waccamaw-Canepatch deposition the region southwest of the river was lower than the region to the northeast; and (2) after Waccamaw-Canepatch deposition the southwestern region was uplifted with respect to the northeastern region. This uplift is tentatively correlated with the initial "reverse" tilting of the Waccamaw-Canepatch plain between Cape Fear and New Rivers.

Late Pleistocene uplift along the Cape Fear fault is further suggested by modern drainage patterns (Fig. 1). The Cape Fear drainage basin is narrow and exhibits a parallel pattern with dominant southeasterly flow. Inland the Cape Fear River flows at the base of a high, northeastward-facing bluff that forms an extremely narrow divide with the Lumber-Big Swamp drainage basin to the southwest. This bluff, actually a receding fault line scarp, loses elevation seaward, and is not distinguishable in the outermost Coastal Plain. Here the divide between the Cape Fear and Waccamaw drainage basins is broad and low, and no abrupt change in elevation marks the divide. Drainage patterns in the Lumber-Big Swamp and Waccamaw basins are dendritic, and flow predominantly to the southwest. The Lumber-Big Swamp and Waccamaw drainage systems are characterized by underfit, poorly integrated, complexly meandering streams occupying very large floodplains, whereas the Cape Fear drainage system is well-integrated and composed of streams in accord with their floodplains.

We propose that the Cape Fear system is younger than drainage systems to the southwest, and developed as a result of uplift of the Coastal Plain southwest of the Cape Fear fault. Prior to uplift, runoff from the inner Coastal Plain and Piedmont flowed southwesterly across what is now the Cape Fear drainage basin, and deposited the sequence of prograded fluviodeltaic sediments described by DuBar et al. (1974) as overlying marine Plio-Pleistocene deposits southwest of the Cape Fear River. Uplift resulted in the beheading of these major drainage systems, causing the formation of underfit streams downstream, and in the deflection of upstream runoff to the southeast, forming the Cape Fear drainage system.

On the basis of the data presented, we propose the following sequence of geologic events:

(1) In the early Pliocene the Duplin sea transgressed over the outer and middle Coastal Plain, cutting and occupying Orangeburg Scarp.

(2) Uplift along the Neuse fault parallel to the modern course of New River occurred after withdrawal of the Duplin sea from the Coastal Plain (circa three million years ago), and resulted in warping of the Duplin plain and a general westward dip to the surface between Cape Fear and New Rivers (Fig. 5A).

(3) In the (?early) Pleistocene, the Waccamaw sea transgressed over unelevated regions of the Coastal Plain southwest of Cape Fear River and northeast of New River, cutting and occupying Surry Scarp. The Waccamaw transgression was insufficient to inundate the uplifted region between Cape Fear and New Rivers, and was limited to the cutting and occupation of Hanover Scarp.

(4) Withdrawal of the Waccamaw sea from the Coastal Plain was followed by the less extensive Cane-

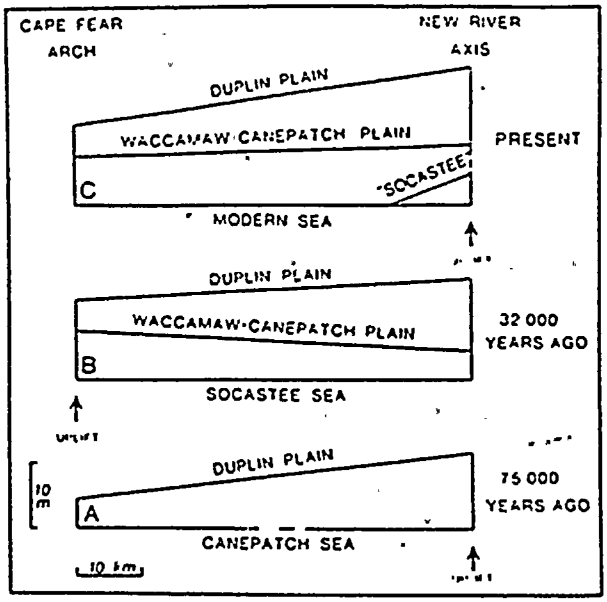


Figure 5. Diagrammatic SW-NE profiles of plains between Cape Fear and New Rivers. (A) Slope of Duplin plain during occupation of Hanover Scarp by Canepatch sea. (B) Slope of Duplin and Waccamaw-Canepatch plains during occupation of Suffolk Scarp by Socastee sea. (C) Present slopes of Duplin, Waccamaw-Canepatch, and "Socastee" plains during modern (Holocene) transgression. Arrows below profiles indicate area of preceding tectonic activity.

patch transgression that re-occupied part of the unelevated outer Coastal Plain and Hanover Scarp during the Sangamon Interglacial.

(5) Withdrawal of the Canepatch sea, presumably at the end of the Sangamon (circa 75,000 years ago) was accompanied by uplift along the Cape Fear fault that resulted in elevation of the region southwest of the Cape Fear River, the development of the Cape Fear drainage system, reduction in the general westward slope of the Duplin plain, and the initial "reverse" or eastward slope of the newly formed Waccamaw-Canepatch plain (Fig. 5B).

(6) The Socastee transgression during the Wisconsin Interstadial (circa 32,000 years ago) occupied coastal regions on both sides of the Cape Fear River and cut and occupied Suffolk Scarp northeast of the river. Further uplift along the Neuse fault occurred after withdrawal of the Socastee sea and resulted in presently observed plain slopes and elevations in the region between Cape Fear and New Rivers (Fig. 5C).

IMPLICATIONS

Recognition of tectonic activity during periods of eustatic sea level change significantly alters interpretations of Plio-Pleistocene history of the North Carolina Coastal Plain. Some determinations of Pleistocene sea level are based on localities now known to have undergone appreciable post-depositional uplift. Regional correlation of scarps and associated relict shoreline features requires re-examination, as it is based on overestimates of maximum sea level and on assumed crustal stability. Furthermore, documentation of Plio-Pleistocene crustal instability in one part of the Atlantic Coastal Plain suggests the possibility that other areas were similarly affected. We offer the suggestion that at least some of the problems encountered in the elucidation of Plio-Pleistocene geologic history of the Coastal Plain may best be solved through abandonment of the stable crust model.

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tional environments within the Santee Limestone reflect the syndepositional-tectonic history within and about the Santee depositional basin; those of the Castle Hayne Limestone reflect the history of its basin. Thus, the initiation of deposition of these formations are, more likely, the products of intrabasinal environmental conditions and are not indicators of contemporeneity. The time-transgressive nature of Santee-Castle Hayne biofacies was alluded to by Cooke and MacNeil (1952, p. 24):

> It is not surprising that the faunas of the Santee, Castle Hayne, and Ocala limestones are somewhat similar, for these three formations represent similar facies. The Santee and Castle Hayne faunas were not recognized as of Claiborne age because no similar bryozoan-bearing limestone facies occurs in the Claiborne west of the Carolinas.

We may not agree with their age assessments, but we agree fully with their philosophical approach (Fig. 3).

ANALYTICAL PROCEDURES AND RADIOMETRIC RESULTS

A composite sample of the glauconitic zone was collected from the lectostratotype of the Castle Hayne Limestone, New Hanover County, North Carolina. Five glauconite concentrates were separated on the basis of grain size and external morphology into samples designated: MM1-100HT; MM1-100HM; MM1-100HF; MM1-70HF; and MM1-70HT. The samples were further prepared for analysis according to the procedure described by Harris and Bottino (1974). The concentrated samples contained less than 1% impurities of pyrite and dolomite. X-ray diffraction analysis of the glauconite samples confirmed that the samples consisted of the well-ordered to disordered glauconite defined by Bentor and Kastner (1965).

The five glauconite samples were analyzed for Rb, Sr, and Sr-isotopic composition using standard chemical and isotopic dilution procedures. A technique using concentrated acids and small ion-exchange columns also was employed for separation of Rb and Sr (Russell, 1978). In addition, Fe was separated from all Sr samples using these small columns. The results are shown in Table 1. Rb and Sr blanks were collected in order to monitor contamination encountered in handling and preparing the samples for analysis. Analysis of the blanks has shown that procedural contamination for the Rb and Sr was negligible. Therefore, no correction for the blanks has been made on the values given in Table 1. On the basis of analyses of the National Bureau of Standards Standard Sam-

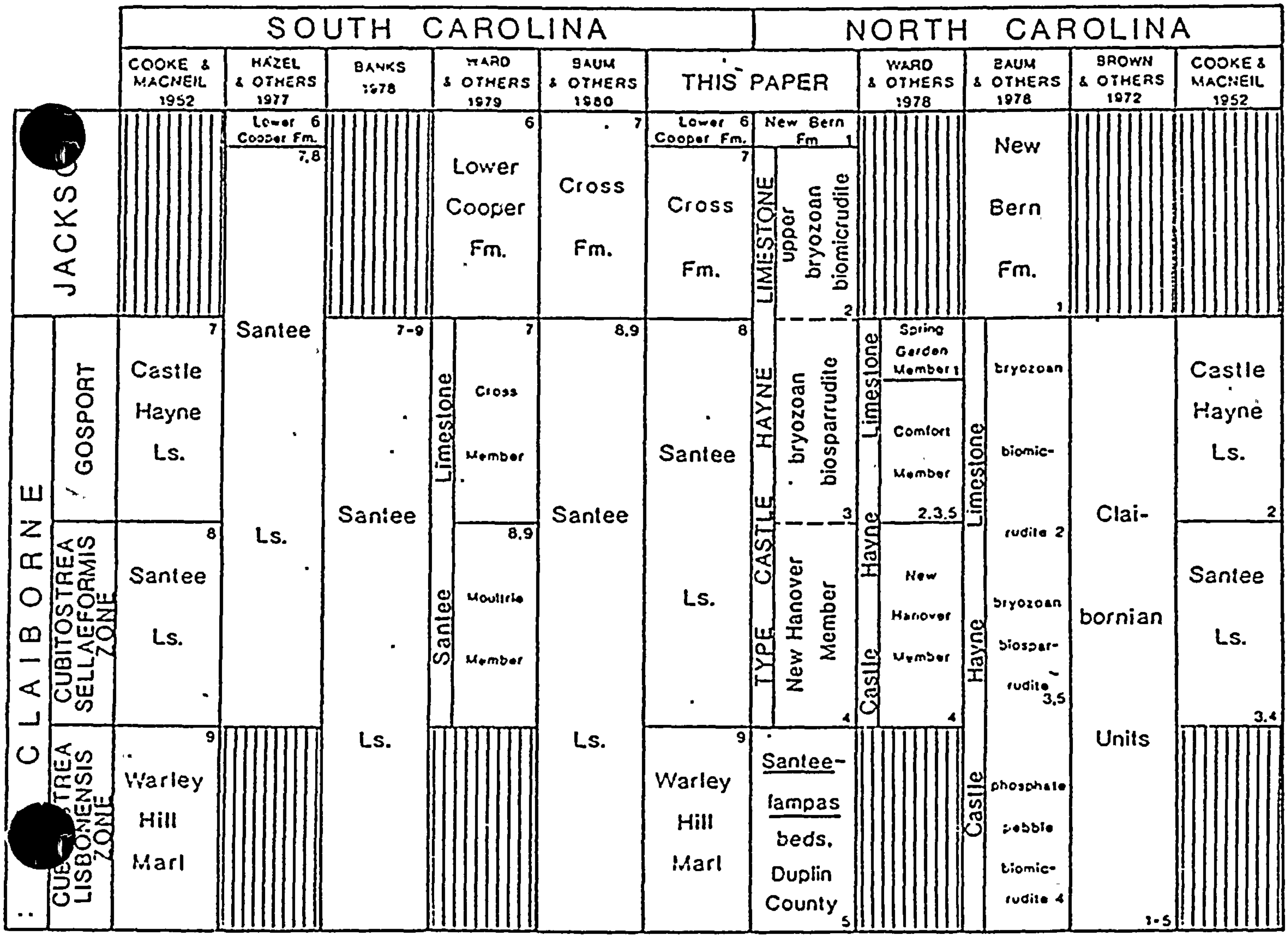

Figure 3. Suggestion correlation of Eocene strata of North and South Carolina. Numbers indicate equivalent rock units.