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TIMING AND DEPOSITIONAL HISTORY OF EUSTATIC SEQUENCES: CONSTRAINTS ON SEISMIC STRATIGRAPHY

Edited by
CHARLES A. ROSS and DREW HAMAN
Chevron, U.S.A., Inc.
P.O. Box 1635
Houston, Texas 77251

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PREFACE

Study of depositional sequences and development of the concepts and principles of sequence stratigraphy owe their beginnings and formulation to many people. Since 1930, those working with Pennsylvanian strata have examined Weller's (1930) 'cyclothems' and Moore's (1958) 'megacyclothems' and argued various origins for them ranging from local tectonism on the one hand to eustatic changes in sea level on the other hand and included numerous other origins based on climatic changes and orogenesis. The idea that systemic boundaries were the result of worldwide orogenies had been discredited and laid to rest only a few years earlier in the late 1920's, so ideas that worldwide sea level might have had repeated major fluctuations wasn't greeted by many geologists with a great amount of enthusiasm.

In the early 1970's, when Ramsbottom (1973) extended the concept of cyclic deposition to explain the depositional facies and stratigraphic relationships in Lower Carboniferous strata, he was met with much skepticism because he related the causes to changes in eustacy. On a much broader and longer time scale, Sloss (1963) identified a number of depositional packages, or sequences, of the magnitude of a geological system or more in size which were separated by times of generally lower sea level.

During the 1960's and 1970's, a new approach to stratigraphy incorporated seismic profiles of high resolution in both deposition strike and dip orientation across the Cenzoic and Cretaceous passive shelves, shelf margins, slopes, and basins. Exxon Production Research Company and many other petroleum companies expended considerable effort in analyzing this new data. In 1977, Exxon released its information in a compendium of papers by Vail, Mitchum, Thompson and many others (Payton, ed., 1977). Most of the early basic concepts in this study of stratigraphy from seismic profiles stressed geometric relations and physical distribution of units as they might be identified and interpreted from modern high resolution seismic profiles. Because similar seismic stratigraphic patterns and depositional packages, or sequences, were found in many late Mesozoic and Cenozoic passive shelves, the inference was made that these depositional patterns were caused by repeated worldwide changes which most likely were the result of changes in sea level. In some examples, paleontological zonation supported these inferences of contemporaneity of these depositional sequences and demonstrated changes in water depths which supported the idea of a series of changes in sea level. In other examples, the paleontological zonation remained to be done. Much of the skepticism about eustatic changes has related to demonstrating the contemporaneity of depositional sequences in widely separated areas.

The development of highly refined planktonic zonations based on several fossil groups also was taking place during the 1960's and 1970's and was largely the outcome of the deep sea drilling programs. This biostratigraphy was closely correlated with radiometric and paleomagnetic time scales. Refinements in the zonation and correlation of late Paleozoic strata also have occurred and made possible correlations of units having about the same magnitude (1 to 3 million years) as the planktonic zonation of the Cretaceous and Cenozoic or the ammonoid zonation of the Jurassic and Triassic.

As recently as the 1985 SEPM Research Conference on Sea Level Fluctuations, it was obvious that additional detailed biostratigraphic correlation of seismicly identified depositional sequences was possible and needed in order to trace these units from the outcrops, across the shelves, shelf margins, slopes, and into the ocean basins. Independently two symposia were organized that explored the utility of biostratigraphy in identifying and correlating depositional sequences. The papers in this volume were selected from these two symposia. One symposium was held by the North American Micropaleontology Section (NAMS) of the Society of Economic Paleontologists and Mineralogists (SEPM) in Raleigh, North Carolina, in September, 1986. The other was held by the Cushman Foundation for Foraminiferal Research in November, 1986, in San Antonio. The organizers and speakers of both symposia agreed that publication of papers from these two symposia would be valuable and timely. The Cushman Foundation agreed to sponsor and edit the publication.

The papers are arranged alphabetically by author and are reproduced from author-prepared camera-ready copy. In the pocket at the back are enlarged versions of the sea level cycle charts for the geological systems for the last 360 million years (Mississippian to Recent).

The editors thank the contributors for their cooperation, enthusiasm, and care in preparing their papers. It has been a pleasure to work closely with R. K. Olson and S. W. Wise, Jr., who organized the NAMS symposium and who helped in the initial work of selecting papers for presentation as a volume. We appreciate the help of B. Kohl, E. B. Picou, Jr., and G. A. Seiglie who reviewed several of the manuscripts. And we thank Chevron U.S.A., Inc. for permitting us to engage in assembling and editing this volume on behalf of the Cushman Foundation.

Charles A. Ross and Drew Haman
Houston, Texas
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PALEONTOLOGIC AND GEOPHYSICAL CORRELATIONS IN BAFFIN BAY AND THE LABRADOR SEA: ODP LEG 105

J.V. FIRTH[1], AND ODP LEG 105 SHIPBOARD SCIENTIFIC STAFF (S. SRIVISTAVA, M. ARTHUR, B. CLEMENT, A. AKSU, J. BALDAUF, G. BOHRMANN, W. BUSH, T. CEDERBERG, M. CREMER, K. DADY, A. DEVERNAL, F. HALL, M. HEAD, R. HISCOTT, R. JARRARD, M. KAMINSKI, D. LAZARUS, A. MONJANEL, O. NIELSEN, R. STEIN, F. THIEBAULT, J. ZACHOS, AND H. ZIMMERMAN)

[1]Department of Geology, Florida State University,
Tallahassee, Florida 32306

ABSTRACT

ODP Leg 105 drilled three sites (645, 646, 647) in Baffin Bay and the Labrador Sea. Paleontologic correlations with regional seismic reflectors established that glacial ice-rafting in the region started about 2.5 Ma (possibly as early as 3.4 Ma in Baffin Bay). Southward directed bottom water circulation in Baffin Bay started in early to mid-Miocene, while major drift sedimentation at Site 646 (Labrador Sea) started in the late Miocene. The opening of Baffin Bay occurred between 55 and 36 Ma. Oceanic crust at Site 647 (Labrador Sea) was dated as early Eocene in age (55-56 Ma, Chron 24).

INTRODUCTION

ODP Leg 105 drilled three sites along a north-south transect from Baffin Bay to the Labrador Sea during September and October of 1985. These two small ocean basins are of geologic interest because of their confinement between two continental crusts (Greenland and Canada), and because of their high latitude position. They presently serve as a conduit for water mass exchange between the Arctic and North Atlantic Oceans, and may have done so since as early as the late Cretaceous. The goals of Leg 105 were to study the early tectonic history of the region, as well as to unravel the paleoclimatic and paleoceanographic conditions through the Cenozoic.

This paper summarizes the paleontologic age constraints placed on several regional seismic reflectors which are important for reconstructing the geologic history of this region. The preliminary data were compiled by the shipboard scientific staff during and shortly after Leg 105 (Srivistava and others, in press), and are presented in Table 1. More detailed investigations are currently in preparation for Part B of the Proceedings of the Initial Reports, Ocean Drilling Program 105.

SITE 645

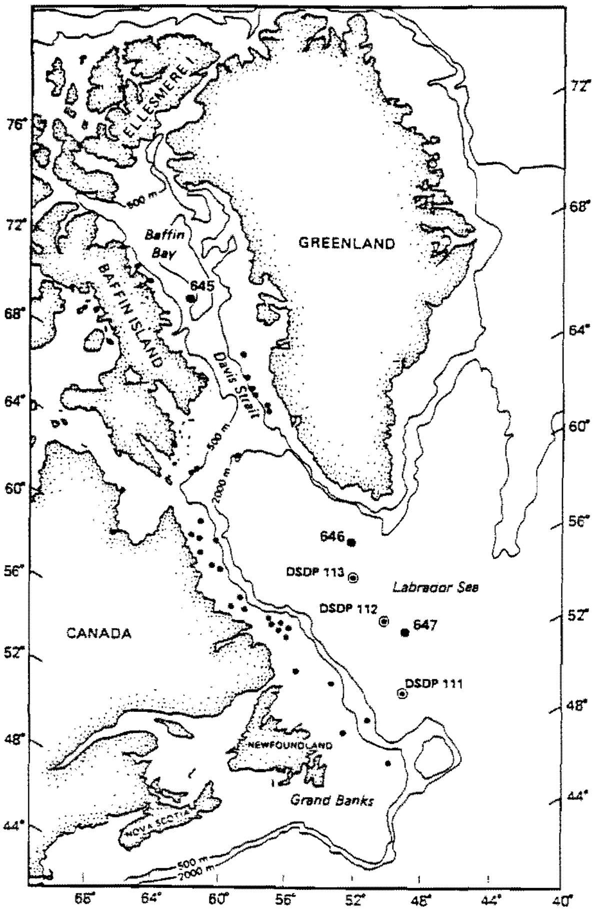

Fig. 1. Bathymetric map of the Labrador Sea and Baffin Bay showing Leg 105 sites (645, 646, and 647), plus DSDP sites and exploratory wells. (Map from the Ocean Drilling Program)

Site 645 (70°27.43′N, 64°39.3′W) was drilled on the continental slope off southern Baffin Island (Fig. 1). To the west of the site, multichannel seismic reflection profiles show down-faulted basement blocks covered by Cenozoic sediments (Fig. 2). The tectonic origin of Baffin Bay has been debated: Keen and Barrett (1972) proposed a model of foundering and thinning of continental crust to explain its origin, while Srivistava and others (1981) instead favored a model of seafloor spreading. One of the purposes of drilling at this site was to establish age control for seismic reflector horizons that pertain to the early tectonic history, as well as the history of deep water

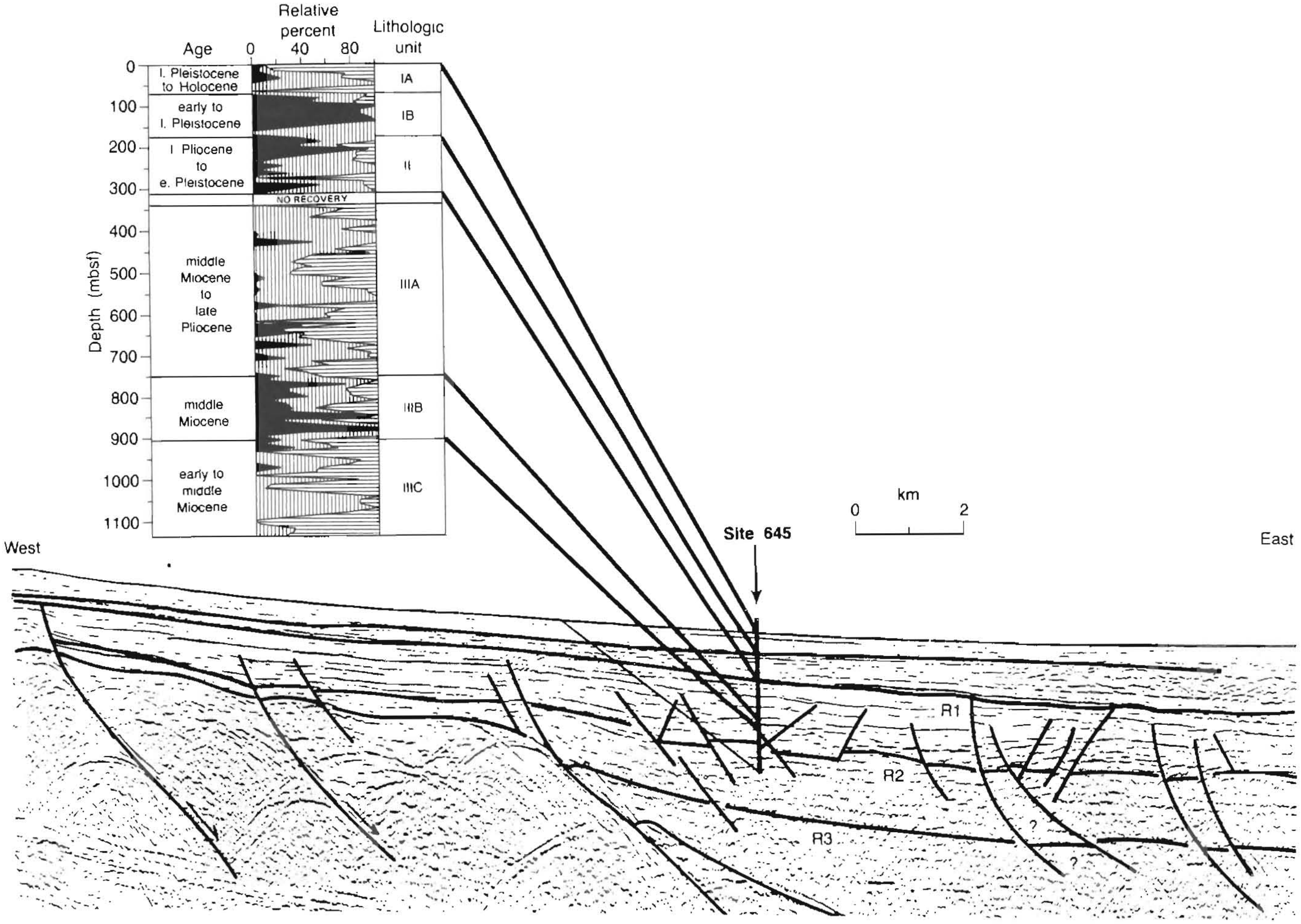

Fig. 2. Correlation of ages and lithologies to reflector horizons on seismic profile at Site 645. (Diagram from the Ocean Drilling Program)

circulation, of the bay. In all, seven holes were drilled, the deepest reaching to 1147.1 meters sub-bottom (msb).

Paleontologic data was unable to establish narrow age constraints on these reflectors because of the scarcity of diatoms, radiolarians, planktonic foraminifers and calcareous nannoplankton throughout the hole. Dinoflagellates and benthic foraminifers, however, were present through most of the section and enabled rough age estimates to be made on some seismic reflectors. Paleomagnetic data furnished additional tie-points with which to construct a sedimentation - rate curve.

Seismic reflector R1 (about 340 msb) lies somewhere around the lower/upper Pliocene boundary as determined by correlation with the sedimentation - rate curve (Fig. 3). The first abundant ice-rafted debris occur at this level, and the overlying sediments of Pliocene to Recent age are primarily of glacial origin. An age of about 2.5 Ma can be assigned to the first abundant ice-rafted debris, but occasional dropstones further down may indicate an age for the earliest ice-rafting in this region of about 3.4 Ma.

Below reflector R1, a change in depositional style is indicated. Sedimentologic and seismic features suggest that strata extending down to seismic reflector R2 were deposited by bottom contour currents. Rare incursions of warmer water North Atlantic planktonic foraminifers suggests that through most of this time interval, these currents were primarily southward directed, possibly derived from an Arctic source.

Reflector R2 (913 msb) is an erosional unconformity over part of the region, which underlies the package of bottom-current formed sediments. It is dated as mid to late Miocene in age, based primarily on dinoflagellates and benthic foraminifers (Table 1). The onset of southward directed bottom water circulation in Baffin Bay appears to have started at about this time.

Reflector R3 was not reached by drilling. This basin-wide reflector is thought by Srivistava and others (1981) to represent the approximate time of cessation of seafloor spreading in the region. The predicted age of this reflector was in the interval from the late Eocene to early Oligocene. An indirect age estimate of this reflector was made by extrapolating the sedimentation rate curve at the bottom of the hole down to the level of R3 (Fig.

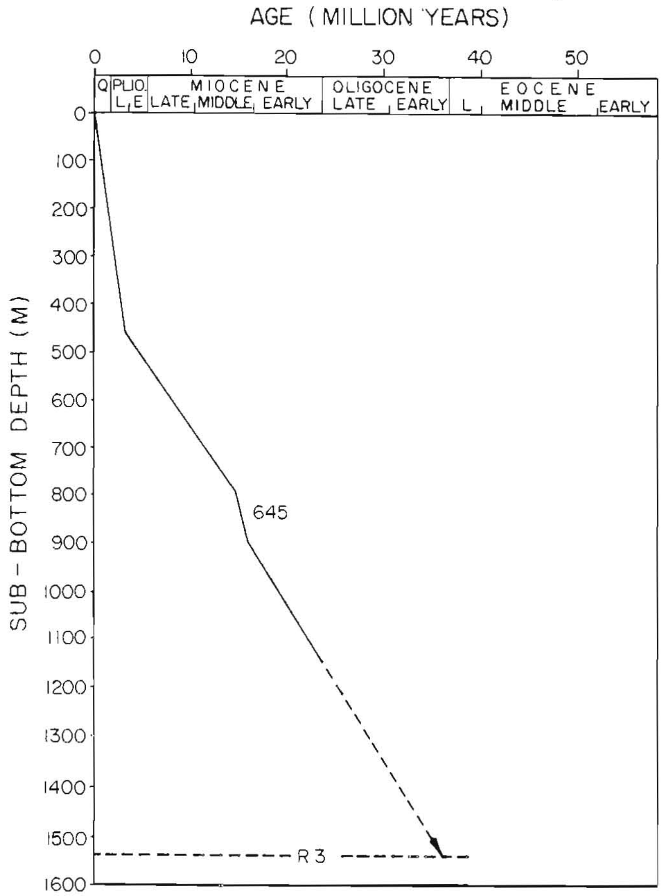

Fig. 3. Sedimentation rate curve for Site 645, showing extrapolation from bottom of hole to reflector R3. (Diagram modified from the Ocean Drilling Program)

3). The resulting age, approximately the Eocene-Oligocene boundary, agrees generally with the predicted age.

SITE 646

Site 646 (58°12.6′N, 48°22.1′W) in the southern Labrador Sea is located to the north of Eirik Ridge, a drift deposit situated in the path of the Norwegian Sea Overflow Water coming through the Denmark strait (Fig. 4). Drilling at this site was expected to provide data on the cold bottom water circulation history, as well as the glacial history, of this region.

A 767 meter section of late Miocene to Recent strata was recovered, which has yielded abundant paleontologic and paleomagnetic data, in contrast to Site 645. Two major seismic reflectors, R2 and R3/R4, were penetrated at this site. Reflector R2 (about 500 msb) had a previously estimated age of late early Miocene. However, both calcareous nannofossil and planktonic foraminiferal data indicate an age for the reflector of about the Miocene/Pliocene boundary (Table 1). Reflector R2 corresponds to a change in the carbonate composition of the sediments: carbonate-rich sediments lie above, and sediments with a variable but generally lower carbonate content lie below.

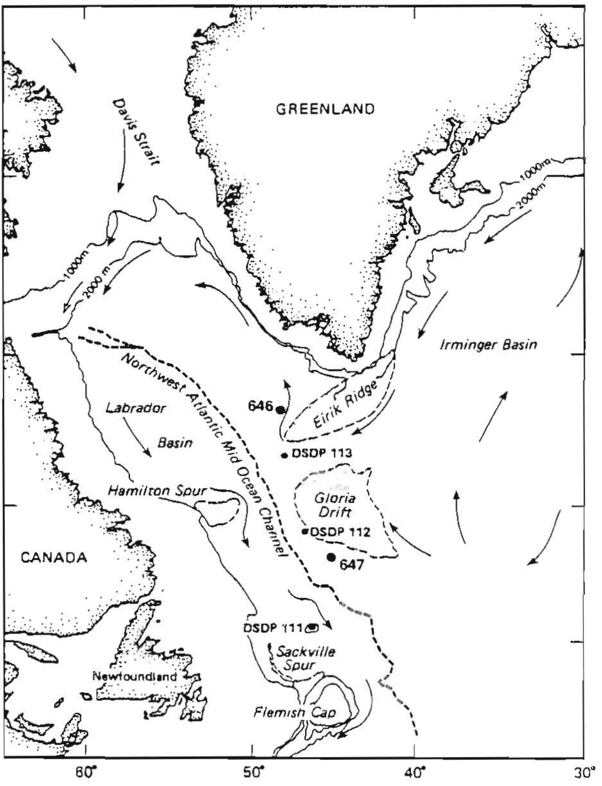

Fig. 4. Bathymetric map of the Labrador Sea, showing the locations of ODP Sites 646 and 647 and DSDP Sites 111, 112, and 113 relative to the Gloria Drift and Eirik Ridge drift deposits. (Map from the Ocean Drilling Program)

The second major reflector, R3/R4, was penetrated at a depth interval of from 680 - 730 msb. This reflector is actually a couplet of two closely spaced reflectors. As with reflector R2, the predicted age for R3/R4 was considerably overestimated. Miller and Tucholke (1983) considered it to be equivalent to a prominent eastern North Atlantic reflector, R4. The R4 reflector marks a period of intensified erosion of the seafloor in the late Eocene to early Oligocene. Both the planktonic foraminifers and the calcareous nannofossils, however, constrain the age of the bottom of Hole 646B, 40-90 meters below R3/R4, to be late Miocene (<10 Ma) in age (Table 1). The age of the reflector itself could not be further constrained based on the paleontologic data alone. A couple of magnetic reversals closely spaced at about 700 msb indicate the presence of a short 0.5 Ma hiatus between the base of Magnetic Chron 7 and the top of Chron 9, or roughly 7.5 Ma ago. No lithologic change is evident across this hiatus, though physical properties data indicate that the density of the sediments decrease below it. This hiatus, therefore, may correspond to the R3/R4 reflector. Seismic profiles show that R3/R4 marks the beginning of major drift sedimentation in the region (Fig. 5), thus indicating that intensified bottom water circulation started in the late Miocene at this site.

Site	Reflector/Depth	Age	Fossil Group / Zone / Other Data
645	R1 340 msb	early/late Pliocene boundary	Diatoms: within assemblage containing *Nitzschia cylindra* and *Porosira glacialis* (Baldauf, 1982; Koizumi, 1973) Other: correlation with sedimentation - rate curve
645	R2 913 msb	mid to late Miocene	Dinoflagellates: presence of *Nematosphaeropsis aquaducta* from 935 - 975 msb (Piasecki, 1980; Edwards, 1984) Benthic forams and incertae sedis: *Melonis zaandamae* (Berggren and Schnitker, 1983) and *Bolboforma metzmacheri* (Murray, 1984) from 772 - 885 msb
645	R3 1540 msb	Eoc./Oligocene boundary	Extrapolation of sedimentation rate curve from bottom of hole
646	R2 500 msb	Mio/Pliocene boundary	Calc. nannos: within NN12; FAD of *Ceratolithus* spp. (base of CN10b) at 465 msb Plank. forams: FAD of *Globorotalia margaritae* (Base of PL1) at 516 msb
646	R3/R4 680 - 730 msb	late Miocene	Calc. nannos: within CN9a - presence of *Discoaster berggrenii*, below FAD of *Amaurolithus primus* at 574 msb Plank. forams: within M11-M12
647	R2 116 msb	hiatus - about 2.5 to 5.5 Ma	Calc. nannos: NN16 directly above (presence of *Discoaster surculus*, absence of *Reticulofenestra pseudoumbilica*); NN11 directly below (presence of *Discoaster quinqueramus*)
647	R4 240 msb	early Oligocene	Calc. nannos: within NP22 (between LAD's of*Reticulofenestra umbilica* and *Ericsonia formosa*) Diatoms: within *Cestodiscus reticulatus* Zone (Fenner, 1984)
647	Basement 700 msb	early Eocene	Calc. nannos: within NP11 (presence of *Tribrachiatus orthostylus*, below FAD of *Discoaster lodoensis* at 675 msb) Dinoflagellates: FAD of *Dracodinium condylus* (Costa and Downie, 1976) at 675 msb.

*NP and NN Zones from Martini (1971); CN Zones from Okada and Bukry (1980); M and PL Zones from Berggren (1977, 1983)

Table 1. Summary of major seismic reflectors intersected at Sites 645, 646 and 647, their ages, and the fossils used to constrain the ages.

SITE 647

Site 647 (53°19.9'N, 45°15.7'W) is located south of the Gloria Drift deposit in the southern Labrador Sea (Fig. 4). Because basement magnetic anomalies are well developed in this region, a major objective of this site was to obtain a direct age correlation with the anomalies by recovery of drilled basement material. The overlying Neogene deposits are fairly thin (120 m) and allowed a quick penetration into the underlying Paleogene sediments and basement.

The first prominent reflector intersected was R2, at 116 msb (Fig. 6). This reflector corresponds to a 3 Ma hiatus separating upper Pliocene from upper Miocene sediments (Table 1). The sediments directly above this hiatus contain the first ice-rafted debris, corresponding in age to ca. 2.5 Ma. This age for the initiation of glacial ice-rafting matches that for Site 646 as well as for other North Atlantic DSDP Sites.

The Miocene interval is represented by a very condensed section, with at least one hiatus present. Calcareous nannofossil rich upper Miocene sediments are separated from siliceous rich sediments, containing lower Miocene radiolarians, by only a few meters of barren sediment in which are found iron-magnesium and

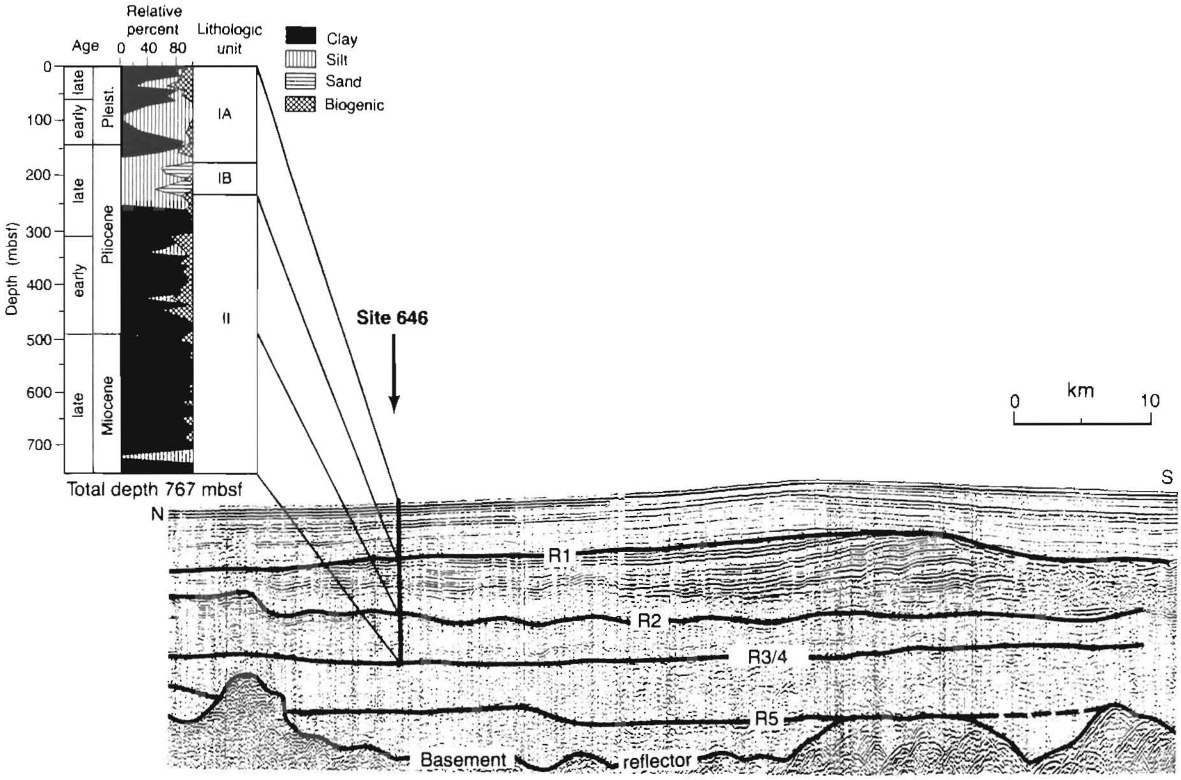

Fig. 5. Correlation of ages and lithologies to reflector horizons on seismic profile at Site 646. (Diagram from the Ocean Drilling Program)

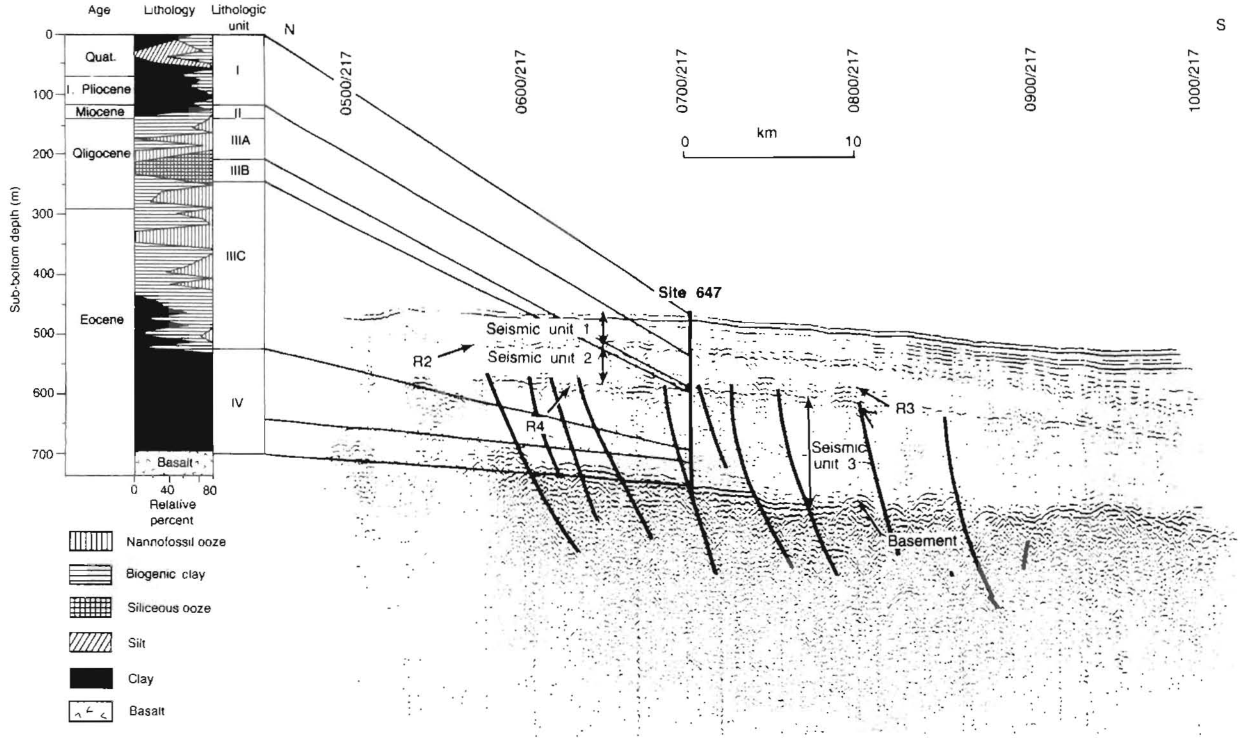

Fig. 6. Correlation of ages and lithologies to reflector horizons on seismic profile at Site 647. (Diagram from the Ocean Drilling Program)

phosphate nodules and banding.

A continuous, high sedimentation-rate sequence of late Eocene through early late Oligocene age contains one major reflector, R4, at about 240 msb (Fig. 6). This reflector was also identified at the nearby DSDP Site 112 (Laughton and others, 1972). It is of early Oligocene age (Table 1), and corresponds to a lithologic change from siliceous rich biogenic sediments above to calcareous rich biogenic sediments below.

Basement was reached at about 700 msb. According to Srivistava and others (1981) this site occurs within magnetic anomaly zone 24, which corresponds to an age of between 55-59 Ma. Fossiliferous sediments immediately overlying the basement contained calcareous nannofossils of early Eocene age (NP11). Dinoflagellates recovered from slightly above this level corroborate this age call, and both microfossil groups restrict the age for the oldest sediments to be from 55-56 Ma (Table 1). This microfossil age agrees with the predicted age based on magnetic anomaly correlations, and supports the tectonic history of the Labrador Sea as proposed by Srivistava and others (1981).

CONCLUSIONS

Important correlations between paleontologic and geophysical (seismic and paleomagnetic) data recovered from ODP Leg 105 yielded the following results on the timing of tectonic, paleoceanographic, and paleoclimatic events in Baffin Bay and the Labrador Sea:

Baffin Bay: first abundant ice-rafting approximately 2.5 Ma; onset of southward directed bottom water circulation in the early to mid-Miocene; time of cessation of seafloor spreading approximately Eocene-Oligocene boundary.

Labrador Sea: the first ice-rafting approximately 2.5 Ma (Sites 646 and 647); beginning of major drift sedimentation in the Eirik Ridge region (Site 646) in the late Miocene; dating of sediments directly above basement magnetic anomaly zone 24 at 55-56 Ma.

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THE NEW CHRONOSTRATIGRAPHIC BASIS OF CENOZOIC AND MESOZOIC SEA LEVEL CYCLES

BILAL U. HAQ, JAN HARDENBOL
Exxon Production Research Company,
Houston, Texas 77252-2189
and
PETER R. VAIL
Rice University, Houston, Texas 77251

A reliable chronostratigraphy is a prerequisite for accurate correlations of sea level events around the world. The chronostratigraphic framework of the new Mesozoic and Cenozoic eustatic cycle charts is based on an integration of geochronologic, and magneto- and biostratigraphic data. The resultant time scales have been in turn reconciled with the cycles of sea level change identified in sedimentary sections of continental margin and interior basins in different parts of the world.

The documentation of sea level events has been facilitated by recent developments in sequence stratigraphy that provides the framework within which depositional history of a margin, in response to changing sea levels can be interpreted. In previous cycle charts, sea level changes were interpreted exclusively from patterns of stratal geometries in seismic sections, with paleontological control from well data. The sequence-stratigraphic concepts have led to a better understanding of genetically-related depositional patterns produced during the sea level cycle (from a sea level fall, to rise, to the subsequent fall). These depositional packages can be identified in outcrop sections around the world for the documentation of sea level events. Public-domain reference sections can be designated for various sea level events, to produce well-dated and well-documented cycle charts.

The chronostratigraphic basis and the procedure followed to integrate various types of stratigraphic data that resulted in the cycle charts has already been discussed in detail elsewhere (Haq, Hardenbol and Vail, 1987a and b) and is not repeated here. Papers on various theoretical and applied aspects of sequence stratigraphy are presented by Jervey and others (1987), Posamentier and others (1987), Sarg (1987) and Vail and others (1987).

At the request of the convenors of the symposium on the "Timing and Depositional History of Eustatic Sequences: Constraints on Seismic Stratigraphy", here we include four cycle charts (Figs. 1-4 & in pocket) that comprise the history of sea level change over the past 250 million years. These cycle charts reconcile the linear time scale with the Mesozoic and Cenozoic magnetochronostratigraphy, standard chronostratigraphy, biochronostratigraphy, sequence chronostratigraphy, and eustatic cycles.

Detailed sources for magnetostratigraphy and magnetobiostratigraphic correlations are included in Haq and others (1987b). Citations included on the cycle charts (Figs. 1-4) are listed in the references below.

The first section on the cycle charts summarizes magnetochronostratigraphy. Both sea floor magnetic anomaly data, where available, and polarity reversal patterns are shown. Pre-Callovian polarity reversal framework (gray and white stripes on Figs. 3 and 4) is a synthesis of the available paleomagnetic data from outcrop sections, and may be subject to modification when additional data becomes available.

The second section on the cycle charts includes standard chronostratigraphic subdivisions (western European stages are accepted as standard for global correlations). The stratigraphic position of most of the stage boundaries (on the Cenozoic cycle chart, indicated by the "extent of the stratotype", see Fig. 1), has been ascertained through sequence-stratigraphic studies of the stratotype or neostratotype sections. This permits more accurate positioning of stage boundaries than that based on paleontologic data alone (see Haq and others, 1987b, for examples). Some of the informal, but commonly used suprastage designations of the Mesozoic are also included in this section of the cycle charts.

The third section on the cycle charts comprises biochronostratigraphy. In each cycle chart the fossil groups that are most useful for the subdivision of that interval are included. This comprises major marine microplankton groups for the Cenozoic. For the Mesozoic only two microfossil groups could be included in each chart (Figs. 2-4), in addition to palynomorph biohorizons and ammonite zones. Palynomorph datum events (mostly dinoflagellates) that are included on all charts have largely been recorded in western European sections and represent the aggregate experience of the palynologists of Exxon production Research Company. The compilers of the palynological data are listed separately on each chart. When direct ties between bio- and magnetostratigraphic events are lacking, as is the case of most ammonite datum events, the zones within individual stages are assigned equal duration.

The fourth section includes sequence chronostratigraphy. This consists of the sequence chronozonal terminology and scaled relative changes of coastal onlap. The ages of sequence chronozone boundaries and downlap surfaces (in Ma) are also indicated, as are the boundaries of depositional systems tracts, i.e. Lowstand and Shelf Margin Wedges (LSW, SMW), and Transgressive (TR) and Highstand (HS) deposits. Sequence boundaries where lowstand fans have been observed are indicated by an 'F' in the systems tracts' column. Unshaded triangular areas within the coastal onlap cycles depict condensed sections, that represent intervals of slow deposition on the shelf, following rapid sea level rise. The relative duration of the condensed section increases basinward. Major, medium and minor sequence boundaries are identified by the relative thickness of the lines drawn through them (see legend under each chart). The dashed lines drawn through the condensed sections ("downlap surfaces" on seismic lines) represent the surfaces of maximum flooding, but their relative thickness indicate the major, medium, or minor magnitude of the condensed sections.

Generally, only the sequence boundaries of major and medium magnitude can be identified at the regional seismic level. Detailed well-log and/or outcrop studies are necessary to discern minor sequences.

The last section contains the eustatic cycles, which include the long- and short-term sea level curves and the cycle nomenclature, which is identical to the sequence-chronozonal terminology. The scale at the top (in meters) represents the best global-average estimates of sea level rises and falls, as compared with the modern global sea level mean.

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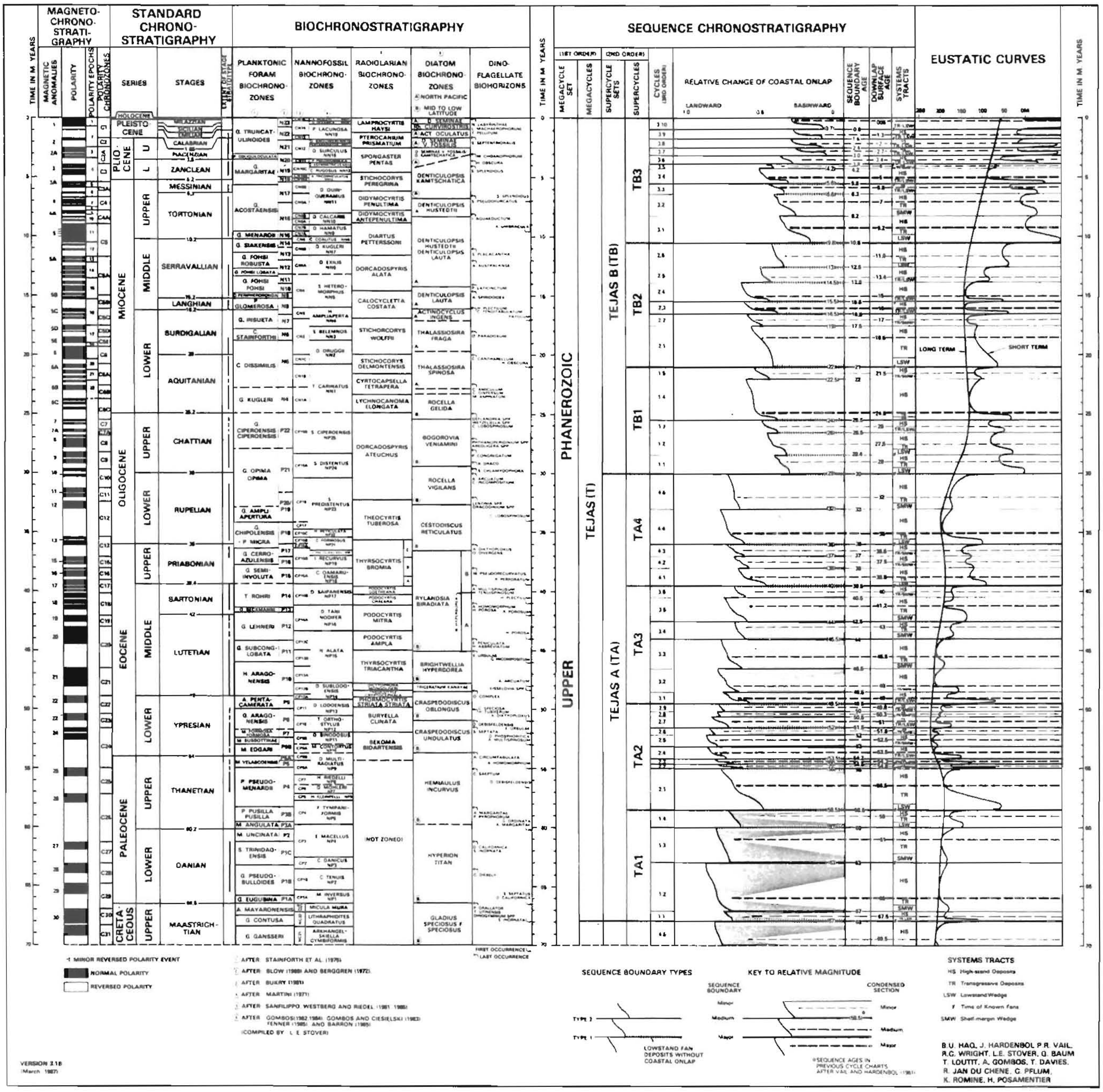

Figure 1. Cenozoic chronostratigraphy and cycles of sea level change. Linear time scale is in millions of years before present. Collaborators for the Cenozoic cycle chart are listed in the lower right corner of the figure. For sources see text and references in Haq and others (1987b).

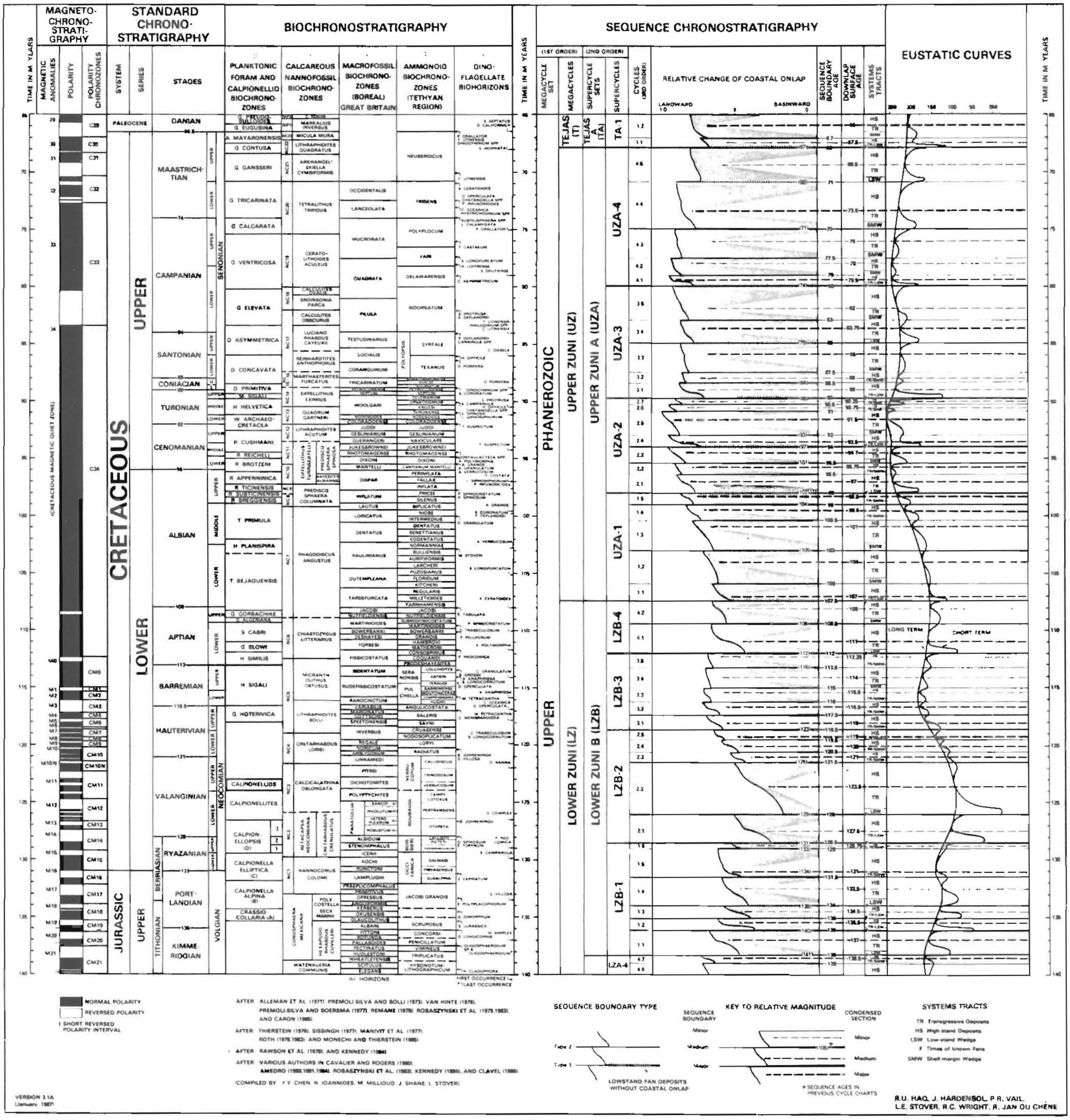

Figure 2. Cretaceous chronostratigraphy and cycles of sea level change. Linear time scale is in millions of years before present. Collaborators for the Cretaceous cycle chart are listed in the lower right corner of the figure. For sources see text and references in Haq and others (1987b). Slightly modified after Haq and others (1987a).

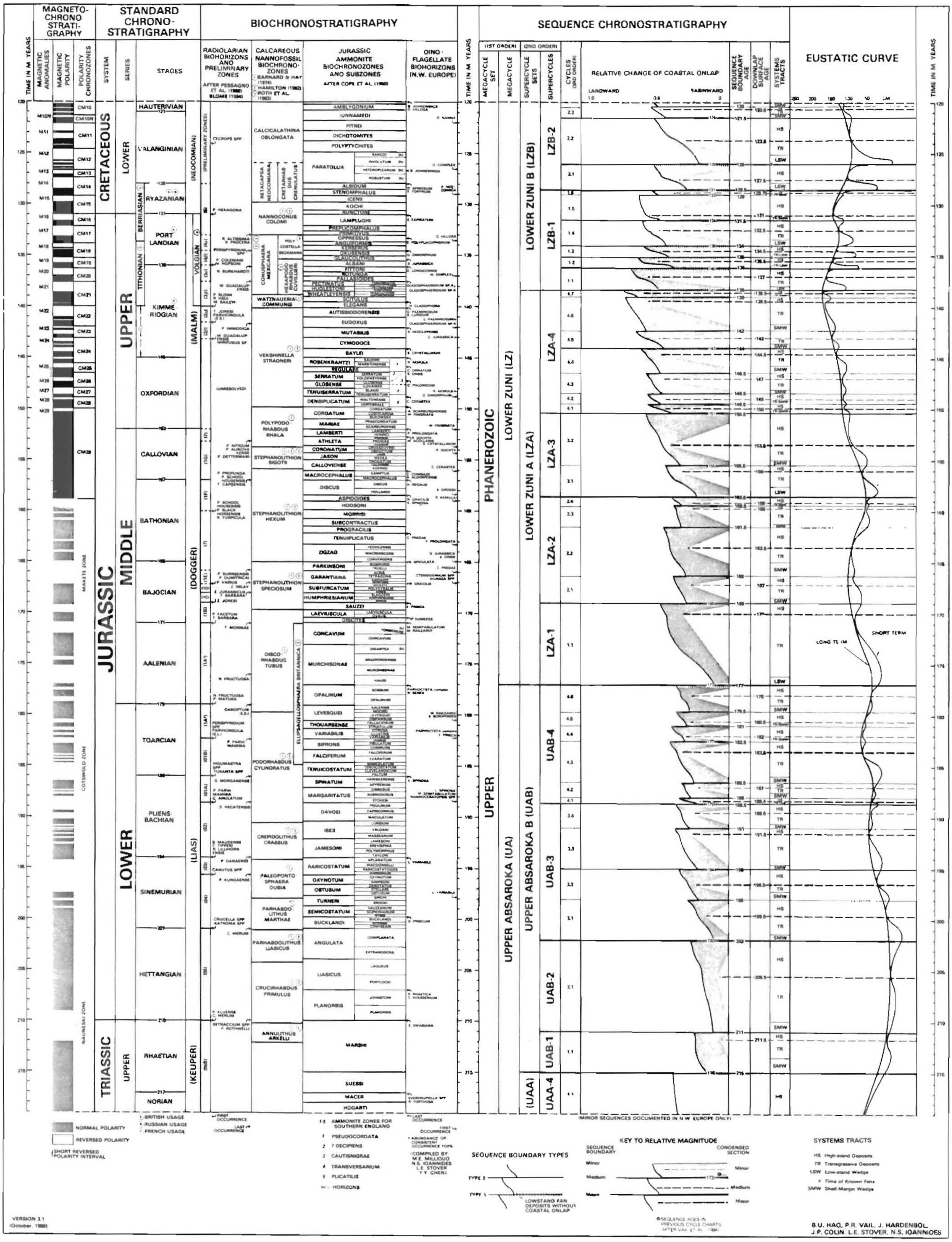

Figure 3. Jurassic chronostratigraphy and cycles of sea level change. Linear time scale is in millions of years before present. Collaborators for the Jurassic cycle chart are listed in the lower right corner of the figure. For sources see text and references in Haq and others (1987b). The pre-Callovian magnetic polarity reversal model is tentative and may be subject to change.

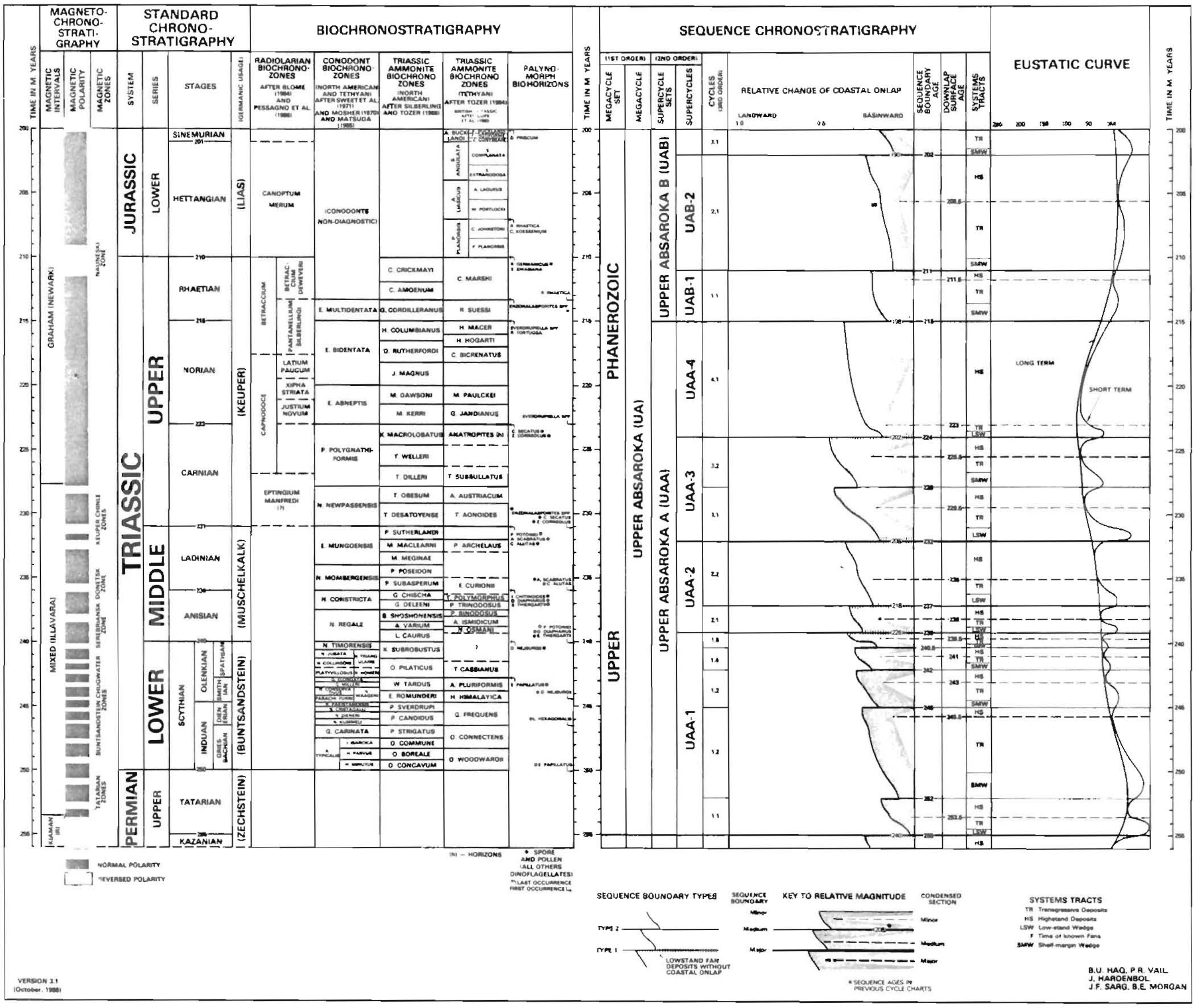

Figure 4. Triassic chronostratigraphy and cycles of sea level change. Linear time scale is in millions of years before present. Collaborators for the Triassic cycle chart are listed in the lower right corner of the figure. For sources see text and references in Haq and others (1987b). The Triassic magnetic polarity reversal model is synthesized from available data and may be subject to change in future as new data become available.

Cushman Foundation for Foraminiferal Research, Special Publication 24, 1987.

MIDDLE CENOZOIC UNCONFORMITY-BOUNDED STRATIGRAPHIC UNITS IN THE CUYAMA AND SOUTHERN SAN JOAQUIN BASINS, CALIFORNIA: RECORD OF EUSTATIC AND TECTONIC EVENTS IN ACTIVE MARGIN BASINS

MARTIN B. LAGOE
Department of Geological Sciences
The University of Texas at Austin
Austin, Texas 78713

ABSTRACT

An examination of the depositional history of two middle Cenozoic active margin basins in central California provides insights into the relative importance of tectonics and eustacy for controlling large scale stratigraphic relationships. Both the Cuyama and southernmost San Joaquin Basin contain prominent middle Cenozoic unconformity-bounded stratigraphic intervals.

The Cuyama Basin exhibits two distinct cycles of basin subsidence and filling--one (late Oligocene/early Miocene) associated with the Vaqueros Formation and the other (early Miocene/late Miocene) associated with the Monterey Formation. The San Emigdio area of the southern San Joaquin Basin exhibits only one major cycle of basin subsidence (late Oligocene/late Miocene) associated with the Temblor and Monterey Formations.

An analysis of the nature and timing of several depositional/stratigraphic events is used to compare basin history in the two areas. These events include distribution of major unconformities, rapid bathymetric deepenings, periods of peak transgression, major shallow marine progradational events, episodes of submarine fan development, changes in foraminiferal biofacies and volcanic rocks.

The timing of these events is not always very well constrained but indicates that both eustacy and tectonics play important roles in shaping basin stratigraphy. Several relationships suggest that tectonics is the relatively more important factor in the two basins studied.

INTRODUCTION

Seismic stratigraphy currently enjoys wide popularity as a means of establishing the depositional geometry and history of marine basins. This approach to stratigraphic analysis concentrates on the definition of depositional sequences based on bounding regional basin-margin unconformities and their basinward disconformities or conformities (Vail and others, 1984). The geometric arrangement of these sequences and their internal geometry (e.g. coastal onlap, basinward downlap, shifts in the position of coastal onlap) are argued to be controlled by globally synchronous changes in sea level. Tectonic effects are not proposed to be an important factor in producing depositional geometries beyond the impact of thermal and/or sediment-loading-induced subsidence providing room for sediment accumulation (Vail and others, 1984). It should be noted that the vast majority of documentation for seismic stratigraphic concepts comes from passive continental margin settings (e.g. Haq and others, 1986) where such subsidence-dominated tectonics would predominate.

The methodology of seismic stratigraphy has provided a new approach to basin analysis that emphasizes the broad geometric relationships of unconformity-bounded depositional sequences. These principles are now being applied to purely outcrop-based stratigraphic studies as well (Haq and others, 1986). The presumed global eustatic control of these sequences and their bounding unconformities lends great predictive power to this approach, if the underlying assumptions are correct. Many stratigraphers now use the global eustatic cycle charts of Vail and co-workers (e.g. Haq and others, 1986) as a primary chronostratigraphic tool in which to date unconformities and sequences with a minimum of primary chronostratigraphic information.

Some concern has recently been expressed regarding seismic stratigraphy's overwhelming dependence on global eustacy for controlling depositional patterns in marine basins (e.g. Miall, 1986; Galloway, in press; Watts, 1982; Watts and Thorne, 1984). These workers suggest that tectonics, sediment supply and eustacy all play a role in the dispersal and accumulation of sediment in marine basins and that any one process may dominate. Thus very similar depositional geometries (sequences) may have a variety of causes. There is also some question regarding the selection of the most fundamental type of genetic sequence boundary for stratigraphic analysis. Seismic stratigraphy emphasizes regional bounding unconformities which enclose regional transgressive (high sea level) episodes (condensed intervals of Vail and others, 1984). Galloway (in press) argues persuasively that the regional transgressive events (intervals of maximum flooding of basin margins), which enclose regional progradational events, are more logical boundaries and define genetically related depositional sequences.

This debate is complicated further when considering the depositional history of basins along active continental margins (such as central California). Such active margin basins are commonly characterized by complicated tectonic histories, more so than many passive margin basins. These complications include rapid vertical (both up and down) and

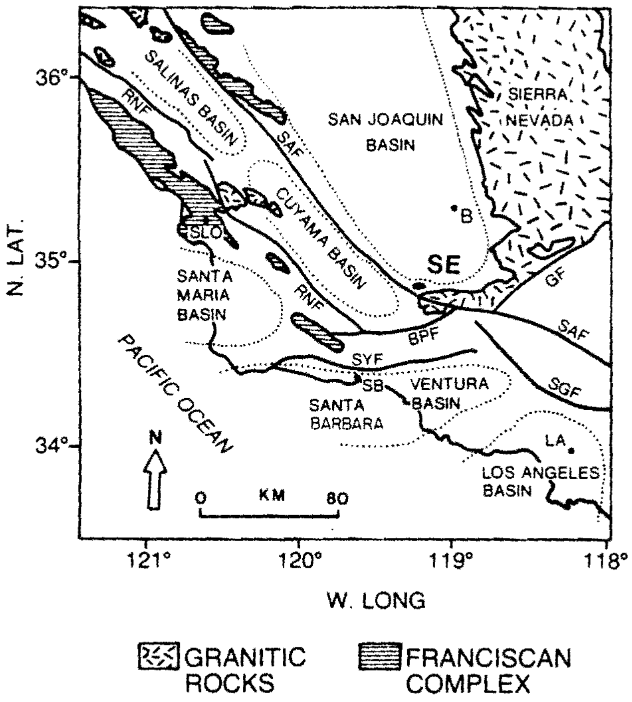

Figure 1. Major middle Cenozoic basins in central California. Areas examined in this study include the Cuyama Basin and the San Emigdio area of the southern San Joaquin Basin (Marked SE). Abbreviations: SAF-San Andreas fault; RNF-Rinconada/Nacimiento fault; GF-Garlock fault; BPF-Big Pine fault; SYF-Santa Ynez fault; SGF-San Gabriel fault; LA-Los Angeles; SB-Santa Barbara; SLO-San Luis Obispo; B-Bakersfield.

lateral movements and periodically intense seismic activity (by earthquakes, not geophysicists). These basins can also contain a variety of basement types which are juxtaposed over short lateral distances (see discussion of Cuyama Basin below). These varied conditions often produce basins with complex bathymetry, which can further influence depositional geometries.

Objectives

This study will examine two middle Cenozoic active margin basins in central California--the Cuyama and southernmost San Joaquin Basins (Fig. 1). The primary objective is an attempt to evaluate whether the principal tenets of seismic stratigraphy can fully explain the depositional characteristics of these basins. Although the two basins are now juxtaposed, removal of Pliocene-Pleistocene right-slip on the San Andreas fault would place them 200-300 km apart during the middle Cenozoic (Oligocene-Miocene). Specific objectives of the study are:

1. A brief description of the middle Cenozoic stratigraphy in the two basins and an evaluation of the importance of unconformities in establishing stratigraphic frameworks.
2. A description of the internal depositional and paleo-environmental architecture within unconformity-bounded stratigraphic intervals. This analysis will focus on paleobathymetric history, lithofacies, depositional environments, timing of events (e.g. periods of maximum transgression) and evidence of tectonic movements.
3. A brief evaluation of tectonic and eustatic controls on the depositional patterns recognized.
4. A comparison of depositional history and timing in the two basins.
5. An evaluation of eustacy as the predominant control of the depositional geometries and timing in the two active margin basins studied.

This analysis is not meant to be a thorough test of seismic stratigraphic principles in active margin settings but should prove useful in pointing out problems with applying these techniques wholesale, as they have been practiced in passive margin settings.

Geologic Setting

The Cuyama Basin

The Cuyama Basin is located in an area now occupied by the southern Coast Ranges of central California (Fig. 1, Fig. 2). This area includes the Sierra Madre, Caliente and La Panza Ranges, Cuyama Valley, Carrizo Plain and Cuyama Badlands. This area was the site of major late Cenozoic deformation, as evidenced by the prominent mountain ranges as well as several major through-going faults--the San Andreas fault to the northeast, the Rinconada-Nacimiento fault to the southwest and the Big Pine fault to the southeast (see Fig. 2 for other important faults).

This area contains a thick accumulation of Mesozoic and Cenozoic sedimentary rocks (Fig. 3). These rocks overlie several basement types, Precambrian metamorphic rocks between the San Andreas fault and San Juan-Chimineas-Morales fault trend, Mesozoic granite between the latter fault trend and the Rinconada-Nacimiento fault and Franciscan subduction complex south of the Rinconada-Nacimiento fault. Stratigraphic information is obtained from extensive outcrops in the mountain ranges and from numerous petroleum exploration and development wells drilled throughout the area.

The middle Cenozoic rocks of this area consist of the Simmler, Vaqueros, Monterey, Santa Margarita, Caliente and Quatal Formations. An extensive stratigraphic database is available for this area, most recently summarized by Lagoe (1981, 1982, 1984, 1985, 1986, 1987).

Southern San Joaquin Basin--San Emigdio Area

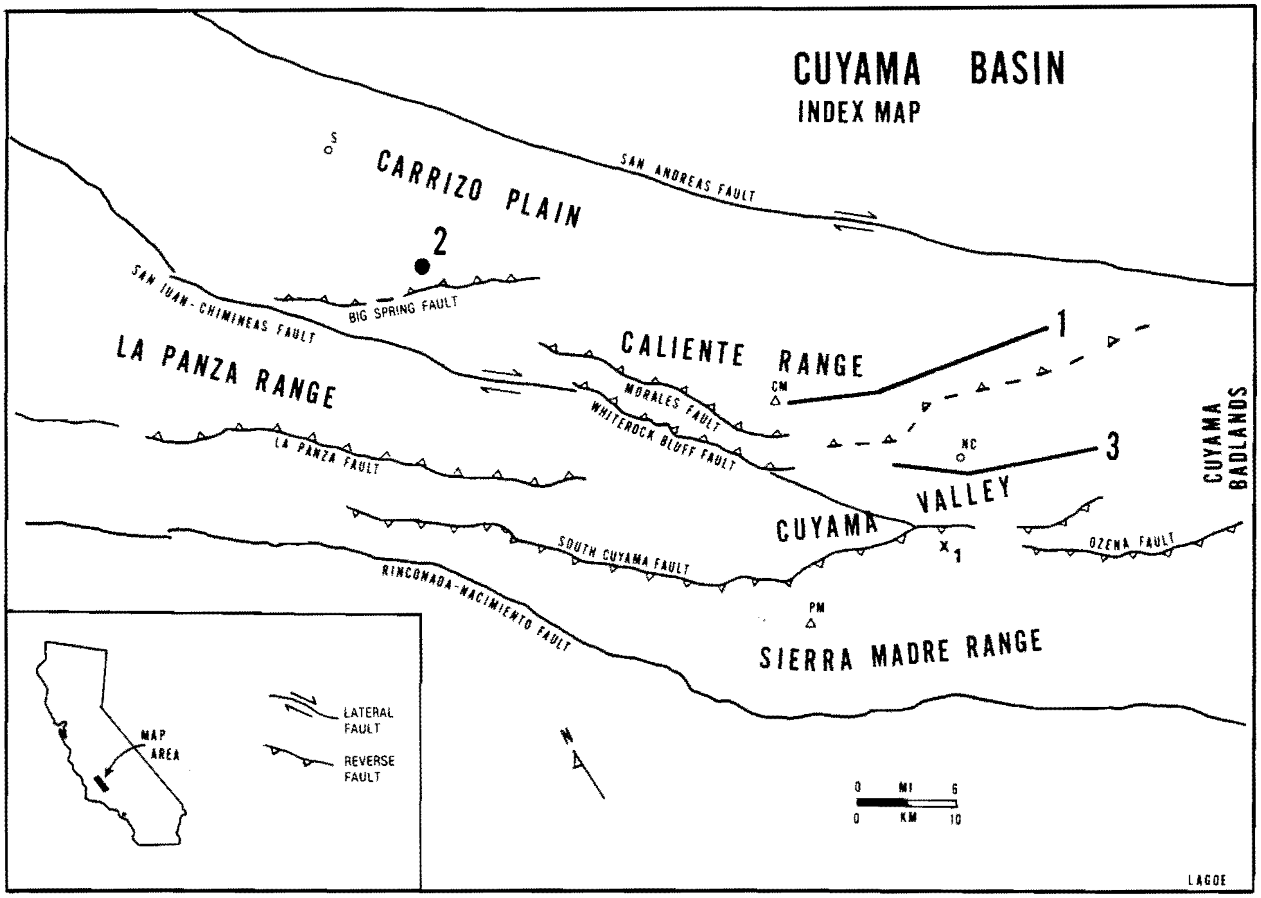

Figure 2. Index map for the Cuyama Basin showing the location of major geographic features and faults. Abbreviations: PM--Peak Mountain; NC--New Cuyama; CM--Caliente Mountain; S--Simmler. Also show are 1--location of stratigraphic section from southeastern Caliente Range; 2--location of type area of Soda Lake Shale Member of the Vaqueros Formation and 3--location of subsurface stratigraphic section from beneath Cuyama Valley (modified from Lagoe, 1984).

Due east of the Cuyama Basin, across the San Andreas fault, is the southern end of the San Joaquin Basin (Fig. 4). As mentioned previously, these two basins, now adjacent to one another, were hundreds of kilometers removed during middle Cenozoic deposition. The southern San Joaquin basin occupies an area now composed of the San Emigdio and western Tehachepi Ranges and the southernmost portions of the San Joaquin Valley. This area was also deformed in the late Cenozoic along the trend of the Pleito Thrust system (Fig. 4; Davis, 1986, Davis and Lagoe, 1984). Middle Cenozoic rocks are prominently exposed in the mountain ranges and, as with the Cuyama Basin, numerous petroleum exploration wells provide subsurface control. An abundant stratigraphic database in this area has recently been summarized by Nilsen (1973, 1978, in press), Nilsen and others (1973), DeCelles (1986, 1987) and Lagoe (1986, in press a) and is utilized in this study.

Basic Approach

The analysis will begin by examining the Cuyama Basin, its middle Cenozoic stratigraphic framework, major unconformities and depositional history. A comparison then is made with coeval rocks in the southern San Joaquin Basin. The relationships are then evaluated in light of the principles of seismic stratigraphy (or sequence stratigraphy as per Vail and others, 1984).

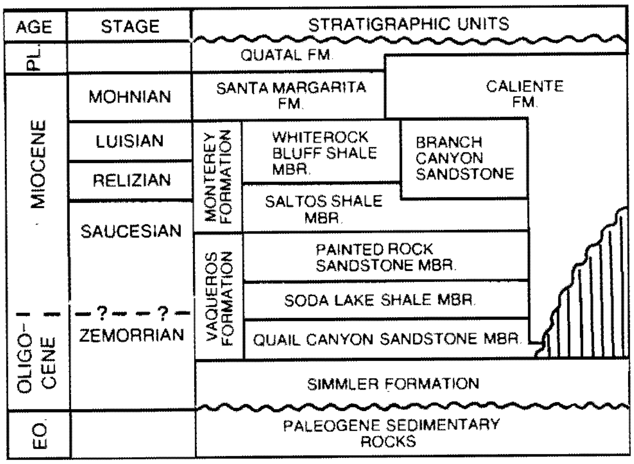

Figure 3. Summary stratigraphy of the middle Cenozoic rocks in the Cuyama Basin. (modified from Lagoe, 1984).

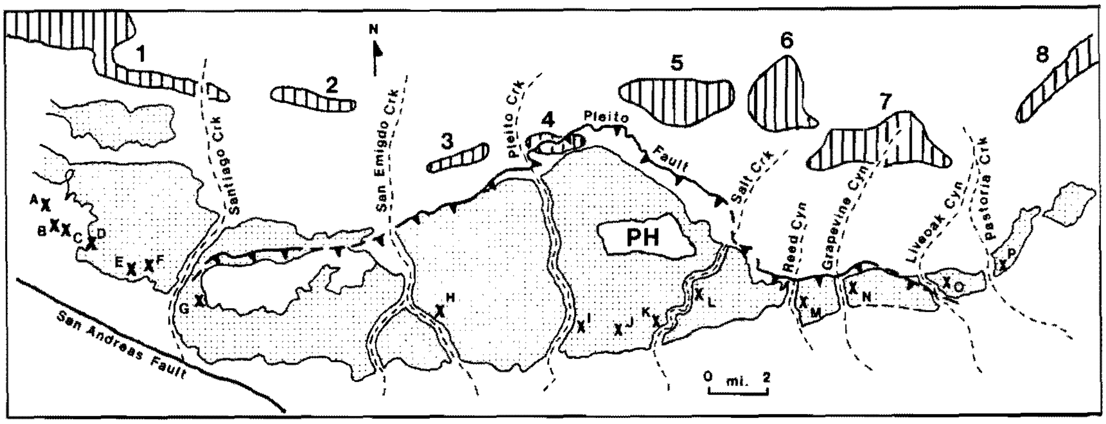

Figure 4. Map of the western Tehachapi and San Emigdio Mountains showing the distribution of Paleogene rocks (stipple) and sources of data for the stratigraphic cross-section of this area (Fig. 12). Oil fields: 1—Midway-Sunset; 2—Los Lobos; 3—White Wolf; 4—Pleito; 5—Wheeler Ridge; 6—North Tejon; 7—Tejon; 8—Tejon Hills. Outcrop sections or well sections: A—Gulf Time Oil S.K. #1; B—Anderson Klipstein A-21; C—Colgrove Klipstein 52; D—Whitehouse and Wilcox McMurray 46; E—Magee Hudson #1; F—Richfield Ramsey #1; G—Upper Twin Creek section; H—Devil's Kitchen Syncline section; I—Pleito Creek section; J—Pleito creek/Salt Creek Divide section; K—Salt Creek section; L—Richfield San Emigdio C-1; M—Reed Canyon section; N—Grapevine Canyon; O—Liveoak Canyon section; P—Pastoria Creek section. Also shown is location of Pleito Hills composite section—PH (Fig. 14). (modified from Lagoe, 1986).

The chronostratigraphic framework used in this study is a compilation from several studies that aim to calibrate provincial biozonations (largely based on benthic foraminifera and molluscs) to global chronostratigraphies (in this case Berggren and others, 1985) by using planktic biozonations and radiometrically-dated rocks. This compilation is documented in Lagoe (in press a). The most comprehensive chronostratigraphic studies useful to this work are those of Vedder (1973), Addicott and others (1978), Poore and others (1981), Barron (1986) and COSUNA (1984). Estimation of paleobathymetries from benthic foraminiferal biofacies were conducted according to methodologies discussed in Lagoe (1984, 1985) and Lagoe and McDougall (1986).

Acknowledgments

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THE CUYAMA BASIN

Stratigraphic Overview

The Cuyama Basin contains a thick accumulation of middle Cenozoic rocks that is bounded by major regional unconformities (Fig. 3). These rocks include the Simmler, Vaqueros, Monterey, Santa Margarita, Branch Canyon, Caliente and Quatal Formations. The middle Cenozoic rocks unconformably overlie Eocene marine sedimentary rocks or crystalline basement and are unconformably overlain by the late Cenozoic, nonmarine Morales Formation (Fig. 3).

A representative section of these middle Cenozoic rocks from beneath Cuyama Valley (Fig. 5) illustrates some important characteristics. Foremost among these is the realization that middle Cenozoic lithofacies and biofacies define two distinct episodes of basin subsidence and subsequent filling. The older of these is associated with the Vaqueros Formation and is called the Vaqueros basin cycle. The younger episode involves the Monterey, Branch Canyon, Santa Margarita and Caliente Formations and is named the Monterey basin cycle (Fig. 5).

The depositional history of middle Cenozoic rocks in the Cuyama Basin will be discussed with reference to these two episodes of basin subsidence and filling.

Vaqueros Basin Cycle

The lowermost rocks of the Vaqueros basin cycle vary within the basin. The nonmarine rocks of the Simmler Formation are the oldest middle Cenozoic rocks in the basin and are the basal beds in many places. Beneath Cuyama Valley (fig. 5) the Simmler Formation is missing and the lowermost, shallow marine rocks of the Vaqueros Formation (Quail Canyon Sandstone Member) are the initial unit of this basin cycle. These units rest unconformably on older rocks throughout the basin—either on Cretaceous to Eocene marine sedimentary rocks (in which case the unconformity is noticably

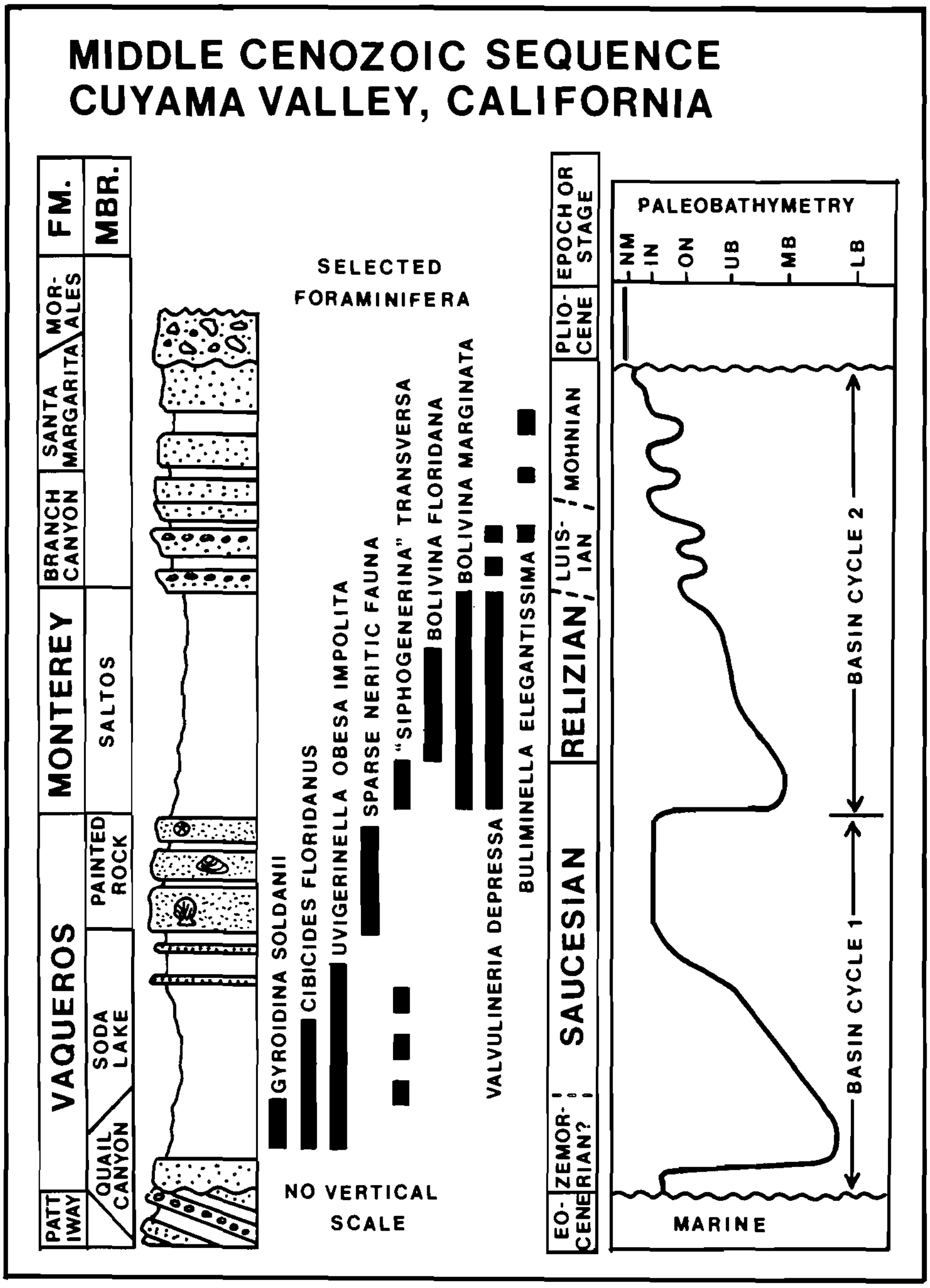

Figure 5. Stratigraphic summary of middle Cenozoic rocks from beneath Cuyama Valley. Note two distinct cycles of basin subsidence and filling: Basin cycle 1 = Vaqueros basin cycle; Basin cycle 2 = Monterey basin cycle. Paleobathymetric abbreviations: NM-nonmarine; IN-inner neritic; ON--outer neritic; UB--upper bathyal; MB--upper middle bathyal; LB--lower middle bathyal. (from Lagoe, 1987).

angular in most cases) or on crystalline basement. The age of the Simmler Formation is constrained by rare Arikareean vertebrate fossils (late Oligocene/early Miocene; Bartow, 1974; 1978; Blake, 1982) and by K-Ar whole rock ages of 23.4 +/- 0.8 and 22.9 +/- 0.7 Ma from basalts interbedded in this formation near Cuyama Gorge (Ballance and others, 1983). The immediately overlying basal Vaqueros Formation contains "Vaqueros" stage molluscs (middle Oligocene/early Miocene). The youngest rocks beneath the basal middle Cenozoic unconformity yielding definitive ages contain Narizian foraminiferal assemblages (middle Eocene; Lagoe, unpublished data). The above biostratigraphic and chronostratigraphic information indicates that the middle Cenozoic regional unconformity represents a minimum gap of ca. 8 Ma (32-40 Ma which equates to the gap between the base of the "Vaqueros" molluscan stage and the top of the middle Eocene). The gap could be, and probably is, substantially larger.

The major characteristics of the Vaqueros basin cycle are well illustrated by a section from the type area of the Soda Lake Shale Member of the Vaqueros Formation (Fig. 6). Following deposition of the nonmarine Simmler

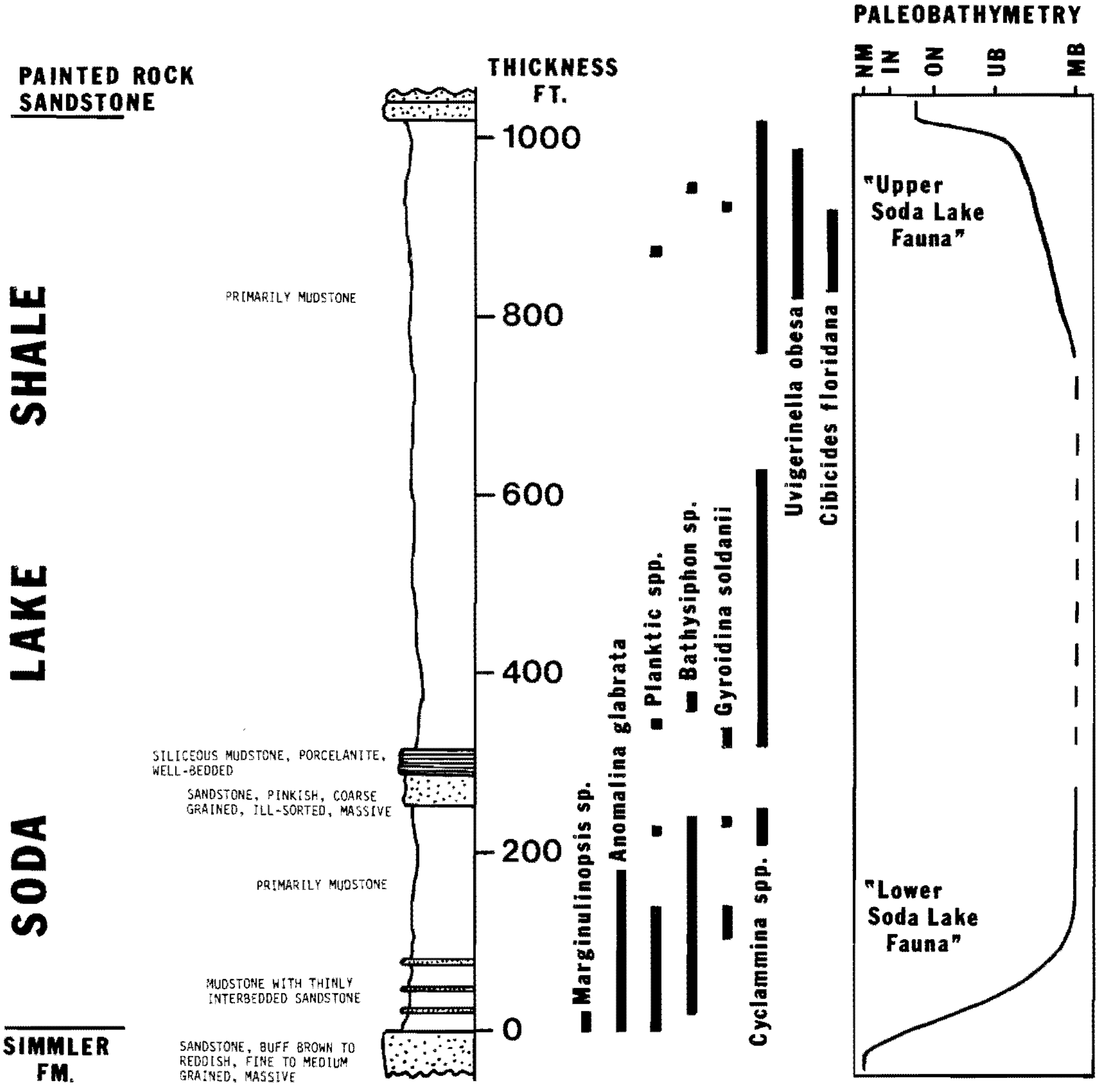

Figure 6. Type section of the Soda Lake Member of the Vaqueros Formation (see Fig. 2 for location). Paleobathymetric abbreviations: NM--nonmarine; IN--inner neritic; ON--outer neritic; UB--upper bathyal; MB--middle bathyal.

Formation the Cuyama Basin rapidly subsided in its central portions to lower middle bathyal water depths (ca. 1500-2000 m). The basal Soda Lake Shale Member of the Vaqueros contains a thin sandy interval at its base containing a neritic microfauna which is abruptly overlain by terrigenous mudstones and thin sandstones characterized by a Saucesian (early Miocene) lower middle bathyal benthic fauna with common planktic foraminifera (Fig. 6). At the top of this lower interval is a prominent accumulation of siliceous mudstone and porcelanite (Fig. 6). These rocks represent former siliceous oozes accumulated during a period of terrigenous sediment-starvation within the basin. The rest of the Soda Lake Shale Member overlying the siliceous interval is composed of mudstone containing a relatively low-diversity Saucesian microfauna commonly dominated by agglutinated species and Uvigerinella obesa s.l. (Fig. 6). These faunas, lacking planktic foraminifera, indicate upper to middle bathyal water depths and possible low-oxygen conditions. The Soda Lake Shale grades upwards into the neritic Painted Rock Sandstone Member of the Vaqueros Formation, which marks the top of the Vaqueros basin cycle.

The Painted Rock Sandstone Member is composed of up to 5500 ft. (1675 m) of interbedded shallow marine sandstone, siltstone and minor conglomerate. It contains "Vaqueros" stage megafossils in the Caliente Range (Vedder, 1973) but the upper part of the Painted Rock extends into the "Temblor" stage (early Miocene/middle Miocene) in other parts of the basin (Addicott and others, 1978). This great thickness of marine sediment, deposited within a narrow bathymetric range (ca. 0-150 m), reflects a delicate balance between sediment supply and subsidence (induced by sediment loading?).

A generalized cross-section from the southeastern Caliente Range illustrates some

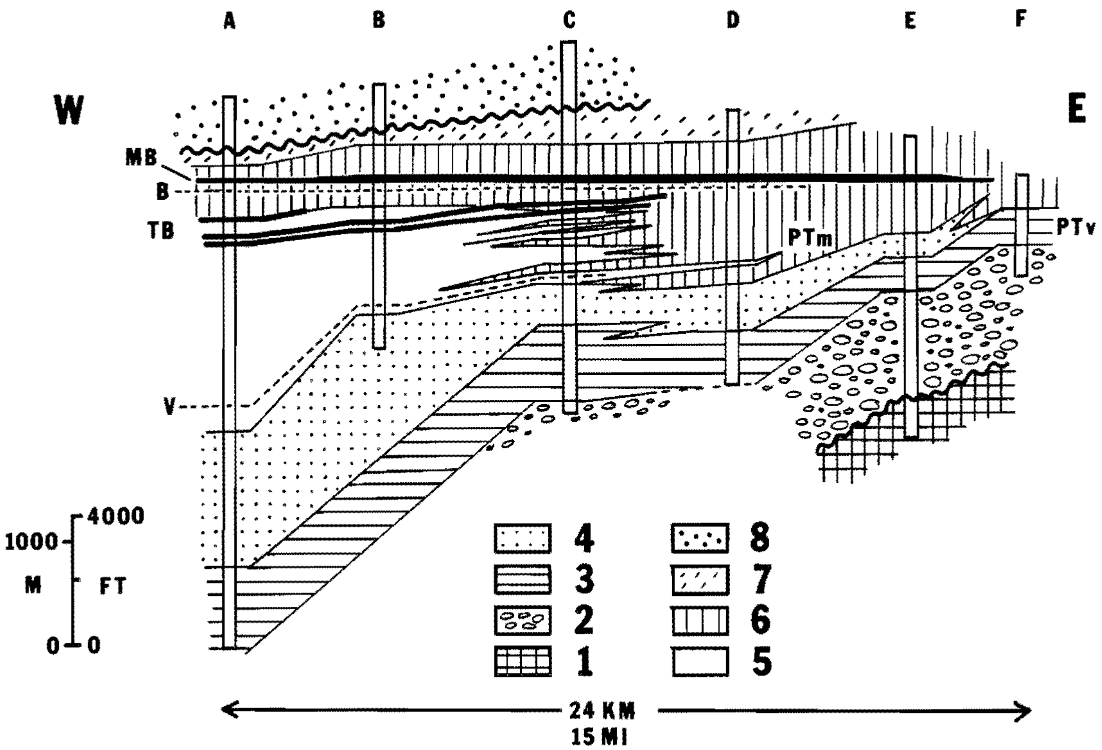

Figure 7. Generalized stratigraphic cross-section across the Cuyama Basin basin-margin. Section is from the southeastern Caliente Range (see Fig. 2 for location). Sections are located at: A--Caliente Mountain-Abbott Canyon area; B--3.5 miles east of A; C--7.5 miles eastsoutheast of A; D--9.8 miles eastsoutheast of A; E--12.0 miles eastsoutheast of A; F--15.0 miles eastsoutheast of A. Stratigraphic units shown are: 1--Pattiway Formation (Eocene marine sedimentary rocks); 2--Simmler Formation (nonmarine); 3--Soda Lake Shale Member Vaqueros Formation (bathyal marine) (contains thin shallow marine Quail Canyon Sandstone Member of Vaqueros Formation at base); 4--Painted Rock Sandstone Member of Vaqueros Formation (shallow marine); 5--Saltos Shale Member of Monterey Formation and Branch Canyon Formation (shallow marine in this area); 6--Caliente Formation (nonmarine); 7--Quatal Formation (nonmarine); 8--Morales Formation (nonmarine). Important stratigraphic horizons/intervals are: V--top of "Vaqueros" molluscan stage (ca. 19 Ma); TB--"Triple Basalts" (ca. 14-16 Ma); B--top of Barstovian vertebrate assemblages (ca. 12 Ma); MB--"Main Basalt". Also shown are peak transgressions for the Vaqueros basin cycle (PTv) and Monterey basin cycle (PTm). (modified in part from Vedder, 1973).

important regional relationships concerning the Vaqueros basin cycle (Fig. 7). In this area the Simmler represents the beginning of middle Cenozoic deposition and unconformably overlies the Eocene Pattiway Formation. Peak transgression for the Vaqueros basin cycle (Ptv) is clearly associated with the greatest eastward extent of the Soda Lake Shale Member (Fig. 7), probably in conjunction with the development of the siliceous rocks observed at the type section of that member (Fig. 6). This peak transgression is roughly coincident with the deepest paleobathymetries observed on the basis of benthic foraminiferal biofacies. Progradation of the Cuyama Basin margin is represented by the northwestward thickening wedge of Painted Rock Sandstone Member. The Vaqueros basin cycle ends in the late Saucesian (early Miocene) with renewed rapid subsidence and re-established bathyal water depths associated with the Monterey Formation.

Monterey Basin Cycle

The generalized section from beneath Cuyama Valley (Fig. 5) shows the rapid subsidence associated with initial deposition of the Monterey Formation. Water depths at this time (latest Saucesian; early Miocene) increased from a maximum of 150 m in the Painted Rock Sandstone Member to a minimum of ca. 1000 m in the Saltos Shale Member of the Monterey Formation, an increase of 800 m or more. A section from the central Caliente Range illustrates some of the faunal trends associated with the initial phase of the Monterey basin cycle (Fig. 8).

Initial Saucesian biofacies are relatively low-diversity middle bathyal assemblages dominated by agglutinated species, <u>Uvigerinella obesa</u> s. l. and <u>Florilus costiferum</u>. These assemblages contain few planktic foraminifera and are very similar to biofacies found in the upper part of the Soda Lake Shale Member of the Vaqueros Formation and may again represent a low oxygen environment. Diversity increases in the latest

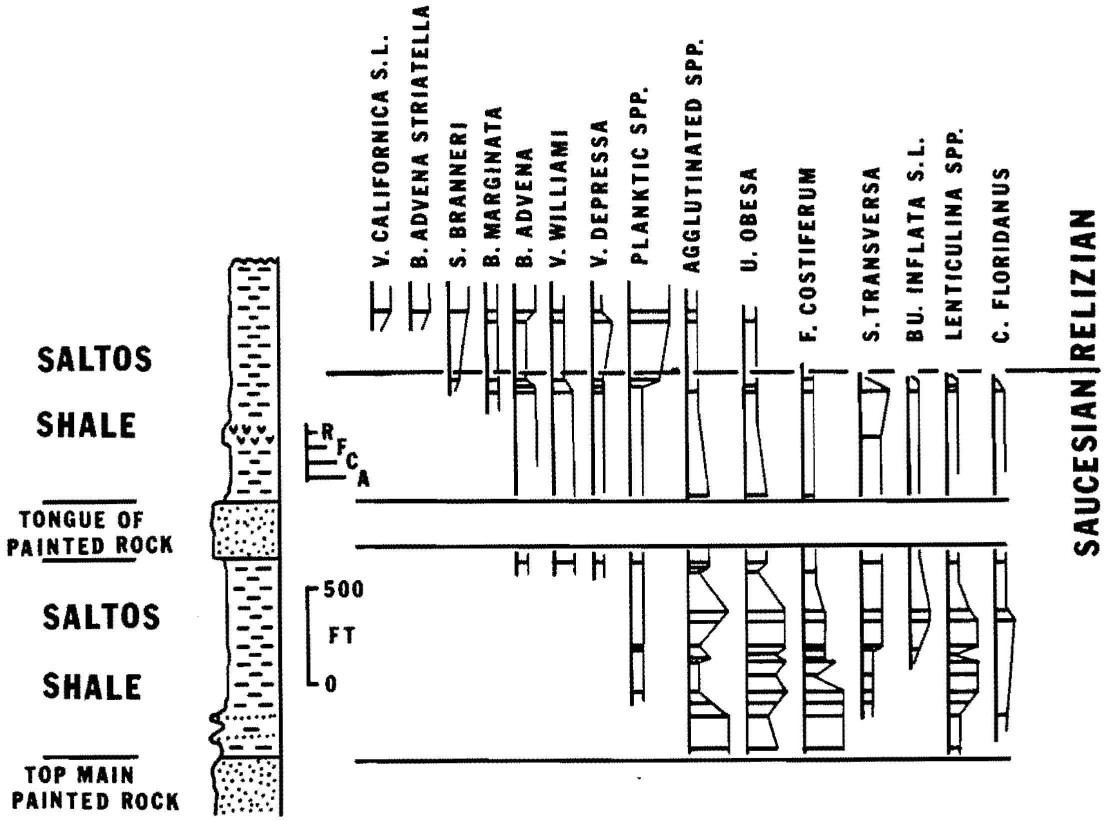

Figure 8. Distribution of foraminifera from a section of lower Saltos Shale Member (Monterey Formation). Section is located in the central Caliente Range, just west of Caliente Mountain (see Fig. 2). Abundance abbreviations: A—abundant; C—common; F—few; R—rare.

Saucesian and at the Saucesian/Relizian boundary (still within the early Miocene, ca. 17.5 Ma) a noticable increase in planktic foraminifera is observed (Fig. 8). Relizian biofacies indicate slightly shallower water depths (upper middle bathyal to upper bathyal, ca. 500 m).

These relationships can be documented in a number of subsurface sections from beneath Cuyama Valley (Fig. 9). In each of these sections a significant increase in planktic foraminifera is observed near the Saucesian/Relizian boundary. This increase could be due either to tectonic/paleogeographic readjustments in the middle Cenozoic borderland leading to better surface water connections to the open ocean or to a decrease in the intensity of low oxygen bottom water environments resulting in the enhanced preservation of carbonate (Lagoe, in press b). Marked lithofacies changes also occur near the Saucesian/Relizian boundary (Fig. 10). A submarine fan complex developed in the southeastern Cuyama Basin during the late Saucesian (Lagoe, 1984, 1987). A significant episode of fan progradation occurs during the Relizian and is followed by a general progradation of the basin margin (Fig. 10) during the Luisian (middle Miocene).

In general, bathyal environments become more areally restricted within the basin during the middle Miocene (Lagoe, 1984, 1985, 1987). Bathyal sedimentation at this time is characterized by a lack of significant submarine fan development, the onset of significant siliceous sedimentation within the Monterey Formation and limited deposition of phosphatic sediments in latest Lusian/early Mohnian (middle Miocene) bathyal and some neritic environments (Lagoe, 1987).

Examination of fully marine sections of the Monterey basin cycle does not reveal any obvious major unconformities. The stratigraphy of coeval rocks in the Cuyama badlands, however, indicate that a significant unconformity is associated with this basin cycle (Fig. 11). The Monterey basin cycle is here represented by the nonmarine Caliente and Quatal Formations. The Caliente Formation rests unconformably on a variety of stratigraphic units. It truncates the older Vaqueros basin cycle sediments and in places rests directly on Precambrian crystalline basement (Fig. 11). Vertebrate fossils found within the lower Caliente Formation indicate a Hemingfordian age (early Miocene; James, 1963), probably upper Hemingfordian which would suggest that these basal beds approximately correlate with the Saucesian/Relizian boundary. The truncation of older rocks and deposition of extensive nonmarine beds in this area suggests a source of sediment for submarine fan progradation observed in the Relizian bathyal rocks of the Cuyama Basin.

The basin margin cross-section from the southeastern Caliente Range provides further details of the Monterey basin cycle (Fig. 7). Peak transgression (PTm) occurs during the Relizian. This correlates with basinwide studies (Lagoe, 1984, 1985, 1987) which indicate that Relizian marine rocks are the most widespread within the Monterey basin cycle. It has been noted for many decades that the Relizian was a major transgressive period during the Miocene in California (Kleinpell, 1938). Unlike the Vaqueros basin cycle, peak transgression during the Monterey basin cycle is not associated with the deepest development of paleobathymetry. Latest Saucesian bathyal environments are normally deeper (lower middle to upper middle bathyal) than overlying, but more widespread, Relizian environments (upper middle to upper bathyal).

The middle Miocene contains evidence of decreased subsidence and major progradation of shallow marine and nonmarine environments (Fig. 7). Decreased subsidence is manifested in reduced accumulation of middle Miocene rocks (see comparative thicknesses in western sections of Fig. 7). Bathyal sections exhibit reduced bulk accumulation rates at this time as well (Lagoe, 1985), not only in the Cuyama Basin but throughout the middle Cenozoic borderland (Isaacs, 1983). The Cuyama Basin was completely filled by the late Miocene and then subjected to late Cenozoic deformation as evidenced by the unconformity at the base of the nonmarine, Pliocene-Pleistocene Morales Formation (Figs. 7 and 10).

Summary of Relationships

The stratigraphic, depositional and paleobathymetric relationships within the middle Cenozoic rocks of the Cuyama Basin are summarized in Table 1. Before evaluating these relationships in light of seismic stratigraphic principles, a comparison with coeval rocks in the southern San Joaquin Basin will indicate which relationships have regional significance.

SOUTHERN SAN JOAQUIN BASIN--SAN EMIGDIO AREA

Stratigraphic overview

The San Emigdio area is located at the southern end of the San Joaquin Valley (Figs. 1 and 4). This area represents the southernmost portion of the middle Cenozoic San Joaquin Basin. Recent work in this area by Nilsen (1973, 1984, in press), Nilsen and others (1973), DeCelles (1986, 1987) and Lagoe (1986, in press a) forms the basis for a summary of Cenozoic stratigraphy in this area (Fig. 12). The middle Cenozoic stratigraphic record is contained in the Temblor, Tecuya and Monterey Formations, plus an unnamed nonmarine conglomeratic unit in the eastern part of the area (Fig. 12). These rocks represent a westward deepening basin-margin transect in this area (Lagoe, 1986, in press a) and provide an excellent opportunity to document middle Cenozoic depositional history in the southern San Joaquin Basin.

Stratigraphic/Depositional Trends

The Temblor Formation represents middle Cenozoic deposition in the San Emigdio area that is roughly coeval with the Vaqueros basin cycle in the Cuyama Basin. The basal Temblor Formation rests unconformably on a variety of older rocks over much of the San Emigdio area (Fig. 13). Near Brush Mountain this unconformity has truncated all older sedimentary rocks to rest directly on crystalline basement. In other areas variable amounts of the older sedimentary units are preserved (Figs. 12 and 13). The geometry of this unconformity is not a simple basin-edge onlap but can be very complex over short distances (Fig. 12). Farther north in the San Joaquin basin the Temblor Formation rests with apparent conformity on the Pleito Formation but a significant paleobathymetric shift (based on benthic foraminiferal biofacies) is associated with the basal Temblor (Fig. 13). The lowermost Temblor Formation often contains neritic foraminiferal assemblages which grade abruptly upwards into late Zemorrian (late Oligocene) lower middle to lower bathyal assemblages (ca. 2000 m water depth). This rapid subsidence is analogous to that documented in the Cuyama Basin during the Vaqueros basin cycle though the latter is slightly younger (occurring near the Oligocene/Miocene boundary).

A preliminary analysis from ongoing work in the San Emigdio area of a section from the Pleito Hills documents additional details about depositional trends in these middle Cenozoic rocks (Fig. 14). Throughout the San Emigdio area the Temblor Formation contains lower to middle bathyal foraminiferal biofacies except locally, as noted, in its basal portion and where it grades eastward into the nonmarine Tecuya Formation (Fig. 12). The fauna from the upper part of the Temblor Formation at Pleito Hills (Fig. 14) is a relatively low diversity Saucesian assemblage dominated by Uvigerinella obesa s.l., similar in many respects to the faunas in the upper part of the Soda Lake Shale Member and lower part of the Saltos Shale Member in the Cuyama Basin. Nowhere in the San Emigdio area is there developed a major shallow marine progradational wedge analogous to the Painted Rock Sandstone Member in the Cuyama Basin.

The Pleito Hills sections contains higher-diversity benthic foraminiferal assemblages in the uppermost Saucesian and a dramatic increase in planktic foraminifera near the Saucesian/Relizian boundary. The Relizian part of the section contains minor turbidite sandstones which do not extend up into the Luisian part of the section. The Luisian (middle Miocene) Monterey Formation at Pleito Hills is unconformably overlain by late

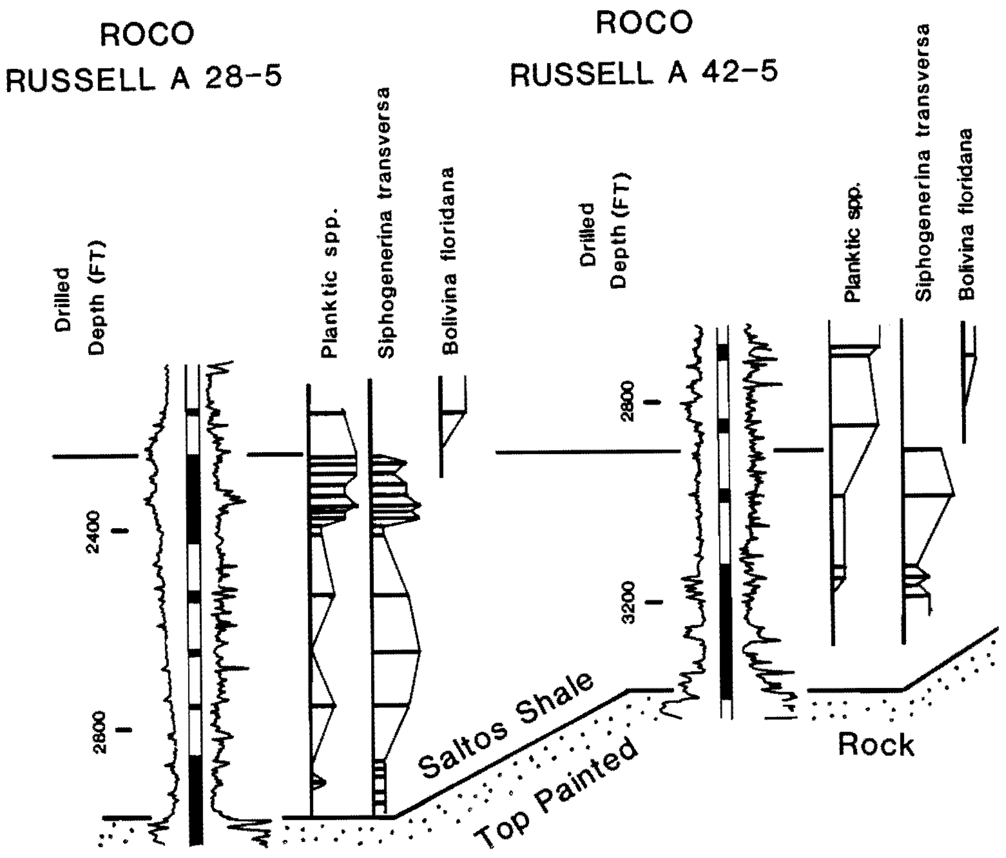

Figure 9. Series of wells from beneath Cuyama Valley (see Fig. 2) showing widespread occurence of the dramatic increase in planktic foraminifera near the Saucesian/Relizian boundary, as observed in Fig. 8. Siphogenerina transversa is a marker species for the Saucesian Stage while Bolivina floridana does not extend below the Relizian. Abundance abbreviations: A--abundant; C--common; F--few; R--rare.

Cenozoic nonmarine rocks (Fig. 14).

The regional stratigraphic summary (Fig. 12) illustrates several other important points. The Monterey Formation is unconformably truncated by a regional unconformity associated with the Etchegoin Formation (late Miocene/earliest Pliocene). Like the basal Temblor unconformity, this late Miocene unconformity truncates variable amounts of older rocks (Fig. 12). The basin-margin transect also identifies two episodes of peak transgression. The oldest (PTt1) is associated with the lower Temblor Formation and occurs below volcanic rocks radiometrically dated at 22.3 +/- 0.7 Ma and 21.5 +/- 0.7 Ma (Turner, 1970), placing this peak transgression near the Oligocene/Miocene boundary (Fig. 12). The younger peak transgression (PTt2) occurs near the top of the Temblor Formation near the Saucesian/Relizian boundary. This transect also illustrates a prominent basin-edge unconformity at the base of the unnamed conglomeratic unit which truncates this younger peak transgression. Interfingering relationships with marine rocks to the west date this unconformity as Relizian in age (Fig. 12). The unnamed conglomeratic unit eventually truncates all older sedimentary units to the east, where it rests directly on crystalline basement.

Summary of relationships

The stratigraphic, depositional and paleobathymetric relationships of the middle Cenozoic rocks in the San Emigdio area are summarized on Table 1. These can now be compared with relationships in the Cuyama Basin.

COMPARATIVE RELATIONSHIPS

The primary obstacle to testing global eustatic control vs. tectonic control of depositional patterns is the inability to precisely constrain the age of all depositional features within a basin. The summary of depositional events in the two areas studied (Table 1) is constrained within the limits of the chronostratigraphic data currently available. Several events remain poorly constrained. The basal Simmler unconformity in the Cuyama Basin is an example.

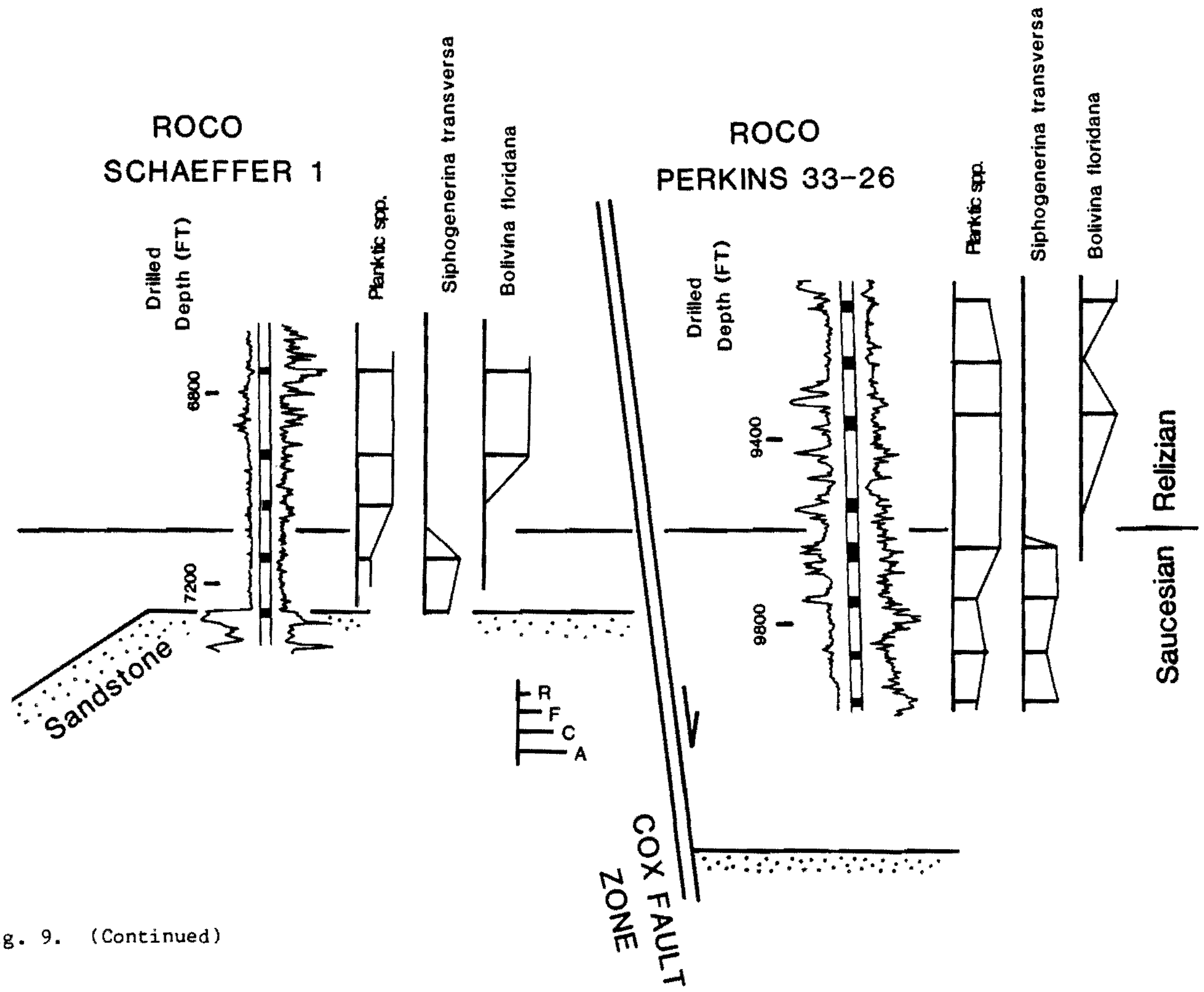

Fig. 9. (Continued)

Sparse chronostratigraphic information from the upper part of the Simmler Formation indicates that unit is no younger than earliest Miocene (radiometric dates cited earlier). The base is Oligocene but is unconstrained otherwise. It is apparent that it will be very difficult to prove global eustatic control by demonstrating the synchroneity of depositional events. Even under the best of conditions in preQuaternary rocks, there will often be a question of absolute synchroneity due to the inherent errors and uncertainties involved in chronostratigraphic dating methods.

A review of the depositional histories of the two areas studied (Table 1 and discussion above) produces evidence of both eustatic and tectonic control of depositional patterns. Relationships suggesting a eustatic control include the following:

1. The apparent synchroneity of Relizian basin edge unconformities in the two areas (although this could be due to regional tectonic activity).

2. The apparent synchroneity of the Relizian increase in planktic foraminifera in bathyal biofacies which could be due to increased surface water circulation with the open ocean caused by sea level adjustments. This, of course, could also be due to tectonic adjustments or a paleoceanographic change favoring enhanced carbonate preservation.

3. The apparent synchronous cessation of Relizian turbidite deposition in the two areas (also observed in other basins of the middle Cenozoic borderland; Lagoe, 1985, 1987) possibly due to a relative rise in sea level.

4. The deposition of phosphatic lithofacies in the Cuyama Basin during the late Lusian/early Mohnian in both bathyal and neritic environments (Lagoe, 1987). This type of deposition requires greatly reduced terrigenous sedimentation rates and could reflect episodes of high sea level (Lagoe, 1987).

5. The coincidence of peak transgression with the development of maximum water depths in the Vaqueros basin cycle of the Cuyama Basin.

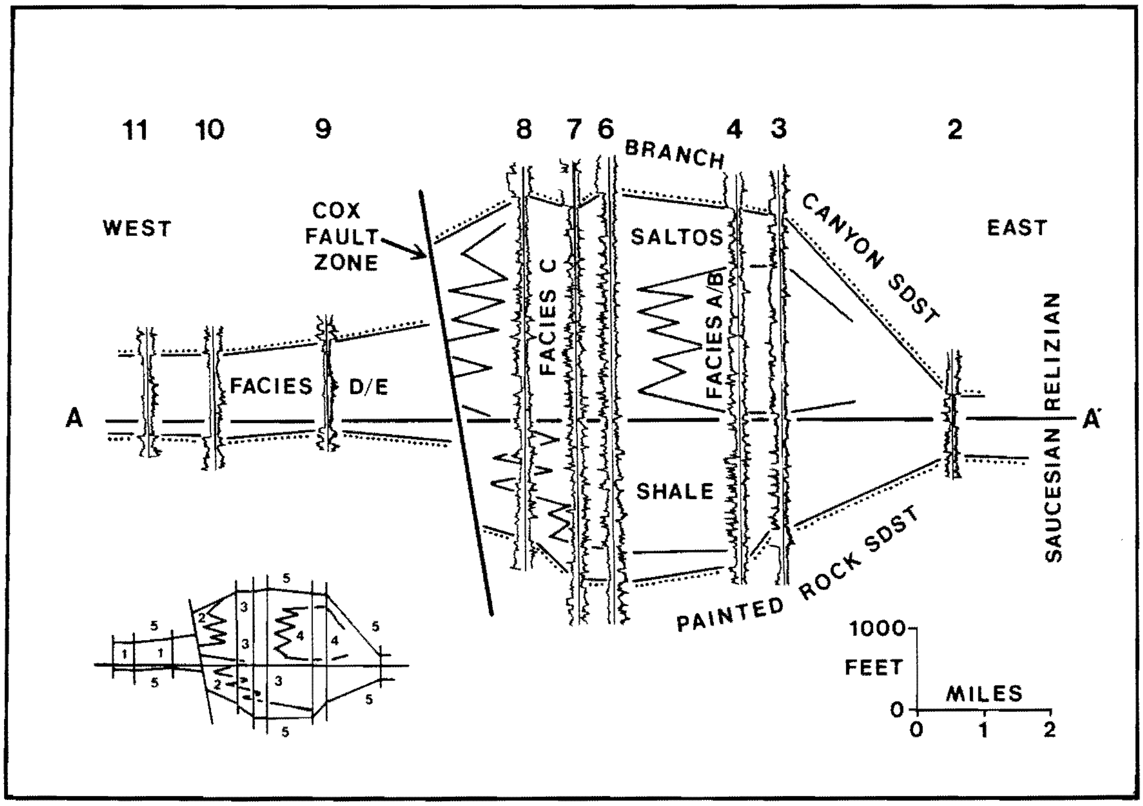

Figure 10. Lithofacies cross-section of Saucesian and Relizian rocks (early Miocene) from beneath Cuyama Valley (see Fig. 2). Lithofacies refer to modified Mutti-Ricci Lucchi lithofacies (see Lagoe, 1985, 1987). Lower left hand corner of diagram shows paleoenvironmental interpretation of lithofacies: 1—basin plain; 2—submarine fan fringe; 3—outer to middle fan; 4—inner fan; 5—shallow marine. Note prominent progradation of fan environments during the early part of the Relizian. Wells used are: 2—Richfield Wegis-Reyes B-1; 3—Richfield Lundstrom Becher; 4—Mohawk Humble Lundstrom 48-2; 6—Richfield Perkins 33-35; 7—Richfield Perkinds 1; 8—Richfield Perkinds 33-26; 9—Richfield Schaeffer 1; 10—Seaboard Kirschenmann 1; 11—Ohio Kirschenmann 1. (from Lagoe, 1987).

Evidence that depositional patterns in the two areas are not controlled by globally synchronous sea level changes but are more sensitive to local/regional tectonics include:

1. The apparent nonsynchroneity of major regional unconformities and peak transgressions in the two areas (although in part this may be due to poorly constrained ages).

2. The apparent synchroneity of latest Oligocene/early Miocene volcanic rocks in the two areas (suggesting regional extension).

3. The noncoincidence of peak transgression (Relizian) and development of maximum water depths (late Saucesian) during the Monterey basin cycle in the Cuyama Basin.

4. The apparent synchroneity of peak transgression, major submarine fan progradation and development of a basin edge unconformity in the Cuyama Basin during the Relizian. If peak transgression is associated with a relative sea level rise only, submarine fan complexes should become starved of coarse clastic sediment. It is clear that sediment availability to the Relizian submarine fan system increased, possibility due to uplift associated with development of the Relizian basin edge unconformity in the Cuyama badlands. This strongly suggests a tectonic control on this particular depositional package.

5. The magnitude of rapid bathymetric deepenings during the middle Cenozoic of both basins can not be accounted for by sea level fluctuations of the magnitude suggested by Vail and others (1984). Three major episodes of rapid basin subsidence (Oligocene/Miocene boundary and early Miocene in the Cuyama Basin, late Oligocene in the San Emigdio area) involve bathymetric deepening of 800-1500 meters. This strongly suggests a tectonic influence of relative sea level in these basins at those times.

6. The massive progradational event associated with the Painted Rock Sandstone Member of the Vaqueros Formation in the Cuyama Basin is missing

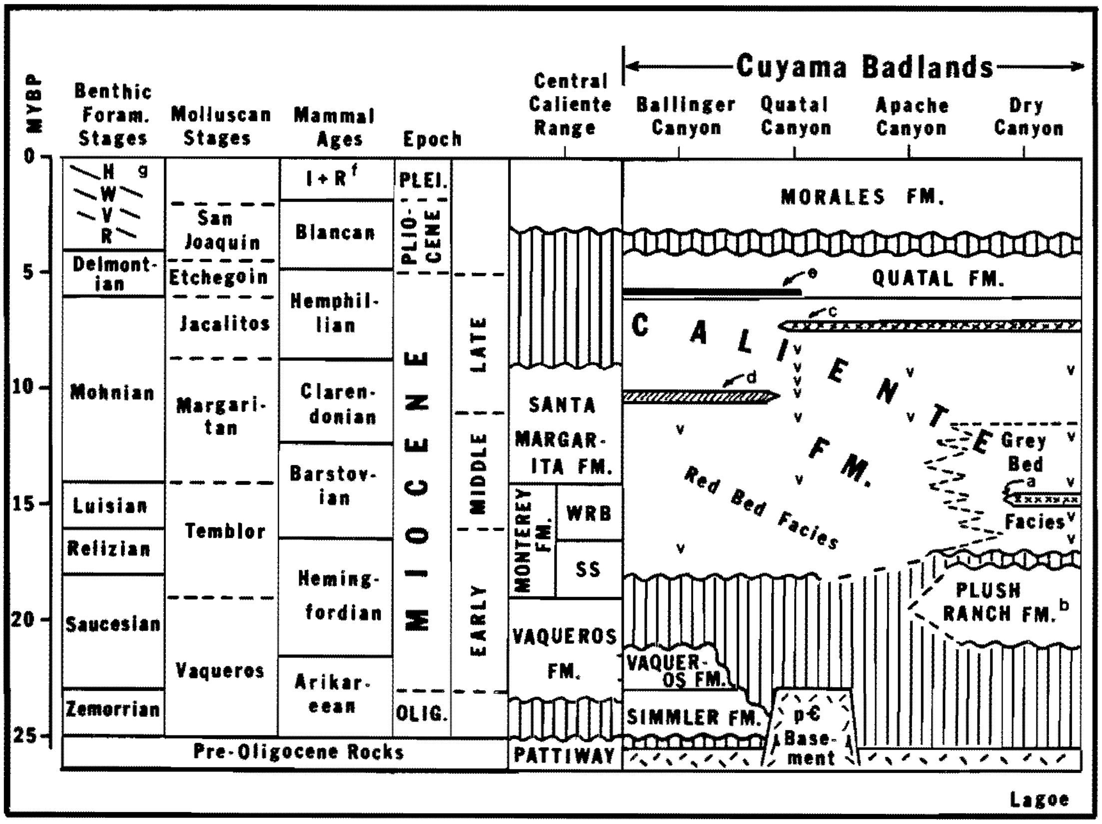

Figure 11. Summary stratigraphy of the Cuyama Badlands. Time scale after COSUNA (1984) with modifications after Poore and others (1981) and Keller and Barron (1981). Stratigraphy after James (1963), Woodburne (1975), Hill and others (1958), and Lagoe (1984). Footnotes: a-Tuff bed in Caliente Formation with a K/Ar date of 15.2 mybp (Evernden and others, 1964); b-K/Ar dates from volcanic rocks in the Plush Ranch Formation of Lockwood Valley are 19.6 ± 1.1 and 17.4 ± 3.7 mybp (Crowell, 1973); C-Lockwood Clay; d-"Main Basalt"; e-Interval of bedded gypsum in lower Quatal Formation: f-Irvingtonian and Rancholabrean Ages; g-Hallian, Wheelerian, Venturian and Repettian Stages. Subdivisions of Monterey Formation in the central Caliente Range are: WRB-Whiterock Bluff Shale and SS-Saltos Shale. Small v's mark vertebrate fossil localities from James (1963).

or very poorly developed in the San Emigdio area. This indicates a major source of sediment (tectonically controlled?) in the Cuyama Basin which was missing in the coeval San Emigdio area.

Large scale geometric relationships (e.g. major shifts in coastal onlap) could not be evaluated within the scope of this study, therefore, a fully conclusive test of seismic stratigraphic principles in this area remains elusive. In particular, higher-precision chronostratigraphic dating of depositional events is needed. It is apparent that both eustatic and tectonic controls are important in shaping the depositional geometries and history of active margin basins. Unconformity bounded stratigraphic units are easily recognized in the two areas and are useful in deciphering basin history. The bulk of the evidence obtained in this study indicates that tectonic controls of basin geometry, bathymetry and clastic source areas were relatively more important in shaping the depositional fabric of these basins than global eustacy. Whether this is the case throughout the middle Cenozoic borderland will be answered by further work.

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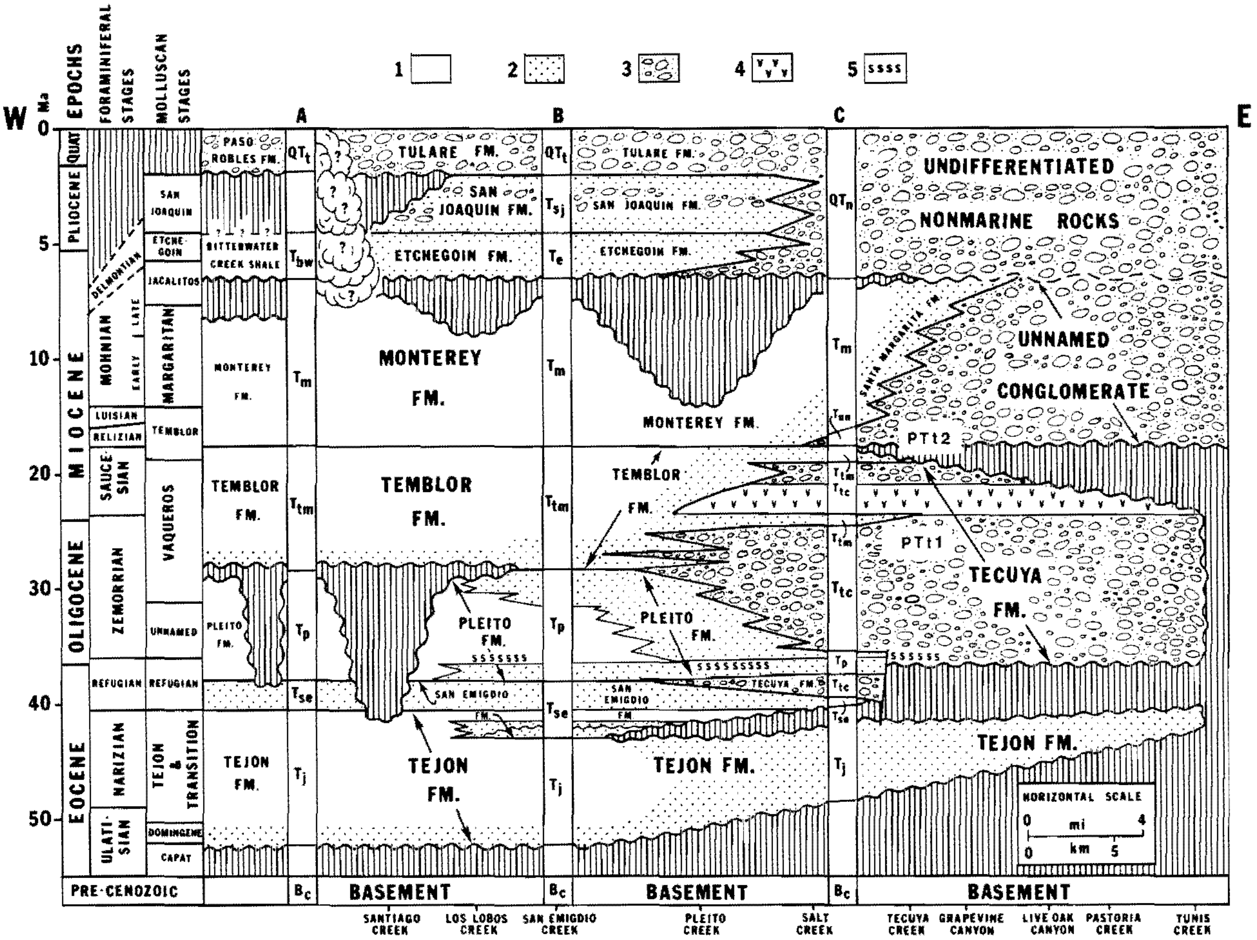

Figure 12. Generalized stratigraphic cross-section of the San Emigdio area of the southernmost San Joaquin Basin. Major unconformities are shown by vertical ruling. Symbols in columns A, B and C are stratigraphic abbreviations used in Figure 13. Paleoenvironmental symbols are as follows: 1--generally bathyal marine; 2--shallow marine; 3--nonmarine. Other symbols are: 4--volcanic rocks; 5--"seismites" of DeCelles (1986). Also shown are two peak transgressions (PTt1 and PTt2) associated with the Temblor Formation. See Figure 4 for location of section.

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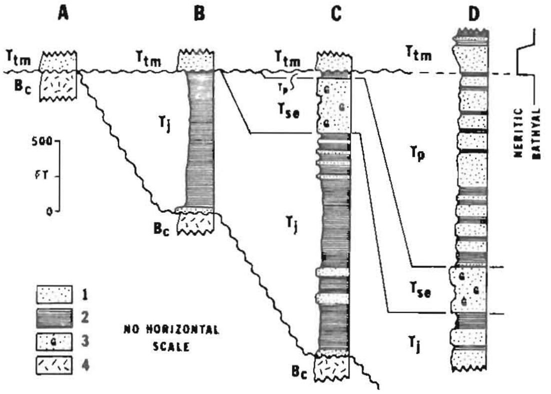

Figure 13. Schematic diagram of the basal Temblor unconformity in the San Emigdio area at: A--Brush Mountain; B--upper East Twin Creek; C--Colgrove Klipstein well and D--North Tejon oil field (see Fig. 4 for location). Stratigraphic symbols as per Fig. 12. Curve to right of column D shows shift in bathymetry from bathyal in Pleito Formation to neritic in basal Temblor Formation. Lithologic symbols: 1--sandstone; 2--mudstone; 3--glauconitic sandstone; 4—crystalline basement.

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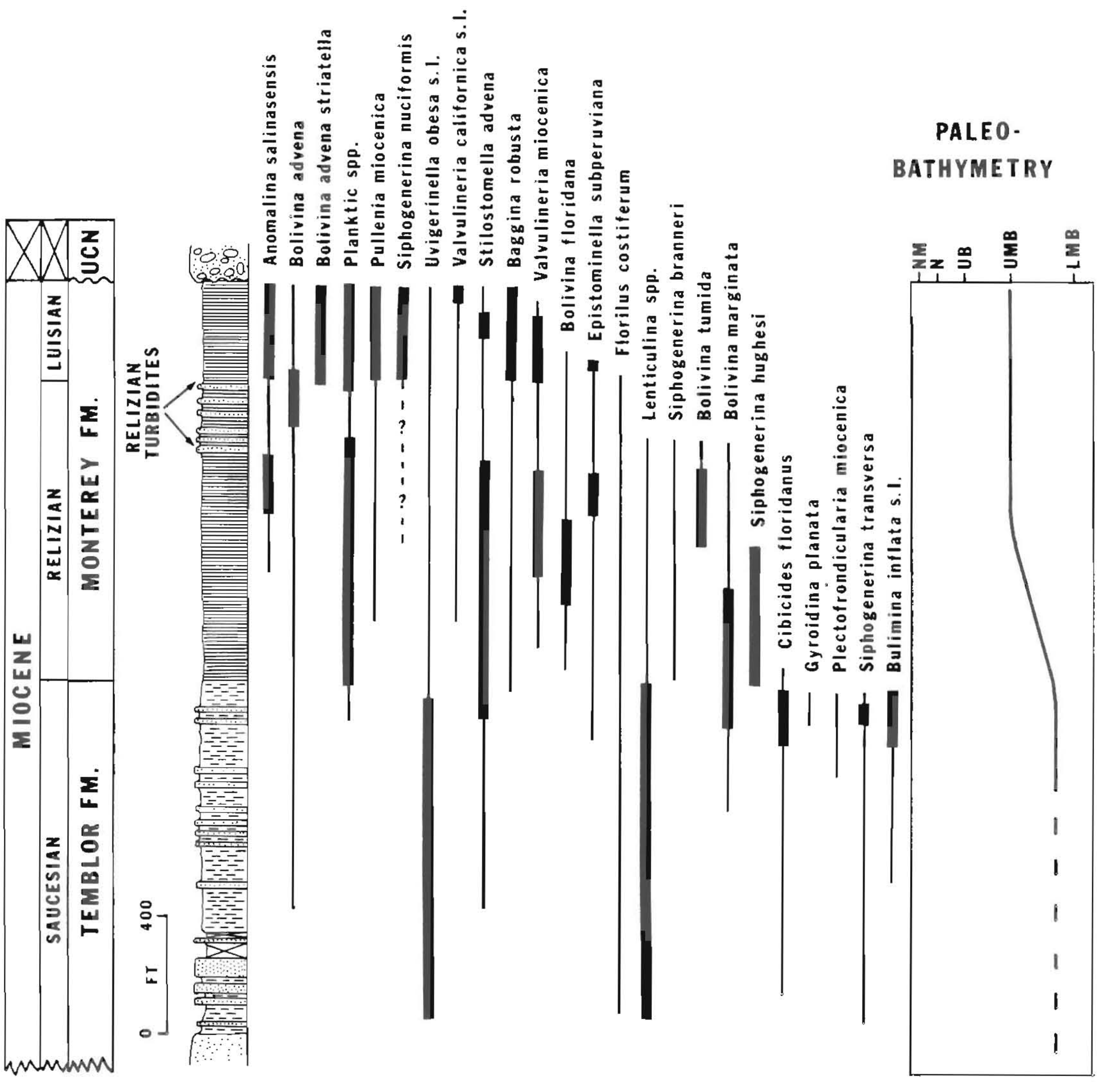

Figure 14. Distribution of foraminifera in a composite section from the Pleito Hills (see Fig. 4 for location). Thick lines indicate common and consistent occurrence while thinner lines indicate less common or sporadic occurrence. Paleobathymetry base on benthic foraminiferal biofacies. Abbreviations: NM--nonmarine; N--neritic; UB--upper bathyal; UMB--upper middle bathyal; LMB--lower middle bathyal; UNC--upper Cenozoic nonmarine rocks.

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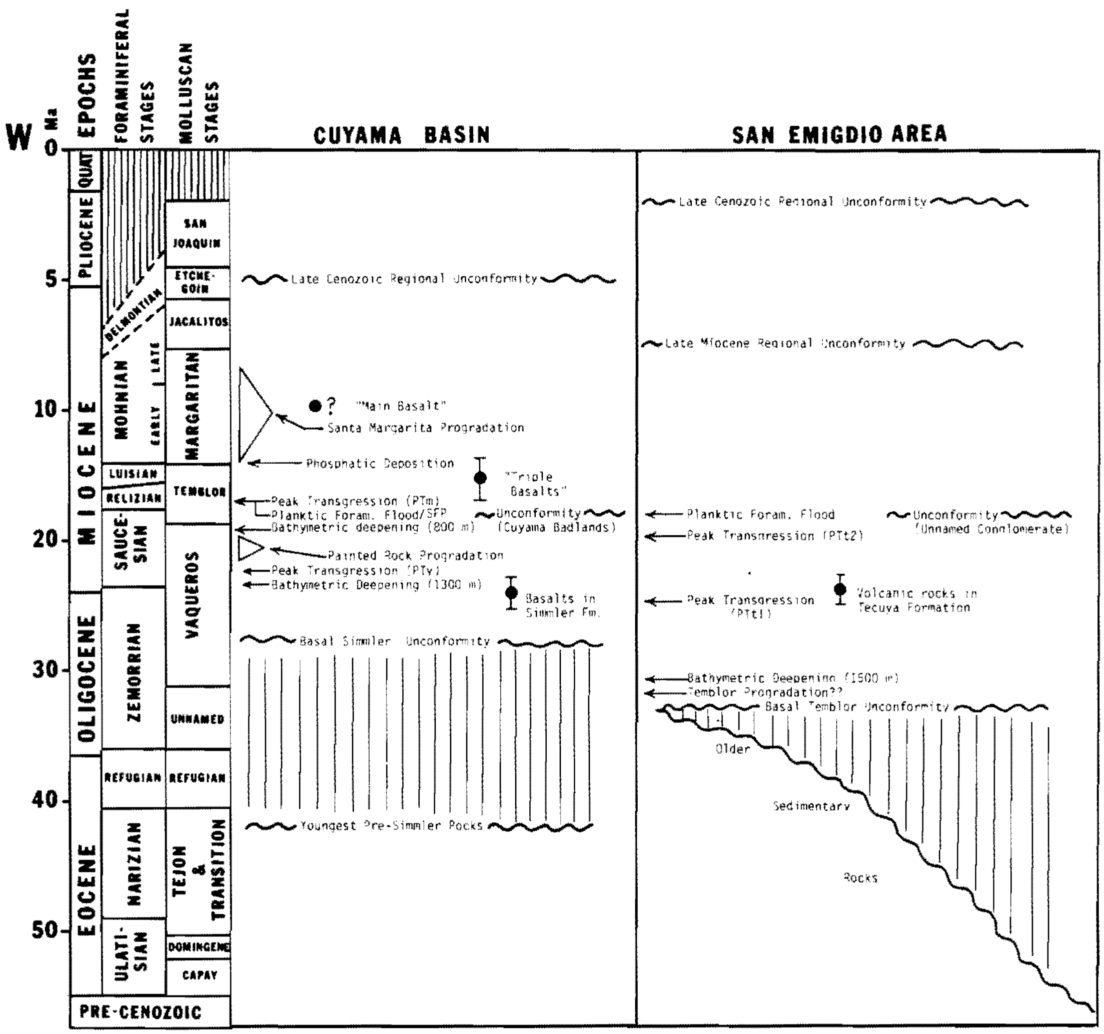

Table 1. Summary of stratigraphic and depositional events in the Cuyama Basin and San Emigdio area. Numbers following bathymetric deepening events are estimates of the minimum deepening involved from analysis of benthic foraminiferal biofacies. Abbreviation following planktic foraminifera flood in Cuyama Basin (SFP) indicates progradation of submarine fan complex at that time.

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PRACTICAL APPLICATION OF PLEISTOCENE EUSTACY IN OFFSHORE GULF OF MEXICO

JAMES L. LAMB[1], WALTER W. WORNARDT[2], TING-CHANG HUANG[1], AND THOMAS E. DUBÉ[3]

1 Micro-Strat Inc., 8401 Westheimer, Suite 110, Houston, Texas, 77063
2 Micro-Strat Inc., 718 17th St., Suite 1500, Denver, Colorado, 80202
3 Department of Geological Sciences, University of Washington, Seattle, Washington, 98195

ABSTRACT

A conceptual model for marine Pleistocene sedimentation in the Gulf of Mexico is described for the past 2.9 million years. It utilizes eight eustatic cycles (i.e., lowstands-highstands of sea level couplets) representing the four major periods of glaciation, planktonic foraminiferal and calcareous nannofossil marker species for establishing datum planes, foraminiferal abundance/diversity plots, and a relative paleotemperature curve that is tied in part to oxygen-isotope stages. Methodology used here to construct this conceptual model is termed multiple-event stratigraphy in which physical happenings and bioevents are applied jointly in delineating Pleistocene Stages rather than reliance on a single datum. Midcontinent conceptual stage names are adapted to Gulf of Mexico stratigraphy and are consistent with the glacial concept for the Pleistocene Epoch having a lower boundary at about 2.5 to 3.0 m.y. B.P.

HISTORY OF STRATIGRAPHIC USAGE

In studies on the Mississippi River system, H. N. Fisk (1940, 1944) described and named alluvial Pleistocene formations of the Louisiana Gulf Coast in ascending order the Williana, Bentley, Montgomery, and Prairie. Each unit has a basal sandstone (the substratum) and an upper clay-siltstone (the topstratum). Later Fisk (1952) and Fisk and McFarlan (1955) related these formations or fluvial terrace deposits to Midcontinent Pleistocene events, whereby the substratums in part reflect lowstand glacial stages and the topstratums reflect highstand interglacial stages. In this manner the concept of glacio-eustatism became established in the Gulf Coast region.

A timely paper by Akers and Holck (1957) described an essentially complete Pleistocene interval in a well located in South Pass Block 41 at the tip of the Mississippi River delta near the edge of the continental shelf. This provided a convincing series of electric-log correlations from the well landward to wells showing the terraces. They also provided a paleobathymetric curve based on benthic organisms. Following the rationale of Fisk and McFarlan (1955) shoaling in the well section was related to glacial intervals and deepening to interglacials.

This pioneer study was refined by Poag and Valentine (1976) based on more precise planktonic information with more refined paleoecological interpretations. Their approach allowed for recognition of previously unrecognized unconformities in the well section, which in turn could be related to glacio-eustatic events.

In the early days of oil exploration in the Mississippi River delta region (during the 1950's), early Pleistocene deposits of southeastern Louisiana and adjacent continental shelf were interpreted commonly as basal regressive sandstones. These, some 5,000 feet thick in the Terrebonne Trough area, were believed to represent sediments deposited during the Nebraskan glacial lowstand. Overlying this sandstone unit is the Terrebonne shale (a prominent unit on electric logs) which led most Gulf Coast stratigraphers to conclude that this transgressive marine shale represents the first interglacial highstand in the Pleistocene (e.g., Sachs and Skinner, 1973). Akers and Holck (1957) interpreted this shale as Aftonian in age and, on the basis of paleobathymetry and lithology, relate the overlying units to the Kansan, Yarmouthian, Illinoian, Sangamonian, and Wisconsinan Pleistocene stages.

Thus an "Ice Age" concept, relating glacial-interglacial cycles, was recognized early in Gulf Coast stratigraphy and oil exploration. We support the tenets of Pleistocene glaciation and eustacy as applied to the Gulf of Mexico and maintain that their application greatly enhances understanding and application of geologic events important to petroleum exploration.

BENTHIC VERSUS PLANKTONIC ZONATIONS

As exploration drilling moved offshore onto the continental shelf during the early 1960s, thick marine Pleistocene intervals were encountered - locally in excess of 15,000 feet. Although some paleontologists found it advantageous to subdivide the Pleistocene interval utilizing detailed lithologic and faunal criteria, others found it expedient to utilize mainly recorded "tops" or last occurrences of particular benthic foraminiferal species (Fig. 1). Faunal datums chosen in this manner seldom are time synchronous over a wide geographical area even on the shelf. Their value as indicators of a precise time datum has validity only for short distances along the strike of the facies because the depth of water and bio- and lithofacies, which obviously control this occurrence, change rapidly up- and downdip.

Theoretically, and as demonstrated by our practical applications, the application of eustatic principles is axiomatic for accurate definition of time-significant datums essential for correlation.

Exploration drilling in the early 1980s moved into deep water off the present continental shelf onto the slope. Minimum depths of water encountered here during the Pleistocene were not less than 600 to 800 feet during the lowstands and were usually deeper. Essentially none of the shelfal benthic markers were able to live in this deep water environment (Fig. 1), but they were at times redeposited downslope usually in younger sediments. It is apparent, therefore, that different paleontological criteria, especially those utilizing the planktonic foraminifers and calcareous nannofossils, are needed to correlate the Pleistocene interval in deep-water regions off the shelf.

THE PLEISTOCENE CONCEPTUAL MODEL

Recently Beard, Sangree, and Smith (1982) introduced a sequence of Quaternary eustatic cycles numbered from Q1 to Q8 (Fig. 3). A lowstand-highstand of sea level couplet constitutes a eustatic cycle. Tectonic or sedimentary events, however, may cause local changes in water depths. In other words, all observed changes in water depth, especially on shallow shelves, are not eustatically controlled. Nondeposition, slumping, and erosion also influence depositional patterns locally. For example, shelfal shallow-water and lowstand glacial sections locally may reflect nondeposition or erosion (Fig. 2).

Presumably much of the deeper water channel deposits were derived from sediments exposed on the shelf and eroded during the glacials or lowstands. Our best estimate, based on benthic foraminiferal paleoecology, is that sea-level changes on the order of 400 to 600 feet occurred during such times. This is supported in part by Poag and Valentine (1976) who found unconformities developed on shallow shelfal areas during essentially all glacial periods.

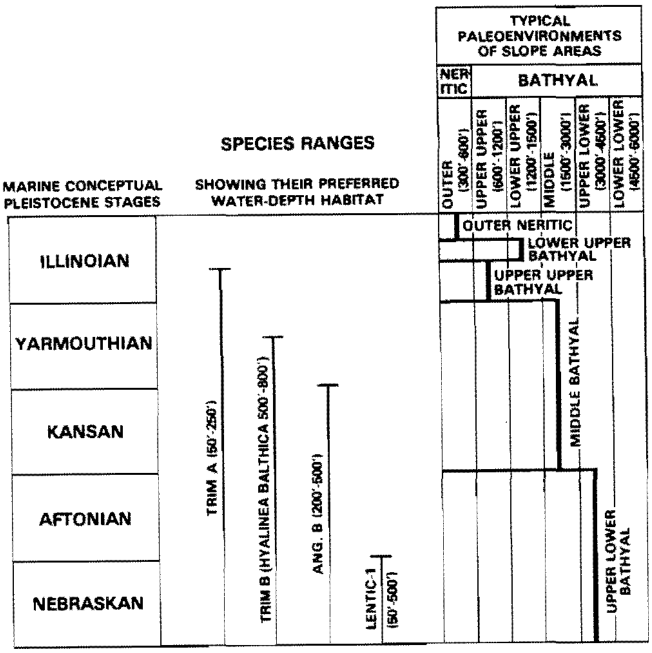

FIGURE 1. Why shelfal shallow water benthic foraminifers are not useful biostratigraphic markers in deep-water regions of the Gulf of Mexico. From actual well data.

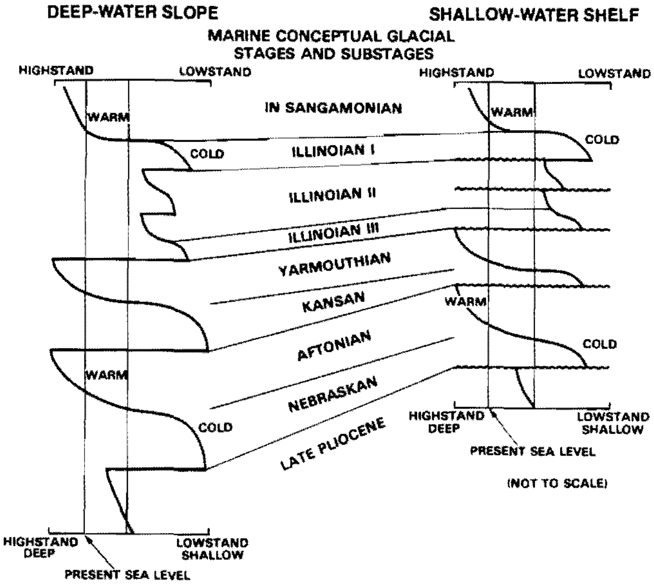

FIGURE 2. Diagrammatic comparison of deep-water versus shallow-water Pleistocene eustatic cycles Gulf of Mexico.

Glacial intervals are indicated in the sediment record by decreasing numbers (low diversity) of warm-water planktonic species and shallowing of benthic biofacies. Interglacial intervals are characterized by increasing numbers (high diversity) of warm-water planktonic species and deepening of benthic biofacies. The actual numbers (abundance) of individual specimens of each species show the same types of trends (Fig. 3).

Other studies support an "Ice Age" model for Pleistocene sediments of the Gulf of Mexico region. For example, Smith (1965) and Beard, Sangree, and Smith (1982) discuss eustacy and climate modeling; Beard (1969, 1973), Kennett and Huddleston (1972), and Thunell (1984) analyze paleoclimatology based on microfossil evidence. Through the efforts of Boellstorff (1978), Beard et al. (1982), Williams (1984), and Dubé (1985) we are able to relate a relative paleotemperature record of the Gulf of Mexico to an absolute time scale that can be compared with the oxygen-isotope scale back to about 0.9 m.y.B.P. (late Yarmouthian Stage). More recently, Gartner, Chen, and Stanton (1983) recorded the Plio-Pleistocene nannofossil biostratigraphy, as well as the carbonate content and coarse fraction of core holes in the northeastern Gulf of Mexico, which show affinity to the oxygen-isotope stages and magnetic stratigraphy of a core hole in the Pacific area. Well-documented planktonic foraminiferal and nannofossil datums, especially those tied to the paleomagnetic time scale, are

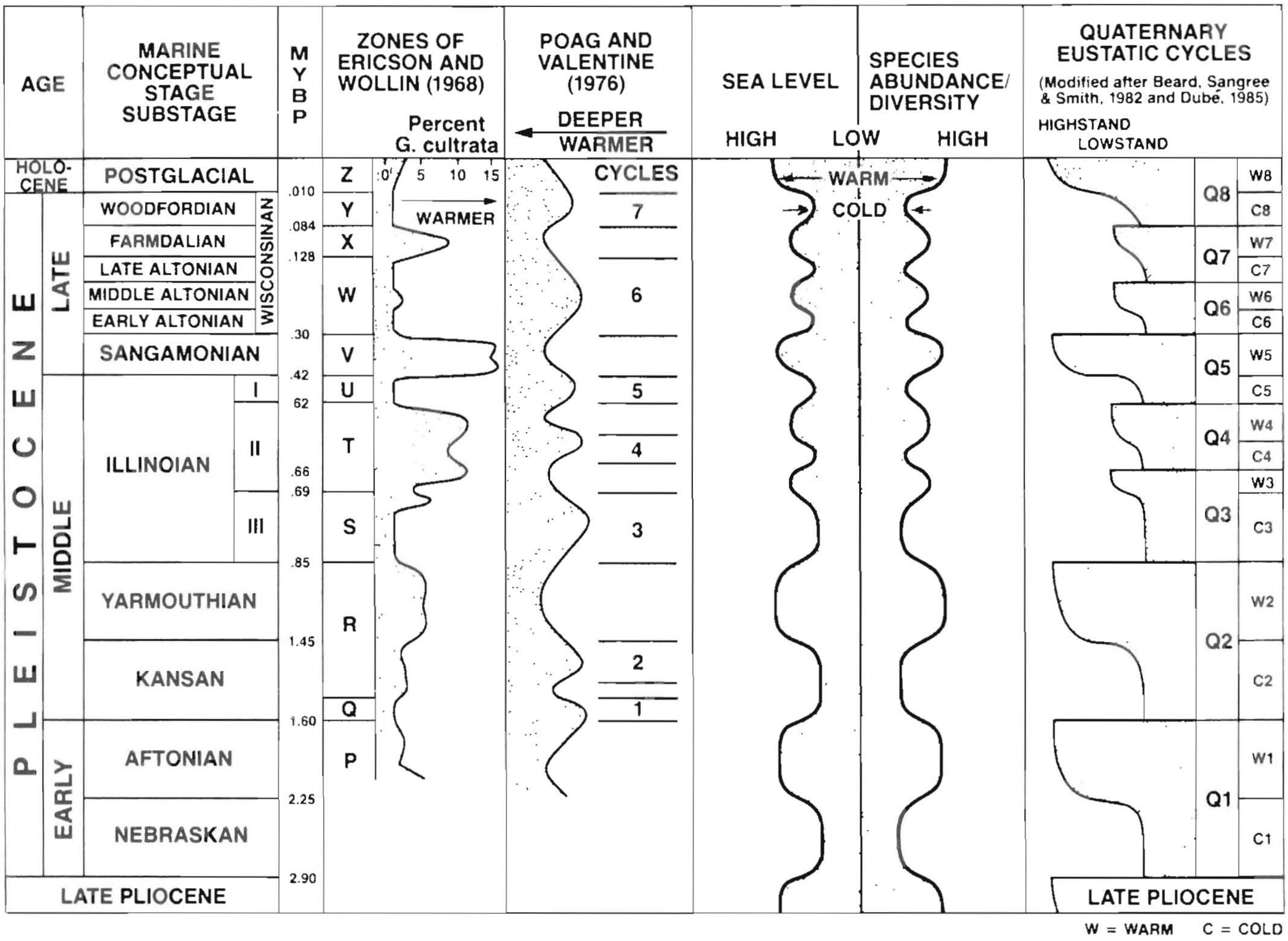

FIGURE 3. Comparison of Pleistocene glacial-interglacial stages, zones of Ericson and Wollin, depositional-climatic cycles of Poag and Valentine, relative sea level and species abundance/diversity curves, and Quaternary eustatic cycles illustrating coastal onlap.

useful for development of a Quaternary chronology in the Gulf of Mexico.

In this discussion, we touch on some of the basic principles that support a conceptual model for the Pleistocene, elements of which were discussed over many years by numerous individuals in the Gulf Coast region (Fig. 4). We believe that this model offers the best approach now available for formulating a reliable zonation and for recognizing local and regional datums needed for accurate stratigraphic correlation. In a broad sense, the chart (Fig. 4) depicts multiple-event stratigraphy. The main elements considered here are:

1. Planktonic foraminiferal and calcareous nannofossil datums (of regional and worldwide significance-tropical and subtropical).
2. Benthic assemblages and datums (reflecting mostly eustacy and paleoclimate).
3. Foraminiferal abundance and diversity (reflecting relative paleotemperature, paleoecology and depositional environment).
4. Eustatic cycles (in the general sense, reflecting changes in sea level, coastal onlap, climate, and sedimentation).
5. Seismic stratigraphy (recognition of reflection patterns on the seismogram; relating paleoclimate to sedimentation).
6. Correlation to oxygen-isotope stages (enhances paleotemperature and absolute-age framework and aids in worldwide stratigraphic correlation of events).

In order to develop and apply the paleontological part of this model, a faunal chart (stratigraphic distribution chart) of the benthic and planktonic foraminifers and calcareous nannofossils must be first prepared. This raw-data base provides a basis for recognizing the Pleistocene stages and eustatic cycles. Moreover these data can be compared with the seismic profile to assist in the

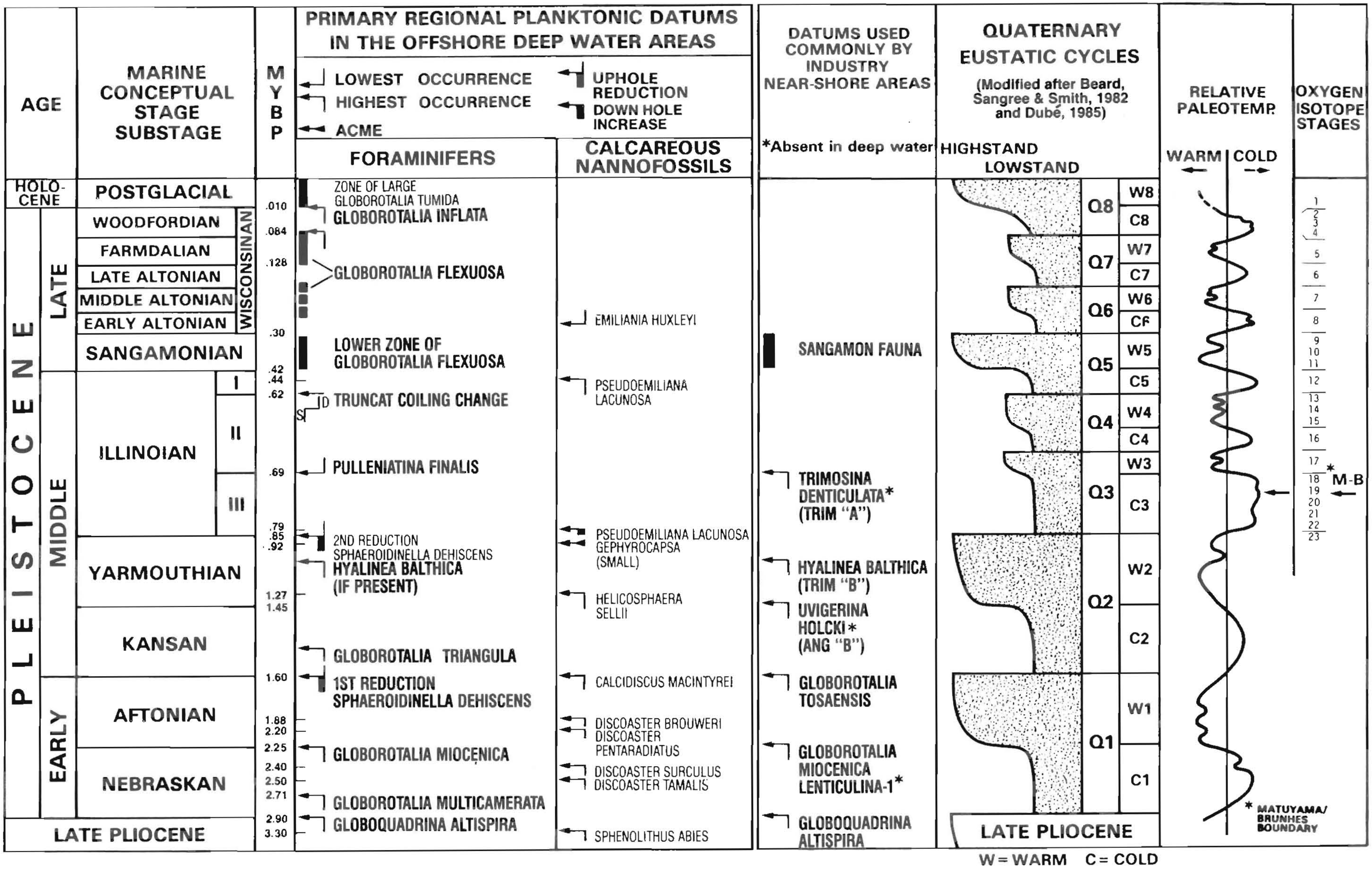

FIGURE 4. Model for Quaternary multiple-event stratigraphy Gulf of Mexico.

interpretation of seismic sequences and reflection patterns (as for example, identification of channeling). Accurate and consistent recognition of Pleistocene stages is important not only for correlation purposes but also for economic analyses.

It is becoming a rule of thumb that glacial intervals generally are more prone to contain potential reservoir sands than interglacial intervals. These sands are deposited during periods of lowstand. Glacial intervals commonly appear as hummocky or chaotic reflection patterns on the seismogram, whereas reflection patterns of interglacials tend to be more parallel. Also, lateral thickening of a glacial interval seen on a seismogram may indicate proximity to channel sandstones in the slope province (Fig. 5).

DISCUSSION

Marine Conceptual Stage Terminology

Since Fisk (1952) and Fisk and McFarlan (1955) related Pleistocene fluvial terrace deposits of the lower Mississippi River system to Midcontinent Pleistocene glacial-interglacial episodes, a number of earth scientists have followed suit [e.g., Akers and Holck (1957), Smith (1965), Beard (1969), Lamb and Beard (1972), Poag and Valentine (1976), Beard, Sangree, and Smith (1984), Stude (1984), and Dubé (1985)].

By dating volcanic ash beds associated with Midcontinent glacial tills, Boellstorff (1978) determined the chronology of the tills and surmised that classic sequences in the Afton, Iowa, region were younger than believed. This resulted in the stage terminology of the Afton, Iowa, region being out of phase with that of the Nebraska region.

Boellstorff (1978) concluded that the presently used North American Pleistocene Stage terms need either to be redefined and stratotypified, or abandoned. He believed, however, that there was a close correlation between the paleotemperature record in the Gulf of Mexico and the chronology of glacial and non-glacial conditions in the central United States. Therefore, a revised stage terminology based on the chronology of major climatic changes could be formulated, which would allow retention of the current North American Stage terms and permit their usage in the conceptual sense. Boellstorff recommended the possibility of adding the prefix "Neo" to the stage name. We agree with Boellstorff's conclusions, but believe a simple statement as to the conceptual nature of the stages is sufficient, as we have done in this paper.

Beard, Sangree, and Smith (1982) reinforced the application of conceptual stage terminology when they introduced eight Quaternary-Holocene eustatic cycles for the Gulf of Mexico region. Each cycle consists of a lowstand and highstand of sea level, much the same as was envisioned by Fisk (1940, 1944) in his substratum-topstratum concept of the fluvial terraces. The basic image of four major glaciations was preserved.

Dubé (1985) formally designated cold and warm subdivisions of the eight Quaternary eustatic cycles, which he matched up with the sequence of oxygen-isotope stages of Morley and Hays (1981), and a relative paleotemperature curve. Because oxygen-isotope stage boundaries are precisely dated back to about 0.9 m.y.B.P. (Imbrie et al., 1984), this allows for a comparative stage chronology back to about the base of the Illinoian Stage (based on paleomagnetic, oxygen-isotopic, paleontologic, or best-estimated ages).

From the base of the Illinoian Stage to the base of the Nebraskan Stage an absolute chronology is based mostly on correlation of planktonic foraminifer and calcareous nannofossil datums with the paleomagnetic time scale. This usually involves the calculation of sedimentation rates between established subchrons in deep-sea coreholes and estimating the age of the planktonic datum from its position. Near Afton, Iowa, Boellstorff (1977a, 1977b) dated an ash overlying two type C tills as 2.2 m.y.B.P. (± 0.2 m.y., estimated). He estimated that continental glaciation in the central United States within his Neo-Nebraskan Stage reached a peak at about 2.5 m.y.B.P.

There is broad concern to preserve the conceptual usage of Midcontinent Pleistocene stages even though, as pointed out by Boellstorff (1978), there is a need to redefine and stratotypify the terms. As Boellstorff inferred, the marine sequences of the Gulf of Mexico seemingly offer the best example and control, and could be used as a model for reevaluating the Midcontinent glacial-interglacial sequences. Some may ask, do we need stage names? Could we not just as easily number the warm and cold intervals and subdivide the section paleontologically using biostratigraphic zones?

It is common practice to subdivide geochronologic units, such as the Pleistocene Epoch, into chronostratigraphic units, such as stages and substages. The International Stratigraphic Guide (Hedberg 1976) defines Quaternary chronostratigraphic units as follows:

> "The basic principles to be used in dividing the Quaternary into chronostratigraphic units (stages) should be the same as for the other Phanerozoic strata, although different emphasis may be placed on the various means (climatic, magnetic, isotopic, etc.) used for time correlation. Carbon-14 dating has been particularly useful in the late Quaternary.
>
> Although it may often be impractical to establish continuous type sections, or comprehensive local unit-stratotypes for Pleistocene and Holocene chronostratigraphic units, the characterization of such units as the intervals between certain designated boundary-stratotypes would seem to be the best means for their definition."

A number of countries use Pleistocene stages, but these are mostly of a provincial nature, and none offer a comparable marine sequence as intensely studied as that of the Gulf of Mexico region. Shallow-water terrace deposits comprise the middle Pleistocene deposits of Italy, and these are overlain in turn mostly by nonmarine late Pleistocene deposits. We can, therefore, expect little more from Italy than a definition for the Tertiary/Quaternary boundary stratotype, which in itself is rife with controversy because

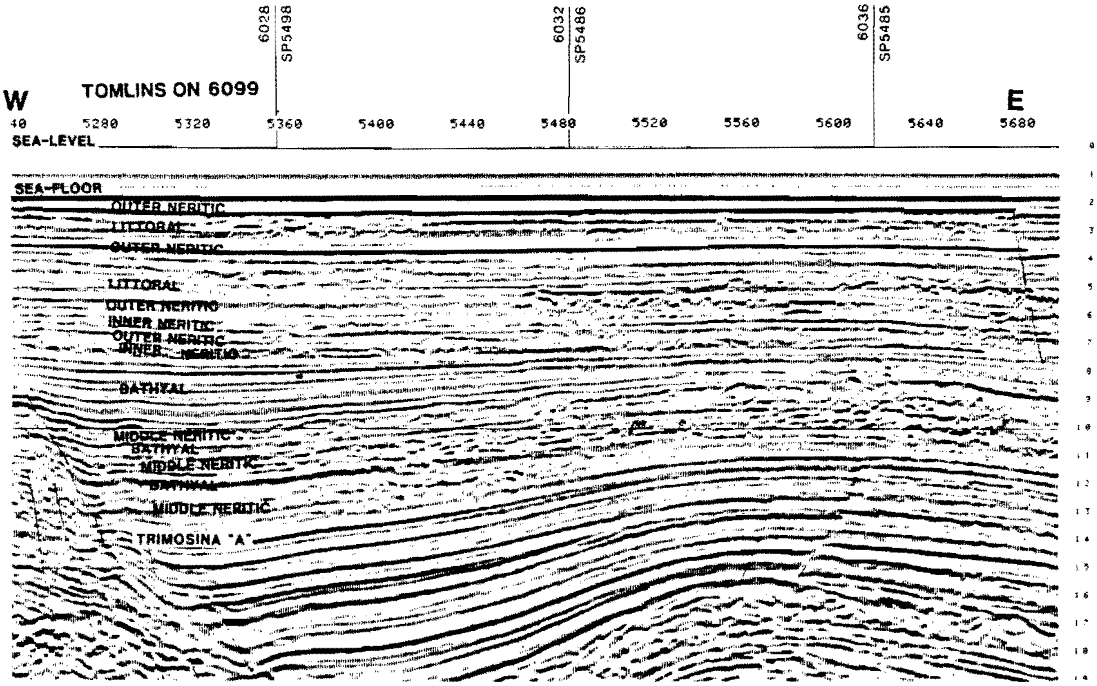

FIGURE 5. Seismic line from the southwest portion of West Cameron Federal lease area. Note alternating sequence of high-amplitude high-continuity reflectors. The former are interpreted to represent the fine-grained terrigenous deposits laid down during a high sea level interglacial and the latter to represent miscellaneous terrigenous sediments deposited during a low sea level. (After Lowrie and McDaniel Lowrie, 1985)

of spurious paleomagnetic events involved in a rather dubious compromise between the "Olduvai" proponents and the "Cold Guest Species" advocates. From our point of view, the Gulf of Mexico offers a desirable place and conditions for defining Pleistocene stages.

While not directly pertinent to this discussion, it should be mentioned that the Tertiary-Quaternary Boundary in the Gulf of Mexico has been positioned at other stratigraphic horizons, the most prominent being the extinction datum of *Discoaster brouweri* at 1.88 Ma, within the Aftonian interglacial stage, and the extinction datum of *Globorotalia miocenica* at 2.25 Ma, near the end of the Nebraskan glacial stage.

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EOCENE-OLIGOCENE BOUNDARY IN SOUTHEASTERN MISSISSIPPI AND SOUTHWESTERN ALABAMA: A STRATIGRAPHICALLY CONDENSED SECTION OF A TYPE 2 DEPOSITIONAL SEQUENCE

ERNEST A. MANCINI[1], BERRY H. TEW[2], and LAURA A. WATERS[3]

[1]Geological Survey of Alabama and University of Alabama, Tuscaloosa, Alabama;
[2]Geological Survey of Alabama, Tuscaloosa, Alabama; and
[3]Amoco Production Company, New Orleans, Louisiana

ABSTRACT

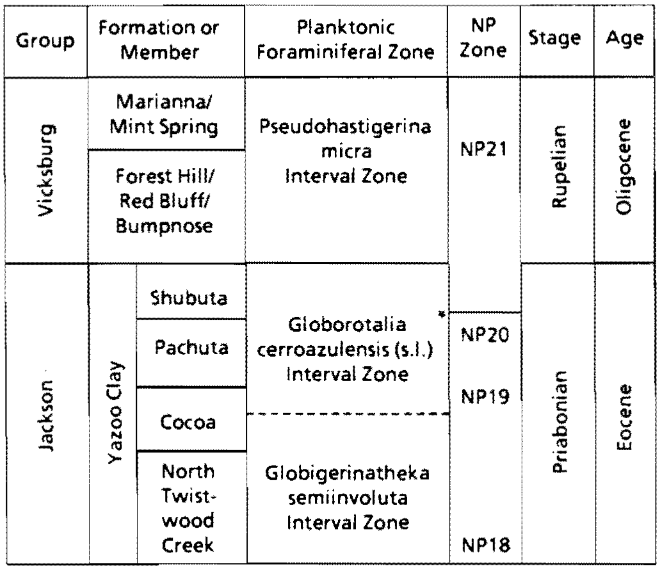
Micropaleontologists using planktonic foraminifera for biostratigraphic age correlation differ on the placement of the Eocene-Oligocene boundary from those workers utilizing calcareous nannoplankton. Employing the stratigraphic distribution of the planktonic foraminifera recovered from the upper Eocene Yazoo Clay and lower Oligocene Red Bluff Clay/Bumpnose Limestone, Forest Hill Sand and Mint Spring Marl Member of the Marianna Limestone in southeastern Mississippi and southwestern Alabama, the epoch boundary is recognized to be at or near the top of the Yazoo Clay. The extinctions of the key calcareous nannoplankton species occur at a lower stratigraphic level. The dilemma regarding the placement of the Eocene-Oligocene boundary exists probably because changes in faunal and floral assemblages were gradual during the late Eocene through the early Oligocene and because this epoch boundary represents a stratigraphically condensed section of a Type 2 depositional sequence. No dramatic faunal or floral changes should be expected at the Eocene-Oligocene boundary because of the absence of a major drop in sea level and because no sudden climatic changes occurred at this time.

In southeastern Mississippi and southwestern Alabama, the Cocoa Sand, Pachuta Marl and Shubuta Clay Members of the Yazoo Clay, the Red Bluff Clay, the Bumpnose Limestone, and the Forest Hill Sand comprise an unconformity-bounded, Type 2 depositional sequence that accumulated during the TE3.3 coastal onlap cycle of the Td supercycle. The contact of the Cocoa/Pachuta with the underlying North Twistwood Creek Clay Member of the Yazoo Clay (highstand regressive deposits of the underlying depositional sequence) is a Type 2 unconformity. The transgressive deposits of the TE3.3 sequence consist of the Cocoa and Pachuta. The condensed section of the sequence includes the Shubuta (lower condensed section deposits) and Red Bluff/Bumpnose (upper condensed section deposits). The Shubuta-Red Bluff/Bumpnose contact, which approximates the Eocene-Oligocene boundary, is a surface of maximum starvation associated with the greatest landward transgression of the coastline during a relative rise in sea level. The Forest Hill Sand overlies the Red Bluff/ Bumpnose and comprises the highstand regressive deposits of the sequence. The contact of the Forest Hill with the overlying Mint Spring is a Type 2 unconformity. The Mint Spring represents the transgressive deposits of the overlying sequence.

The use of stratigraphically condensed sections, sequence boundaries and genetic depositional sequences has the potential to be a useful correlation tool for resolving stratigraphic and depositional problems in the Gulf Coastal Plain. The age of a condensed section within a specific depositional sequence should be synchronous worldwide.

INTRODUCTION

Micropaleontologists using planktonic foraminifera for biostratigraphic age correlation differ on the placement of the Eocene-Oligocene boundary from those workers utilizing calcareous nannoplankton. This epoch boundary is defined worldwide on the basis of the extinction of the planktonic foraminifera, *Hantkenina* and subspecies of *Globorotalia cerroazulensis* (Toumarkine and Bolli, 1970; Stainforth and others, 1975; Blow, 1979; Stainforth and Lamb, 1981; Van Couvering and others, 1981; Snyder and others, 1984; Berggren and others, 1985). The Eocene-Oligocene boundary is drawn at the top of the planktonic foraminiferal *Globorotalia cerroazulensis* (s.l.) Interval Zone of Stainforth and others (1975) by many of these micropaleontologists (Fig. 1). However, the extinctions of the key calcareous nannoplankton species, *Discoaster saipanensis* Bramlette and Riedel, *Discoaster barbadiensis* Tan Sin Hok, and *Reticulofenestra reticulata* (Gartner and Smith), occur at a lower stratigraphic elevation than the last occurrence of diagnostic planktonic foraminifera in surface exposures and in coreholes drilled in the Gulf and Atlantic Coastal Plains, Gulf of Mexico, and Atlantic Ocean (Gartner, 1971; Stainforth and Lamb, 1981; Bybell, 1982; Poore and others, 1982; Snyder and others, 1984; Bybell and



Group	Formation or Member		Planktonic Foraminiferal Zone	NP Zone	Stage	Age
Vicksburg	Marianna/ Mint Spring		Pseudohastigerina micra Interval Zone	NP21	Rupelian	Oligocene
	Forest Hill/ Red Bluff/ Bumpnose					
Jackson	Yazoo Clay	Shubuta	Globorotalia cerroazulensis (s.l.) Interval Zone *		Priabonian	Eocene
		Pachuta		NP20		
		Cocoa	Globigerinatheka semiinvoluta Interval Zone	NP19		
		North Twist-wood Creek		NP18		

* Eocene-Oligocene boundary recognized on basis of calcareous nannoplankton

Figure 1. Upper Eocene and lower Oligocene lithostratigraphy, biostratigraphy and chronostratigraphy in southeastern Mississippi and southwestern Alabama.

others, 1986; Mancini and Waters, 1986) leading to a difference of opinion among micropaleontologists.

To resolve this difference in the placement of the Eocene-Oligocene boundary, Hardenbol and Berggren (1978) and Van Couvering and others (1981) have suggested using the last occurrence of *D. barbadiensis* and *D. saipanensis* for defining the calcareous nannoplankton NP20/NP21 zonal boundary of Martini (1971) and point out that these extinctions occur stratigraphically below the Eocene-Oligocene boundary. Therefore, the epoch boundary is within the NP21 Zone since these authors recognized the boundary based on the extinction of the planktonic foraminifera, *Hantkenina* and subspecies of *Globorotalia cerroazulensis*.

It has been suggested that the dilemma regarding the placement of the Eocene-Oligocene boundary exists probably because no major floral or faunal changes occurred during the latest Eocene (Snyder and others, 1984; Frederiksen, 1986). Because changes in faunal and floral assemblages were gradual during the late Eocene through the early Oligocene, Snyder and others (1984) concluded that no sudden climatic change took place at this time. Loutit and others (1983) described the Eocene-Oligocene boundary as a stratigraphically condensed section and contend that no dramatic faunal or floral changes should be evident at this epoch boundary because of the absence of a major drop in sea level.

The objectives of this paper are to review the lithostratigraphy and biostratigraphy of the upper Eocene and lower Oligocene strata as exposed in southeastern Mississippi and southwestern Alabama (Fig. 2), to determine the stratigraphic elevation at which faunal and sedimentological changes occur, to study the spatial distribution of faunal and sedimentological changes, and to evaluate the usefulness of sequence stratigraphy in correlating and interpreting Gulf Coastal Plain strata.

LITHOSTRATIGRAPHY AND LITHOLOG

The Yazoo Clay, which is included in the
Group, consists of four members (Fig. 1). Thes
ascending order, the North Twistwood Cre
Member, the Cocoa Sand Member, the Pachu
Member, and the Shubuta Clay Member. Th
Twistwood Creek Clay Member consists of g
gray, fossiliferous, calcareous, micaceous, silty
marl. Its lower contact is conformable with the
Branch. The Moodys Branch is comprised of g
gray, glauconitic, fossiliferous, calcareous, fine
sand and silty, sandy marl. The Cocoa Sand M
comprised of grayish-green, fossiliferous, cal
medium- to fine-grained sand and sandy marl. T
contact of the Cocoa with the North Twistwood
sharp, disconformable and burrowed. The Pach
Member includes greenish-gray, glauconitic,
ferous, phosphatic, argillaceous, silty, sandy
limestone. Its contact with the Cocoa is gra
The Cocoa is difficult to distinguish from the P
southern Clarke County, Alabama, and, in fact,
be present in this area (Fig. 2). The Shubuta co
greenish-gray, glauconitic, fossiliferous, pho
argillaceous, silty marl and calcareous clay. T
contact of the Shubuta is gradational with the
The Shubuta is difficult to recognize from the
in southern Clarke and Washington Counties,
and therefore only the massive marl ("blue clay
top of the Yazoo Clay is assigned to the Shub
3). The Shubuta ("blue clay") includes only the u
1.2 m (4 ft) of marl in the Yazoo Clay at St.
Quarry (Fig. 4) and the uppermost 0.9 to 1.2 m (
of marl in the Yazoo Clay at Little Stave Creek
to 0.6 m (1 to 2 ft) of argillaceous, fossilifer
above the massive marl ("blue clay") and below

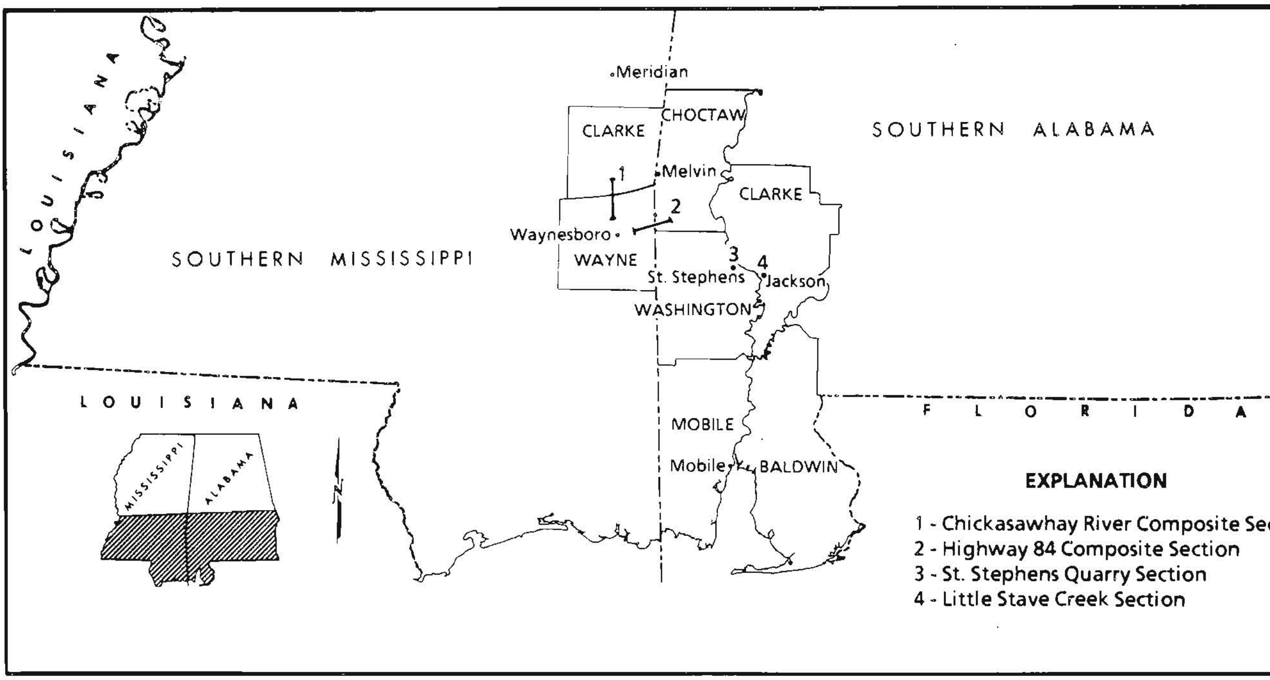


Figure 2. Location map for upper Eocene and lower Oligocene sections studied in southeastern Mississippi and southwestern Alabama.

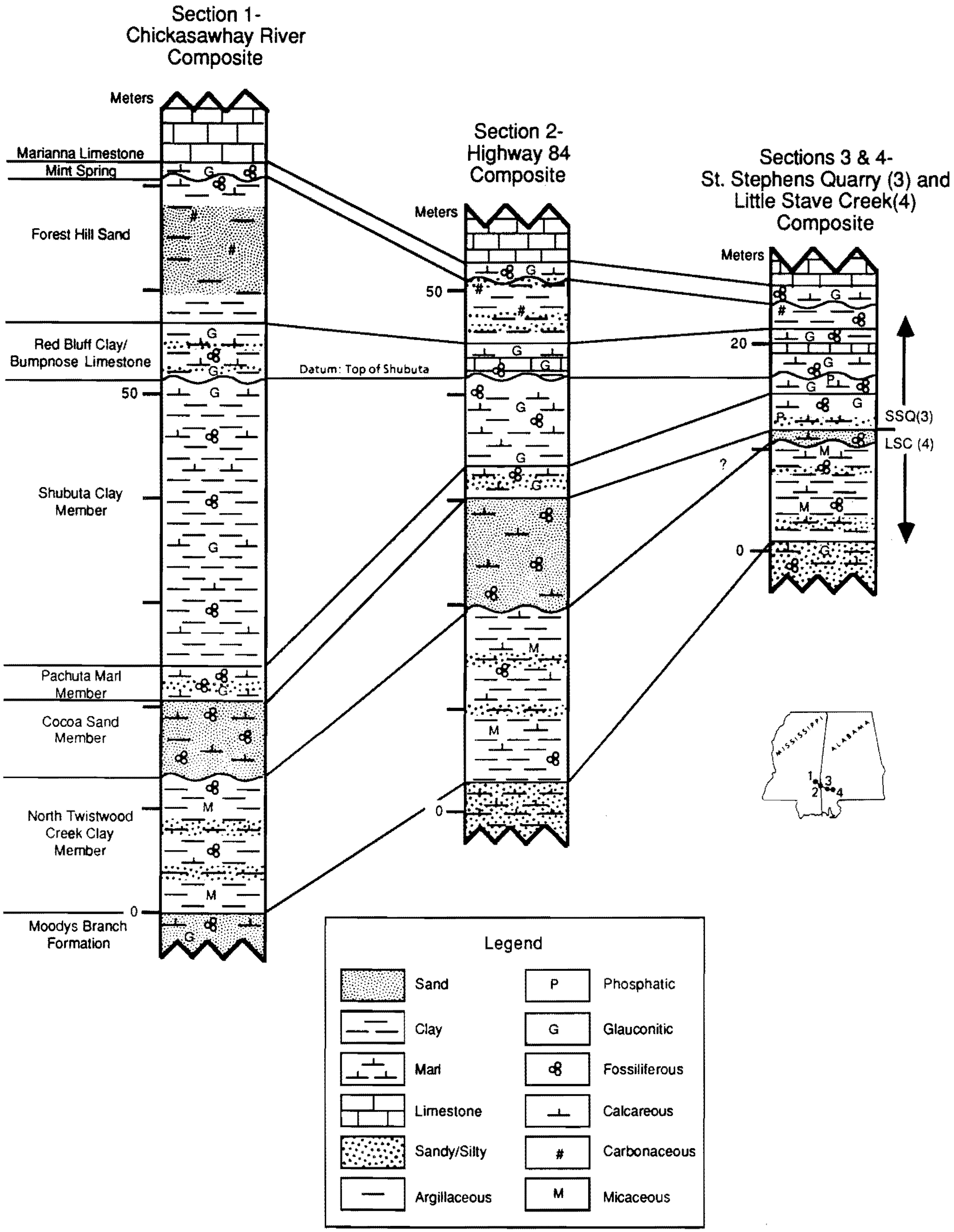
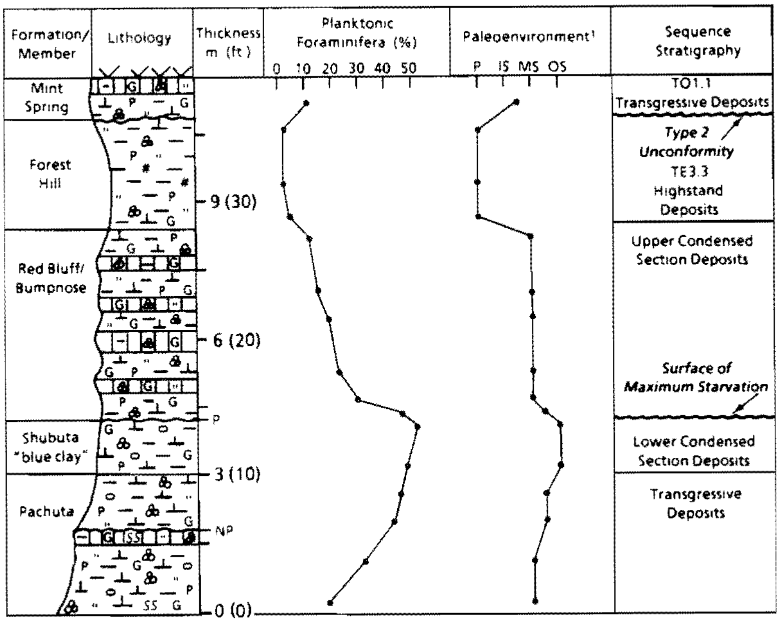


Figure 3. Correlation of the upper Eocene and lower Oligocene sections studied in southeastern Mississippi and southwestern Alabama.



¹ Prodelta (P), Inner Shelf (IS), Middle Shelf (MS), Outer Shelf (OS)
NP = Eocene-Oligocene boundary recognized on the basis of calcareous nannoplankton
P = Eocene-Oligocene boundary recognized on the basis of planktonic foraminifera.

Figure 4. Upper Eocene and lower Oligocene lithostratigraphy, paleoenvironments, vertical changes in foraminiferal assemblages, and sequence stratigraphy at the St. Stephens Quarry section, Washington County, Alabama. See Figure 2 for section location.

limestone bed is assigned to the Red Bluff/Bumpnose (Fig. 4). The lower contact of the basal Red Bluff/ Bumpnose fossiliferous marl with the Shubuta massive marl ("blue clay") is disconformable and can be marked by burrowing, quartz, glauconite, phosphate grains, and shell hash.

The Red Bluff Clay, Forest Hill Sand, Bumpnose Limestone, and Mint Spring Marl Member of the Marianna Limestone are included in the Vicksburg Group (Fig. 1). The Red Bluff consists of olive-gray, glauconitic, fossiliferous, calcareous silty clay and silty limestone. The Red Bluff and Bumpnose are time-equivalent lithofacies which intertongue in the study area and are therefore considered as one unit. The Bumpnose Limestone includes white, glauconitic, fossiliferous, argillaceous, silty marl and limestone. The Forest Hill is a time-equivalent lithofacies of the Red Bluff and Bumpnose; however, its distinct lithology make it recognizable as a separate unit in the study area. It is comprised of dark gray, carbonaceous, slightly fossiliferous and calcareous, laminated silty clay and fine-grained sand. Its contact with the Red Bluff/ Bumpnose is gradational. The Mint Spring Marl Member is the basal unit of the Marianna Limestone. The Mint Spring consists of greenish-gray, glauconitic, fossiliferous, argillaceous marl. The lower contact of the Mint Spring with the Forest Hill is sharp, disconformable, burrowed, and can be marked by the presence of rounded clay clasts. The Marianna is comprised of pale orange to white, fossiliferous, argillaceous, silty marl and limestone.

The upper Eocene and lower Oligocene strata of southeastern Mississippi are recognized as a terrigenous clastic dominated sequence, while the time-equivalent lithofacies of south-central Alabama have been described as part of a carbonate dominated sequence (Cooke, 1918; MacNeil, 1944; Huddlestun and Toulmin, 1965; Hazel and others, 1980; Mancini and Waters, 1986). The strata intertongue in southwestern Alabama, making recognition of these lithofacies difficult in this area. Overall, there is an increase in the amounts of glauconite, phosphate and calcium carbonate and a decrease in terrigenous material in the upper Eocene and lower Oligocene strata from southeastern Mississippi to southwestern Alabama. In addition, there is an overall increase in the amounts of calcium carbonate, glauconite, and phosphate and a decrease in terrigenous material stratigraphically within the Yazoo Clay in the study area. On the other hand, there is an overall decrease in the amounts of calcium carbonate, glauconite, and phosphate and an increase in terrigenous material stratigraphically within the Red Bluff/ Bumpnose-Forest Hill sequence.

There is also a change in the thickness of the upper Eocene and lower Oligocene section from southeastern Mississippi to southwestern Alabama (Fig. 3). The Yazoo Clay is about 61 m (200 ft) thick in Yazoo County, west-central Mississippi and thins eastward to 22 m (72 ft) in Clarke County, southwestern Alabama (May, 1974; Toulmin, 1977). Stratigraphic thickness changes in the Yazoo members from Clarke and Wayne Counties, Mississippi, to southern Clarke and Washington Counties, Alabama, are as follows: North Twistwood Creek (13 to 10 m (43 to 32 ft)), Cocoa/Pachuta (10 to 3.6+ m (33 to 12+ ft)), and Shubuta (28 to 1.2 m (92 to 4 ft)) (Deboo, 1965; May, 1974; Toulmin, 1977; Mancini and Waters, 1986). The Red Bluff/Bumpnose thins from 6 m (20 ft) in Wayne County, Mississippi, to 4.3 m (14 ft) in Washington County and 3.6 m (12 ft) in southern Clarke County, Alabama; and the Forest Hill is 14 m (46 ft) thick in Wayne County, Mississippi, and thins to 2.4 m (8 ft) in Washington County and pinches out in southern Clarke County, Alabama (May, 1974; Mancini and Waters, 1986).

BIOSTRATIGRAPHY AND PALEOECOLOGY

The upper Eocene and lower Oligocene planktonic foraminiferal zonation utilized in this study was first established by Bolli (1957, 1966, 1972) and later modified by Stainforth and others (1975) and Stainforth and Lamb (1981). This zonation and the planktonic foraminiferal zonation of Blow (1979) have been used widely as accepted biostratigraphic standards for warm-water areas of the world, including the Gulf Coastal Plain region.

The North Twistwood Creek, in part, and the Cocoa were assigned by Barker (in Blow, 1979) to the *Porticulasphaera semiinvoluta* Partial Range Zone of Blow (1979), which is approximately equivalent to the *Globigerinatheka semiinvoluta* Interval Zone of Stainforth and others (1975) (Fig. 1). This assignment was based on the occurrence of *Globorotalia cerroazulensis pomeroli* Toumarkine and Bolli, *Globorotalia cerroazulensis cerroazulensis* (Cole), *Hantkenina alabamensis* Cushman and the absence of *Cribrohantkenina inflata* (Howe) and *Globorotalia cerroazulensis cocoaensis* Cushman (Mancini and Waters, 1986). The Pachuta and Shubuta have been placed in the *Globorotalia cerroazulensis* (s.l.) Interval Zone of Stainforth and others (1975) by Mancini (1979) and Mancini and Waters (1986) based on the presence of *G. cerroazulensis cerroazulensis*, *G. cerroazulensis cocoaensis*, *H. alabamensis*, *C. inflata*, *Globigerina ampliapertura* Bolli, and *Globigerina gortanii* (Borsetti). The Red Bluff, Bumpnose, Forest Hill, Mint Spring, and basal Marianna have been assigned to the *Pseudohastigerina micra* Interval Zone of Stainforth and Lamb (1981) by Mancini and Waters (1986) based on the occurrence of *Pseudohastigerina micra* (Cole), *G. ampliapertura*, *Globigerina ciperoensis* Bolli, and *Globigerina tapuriensis* Blow and Banner and the absence of subspecies of *Globorotalia cerroazulensis* and *Hantkenina* that are considered autochthonous.

Utilizing the planktonic foraminiferal zones recognized in these upper Eocene and lower Oligocene strata and the Cenozoic geochronology published by Berggren and others (1985), geologic age determinations can be made for the respective members and formations (Fig. 1). The North Twistwood Creek Clay, Cocoa Sand, Pachuta Marl, and Shubuta Clay Members of the Yazoo Clay are Priabonian in age. The Red Bluff Clay, Bumpnose Limestone, Forest Hill Sand, Mint Spring Marl Member and at least the basal Marianna Limestone are Rupelian in age.

The foraminiferal assemblages present in the upper Eocene and lower Oligocene strata in southeastern Mississippi and southwestern Alabama are useful in determining probable paleoenvironments and paleobathymetries of these units. These determinations are made utilizing the works of Lowman (1949), Phleger (1965), Bandy (1964), Walton (1964), Loep (1965), and Murray (1973) which concern modern foraminiferal distributions. In southeastern Mississippi and southwestern Alabama, the North Twistwood Creek has a foraminiferal assemblage dominated by nonionids, planulinids, cibicidids and agglutinates indicating inner to middle shelf depositional conditions. Although the Cocoa in southeastern Mississippi and southwestern Alabama contains a benthonic foraminiferal assemblage similar to that of the North Twistwood Creek, it has the lowest percentage of planktonic foraminifera of any of the Yazoo units, which suggests an inner shelf environment (Fig. 5). The Pachuta is characterized by a middle to outer shelf assemblage having abundant nonionids, cibicidids, buliminids and uvigerinids and a high percentage of planktonic foraminifera in southeastern Mississippi and southwestern Alabama. The Shubuta foraminiferal assemblage in southeastern Mississippi and southwestern Alabama is dominated by bolivinids, buliminids and uvigerinids and the highest percentage of planktonic foraminifera indicating outer shelf depositional conditions. In southeastern Mississippi and southwestern Alabama, the Red Bluff/Bumpnose is

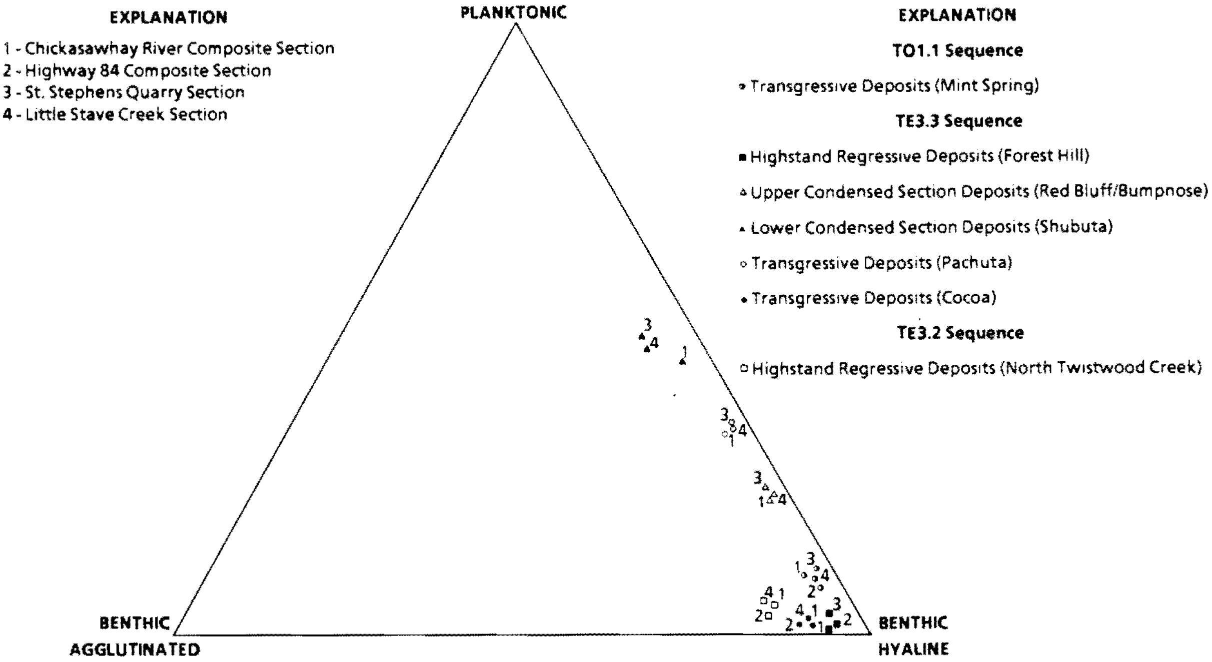


Figure 5. Diagram illustrating the change in composition of foraminiferal assemblages in southeastern Mississippi and southwestern Alabama. See Figure 2 for location of sections studied.

typified by a middle shelf assemblage having abundant uvigerinids, buliminids, nodosariids, discorbids and cibicidids and a moderate percentage of planktonic foraminifera. The Forest Hill represents deltaic deposition and in southeastern Mississippi and southwestern Alabama a sparse assemblage of foraminifera dominated by benthonics was recovered from this unit, which suggests a prodelta environment. The Mint Spring in southeastern Mississippi and southwestern Alabama has an inner to middle shelf assemblage of nodosariids and cibicidids and a low percentage of planktonic foraminifera.

Huff (1970), from his work with ostracodes and foraminifera recovered from upper Eocene strata of Mississippi, reported the following paleoenvironments for the upper Eocene strata: Moodys Branch, inner to middle shelf; North Twistwood Creek, inner to middle shelf; Cocoa, inner shelf; Pachuta, middle to outer shelf; and Shubuta, outer shelf. Hazel and others (1980), in their study of the ostracode assemblages of the lower Oligocene strata of southeastern Mississippi and southwestern Alabama, concluded that the Forest Hill in southwestern Alabama accumulated in a prodelta environment and the Red Bluff/Bumpnose, Mint Spring and Marianna were deposited in deeper water environments.

Utilizing the assemblages of microfossils in these strata, bathymetric trends can be recognized. In general, there is an increase in the percentage of planktonic foraminifera from southeastern Mississippi to southwestern Alabama indicating a change in depositional conditions (Fig. 5). This change is probably the result of the combined effects of an increase in water depth and a reduction in terrigenous influx at the depositional site. In addition, there was an overall increase in water depth in the Priabonian (Cocoa to Shubuta) (Fig. 4). Cocoa deposition records the initiation of a relative rise in sea level in the Priabonian. A relative fall in sea level began at the close of the Priabonian and continued through the early Rupelian with deposition of the Red Bluff/Bumpnose and Forest Hill. The Mint Spring represents a return to middle shelf depositional conditions in the Rupelian during a relative rise in sea level that began after the Forest Hill delta prograded across Mississippi and into southwestern Alabama.

DEPOSITIONAL SEQUENCES, GLOBAL SEA LEVEL AND COASTAL ONLAP CYCLES

Cyclic global changes in eustatic sea level and coastal onlap (progressive landward encroachment of coastal deposits) during the Cenozoic have been recognized by Vail and others (1977) and Baum (1986). Baum (1986) reports 19 global unconformities, which he uses to divide the Paleogene into 18 depositional sequences associated with eustatic sea level cycles. His TE3.3 cycle (38 to 36 ma) of the Td supercycle (40.5 to 29 ma) is bounded by a basal Type 2 unconformity in the Priabonian and an upper Type 2 unconformity in the Rupelian. A stratigraphically condensed section is recognized near the Priabonian-Rupelian boundary (Loutit and others, 1983; Baum, 1986). According to Vail and others (1977) and Baum (1986), a rise in sea level occurred worldwide during the Priabonian, and a fall in sea level is apparent beginning at the close of the Priabonian and continuing into the Rupelian. A subsequent major rise in sea level is evident worldwide during the Rupelian (Vail and others, 1977; Baum, 1986). The TE3.3 depositional cycle is a function of changes in sea level, subsidence, and sedimentation rates.

The strata deposited during the TE3.3 coastal onlap cycle have been recognized as an unconformity-bounded, Type 2 depositional sequence by Baum (1986). A Type 2 depositional sequence represents a relatively conformable succession of genetically related strata bounded at the base by a Type 2 unconformity and at the top by a Type 1 or Type 2 unconformity (Mitchum and others, 1977; Baum, 1986). The sequence (Fig. 6) typically consists of a composite of the following: basal Type 2 unconformity, regressive or aggradational shelf margin deposits, transgressive surface, transgressive deposits, condensed section with a surface of maximum starvation, regressive highstand deposits, and Type 1 or 2 unconformity (Baum, 1986).

A Type 2 unconformity is formed when the rate of sea level fall is less than subsidence at the shelf edge but exceeds the rate of subsidence on the inner portion of the shelf (Vail and others, 1984). These authors characterize a Type 2 unconformity as a downward shift of coastal onlap to a position at or landward of the shelf edge, with subaerial exposure of the landward portion of the shelf. Unlike a Type 1 unconformity, which is characterized by an abrupt downward shift in coastal onlap below the shelf edge, a Type 2 unconformity exhibits no evidence of canyon cutting along the shelf edge or valley entrenchment on the shelf (Vail and others, 1984).

Stratigraphically condensed sections are thin, fine-grained (shales or micrites) marine units which are the result of slow deposition on the middle to outer shelf and in basinal areas (Loutit and others, 1983; Vail and others, 1984). A condensed section, along with the associated surface of maximum starvation, commonly marks the greatest water depth and maximum landward transgression of the coastline during sea level rise and usually develops when the rate of sea level rise is greater than the rate of sediment accumulation, resulting in the depositional site shifting landward, producing low sedimentation rates seaward (Vail and others, 1984). Condensed sections are commonly associated with fairly continuous beds of burrowed or lithified sediment (hardgrounds or omission surfaces), which exhibit high concentrations of pelletal glauconite, phosphate, volcanic ash, planktonic organisms, and/or radioactive minerals (Vail and others, 1984).

DISCUSSION

As mentioned previously, there exists a dilemma regarding the placement of the Eocene-Oligocene boundary in the Gulf Coastal Plain based on biostratigraphic zonations. Snyder and others (1984) suggest that this problem exists because no major floral or faunal changes

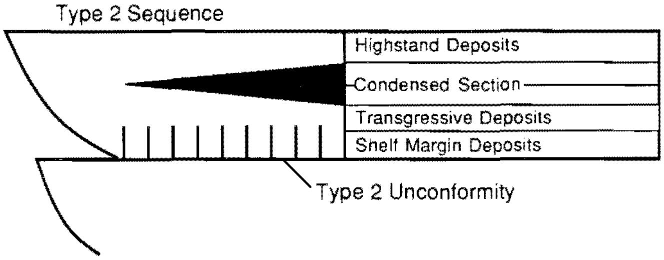


Figure 6. Sequence stratigraphy of a Type 2 depositional sequence modified from Baum (1986).

occurred during the latest Eocene. Loutit and others (1983) described the Eocene-Oligocene boundary as a condensed section and contend that no dramatic faunal or floral changes should be evident at this epoch boundary since no major drop in sea level occurred. This study of upper Eocene and lower Oligocene strata and associated microfaunas in southeastern Mississippi and southwestern Alabama indicates a relative fall in sea level after deposition of the North Twistwood Creek, followed by an overall increase in water depths through the remainder of the Priabonian (Cocoa to Shubuta). Subsequent shallowing of water depths at the close of the Priabonian and continuing into the early Rupelian is represented by the shelf deposits of the Red Bluff/Bumpnose and deltaic deposits of the Forest Hill (Fig. 4). A relative fall in sea level is indicated after the deposition of the Forest Hill, followed by a relative rise in sea level resulting in extensive carbonate shelf sedimentation (Mint Spring and Marianna). The observed water depth variations are consistent with worldwide sea level changes in the Priabonian and Rupelian as reported by Vail and others (1977) and later refined for the Gulf Coastal Plain by Baum (1986). The vertical variations in faunal composition and lithology observed in these strata can be explained by changes in sea level, subsidence, and sedimentation rates through time. In addition, spatial variations in fauna, lithology, and stratigraphic thickness across the study area are probably due to bathymetric changes and differential sedimentation rates at the depositional sites.

The concepts of sequence stratigraphy developed by Vail and others (1977), Mitchum and others (1977), Vail and others (1984), and Baum (1986) can be used to better understand the stratigraphic, lithologic, and paleontologic changes observed in the upper Eocene and lower Oligocene strata in southeastern Mississippi and southwestern Alabama. Baum (1986) reported that the Cocoa Sand, Pachuta Marl, and Shubuta Clay Members of the Yazoo Clay, along with the Red Bluff Clay/Bumpnose Limestone and Forest Hill Sand comprise an unconformity-bounded, Type 2 depositional sequence that was deposited during the TE3.3 coastal onlap cycle of the Td supercycle. In the study area, this depositional sequence consists of a basal Type 2 unconformity, transgressive deposits (Cocoa and Pachuta), lower condensed section deposits (Shubuta), upper condensed section deposits (Red Bluff/Bumpnose), highstand regressive deposits (Forest Hill), and an upper Type 2 unconformity. The North Twistwood Creek Clay Member of the Yazoo comprises the highstand regressive deposits of the underlying TE3.2 sequence, and the Mint Spring represents the transgressive deposits of the overlying TO1.1 sequence (Fig. 7).

The contact of the inner to middle shelf clays and marls of the North Twistwood Creek with the sands and glauconitic, sandy marls of the Cocoa/Pachuta is the basal Type 2 unconformity of the TE3.3 sequence. This contact is sharp, disconformable and burrowed. The presence of rounded phosphate grains, quartz pebbles, and shell hash at the base of the Cocoa/Pachuta in parts of southwestern Alabama indicates that the transgressive surface of the TE3.3 sequence is coincident with the unconformity. The inner shelf, sands and sandy marls of the Cocoa and the middle to outer shelf, glauconitic, sandy marls and limestones of the Pachuta represent transgressive deposits. These are overlain by the outer shelf, glauconitic, phosphatic marls and calcareous clays of the Shubuta, which comprise the lower condensed section deposits. The Cocoa-Pachuta and Pachuta-Shubuta contacts are gradational, and these strata exhibit a progressive upward increase in water depths. The middle shelf, glauconitic clays, marls and limestones of the Red Bluff/Bumpnose represent the upper condensed section deposits. The Shubuta-Red Bluff/Bumpnose contact is disconformable and may be marked by burrowing, glauconite, phosphate grains, quartz and shell hash. This contact represents the surface of maximum starvation (time of minimum sedimentation) within the condensed section and is associated with maximum transgression of the shoreline during a relative rise in sea level. Above this surface, which represents a marine hiatus, the depositional sequence is characterized by regressive deposits associated with a relative fall in sea level. The change in stratigraphic thickness of the Shubuta and Red Bluff/Bumpnose and lithofacies changes in these units from southeastern Mississippi to southwestern Alabama (Fig. 3) are also characteristic of a stratigraphically

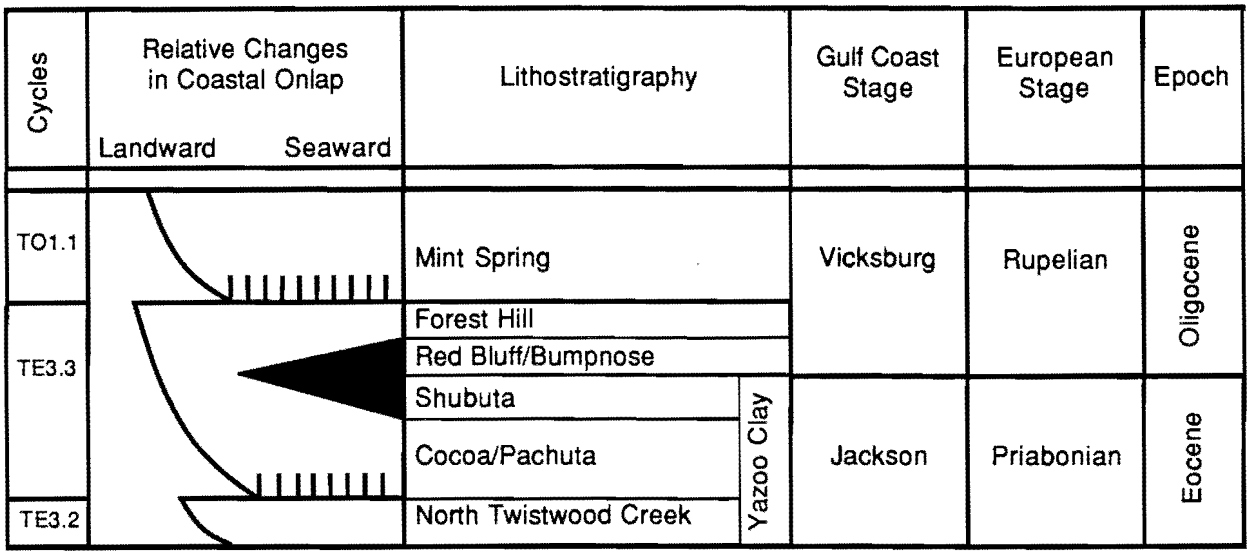


Figure 7. Sequence stratigraphy of upper Eocene and lower Oligocene strata in southeastern Mississippi and southwestern Alabama.

condensed section. Stratigraphic thinning, increases in the amounts of calcium carbonate, glauconite and phosphate, and decreases in the amounts of terrigenous material in the Cocoa/Pachuta, Shubuta and Red Bluff/Bumpnose to the southeast from Mississippi into Alabama indicate that the shelf edge is to the southeast of the surface exposures in southwestern Alabama. These stratigraphic and sedimentologic changes show progressive sediment starvation in that direction as a result of increased water depths. The deltaic carbonaceous clays and sands of the Forest Hill represent the highstand regressive deposits of the TE3.3 depositional sequence. These deposits prograded to the southeast from Mississippi into Alabama. The Red Bluff/Bumpnose contact with the Forest Hill is gradational. The middle shelf, glauconitic marls of the Mint Spring represent transgressive deposits of the TO1.1 depositional sequence, and the contact of the Forest Hill with the Mint Spring is the basal Type 2 unconformity of this sequence. This contact is sharp, disconformable and burrowed. The presence of rounded clay clasts at the base of the Mint Spring in parts of southeastern Mississippi indicate that the transgressive surface of the TO1.1 sequence is coincident with the unconformity.

The stratigraphic, lithologic and paleontologic relationships observed in the upper Eocene (Priabonian) and lower Oligocene (Rupelian) strata of southeastern Mississippi and southwestern Alabama indicate that these sediments represent a Type 2 depositional sequence (TE3.3) with the Eocene-Oligocene boundary corresponding to the surface of maximum starvation. No major drop in sea level occurred at the epoch boundary. Rather, the Priabonian sediments were deposited during a relative rise in sea level, so that no dramatic faunal or floral changes should be evident at the boundary. Without dramatic environmental changes, extinctions and appearances of organisms were gradual through the late Eocene and into the early Oligocene. Therefore, late Eocene calcareous nannoplankton species became extinct prior to certain late Eocene planktonic foraminiferal species. Such extinction patterns should be expected under gradually changing environmental conditions with environmentally sensitive organisms disappearing first.

Based on the characteristics of this Type 2 depositional sequence (TE3.3), it would appear that chronostratigraphic (stage) boundaries which are commonly defined on the basis of biostratigraphic criteria may not correspond to depositional sequence or lithostratigraphic boundaries. For example, within the TE3.3 sequence, the Cocoa-Pachuta-Shubuta comprise a transgressive genetic unit and the Red Bluff-Bumpnose-Forest Hill comprise a regressive genetic unit. The North Twistwood Creek is a part of a preceding regressive genetic unit. Further, the Cocoa and Pachuta represent the transgressive deposits of the TE3.3 sequence; as such, they are, in part, lithofacies equivalents. The Cocoa has been assigned to the early Priabonian *Globigerinatheka semiinvoluta* Interval Zone of Stainforth and others (1975) based primarily on the absence of *Cribrohantkenina inflata* and *Globorotalia cerroazulensis cocoaensis*. It is highly probable, however, that the Cocoa accumulated in the early through late Priabonian. The calcareous nannoplankton species, *Isthmolithus recurvus* Deflandre, has been reported from the Cocoa and Pachuta by Levin and Joerger (1967) and Siesser (1983). This species is recognized worldwide as an upper Eocene index form (Gartner, 1971). Although this species occurs in the *Globorotalia cerroazulensis* (s.l.) Interval Zone, its first appearance is in the *Globigerinatheka semiinvoluta* Interval Zone (Gartner, 1971). Therefore, its occurrence in the Cocoa is not conclusive as to the time-equivalency of these units; however, the observed lithofacies relations of these units implies that the Cocoa and Pachuta are at least, in part, time-equivalent. The absence of the key planktonic foraminifera in the Cocoa may be a result of depositional conditions inherent in inner shelf environments rather than evolution. In addition, it might be the case that the North Twistwood Creek should be recognized as a formation rather than as a member of the Yazoo Clay. This observation is based on its disconformable contact with the Cocoa and the lithology and relative uniform thickness of the North Twistwood Creek throughout the study area, which make it easily recognizable and mapped in the area.

The use of stratigraphically condensed sections, sequence boundaries, and genetic depositional sequences has the potential to be a useful correlation tool for resolving stratigraphic and depositional problems in the Gulf Coastal Plain. The age of the condensed section is useful in identifying the depositional sequence, and the nature of the sequence boundaries and characteristics of the units comprising the sequence can be utilized in deciphering the depositional history of the strata. Condensed sections appear to have considerable potential for recognition of other epoch boundaries and for chronostratigraphic correlation in the Gulf Coastal Plain. The age of a condensed section within a specific depositional sequence should be synchronous worldwide.

CONCLUSIONS

1. Employing the stratigraphic distribution of the planktonic foraminifera recovered from the upper Eocene (Priabonian) Yazoo Clay and lower Oligocene (Rupelian) Red Bluff Clay/Bumpnose Limestone, Forest Hill Sand and Mint Spring Marl Member of the Marianna Limestone in southeastern Mississippi and southwestern Alabama, the epoch boundary is recognized to be at or near the top of the Yazoo Clay. The extinctions of the key calcareous nannoplankton species occur at a lower stratigraphic level.
2. The dilemma regarding the placement of the Eocene-Oligocene boundary exists probably because changes in faunal and floral assemblages were gradual during the late Eocene through the early Oligocene and because this epoch boundary represents a stratigraphically condensed section of a Type 2 depositional sequence. No dramatic faunal or floral changes should be expected at the Eocene-Oligocene boundary because of the absence of a major drop in sea level and because no sudden climatic changes occurred at this time.
3. These upper Eocene and lower Oligocene strata and associated microfaunas reflect a relative fall in sea level after the deposition of the late Eocene North Twistwood Creek Member of the Yazoo Clay, followed by an overall increase in water depths through the remainder of the late Eocene (Cocoa Sand, Pachuta Marl and Shubuta Clay Members of the Yazoo Clay). Subsequent shallowing of water depths beginning at the close of the late Eocene and continuing into the early Oligocene is represented by the shelf deposits of the Red Bluff/Bumpnose and deltaic deposits of the Forest Hill. A relative fall in sea level is indicated after the deposition of the Forest Hill, followed by a relative rise in sea level

resulting in extensive carbonate shelf sedimentation (Mint Spring and Marianna).

4. In southeastern Mississippi and southwestern Alabama, the Cocoa, Pachuta and Shubuta Members of the Yazoo Clay, the Red Bluff Clay, the Bumpnose Limestone and the Forest Hill Sand comprise an unconformity-bounded, Type 2 depositional sequence that accumulated during the TE3.3 coastal onlap cycle of the Td supercycle. The contact of the Cocoa/Pachuta with the North Twistwood Creek (highstand regressive deposits of the underlying depositional sequence) is a Type 2 unconformity. The transgressive deposits of the TE3.3 sequence consist of the Cocoa and Pachuta. The condensed section of the sequence includes the Shubuta (lower condensed section deposits) and Red Bluff/Bumpnose (upper condensed section deposits). The Shubuta-Red Bluff/Bumpnose contact which approximates the Eocene-Oligocene boundary is a surface of maximum starvation associated with the greatest landward transgression of the coastline during a relative rise in sea level. The Forest Hill Sand overlies the Red Bluff/Bumpnose and comprises the highstand regressive deposits of the sequence. The contact of the Forest Hill with the overlying Mint Spring is a Type 2 unconformity. The Mint Spring represents the transgressive deposits of the overlying sequence.
5. The use of stratigraphically condensed sections, sequence boundaries, and genetic depositional sequences has the potential to be a useful correlation tool for resolving stratigraphic and depositional problems in the Gulf Coastal Plain. The age of the condensed section is useful in identifying the depositional sequence, and the nature of the sequence boundaries and characteristics of the units comprising the sequence can be utilized in deciphering the depositional history of the strata. The age of a condensed section within a specific depositional sequence should be synchronous worldwide.

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TESTING CENOZOIC EUSTATIC CHANGES:
THE CRITICAL ROLE OF STRATIGRAPHIC RESOLUTION

KENNETH G. MILLER AND DENNIS V. KENT
Lamont-Doherty Geological Observatory of Columbia University,
Palisades, NY USA 10964

Integration of magnetostratigraphy, biostratigraphy, sequence stratigraphy, and isotope stratigraphy provides a means to decipher the temporal relationships among lithological, geophysical, and geochemical indications of sea-level changes. However, the importance of chronology in testing cause and effect relationships among inferred sea-level changes is often underestimated. Firm chronostratigraphic control is required to establish equivalency among events. The establishment of age equivalency between two events does not prove cause, but rather constitutes a necessary condition for inferring causal relationships.

The greatest potential for testing causes and effects of sea-level changes lies in linking deep-water pelagic sections with shallow-marine sections. Pelagic sections are more likely to contain relatively complete records of paleoceanographic and climatic changes. Coastal plain and continental shelf sections are particularly sensitive to sea-level fluctuations, containing physical evidence of transgressions/regressions and erosional events, but are notoriously difficult to correlate to a standard chronology. We have been particularly successful in applying integrated stratigraphic techniques (magnetostratigraphy, biostratigraphy, and Sr-isotope ($^{87/86}$Sr) stratigraphy) to boreholes drilled onshore in the coastal plain (Miller and others, in press). Drilling minimizes problems with weathering which often inhibits magnetostratigraphic and stable isotope studies of outcrops. We suggest that future drilling designed to test sea-level changes should concentrate not only on the shelf and slope, but also on emergent coastal plains. Relationships among Cenozoic oxygen isotope, seismic stratigraphic, and rock stratigraphic records illustrate the importance of chronologic ties between pelagic and shallow-water records. We have correlated Cenozoic foraminiferal δ^{18}O records to a standard geomagnetic polarity time scale (Berggren and others, 1985), and inferred that the Oligocene-Recent δ^{18}O record provides a history of glacio-eustatic changes. Such inferred glacio-eustatic lowerings correlate with erosion observed as unconformities on passive continental margins using chronostratigraphic and seismic stratigraphic records (Miller and others, 1987a). The two best examples of these correlations are provided by events in the "middle" Oligocene and earliest late Miocene.

1) Miller and others (1985a) noted that a hiatus occurred between 30-34 Ma in the "middle" Oligocene on the Irish and U.S. east coast margins, apparently correlating with seismic stratigraphic evidence for downslope erosion of canyons and a major downward shift in coastal onlap (Vail and others, 1977; Haq and others, 1987). Sr-isotope stratigraphy has confirmed and refined previous biostratigraphic estimates of the timing of erosional events (Miller and others, in press). Our synthesis of Oligocene benthic foraminiferal δ^{18}O records suggests that a major δ^{18}O increase occurred at about 32 Ma (i.e., within Chron C11)(Fig. 1); a similar δ^{18}O increase has been noted in planktonic foraminiferal records (e.g., Miller and others, 1987a, fig. 4). We interpret this as evidence of a glacio-eustatic lowering which caused erosion of unconformities and canyons and a downward shift in coastal onlap.

2) A major erosional event was noted in the basal upper Miocene on the New Jersey continental slope in the form of a downslope canyon (e.g., Miller and others, 1987b). This erosional event correlates with increased benthic (Fig. 1) and

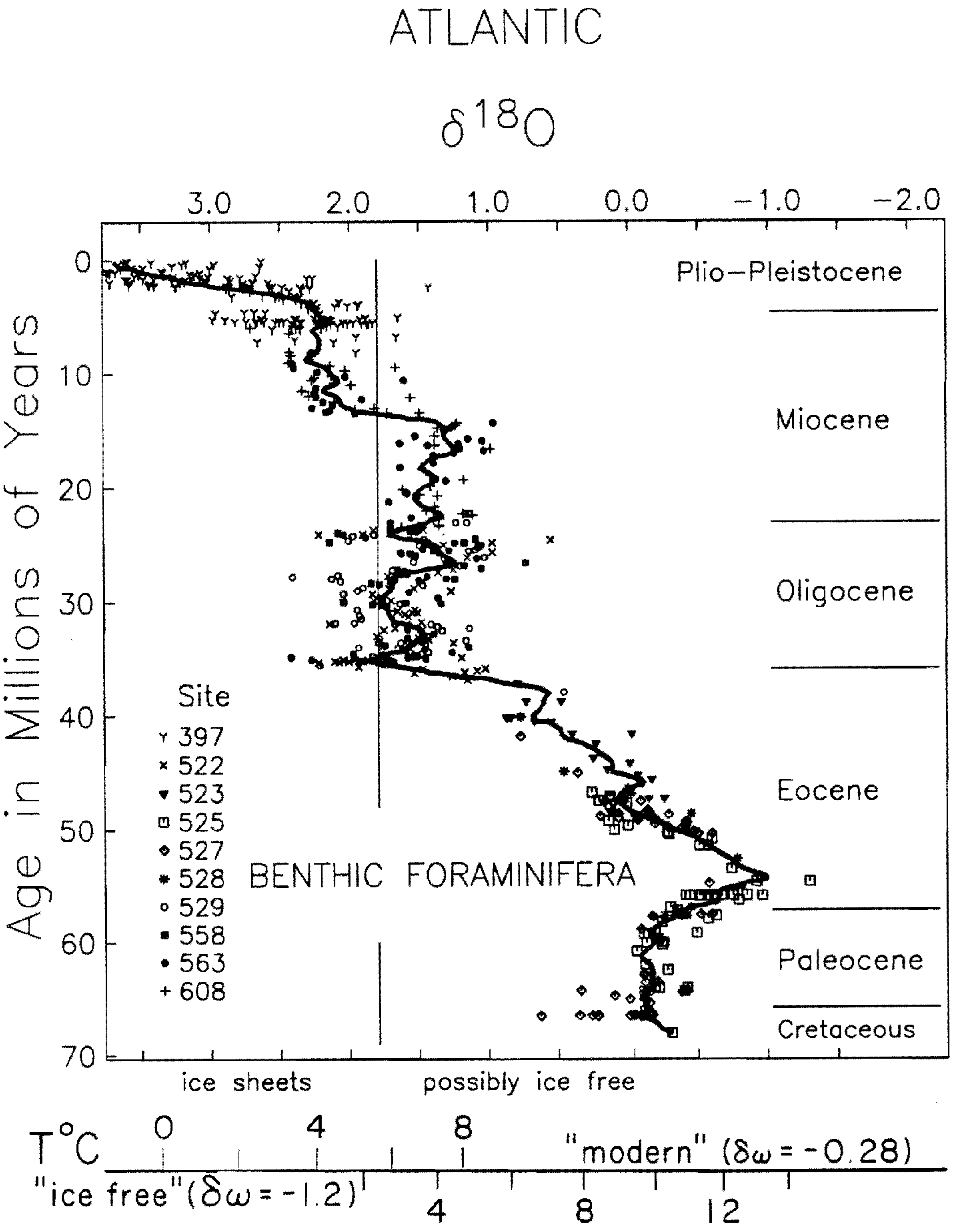

Figure 1. This composite benthic foraminiferal $\delta^{18}O$ record for Atlantic DSDP sites is corrected to *Cibicidoides* and reported to PDB standard. In general, data have been calibrated to the geomagnetic polarity time scale of Berggren and others (1985) by first-order magnetostratigraphic correlations. The vertical line is drawn through 1.8 $^{0}/oo$; values greater than this provide evidence for significant continental ice sheets. The lower temperature scale assumes no significant ice sheets and can be used for the pre-Oligocene, while the upper scale assumes ice volume equivalent to modern values. Intervals of high $\delta^{18}O$ values at ca. 35 Ma, 32-28 Ma, 25-24 Ma, 15-13 Ma, and 10 Ma are are associated with increases in planktonic $\delta^{18}O$ values, and interpreted as reflecting ice growth events at these times. After Miller and others (1987a).

planktonic (fig. 4 in Miller and others, 1987b) $\delta^{18}O$ values in the earliest late Miocene (ca. 10 Ma), and with evidence of a major downward shift in coastal onlap (Vail and others, 1977; Haq and others, 1987). These observations are consistent with a cause and effect relationship between glacio-eustatic lowerings at about 32 Ma and near 10 Ma, erosion on passive continental margins, and downward shifts in coastal onlap.

Cenozoic stratigraphic resolution is the limiting factor for testing global sea-level changes. In the case of the second-order cycles, seismic stratigraphy can be used to establish their regional extent, while chronostratigraphy must be used to establish their interregional extent and possible eustatic cause. We have shown in previous examples that there is a good correlation between the major downward shifts in coastal onlap (the second-order sequence boundaries of Haq and others (1987)) and erosional events noted on different passive continental margins (see Poag and Mountain, 1987, for other examples). This supports the interpretation that most of the downward shifts in onlap reported by Exxon were caused by eustatic changes. As noted by Pitman (1978) and recognized by Miller and others (1985b) and Haq and others (1987), the timing of these downward shifts in coastal onlap correspond with the maximum *rates* of eustatic lowering.

Although we agree that the major (second-order) downward shifts in onlap correspond with eustatic changes, we question the ability to decipher readily the higher-order sea-level cycles of Haq and others (1987). The third-order cycles were developed from analyses of outcrop and well-log sections rather than from seismic stratigraphy like the second-order cycles (Haq and others, 1987). The recognition of the third-order cycles hinges upon the identification of lithologic, well-log, or paleodepth changes which may be locally controlled. Since they are not recognized using seismic stratigraphy, chronostratigraphy is needed to establish both the regional and interregional extent of the third-order cycles. Thus, the cycles must be proven to be synchronous in different locations in order to establish first that they are regional events and second that they are eustatic events. However, the mean duration of the third-order Cenozoic cycles is about 1.5 m.y., which is approaching the limits of biostratigraphic resolution. The average length of Cenozoic planktonic foraminiferal and nannofossil zones is about 1.5 m.y. Although biostratigraphic resolution as fine as 0.5 m.y. may be possible in some Cenozoic sections (e.g., Aubry, 1985), it is usually coarser, particularly in shallow-water sections such as those studied by Haq and others (1987). Since the durations of the third-order cycles are at the limit of biostratigraphic resolution, they generally are testable with conventional biostratigraphy only if multiple microfossil correlations are used or if an independent chronological constraint (e.g., magnetochronology, Sr-isotope stratigraphy) is obtained.

We agree that in order to test the validity of the third-order cycles it is not necessary to establish that *every* third-order cycle is precisely the same age on different margins. Haq and others' (1987) utilized a sequence approach to recognize third-order events above known datum levels. Assuming that they observed the same patterns on different margins, their observation of the same ordinal heirarchy of events within a given time window on different margins argues against a local cause and points to eustatic control. Presumably datum levels may be biostratigraphic or seismic stratigraphic (i.e., the second-order cycle boundaries are assumed to have chronostratigraphic significance). However, the simple matching of third-order cycles between locations is complicated by gaps in the records, uncertainties in establishing datum planes, and the ability to discriminate between these cycles at the outcrop level. We believe that it is ultimately necessary to establish that the datum levels used to fix the ordinal heirarchy at any place should be correlatable to within at least 1/2 the cycle length to be tested. In the case of the third-order cycles, this requires resolution of better than 0.75 m.y. at critical levels.

Using multiple microfossil zonations to improve resolution has promise, but the uncertainties in microfossil calibrations are often understated. For example, it has been assumed that calibrations among various groups are well known, and that by simply integrating biostratigraphic zonations or ranges of two or more groups (e.g., planktonic foraminifera and nannofossils), the stratigraphic resolution is substantially improved beyond the 1-1.5 m.y. level afforded by any single group (Barron and others, 1985; Haq and others, 1987). This is not true in many cases. A good case in point is afforded by the middle/late Miocene boundary. Estimates of the boundary vary from 9.5-11.5 Ma (i.e., nearly 20 percent difference), and the relative placements of the magnetochrons, planktonic foraminiferal zones, and nannofossil zones vary by more than 1 m.y. among different authors (Fig. 2). We have discussed the problems with intercalibrations of microfossils and the geomagnetic polarity time scale near this boundary (Miller and others, 1985b) and believe that resolution in this interval is now improved for low- to middle-latitude locations.

Other potential problems limit biostratigraphic resolution and our ability to test the synchrony of sea-level events. Biozonal boundaries are often geographically limited and diachronous across latitude (e.g., Johnson and Nigrini, 1985). Biostratigraphic resolution in continental margin sequences is hindered by complications due to absence of marker microfossils in shallow-water facies, reworking, and reliance upon cuttings. We caution that attempts to resolve events finer than 1-2 m.y. in the Cenozoic based upon biostratigraphy alone may often fail because of these problems. Only by integrating biostratigraphy with independent estimates afforded by magnetostratigraphy and Sr-isotope stratigraphy is stratigraphic resolution sufficient to test sea-level changes on the million year scale.

Acknowledgments

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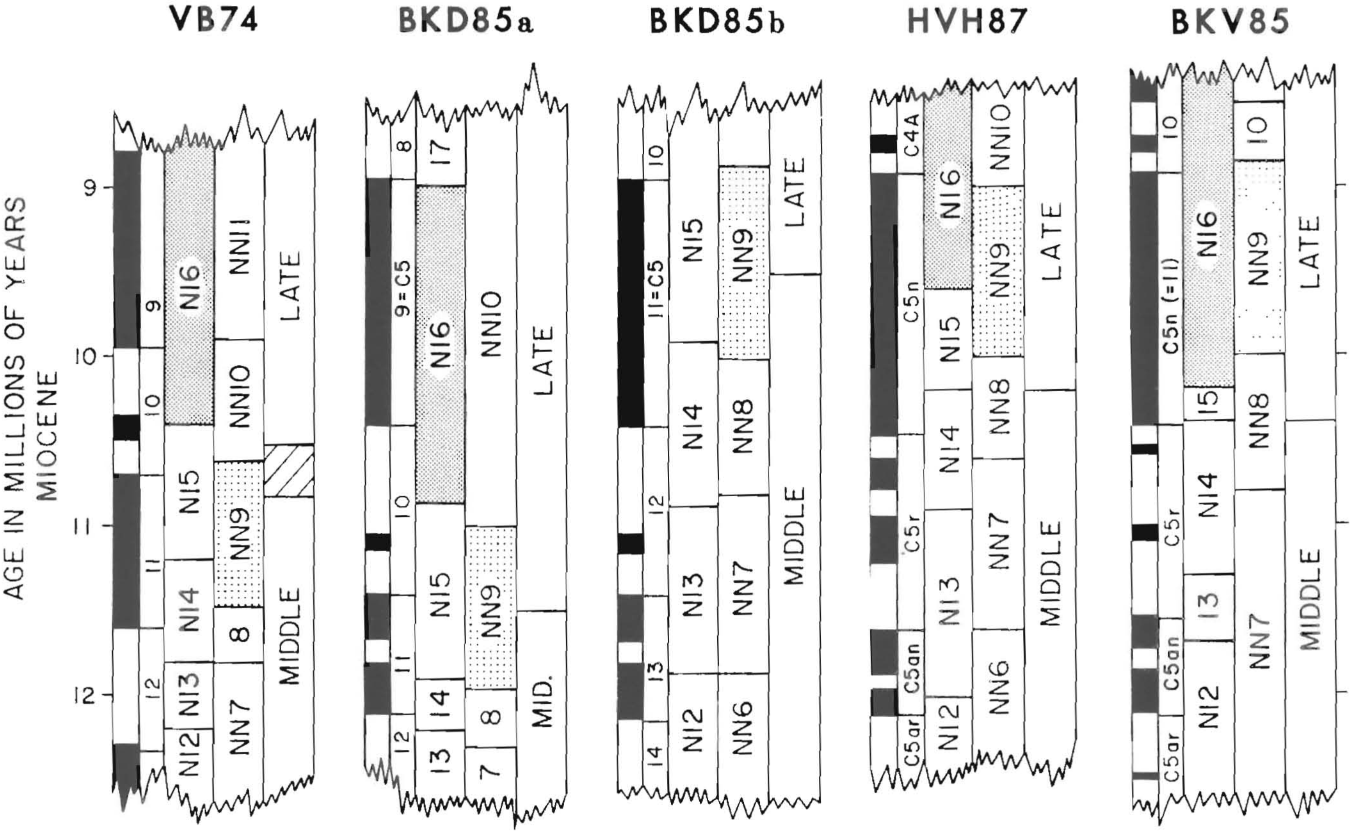

Figure 2. Comparison of time scales for the late middle to early late Miocene points out the variable position of nannofossil Zone NN9 (light stipple) relative to planktonic foraminiferal Zone N16 and the geomagnetic chrons (black normal, white reversed polarity).

Symbols:

VB74 = Van Couvering and Berggren (1974) based upon biochronology.

BKD85a = Barron and others, (1985) based upon second-order magnetobiostratigraphic correlations and assuming that Epoch 9 is equivalent to Chron C5n;

BKD85b = Barron and others (1985) assuming correlation of Epoch 11 with Chron C5n;

HVH87 = Haq and others (1987) based upon second-order magnetobiostratigraphic correlations in this time interval.

BVK85 = Berggren and others (1985) based upon first-order magnetobiostratigraphic correlations of Miller and others (1985b).

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CENOZOIC MARGIN CONSTRUCTION AND DESTRUCTION OFFSHORE NEW JERSEY

GREGORY MOUNTAIN

Lamont-Doherty Geological Observatory, Palisades, N.Y. 01964

ABSTRACT

Seismic reflection profiles have been correlated to eleven wells drilled between the upper continental slope and upper continental rise in a 2100 km^2 grid, offshore New Jersey. This high density of publicly available data represents a unique opportunity for examining processes that control the buildup and erosion of passive continental margins. The conclusions drawn from this study have implications relating to the formation of submarine canyons, mechanisms of down-slope sediment transport, and the validity of interpreting seismic sequence boundaries along continental slopes as evidence of past sea-level changes.

One Upper Cretaceous and four Paleogene episodes of slope failure, slumping, and infilling have been documented. Each event is concentrated in broad, channel-like depressions beneath today's middle and lower slope. Though a hiatus of from one to several million years is associated with most episodes, very little shallow water debris rests on each erosional surface. In general, the channels are partly filled and smoothed by locally derived slumps. A common origin for all of these channels is strongly implied by the fact that each is stacked above the preceding one. It is proposed that these unconformities resulted from slope failure during episodic collapse of the underlying Mesozoic carbonate margin. Headward erosion may have lengthened many channels, but none reached a major source of shallow-water clastic sediment.

Three events occurred during or shortly after the Oligocene that influenced the nature of subsequent margin processes. First, the slope was cut landward between 5 and 25 km, interrupting the stacked arrangement of channels that had been maintained throughout the Paleogene, and this allowed a new pattern of subsequent slope failures to develop. Second, biogenic carbonate and siliceous sedimentation was replaced by detrital accumulation; these younger sediments experienced less early diagensis and more readily failed on gently dipping slopes. Lastly, the polar regions entered into the first of numerous ice-dominated climates. Consequently, oscillatory changes of global sea level became especially rapid, and contributed to the transport of large volumes of clastic sediment to the heads of pre-existing slump scars indenting the shelf edge.

Unfortunately, drill core data available for this study cannot provide information relating to the history of the margin from Oligocene to late middle Miocene time. Seismic profiles show that sediment accumulated in a thick deposit off the mid-Atlantic states during this interval, but any canyons through which these sediments may have been transported from the New Jersey margin are not within the study area.

An era of significant slope erosion and sediment by-pass occurred between 11.7 and 9.5 MY when a narrow canyon was cut directly beneath DSDP Site 612. During this interval a variety of shallow-water and reworked debris by-passed the slope and accumulated on the upper rise.

Pliocene sediments at the foot of the slope were particularly unstable because of layers of glauconite-rich sand. These unconsolidated layers served as detachment surfaces for large Pleistocene debris flows that came down off the slope and spread out onto the upper rise.

At present, the lower and middle slope offshore New Jersey continues to be a zone of sediment failure. Much of the material detached from this region is deposited by debris flows on the upper rise; a few major canyon systems provide channels for turbidity currents to continue beyond the 2500 m isobath. Most of these especially lengthy canyons are incised into the upper slope as well, and consequently are capable of delivering shelf-edge sediment to the deep basin.

INTRODUCTION

Unconformities Along the Margin

The stratigraphic record on the shelf, slope and rise of eastern North America is far from complete (Jansa and Wade, 1975; Tucholke and Mountain, 1979; Poag, 1980; Poag and Schlee, 1984; Poag, 1985a; Mountain and Tucholke, 1985). Three regional hiatuses are common to most of the margin depocenters (Fig. 1). The oldest of these occurs in the lower Tertiary of the major sedimentary basins between southeastern United States and eastern Canada. In the shelf and slope of Baltimore Canyon Trough, this is a gap of approximately 10 MY, separating middle Maestrichtian from uppermost Paleocene to lowermost Eocene (Fig. 2). Despite no deep sampling of the upper rise prior to Deep Sea Drilling Project (DSDP) Legs 93 and 95 (van Hinte, Wise and others, 1987; Poag, Watts and others, 1987), it had been proposed on the basis of seismic evidence that this gap narrows in the seaward direction, and that thick Paleocene sediments onlap the base of the slope (Schlee and Grow, 1980; Mountain, 1981; Mountain and Tucholke, 1985).

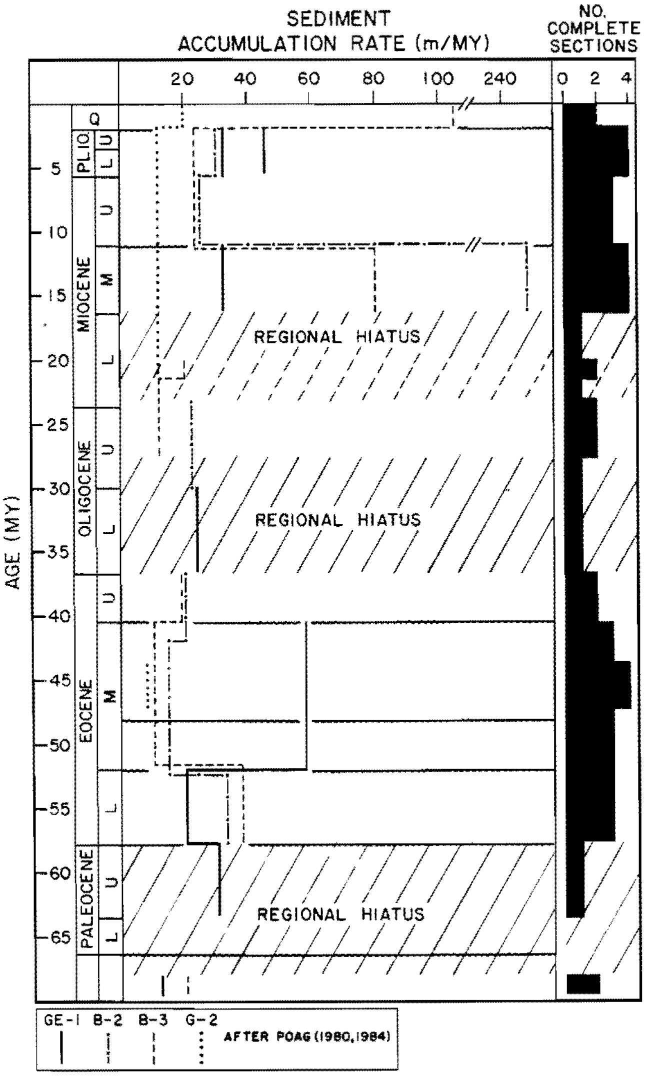

FIGURE 1 - Cenozoic accumulation rates (Poag, 1980; 1984) measured in Continental Offshore Stratigraphic Test (COST) wells from major depocenters along eastern North America ("GE-1" = Southeast Georgia Embayment, "B-2" and "B-3" = Baltimore Canyon Trough, and "G-2" = Georges Bank Basin). Regional hiatuses are recognized during three intervals when no sediment was preserved in at least two basins. Seismic discontinuities identified in profiles from the uppermost continental rise, confirmed by drilling, and discussed in this report are shown as wavy lines at roughly 66, 58, 52, 48, 40, 11 and 2 MY before present (time scale of Berggren and others, 1985).

A second regional hiatus off eastern North America separates upper Eocene from upper Oligocene strata (Fig. 1). In the Baltimore Canyon Trough this contact represents an 8 MY gap, and can be traced in profiles seaward to where it crops out on the middle continental slope (Fig. 2). Again, no deep samples from the upper rise had been collected before DSDP Legs 93 and 95, but earlier seismic tie-ins from the lower rise predicted that a hiatus (Reflector A^u; Fig. 2) at least as long (from 7 to 30 MY) as that in shelf and slope wells occurs across the Eocene/Oligocene boundary beneath the upper rise (Tucholke and Mountain, 1979).

The widespread lack of most of the lower Miocene section constitutes a third regional unconformity (Fig. 1). In general, this hiatus represents a gap on the shelf of approximately 7 MY, spanning the upper Oligocene and the lowermost Miocene; very little hard information was available before DSDP Legs 93 and 95 to determine how far seaward this unconformity continues. By tracing reflectors landward from the lower rise, a locally erosional unconformity (Reflector Merlin) had been estimated as late middle Miocene on the upper rise, and thick fan and current-controlled deposits of lower Miocene age were speculated to have accumulated on the rise off New Jersey between Merlin and the underlying Reflector A^u (Mountain and Tucholke, 1985).

Controversy of Global Sea-level Changes

Because the three unconformities described above are found on other continental margins as well, their origins have been attributed to changing global sea level (Vail and others, 1977; Poag, 1980; Poag, 1985a). Unconformities that are matched by coeval units of sediment onlappping the foot of the slope, for example the basal Paleocene event just described (Fig. 2), indeed suggest shelf erosion and sediment transport to the deep sea during lowstands. However, it is unclear how this process could be responsible for unconformities that appear to have developed simultaneously on the shelf and the rise. Ironically, these latter gaps have been explained by sea-level highstands, when it is argued that terrestrial sediment remained trapped in estuaries and never reached the outershelf or beyond; the result was not an erosional unconformity, but rather a "marine condensed section" (Vail and others, 1984.) This explanation ignores pelagic contribution to outer margin sediments, which should continue unabated during sea-level highstands.

As an alternative explanation, erosion by ocean currents (both shallow and deep) has been proposed as a cause for hiatuses along eastern North America. It is known, for example, that the modern Gulf Stream is responsible for the widespread exposure of relict Miocene sediments on the Blake Plateau. A predominantly wind-driven predecessor to the Gulf Stream has been flowing along this same path throughout the Cenozoic, and at times has deeply scoured the shelf south of Cape Hatteras (Pinet and Popenoe, 1982.) A south-flowing surface current has been proposed to explain gaps on the shelf

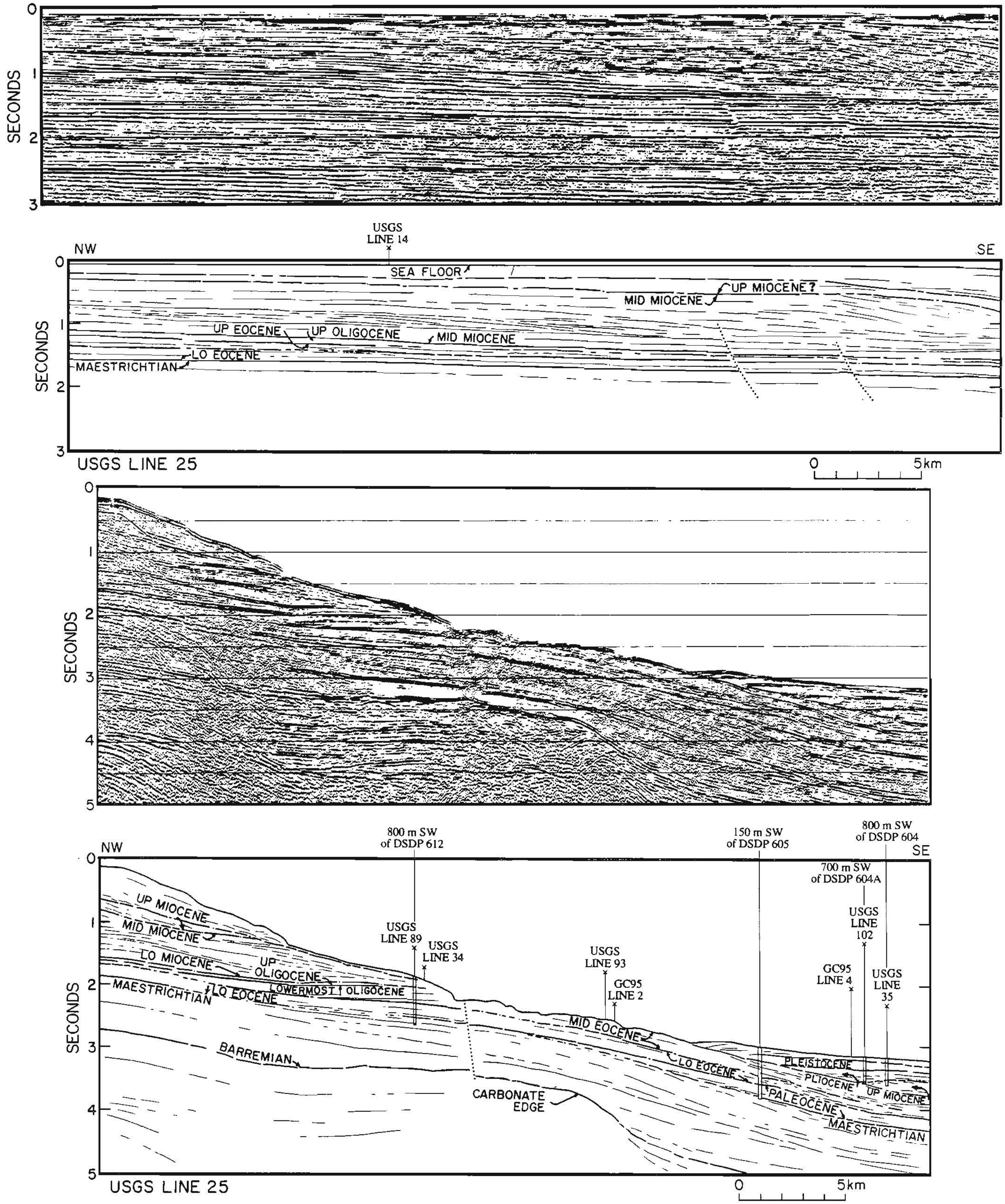

FIGURE 2 - UPPER : Original and line drawing interpretation of a portion of USGS Line 25 on the outer continental shelf showing a notably incomplete Paleogene section and prograding Miocene. The COST B-2 well is 58 km NE on USGS Line 14. LOWER : Original and line drawing interpretation of the continuation of USGS Line 25 from the shelfedge to the uppermost continental rise. One of many faults is highlighted near the buried carbonate edge, and cuts into Eocene strata exposed on the middle and lower slope. The Neogene section onlaps the uppermost rise (profile location in Fig. 11).

north of Cape Hatteras (Olssen and Wise, 1985); its modern analog, the Labrador Current, is not sufficiently developed to erode the shelf at present.

Although affecting deeper and more remote regions, thermo-haline circulation of the Western Boundary Undercurrent is important in the shaping of the North American continental rise (Heezen and others, 1966). Predecessors to this current have been linked to the angular unconformities marked by Reflectors A^u, Merlin and Blue (Mountain and Tucholke, 1985.) The processes that maintain thermo-haline circulation, however, are not directly linked to those processes that maintain wind-driven circulation. Therefore, as with the mechanism of eustatic changes, *ad hoc* assumptions must be invoked to argue effectively that ocean currents can erode synchronous hiatuses on the shelf, slope and rise.

Longstanding Model for the Origin of Submarine Canyons

An essential factor in understanding erosional mechanisms along the eastern North American margin is determining how sediment eroded from the continental shelf actually gets to the deep sea. Unless unchannellized sediment transport is the dominant process, the origin and maintenance of slope canyons is critical to this problem.

The largest slope canyons off the eastern United States begin at the shelfedge. Their similarities to alluvial drainage systems led Daly (1936) and Kuenen (1937) to argue for a submarine origin that has come to be widely accepted. In this view, much if not all of the continental shelf was exposed during the lowstands of Pleistocene glacial maxima, and rivers flowed as far seaward as the present shelfedge. Large volumes of river-borne sediment accumulated along the uppermost slope, and massive slope failures occurred periodically. These failed sediments, in turn, generated turbidity currents that carved canyons into the slope. Continued canyon erosion and inter-canyon buildup have resulted in entrenched features (Shephard, 1981).

Buried Pleistocene valleys of the Hudson and Delaware Rivers have indeed been traced across the shelf (Ewing and others, 1963; Twichell and others, 1977; Knebel and others, 1979), and clearly they once transported sediment to the heads of the Hudson and Wilmington Canyons. Daly's model has been substantiated further by the recovery of turbidites on the Hatteras Abyssal Plain that are similar in composition to the relict sands found near the heads of slope canyons (Horn and others, 1971).

Recent Studies of U.S. Continental Slope

However, it is possible that turbidity currents crossing the slope are strictly opportunistic, and do not contribute significantly to the initiation of slope canyons. Perhaps turbidity currents utilize canyons only after the canyons have been formed by some other process. This arguement has been strengthened by recent studies that reveal far more complexity to the continental slope than had previously been recognized (Robb and others, 1981; Twichell and Roberts, 1982; Farre and others, 1983). It is now known that large canyons indenting the shelfbreak are far outnumbered by smaller, relatively straight canyons that begin entirely on the upper and middle slope. Slumping causes headward erosion in these smaller canyons, and consequently the slope is eroding from the bottom up, in water depths below all reasonable levels of Pleistocene subaerial exposure.

These recent studies have shown further that slope failure is not restricted to the heads of canyons: "tributary gullies" incise the flanks of canyons in a trellice-like pattern, and are themselves eroding inter-canyon areas. Some systems are so dense that gullies from adjacent canyons meet in sharp-crested ridges. The absence of debris at the junction of gullies and their host canyon has been cited as evidence that these canyons are being flushed by Holocene turbidity currents (McGregor and others, 1982), and strongly suggests that efficient down-slope transport is occurring at present.

In those instances where headward erosion has breached the shelf edge, outershelf clastic sediments have been provided with a conduit that crosses the slope (Twichell and Roberts, 1982). At this stage, a canyon enters a "mature" phase (Farre and others, 1983), and abrasive turbidity currents deeply incise the canyon axis and steepen the canyon walls. The occurrence of "hanging" tributary gullies result from the envigorated erosion that occurs during this stage of canyon development.

An especially significant aspect of this latter model is that turbidity currents crossing the slope may indeed reach a maximum during sea-level lowstands, but they did not necessarily contribute to the formation of the slope canyons. It is entirely possible that canyon formation is related to processes other than sea-level lowerings.

Some of these other processes that have been recognized as contributors to slope failure include 1) seismicity (Heezen and Ewing, 1952), 2) along shelf transport and sediment buildup at the shelf edge (May and others, 1983); 3) activity of bottom-dwelling fauna (Shepard and Dill, 1966), 4) undercutting by erosive bottom currents (Paull and Dillon, 1980, Mountain and Tucholke, 1985), and 5) changing *in situ* pore pressure due to sea-level change or gas generation (Hathaway and others, 1976; Embley and Jacobi, 1977; Robb and others, 1981).

It is the purpose of this paper to describe new lithologic and seismic evidence of unconformities on the New Jersey margin, and to discuss the evidence that bears on the problem of how slope canyons are formed, maintained and later recognized in the the geologic record. It is hoped that sorting out the relevant facts will contribute to a better understanding of which processes are significant in the development of slope unconformities. This analysis is based on drilling results and their correlation with seismic data collected prior to and during DSDP Leg 95 (Poag, Watts and others, 1987).

DATA BASE

Seismics

DSDP sampling during the "New Jersey Transect" (Legs 93 and 95; van Hinte, Wise and others, 1987; Poag, Watts, and others, 1987) provided valuable chrono- and lithostratigraphic control to an already excellent seismic data set along the U.S. eastern margin (Fig. 3). Furthermore, seismic data were collected during Leg 95 that firmly tie these samples into the seismic network (Mountain, 1987).

High-resolution, shallow-penetration single channel seismic (SCS) air gun records collected during the USGS geohazards analysis program (Robb and others, 1981) constitute a dense grid of data that were an important component of this and other related studies (see Poag, 1985b; Poag and Mountain, 1987). Additional multichannel seismic (MCS) profiles contracted by the USGS were a valuable resource. An additional SCS line collected by the author during cruise 2502 of the *R/V CONRAD* was useful because it represents the only seismic line that directly ties DSDP Sites 612 and 613.

USGS Lines 25, 34 and 35 (Figs. 2 and 4) provided the basis for the original siting of the Transect drill sites, and to meet safety considerations, holes were to be located as close as possible to the crossings of these profiles. Site 612 was therefore positioned near the intersection of Lines 25 and 34. Unfortunately, poor hole conditions forced the early abandonment of Site 604, located on the upper rise near the crossing of Lines 25 and 35. To avoid similar problems, Site 605 was located upslope from Site 604, and Site 613 was farther northeast along slope at a location slightly landward of Line 35 (Fig. 3).

Navigation

Internal consitency of the existing SCS data navigated by LORAN-C (Robb and others, 1981) proved very reliable, and consequently the location of all other seismic data has been adjusted where necessary to agree at crossings with this grid (Mountain, 1987). Without adjustments of satellite-navigated data (such as that from *Glomar Challenger,* Leg 95), location errors of as much as 1 km can be expected. Along the continental slope where both seafloor and sub-bottom reflectors frequently have dips of 1:10 or greater, this uncertainty in location can lead to incorrect seismic interpretations. In this study, repositioning of satellite-navigated profiles amounted in occasional lateral shifts of a few hundred meters or less.

Drill Samples

Shipboard and shorebased analyses of Sites 612 and 613 (Leg 95, Poag, Watts and others, 1987) and shipboard reports from Sites 604 and 605 (Leg 93, van Hinte, Wise and others, Shipboard Report) have been incorporated in this study. The data from Leg 95 are especially unique because both holes were logged. Additional data were used from DSDP Site 108 (Hollister, Ewing and others, 1972), COST B-2 and B-3 (Scholle, 1977; Scholle, 1980) and ASP 14 and 15 (Poag, 1978; Poag, 1985a). Though these latter data were essential in evaluating the observations reported here, the following discussion focuses on the significance of the new Transect data.

DISCUSSION

Discussion of the unconformities on the New Jersey margin is structured around the packages of sediment separated by the three major unconformities described in the Introduction. This is not intended to be a synthesis of all information relevant to these unconformities; numerous previous studies have covered this thoroughly, calling on much more data than can be presented here (see Poag, 1985a and Poag, 1985b.) Rather, the attempt is made to focus on those drilling results that add new information to the unresolved problem of how passive margin unconformities are formed. It necessarily begins with inferences drawn from the seismic character of structures beneath the deepest samples recovered by DSDP drilling.

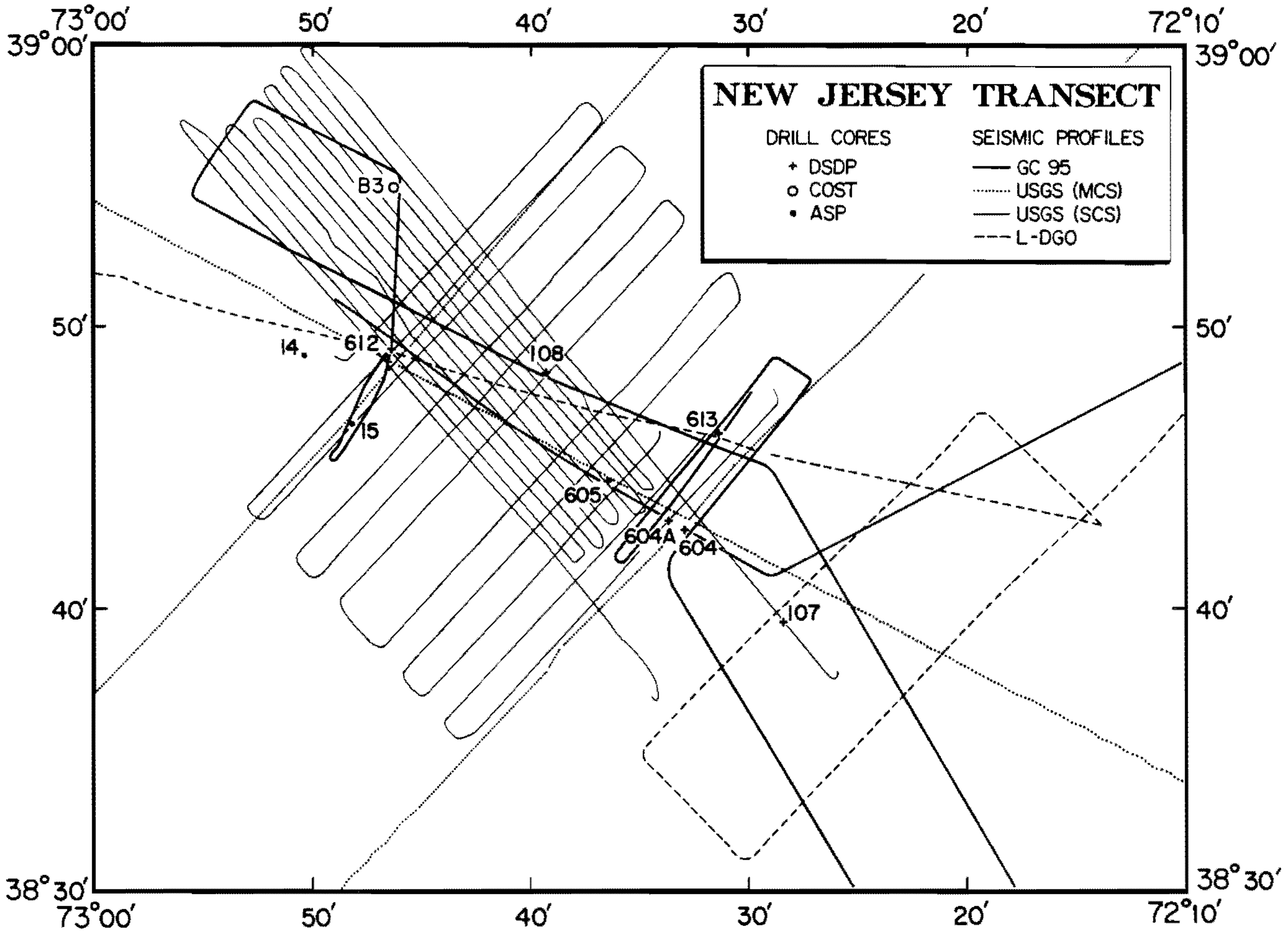

FIGURE 3 - Location map of drill core samples and seismic profiles used in this study. Cores are from Deep Sea Drilling Project (DSDP) Legs 11, 93 and 95 (Hollister, Ewing and others, 1972; van Hinte, Wise and others, 1987; Poag, Watts, and others, 1987); Continental Offshore Stratigraphic Test (COST) B-2 and B-3 (Scholle, 1977; Scholle, 1980); and Atlantic Slope Project (ASP) 14 and 15 (Poag, 1978). Profiles include Glomar Challenger (GC) 95 collected during Leg 95 (Mountain, 1987); U.S. Geological Survey (USGS) multichannel (MCS) and single-channel (SCS) seismic lines (Schlee and Grow, 1980; Robb and others, 1981) and a single-channel profile collected during Lamont-Doherty Geological Observatory (L-DGO) cruise C2502 (Shor and others, 1986).

Cretaceous

A carbonate structure fringed the eastern margin of North America throughout much of the Mesozoic (Jansa, 1981.) Although back-reef facies have been recovered in the off-structure COST B-3 well (Scholle, 1980), the existence of a true framework reef remains to be proven. Regardless of this detail, a steep margin was maintained throughout the Jurassic and into the Cretaceous. Eventual burial proceeded from north to south; off Nova Scotia this occurred in the Late Jurassic to Early Cretaceous (Jansa and Wade, 1975), off the northern and central east coast of the U.S. it occurred in the Early Cretaceous (Grow and others, 1979), while along the Blake Escarpment it occurred in the middle Cretaceous (Heezen and Sheridan, 1966). The carbonate banks and reefs of the Bahamas are the modern survivors.

The carbonate margin prograded seaward off New Jersey throughout the Early Cretaceous, but eventually was buried by sands and shales in Barremian time (Figs. 2 and 4; Poag, 1985a). This was followed by slope-front fill for most of the Late Cretaceous, and as a result, topographic relief was gradually reduced. Near the end of the Cretaceous, the margin was blanketed by mudstones, sandstones and occasional limestones (Pollack, 1980). These strata correlate to the nearly level reflectors beneath the upper and middle slope, and to the seaward dipping correlatives seaward of the buried carbonate structure (Figs. 2 and 4). DSDP Site 612 bottomed in these strata, 750 m above and 6 km landward of the seaward edge of this ancient carbonate margin.

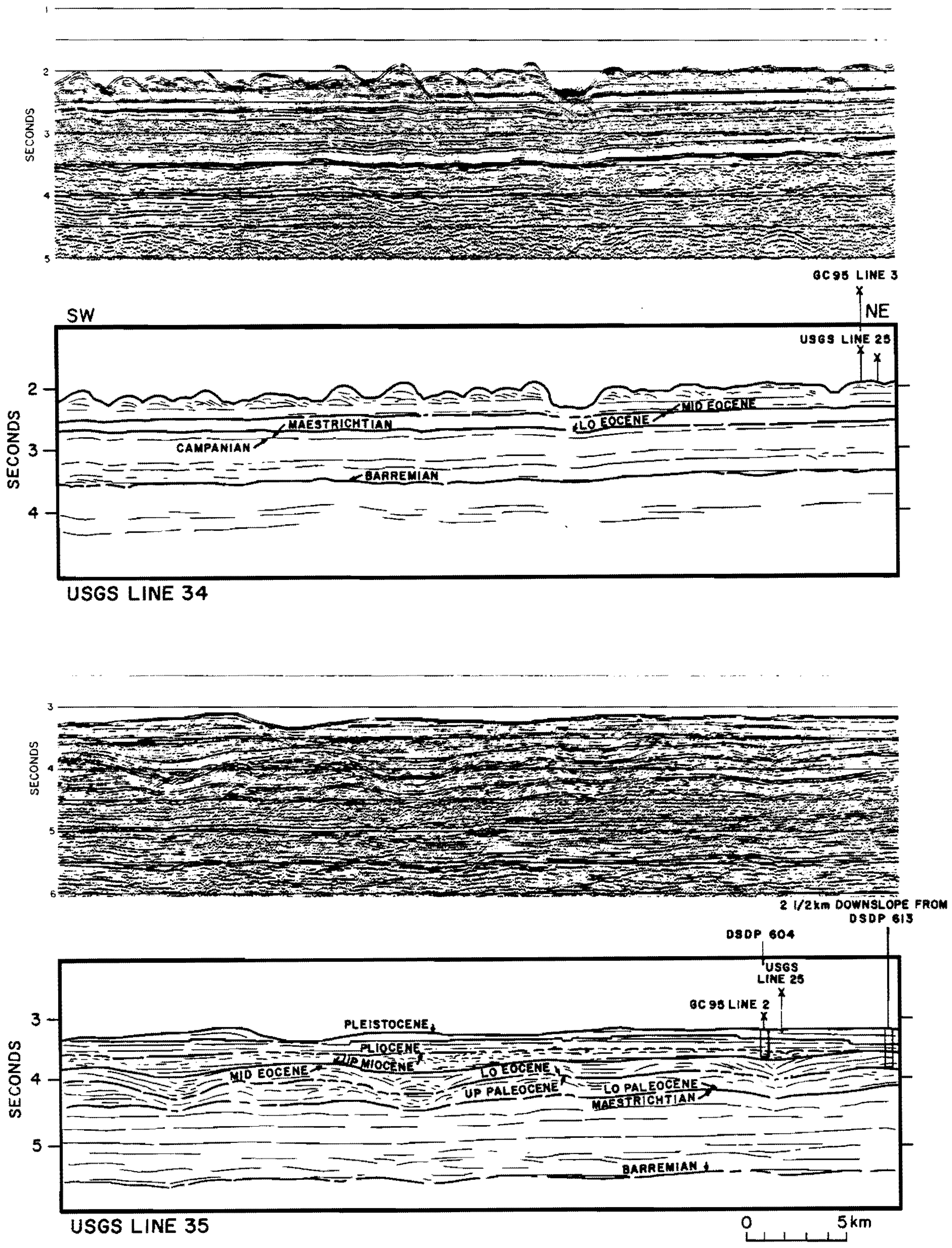

FIGURE 4 - UPPER : Original and line drawing interpretation of USGS Line 34 on the continental slope showing even layering of Eocene reflectors. LOWER : Original and line drawing interpretation of USGS Line 35 on the continental rise passing over DSDP Site 604, showing the stacked arrangement of Paleogene cut-and-fill channels (profile locations in Fig. 11).

The seaward dipping reflectors in the Upper Cretaceous slope-front fill are relatively uniform, parallel and continuous in strike lines (Fig. 4). An undulating erosional surface overlain by erratically dipping reflectors marks a cut-and-fill event representing a significant change in margin sedimentation shortly before the K/T boundary. Several more episodes of erosion, chaotic fill and partial re-excavation can be seen above this level (Table 1). Most of these features are stacked above preceding ones, suggesting that the occurrence of one controlled the location of the next. Presumably, all represent episodes of canyon formation, burial and partial re-exposure. Each will be discussed in stratigraphic order, and arguements will be presented to suggest the geometry and origin of these features are distinctly different from those of overlying Pleistocene canyons.

The oldest of the cut-and-fill episodes considered in this study occurred within the Upper Cretaceous. A possible 1 MY hiatus between Campanian shales and Maestrichtian chalks was detected on the upper slope at Site 612 (Poag, Watts and others, 1987); a similar gap was found in the COST B-2 and B-3 wells (Poag, 1980; Poag and Low, 1987). This unconformity has not been sampled on the lower slope, but a likely match is observed in profiles where a reflector marking an erosional surface cuts into acoustically laminated Upper Cretaceous strata (Fig. 4).

Paleocene and Eocene

Previous biostratigraphic studies had shown that the Cretaceous/Tertiary (K/T) boundary is an unconformable contact along the New Jersey margin as far seaward as the upper slope (Poag, 1980; Poag, 1985a); this was substantiated at Site 612. Furthermore, seismic stratigraphic analyses along USGS Line 25 (Fig. 2) had identified this as a major discontinuity (Schlee and Grow, 1980; Mountain and Tucholke, 1985). A strong, nearly level, and widely recognized reflector truncates underlying features at a shallow angle on the shelf, matching the eroded top of Maestrichtian sediments at several locations (Island Beach Well, COST B-2, COST B-3; Poag, 1985a). Reflectors of Eocene age downlap onto this surface on the inner shelf, but the pattern gradually changes towards the outer shelf where overlying reflectors become nearly parallel to this unconformable surface. Marine onlap becomes detectable beneath the upper slope (Figs. 2 and 5), and is very much more obvious seaward of this point. At the base of the slope, as much as 0.25 seconds (reflection time) constitutes a wedge of sediment that has no correlative on the shelf; it matches 200 m of Paleocene strata recovered at DSDP Site 605 (van Hinte, Wise, and others, 1987). Paradoxically, erosion into the Maestrichtian section is apparent at the projection of 605 onto Line 25, despite the report of a conformable K/T boundary at this drillsite (van Hinte, Wise, and others, 1987).

This questionably conformable K/T boundary at Site 605 matches the oldest of several Paleogene cut-and-fill episodes identified in seismic profiles (Fig. 4; Table 1). That no biostratigraphic break was detected by shipboard analysis implies one of three things: 1) the seismic correlation is wrong; 2) the paleontology is wrong; or 3) the gap is below biostratigraphic resolution. No logs were collected at Site 605, but the extrapolation of ages, lithologies and seismic velocities determined with a high degree of certainty by logging at Site 613 (8 km to the NE) makes it unlikely that the seismic correlation is at fault. It is also unlikely that shorebased re-evaluation will reveal that a major age refinement is necessary. The most probable explanation to this inconsistency is that the cut-and-fill process was geologically rapid, on the order of a one or two million years at most.

The basal Paleocene strata resting on the K/T boundary at 605 are glauconite-bearing, silty marls that become increasingly calcareous upsection. The corresponding seismic sequence progressively onlaps the Cretaceous section, and feathers out beneath the middle slope; Eocene strata rest directly on the Cretaceous section landward of this point (Fig. 2). It is clear that the Paleocene sediments at Site 605 originated farther upslope, perhaps on the shelf itself. However, no primary structures or macroscopic debris were reported at 605 that would lend strong support to down-slope displacement.

Samples close to the Paleocene/Eocene boundary were recovered at Site 605. This boundary matches the second Paleogene cut-and-fill surface detected in seismic profiles (Table 1; Fig. 4). As with the K/T boundary, shipboard paleontologists did not detect any missing section across this boundary, and it is concluded again that any cut-and-fill episode must have been geologically brief. However, in support of this small hiatus, sedimentation rates suggest that a small amount of basal Eocene section is missing at 605. Site 613 bottomed in nannofossil zone CP9b, 1 to 2 MY above the Paleocene boundary (Berggren and others, 1985). The lower Eocene accumulation rate based on nannofossils at this Site is 45 m/MY, and assuming no breaks, downward extrapolation of this rate places the Paleocene/Eocene boundary 45 to 90 meters below the bottom

Table 1

SEISMIC DISCONTINUITIES SAMPLED OFFSHORE NEW JESERY

AGE, MY	STAGE	SEISMIC CHARACTER ACROSS DISCONTINUITY		SEDIMENT TYPE ACROSS DISCONTINUITY		REMARKS
		DIP LINE	STRIKE LINE	SLOPE	RISE	
2	Pleistocene ------------ Pliocene	moderate slope by-pass; rise onlap changing laterally from reflectorless to chaotic to irregular to laminated seaward from base of slope	reflectorless to weakly reflective and chaotic ------------------------ evenly laminated, highly reflective unit (on rise)	mud ------ mud (upper slope only)	homogenous mud -------------------- mud alternating with glauconitic sandy mud	result of Pleistocene failures along Pliocene surfaces
11	Up Miocene ------------- Mid Miocene	considerble slope by-pass; especially irregular, hummocky character on rise	highly reflective, chaotic fill in broad upper rise channels	mud ------ mud	glauconitic sandy mudstone -------------------------------- chalk (Eocene)	canyon-cutting event; approximately correlative to age of Reflector Merlin along rise
40	Up Eocene ------------- Mid Eocene	very subtle slope onlap	conformable, parallel reflectors on slope; eroded on rise	chalk -------- chalk	(large hiatus)	5 MY hiatus on slope; large erosional hiatus on rise caused by bottom currents
48	Mid Eocene -------------- Mid Eocene	clearly rotated slump blocks on slope	cut-and-fill channels stacked above older ones on uppermost rise	chalk -------- chalk	chalk slump ----------- chalk	no biostratigraphically resolvable hiatus
52	Mid Eocene -------------- Lo Eocene	upper slope by-pass; onlapping mid-slope fill	cut-and-fill channels stacked above older ones on uppermost rise	(large hiatus landward of mid-slope)	limestone slumps ---------------------------- porcellantic limestone slumps	1-5 MY hiatus on slope; 1-2 MY hiatus on rise
58	Lo Eocene -------------- Up Paleocene	shelf + slope by-pass; onlapping lower-slope fill	cut-and-fill channels stacked above older ones on uppermost rise	(large hiatus landward of lowermost slope)	porcellanitic limestone slumps --------------------------- (porcellantic?) limestone	1 MY hiatus on rise
66	Lo Paleocene --------------- Maestrichtian	shelf + slope by-pass; onlapping base-of-slope fill	cut-and-fill channels stacked above older ones on uppermost rise; faulted above edge of underlying carbonate structure	(large hiatus)	glauconitic mudstone ------------------------ limestone	no biostratigraphically resolvable hiatus on rise
70	Maestrichtian ---------------- Campanian	mid-slope by-pass? onlapping base-of-slope fill	cut-and-fill channels cut into blanket of laminated, draping reflectors	chalk ------------- mudstone	(not sampled)	1 MY hiatus on slope

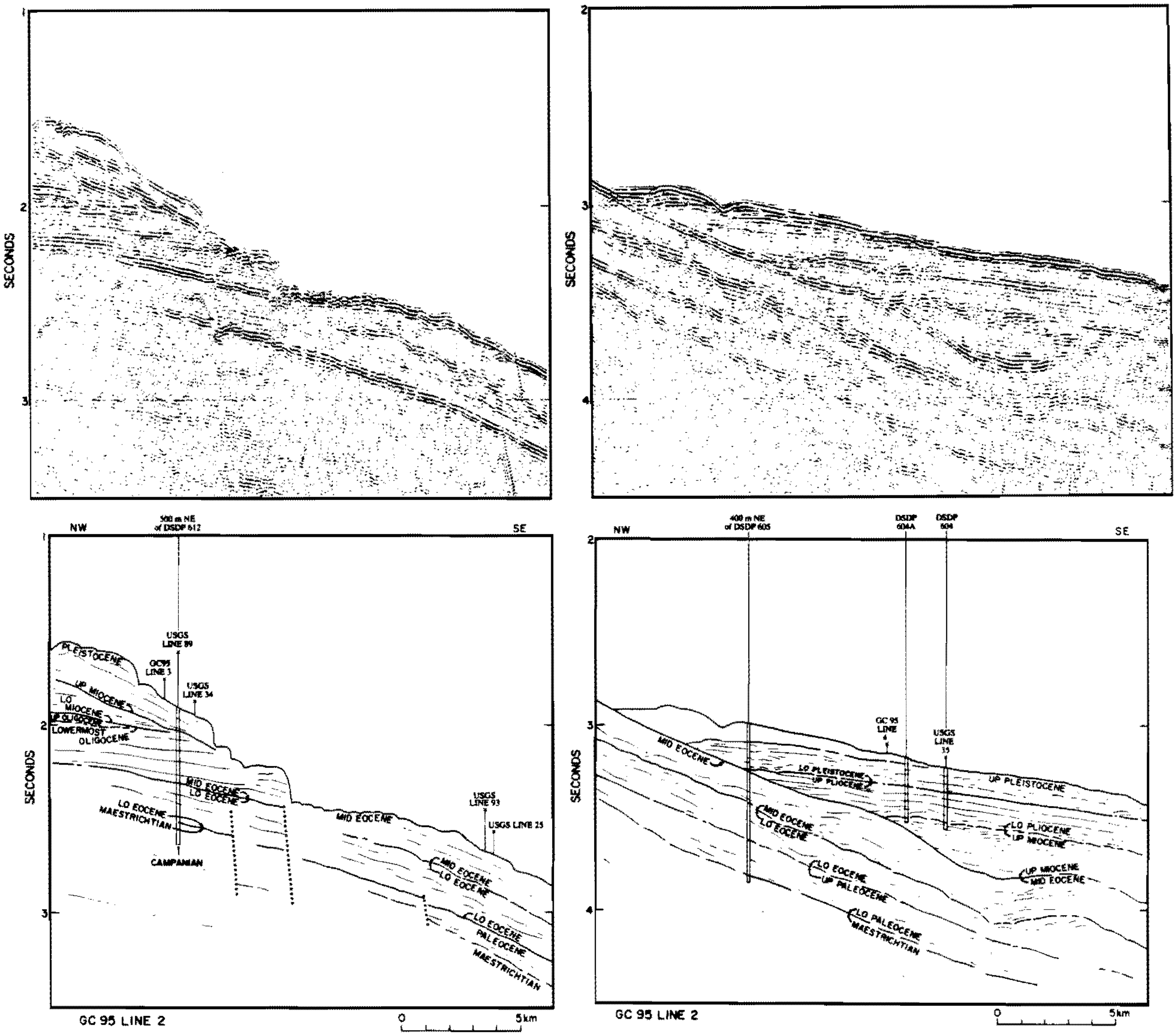

FIGURE 5 - LEFT : Original and line drawing interpretation of Glomar Challenger Leg 95 Line 2 approaching DSDP Site 612. RIGHT : Continuation of GC Line 2 onto uppermost rise near DSDP Site 605 and through Sites 604 and 604A. Rotated Eocene slumps are visible downslope from Site 612 (profile location in Fig. 11).

of the hole. At Site 605, the paleontologic equivalent to the bottom of hole 613 is less than 34 m above the Paleocene/Eocene boundary. This 11 to 56 meter difference suggests that an earliest Eocene gap of roughly 1 MY may exist at Site 605; the same conclusion was arrived at by Poag (1985b).

Features consistent with the irregular seismic character of the basal Eocene section were observed at Site 613. These strata had a particularly fissle and crenulated texture that became very obvious as they dried. Presumably, once these cores were unburdened, split, and rinsed in fresh water, their clay components swelled and revealed incipient deformation that otherwise was visually undetectable. Horizontal elongation of burrow mottles was noted at both 613 and 605, and though no overturned folds were observed, down-slope creep could explain all of these characteristics.

However, none of the lower Eocene strata at either Site 605 or Site 613 show compositional evidence of "chaotic fill". No terrestrial material other than what can reasonably be asssumed as aeolian was reported in smear slides. There are occasional cm-scale faults in these sediments, but their significance is uncertain due to the fractured nature of the cores.

The reflector marking the Paleocene/Eocene boundary cut-and-fill-surface continues upslope, and the onlap of progressively younger Eocene strata can be traced landward of

Site 612 (Fig. 2). As described above, there is 45 to 90 meters of basal Eocene below nannofossil zone CP9b on the upper rise at Site 613; there is less than 34 meters at 605, and less than 18 meters of this same interval on the slope at Site 612. This pattern of decreasing thickness of the basal Eocene section continues towards shallower water (Poag, 1980; Poag, 1985a).

The most dramatic seismic evidence of chaotically deposited slope sediments along the New Jersey margin occurs near the top of the lower Eocene (Figs. 4 and 6). This observation is probably biased by two facts: 1) variable silica diagenesis in these strata has led to abrupt and large impedance contrasts that generate especially strong reflections (see Goldberg and others, 1987); and 2) few SCS profiles penetrate below these strata, so observations are necessarily focused on these Eocene examples. There could be other, more deeply buried occurrences that have not been detected seismically.

The most obvious visual example of deformation in all of the samples from the New Jersey Transect occurs a few meters to either side of the lower/middle Eocene boundary at Site 613. This level marks the third Paleogene cut-and-fill horizon (Table 1; Fig. 4). Cores 613-39 to -37 contain several soft-sediment folds on scales from centimeters to meters. No exotic material has been detected in these deposits, and no biozones appear to be missing. The only suggestion of a gap is based on foraminiferal sedimentation rates which indicate a possible hiatus of 1 to 2 MY across the boundary. Similarly, sedimentation rates at Site 612 suggest a 1 to 5 MY gap across this same boundary (Miller and Hart, 1987), but this is not accompanied by deformed sediments. A clast of Maestrichtian limestone rests directly on the eroded surface in core 612-37 (Poag and Low, 1987); it constitutes the only evidence for displaced material at this gap, and its significance is not well understood.

Detailed sedimentation rates are not available for a similar estimate at Site 605. However, the same lower/middle Eocene gap is inferred as follows. At 605 a 3-cm ash layer occurs 10 m above the lower/middle contact; at 613 (roughly 2 km downslope from 605, Fig. 3) what is assumed to be the same ash layer is 37 m above the contact. Only 9 m of this 27 m difference is accounted for by slumps at 613, implying that if there is a gap at the lower/middle Eocene boundary then it must be greater at Site 605. As with all of the older cut-and-fill horizons (Table 1), this suggests that as a rule the most nearly complete depositional record occurs at the base of the slope. Furthermore, it suggests that after each brief erosional episode, the slope is covered again from the base of the slope upwards.

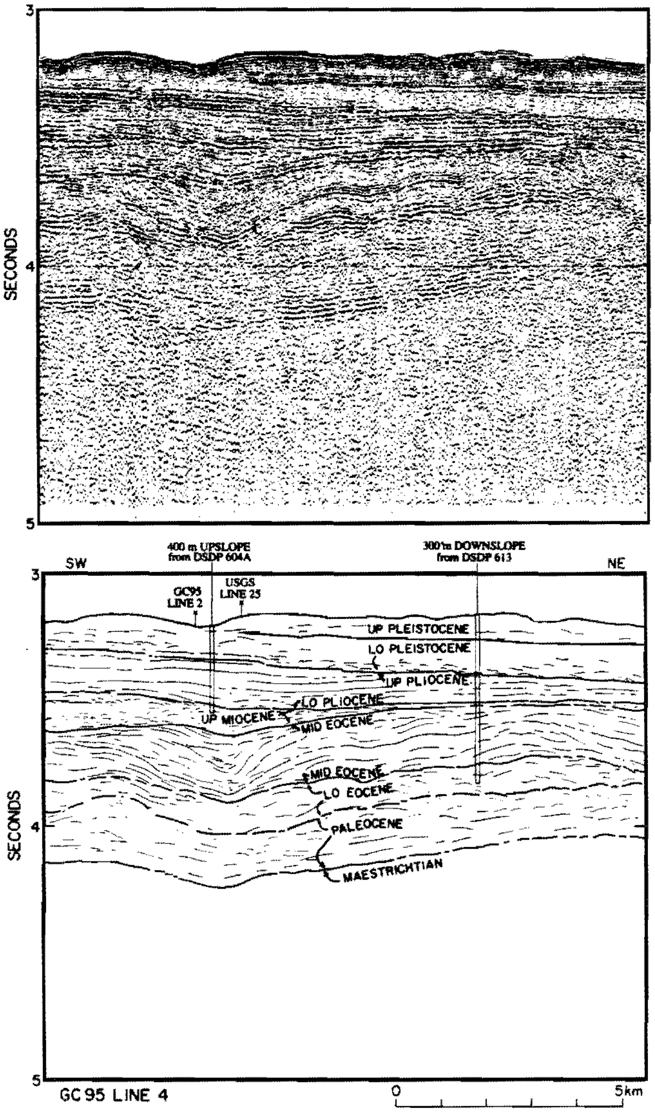

FIGURE 6 - Original and line drawing interpretation of Glomar Challenger Leg 95 Line 4 near DSDP Site 613, where the following was recovered: a) acoustically featureless lower Pleistocene debris flow deposit unconformably overlying b) stratified Pliocene glauconitic mud, c) feather edge of upper Miocene conglomeratic mud unconformably resting on d) middle Eocene chalk with slumped intervals conformably on e) lower Eocene porcellantic chalk with additional slump intervals. The axis of a set of stacked Paleogene cut-and-fill channels is 5.5 km SW of Site 613 (profile location in Fig. 11).

A surprising fact, however, is that the more nearly complete base-of-slope sedimentary record does not contain material eroded from the coastal plain or the continental shelf. Even the slump deposits at the lower/middle Eocene boundary at Site 613 are compositionally identical to the enclosing sediment;

their only detectable difference is slightly decreased bioturbation, and a consequent increase in the preservation of degradable organic carbon. This is interpreted to mean that these slumps originated farther upslope in a zone of relatively oxygen-depleted bottom water (Tarrafa, and others, 1987). That original bedding planes are generally intact, that little to no section is missing, and that these strata show soft-sediment deformation is consistent with the conclusion that these slumps came to rest at Site 613 after travelling no more than a few kilometers downslope.

The fourth Paleogene cut-and-fill episode (Table 1) was cored at Site 613. Logging confirms the match of this surface to the isolated slump in core 613-29. The correlation at 605 is less certain, but extrapolating sonic log velocities from 613 places the reflector in core 605-11, where horizontally deformed burrow mottles and laminae were observed (Leg 93, unpublished Shipboard Report). At both sites these sediments are radiolarian-nannofossil limestones with only a trace of displaced debris. Shipboard nannofossil analysis placed these sediments within zone CP13 at 605, and with more detailed analysis at 613, within CP13b. The pattern from the underlying cut-and-fill horizons would predict a small hiatus at this surface, but none has been detected. The reflector marking this surface can be traced upslope to Site 612 (Figs. 2 and 5), where it again occurs within zone CP13b. No breaks, slumps or deformed strata were found at this latter site, and none are implied from seismic data.

Slumping continued on the lower slope after the time of CP13b, but apparently was too localized to have been sampled at Site 613 (Figs. 4 and 6). Later Paleogene erosion and slumping may have occurred, but because much of the middle and all of the upper Eocene has been removed from the middle slope to upper rise, this cannot be determined. Based on sedimentation rates, a 5 MY hiatus was detected on the slope at the middle/upper Eocene contact in core 612-21. If the pattern of the underlying cut-and-fill episodes has validity, this hiatus probably was matched on the upper rise by a shorter gap separating undeformed middle Eocene deposits from upper Eocene slumps.

Late Oligocene to Miocene

The Paleogene cut-and-fill episodes just described occur at and above the oldest of three margin-wide unconformities described in the Introduction (Fig. 1). The next youngest major unconformity separates the upper Eocene from the upper Oligocene. At Site 612, however, a few tens of cm of lowermost Oligocene were found above the upper Eocene; despite the fact that no biozones are missing across this boundary, subtle compositional contrasts have been cited as support for a minor Eocene/Oligocene hiatus beneath the very much more significant lowermost Oligocene/middle Miocene hiatus (Poag and Low, 1987).

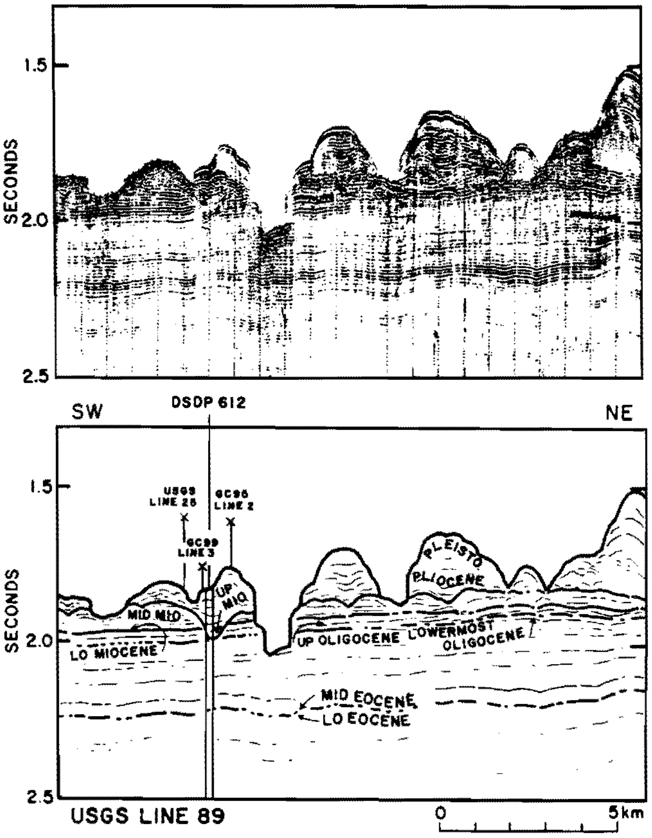

FIGURE 7 - Original and line drawing interpretation of USGS Line 89 passing over DSDP Site 612 showing acoustically layered pre-Upper Miocene sediment cut by an upper Miocene canyon directly beneath the Site. None of the channelling or slumping that is characteristic of the Eocene section on the lower slope/upper rise (Figs. 4 and 6) is observed here on the middle slope (profile location in Fig. 11)

A substantial lithologic change occurs across the 25 MY gap separating lowermost Oligocene from middle Miocene sediments in core 612-16; this contact spans both the second and the third of the regional unconformities cited in the Introduction (Fig. 1) because of the unintended Site location within a narrow slope canyon (Fig. 7). Unfortunately, Early Neogene slope by-pass plus bottom current erosion resulted in an even longer hiatus on the uppermost rise (Sites 605 and 613; Figs. 4 and 6). Consequently, the following discussion of earliest Oligocene to middle Miocene is based largely on inferences drawn from seismic data.

Erosion into the uppermost Eocene/lowermost Oligocene shelf section is difficult to detect in MCS profiles because the angular discordance is so sublte (Fig. 2). A few tens of meters of strata are banked against prograding Eocene clinoforms on the innermost shelf. Traced to COST B-2 and COST B-3, this unit correlates to mudstones with variable amounts of siliceous and calcareous biogenic components that contrast sharply to the underlying Eocene/lowermost Oligocene argillaceous limestones. At its maximum thickness, this section measures only 70 to 90 meters because, in turn, it has been partially eroded and unconformably overlain by middle Miocene and younger strata on the shelf and upper slope. Nannofossil stratigraphy places this thin unit entirely within the upper Oligocene (Valentine, 1980); based on foraminfera, it extends across a conformable boundary and up into the lowermost Miocene (Poag, 1980).

The top of the Eocene beneath the continental rise is also marked by an erosional unconformity (Reflector A^u; Tucholke and Mountain, 1979.) This is an extensive hiatus throughout the western basins of the North Atlantic, documenting a pulse of swiftly flowing bottom water near the Eocene/Oligocene boundary (Miller and Tucholke, 1983). The oldest sediment yet sampled above this gap is lower Miocene (?)/middle Miocene (Hart and Mountain, 1987), and is found at Site 603 on the lower rise, 250 km seaward of the New Jersey slope. Speculations concerning the age and depositional history of post-Eocene sediments on the intervening central and upper continental rise is drawn solely from seismic profiles.

It has been proposed that "mid"-Oligocene turbidites may rest on A^u beneath the upper rise off New Jersey (Mountain and Tucholke, 1985; Miller and others, 1985). These workers speculated that the margin-wide lack of a lower Oligocene shelf section was caused by subaerial exposure during a "mid"-Oligocene eustatic lowstand (Vail and others, 1977), and that shallow-water debris eroded from the shelf now rests on Reflector A^u beneath the rise. Deeply incised, buried canyons on the margin off South Carolina support this arguement, and though not sampled, their stratigraphic position is consistent with a "mid"-Oligocene age. In contrast, the equivalent (and more conclusively dated) surface beneath the New Jersey slope does not show evidence of canyon erosion (Fig. 7). Instead, this unconformity is relatively smooth and shows only very slight angular truncation in strike direction, cutting more deeply into the top of the Eocene towards the southwest.

Oligocene canyons along the lower slope cannot be discounted off New Jersey; any that may have formed in this region have been removed by subsequent erosion. Strike-oriented profiles show that the top of the Eocene is a highly irregular surface along the uppermost rise, as would be expected of canyons (Fig. 8). However, with the quality of available profiles in this region, it is not certain that the observed irregularity results from post-Eocene erosion or from inherited Eocene structures.

Late Middle Miocene to Present

The third and youngest margin-wide unconformity described in the Introduction constitutes a 4 to 7 MY gap across most of the lower Miocene (Fig. 1). Along New Jersey it marks the base of a sharply changed depositional environment: a large delta advanced across the shelf, and the eroded upper Oligocene/lowermost Miocene section was covered by a thick wedge of prograding sands and shales. Accumulation rates were as high as 248 m/MY (Poag, 1980), but because of the localized nature of deltaic lobes, calculated rates can vary over short distances and short time averages. Nonetheless, while the 40+ million years of Paleogene are represented on the shelf by roughly 500 meters of sediment, these last 10 million years of the Miocene are represented by 800 meters of sediment.

A similarly disproportionate amount of Miocene is found on the continental rise. It has been shown (Mountain and Tucholke, 1985) that the high influx of terrestrial sediment from the North American continent (for example, from units that prograded out onto the New Jersey slope and beyond, Fig. 2), coupled with the margin-parallel flow of the Western Boundary Undercurrent (Heezen and others, 1966) resulted in thick sediment drifts on the continental rise. The region off the mid-Atlantic states was an especially large depocenter, and as much as 2000 meters of sediment accumulated in the Miocene alone. However, there is no drill core data to determine if there was a significant increase in accumulation rate on the rise corresponding to the age of the lower Miocene section that is missing from the shelf.

Prograding middle Miocene strata built up, out and across the New Jersey outershelf. Several sets of clinoforms reached the location of the modern slope, suggesting episodes of accelerated cross-slope transport. Paradoxically, there is no suggestion in strike line profiles that canyons cut across the slope before late Miocene time (Figs. 4 and 7).

Features proposed above as "mid-Oligocene" canyons are observed in profiles from the uppermost rise (Fig. 8), but until there are better quality strike line profiles and drilling on the uppermost rise there remains the

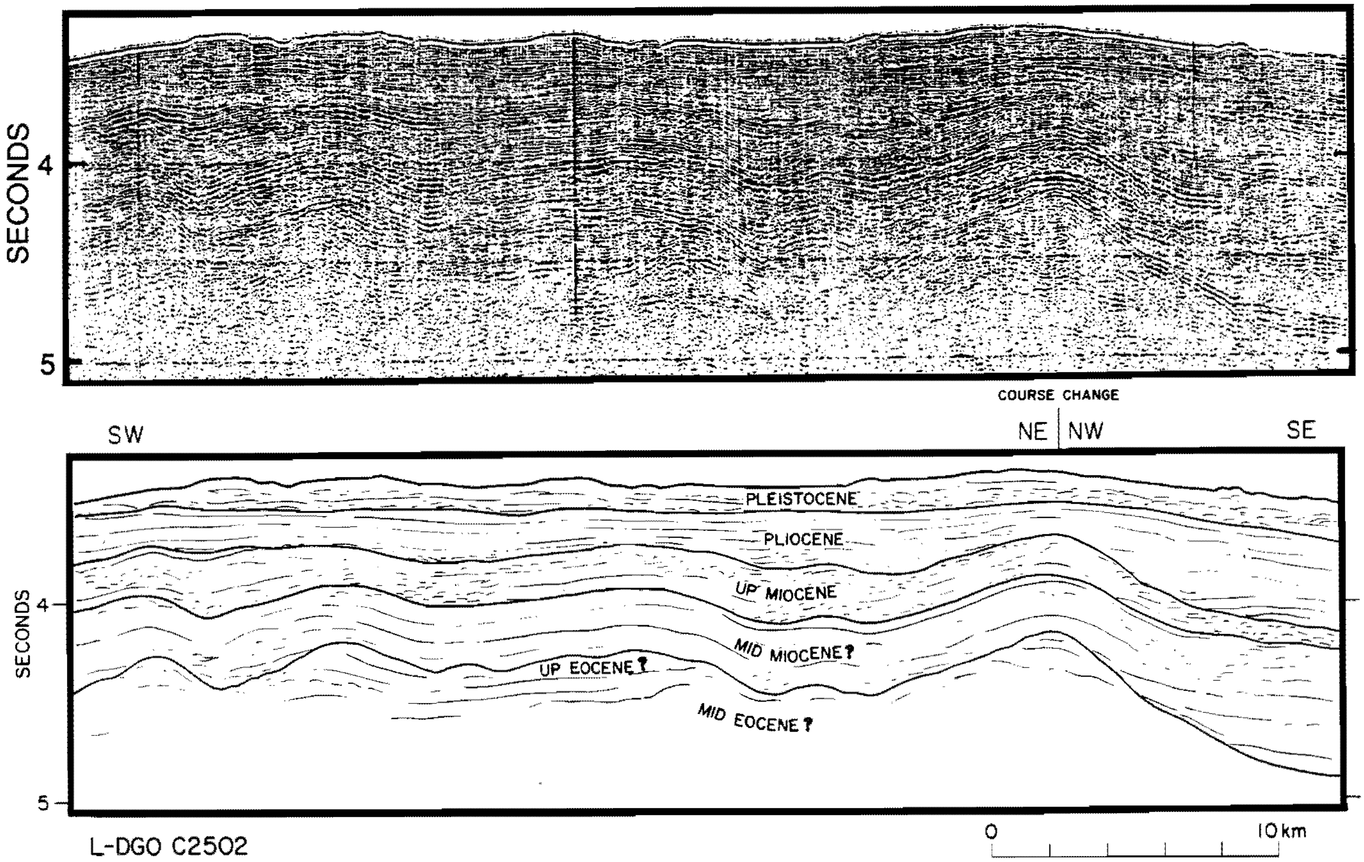

FIGURE 8 - Original and line drawing interpretation of C2502 showing unsampled section of probable middle Miocene age that suggests down-slope erosion cut into Eocene strata before the major late Miocene canyon-cutting event encountered at Site 612 (Fig. 7; profile location in Fig. 11).

possibility that these are middle Miocene canyons.

One of the unanticipated discoveries of drilling on the New Jersey margin was the evidence for severe canyon erosion and down-slope sediment transport near the middle/late Miocene boundary (Fig. 7). The disappointment of missing most of the Oligocene section because of one of these narrow canyons at Site 612 was compensated by the fortunate documentation of this Miocene erosional event. A 5-cm glauconitic sand layer rests on the middle Miocene/basal Oligocene contact in core 612-16. However, 30 meters of overlying Miocene fill is relatively fine-grained, homogeneous mud. The foraminiferal assemblages within this interval have not been displaced significantly (Katz and Miller, 1987), suggesting that the canyon was filled with slope material, probably from the canyon walls. These sediments provide a minimum age for the canyon-cutting event of 9.5 ± 0.5 MY (CN7, time scale of Berggren and others, 1985). The youngest sediments not cut by this canyon can be traced seismically to where they correlate to 80 meters above the shallowest sample at the COST B-3 well (Fig. 9); this shallowest sample has been dated as 12.7 ± 1.2 MY (Mellilo, 1985 Miller and others, 1987). Upward extrapolation of the middle Miocene sedimentation rate of 80 m/MY measured at B-3 (Poag, 1980) provides an estimate of 11.7 ± 1.2 MY for the maximum age of this canyon-cutting event. Consequently, the narrowest interval constrained by these data places the canyon-cutting event between 10.5 and 10.0 MY; the more conservative estimate of between 11.7 and 9.5 MY, however, is adopted here (see Miller and others, 1987).

This canyon beneath Site 612 is one of several other Miocene canyons cut into the upper slope (Fig. 7). The precise down-slope extension of any of them cannot be documented because of the Eocene outcrop along the New Jersey lower slope (Figs. 2 and 4). Strike line profiles, however, show equivalent canyons on the uppermost rise (Figs. 4 and 6). Site 604 was abandoned in loose sands that fill one of these rise canyons; nannofossil ages in these deposits (van Hinte, Wise, and others, Shipboard Report) are identical to those from the canyon thalweg at 612. In lithologic contrast to those at 612, however, these rise deposits commonly contain displaced material such as glauconitic sands, shell fragments, inner shelf microfauna, quartz pebbles, and

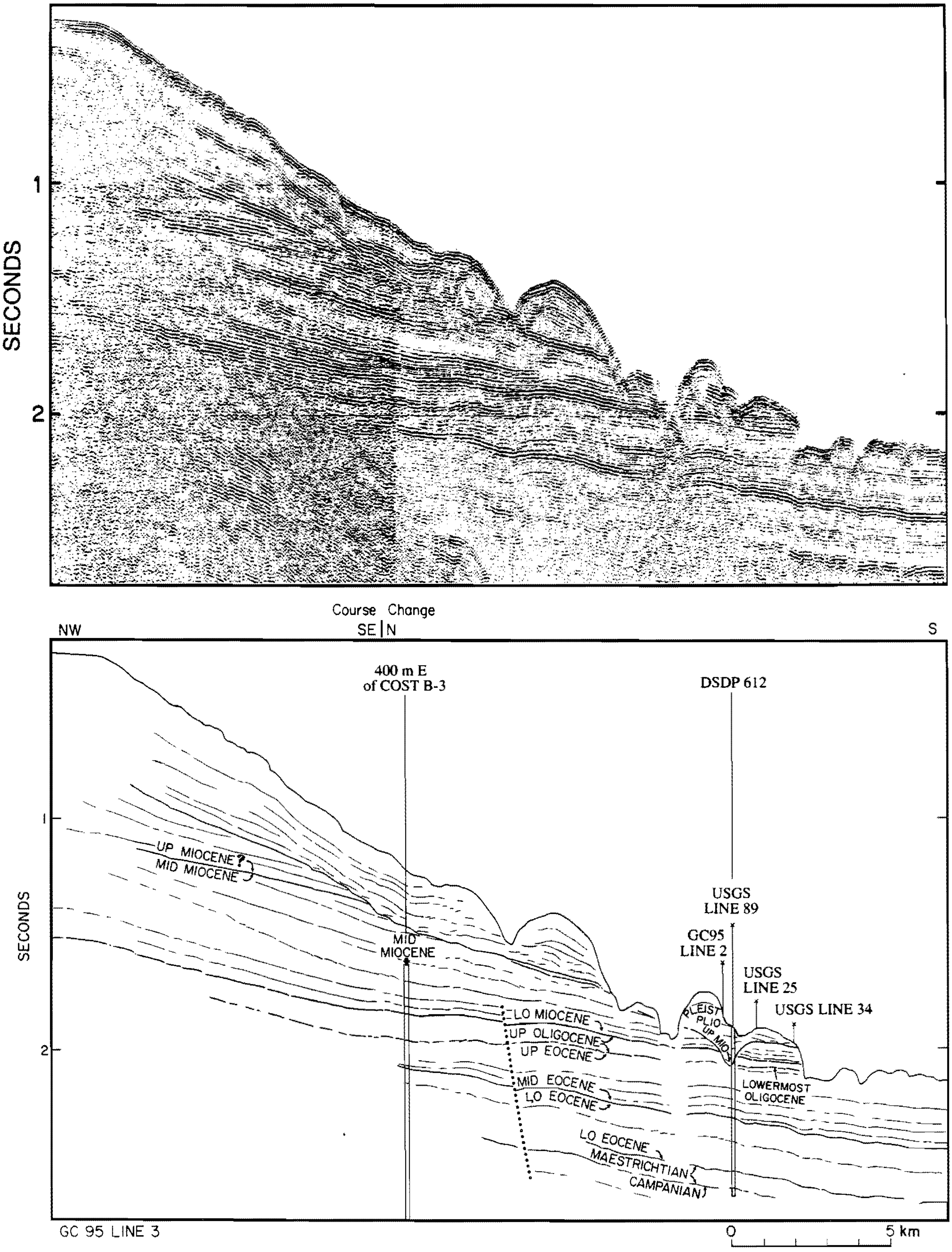

FIGURE 9 - Original and line drawing interpretation of Glomar Challenger Line 3 crossing over DSDP Site 612 and passing near COST B-3. The oldest canyon-fill at 612 and the youngest sediments extrapolated upwards from the shallowest sample at B-3 bracket the canyon-cutting event between 9.5 and 11.7 MY (profile location in Fig. 11).

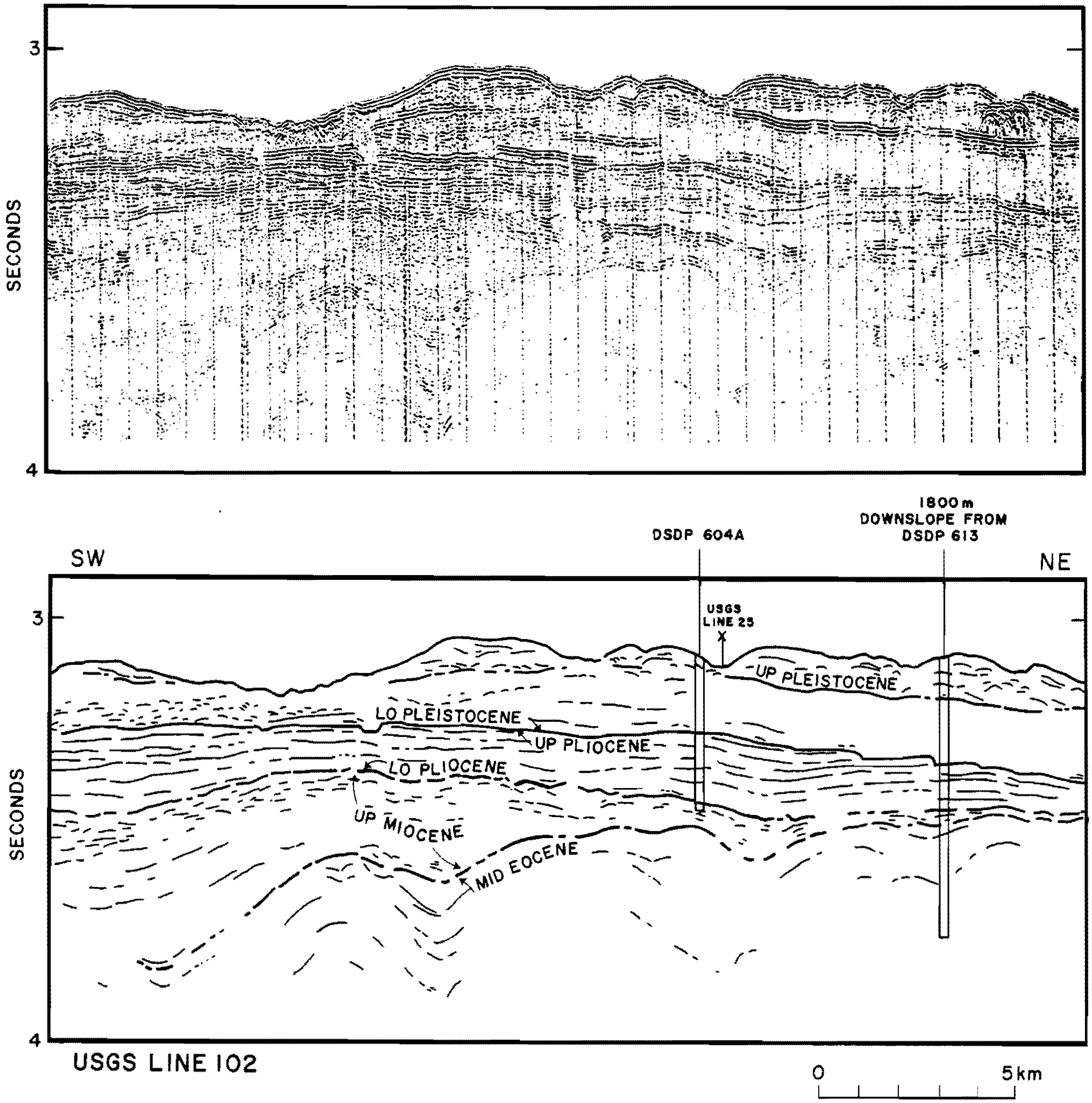

FIGURE 10 - Original and line drawing interpretation of USGS Line 102 passing over DSDP Site 604A and near Site 613. The profile shows an acoustically featureless lower Pleistocene debris flow unit unconformably resting on a seismically laminated Pliocene section. The latter is interpreted to be alternating mud and glauconitic sand. These sandy intervals probably served as detachment surfaces for the bedding plane slides that separate the Pliocene and Pleistocene units (profile location in Fig. 11).

metamorphic rock fragments; even limestone fragments from the Eocene outcrop belt were found (van Hinte, Wise, and others, Shipboard Report; Poag, 1985b). To avoid the thickest and coarsest-grained part of these deposits, Site 613 was located near the edge of this canyon fill. A few meters of conglomeratic sand were successfully penetrated at this site and determined to be identical in age to the canyon deposits at the other two drill sites.

Older, high-amplitude reflectors onlap the eroded top of the Eocene seaward of Site 604 and 613 (Figs. 2, 5 and 8), suggesting that earlier events like the one between 11.7 and 9.5 MY transported coarse-grained, shallow water debris to the continental rise. Post-Eocene canyons have been tentatively identified on the rise seaward of Site 613 (Fig. 8; Poag and Mountain, 1987.) As has been noted, however, there is no evidence that canyons cut into the New Jersey slope before late middle Miocene time.

The occurence of Upper Miocene sediment younger than CN7 on the uppermost rise has not determined clearly at Sites 613 and 604 (compare van Hinte, Wise *et al*, Shipboard Report with Poag, 1985b). As much as 3 MY of late Miocene may be missing, making the Miocene/Pliocene boundary and erosional contact.

The Pliocene sediments on the upper rise correlate to a particularly unique seismic interval (Fig. 6). This unit displays marked acoustic layering that in strike direction is level and laterally continuous over many kilometers (Fig. 10). This seismic character suggests the interbedding of contrasting strata, but paradoxically, the cores from both Site 604 and 613 are predominantly fine-grained muds. Only one interval, core 604-19, recovered any graded sand units. It may be argued, however, that core recovery from this Pliocene section was not representative. The muds recovered at 613 were firm enough to have been broken into "biscuits", and the drilling slurry between them was consistently rich in sand-sized glauconite; these sands were rarely observed within the intact biscuits. The conclusion is drawn here that the Pliocene sediments at this location are actually fine-grained, dense muds interbedded with layers of porous, loose glauconitic sands, and few of these latter beds were recovered intact.

For these or any sands to have reached the upper rise in such laterally uniform blankets (Fig. 10), channelized transport across the slope is highly unlikely; deposits emanating from such canyons would more likely occur as localized and acoustically chaotic units, such as the late Miocene sands found at the bottom of Site 604. Consequently, it is speculated that the canyons that had been cut into the upper slope between 11.7 and 9.5 MY were filled and no longer topographic features during the Pliocene. This is supported by correlation to Site 612 (Fig. 7).

The inferred lithologic heterogeneity of the Pliocene section on the upper rise may have contributed to another seismic feature unique to this interval. The top of the acoustically laminated Pliocene unit ends very abruptly, often as box-shaped erosional channels filled with faint, chaotic reflectors (Fig. 5). This contact has been interpreted as evidence of bedding plane slides (Farre, 1985); the unconsolidated sands speculated in the Pliocene section may have provided the necessary detachment surfaces.

These slides need not have occurred during the Pliocene. The Pleistocene muds at Site 613 have a particularly homogenous gamma-ray log character that suggests uniformly high clay content and which correlates to the interval of slumps reported in visual descriptions. It is speculated here that very large slope failures occurred during the early to middle Pleistocene, and that the loose Pliocene sands were often the weak, underlying failure surfaces. Only thoroughly homogenized, fine-grained muds were left behind to cover the slide scars. The diagnostic Pliocene/Pleistocene contact (Figs. 5 and 10) can easily be traced in profiles several tens of kilometers farther onto the rise. The weakly reflective Pleistocene section grades downslope into more highly-reflective debris flow deposits that are tens of meters thick and tens of kilometers in cross-sectional area (C2502 unpublished seismic data.)

No younger detachment surfaces can be recognized in profiles, and no slumps were reported throughout the remaining Pleistocene section. Nonetheless, glauconitic and quartzose sands plus Eocene limestone fragments eroded from the lower slope outcrop belt continued to reach the rise at Sites 604, 605 and 613, indicating that erosional processes continued to remove and transport large amounts of sediment across the New Jersey margin. Without additional strike line profiles updip from Site 612, it cannot be determined how many episodes of Pleistocene canyons cut into the upper slope. The upper Pleistocene rests unconformably on the upper Pliocene at Site 612, suggesting that many of the canyons now on the New Jersey margin may have formed during the intervening 2 MY gap.

SYNTHESIS

Paleogene vs. Pleistocene Margin Erosion

The four episodes of Paleogene margin erosion described in this report contrast with major Pleistocene analogues off New Jersey and elsewhere in terms of both geometry and location. Furthermore, the character of the sediment filling these ancient and modern features is distinctly different. These facts suggest contrasting processes of slope defacement.

The geometry of the channels buried off New Jersey were measured by transforming tracings of seismic profiles to depth sections using the sonic log velocities measured at DSDP Site 613 (Poag, Watts and others, 1987). The best examples of these channels are broad, U-shaped features now beneath the uppermost rise (e.g. Figs. 4 and 6). Each widens seaward from less than 1 km to more than 15 km between the middle and lower slope; similarly, each deepens from a few tens of meters to roughly 200 m across this same distance. Channel axes have gradients of 1:10, while their walls face inward at a more gentle gradient of 1:17 or less.

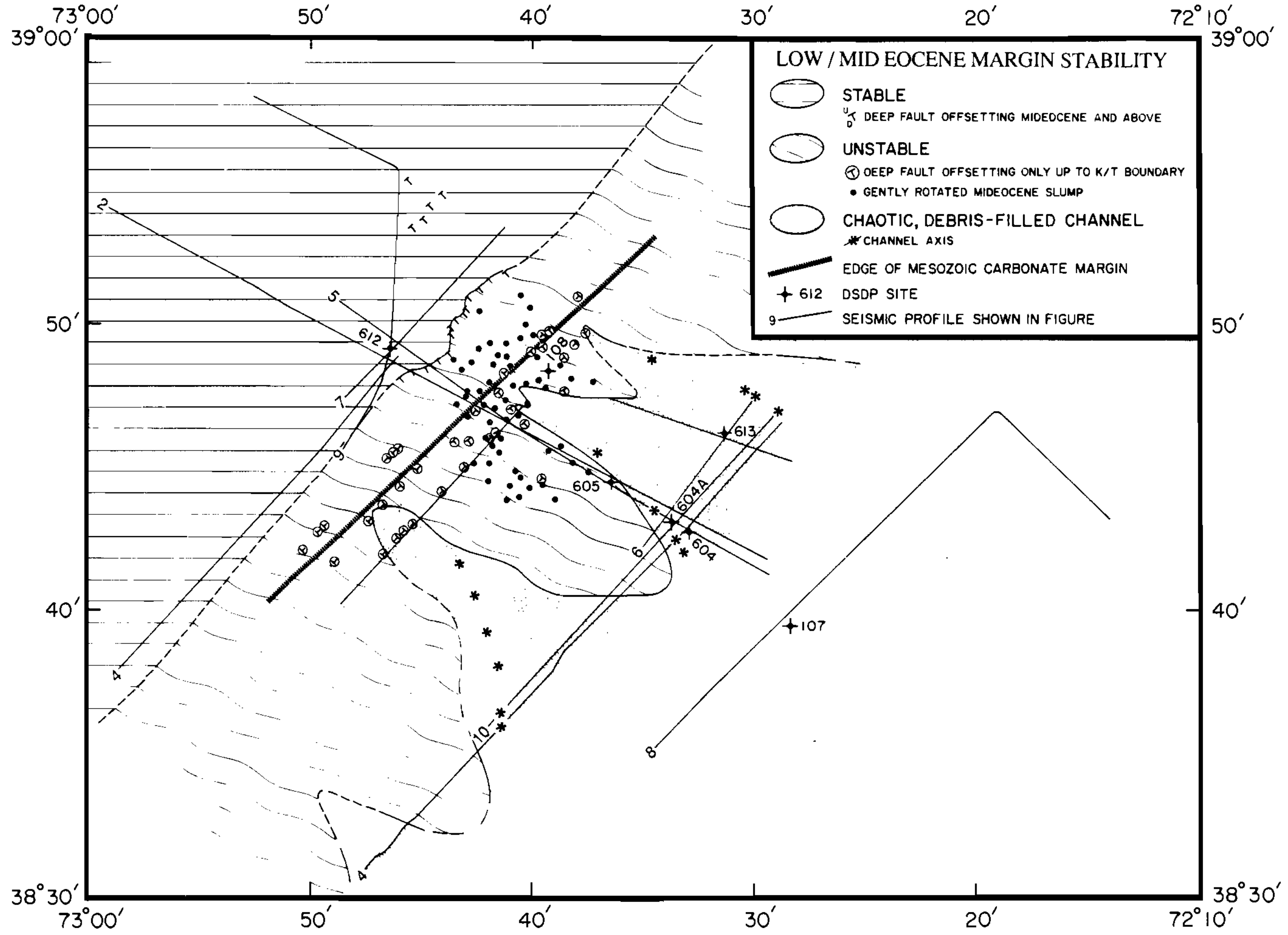

FIGURE 11 - Summary map of features near the lower/middle Eocene boundary. Three seismic provences are recognized: 1) stable inner region where reflectors are continuous, parallel and undeformed except for a few deep, down-to-the-basin growth faults; 2) unstable middle region of dominantly intact but rotated slump blocks along with numerous, irregularly oriented normal faults cutting the K/T boundary; and 3) discontinuous, chaotic reflectors that partly fill channels originating above the edge of the underlying Mesozoic carbonate margin. Numbers next to profiles refer to their figure numbers in this report.

In contrast, major Pleistocene analogues offshore New Jersey are dramatic V-shaped incisions that maintain more nearly uniform width down the continental slope. Admittedly their cross sectional dimensions depend on the physiographic province in which they are measured: canyons on the upper slope (Fig. 7) are far narrower and deeper than they are on the upper rise (Fig. 10). Nonetheless, there are clear differences between the sharpest Pleistocene features and the sharpest Paleogene analogues. Major upper slope Pleistocene canyons in this region are typically 3 km wide, 400 m deep and have down-slope axial gradients of 1:10. Their walls dip inward at steeper gradients, often as high as 1:5.

While Pleistocene slope canyons can be traced landward to the edge of the continental shelf, Paleogene examples terminate farther seaward. Of all the ancient canyons, those cut near the lower/middle Eocene boundary can be traced farthest landward, but even these reach only as far as today's middle slope, disappearing near the 1700 m isobath (Fig. 11). In strike lines passing along the 1300 m isobath near Site 612 (Figs. 4 and 7), lower/middle Eocene reflectors show no signs of channeling. Backstripping Site 612 returns this location to a middle Eocene water depth of slightly more than 1 km (Steckler, 1987; Miller and Katz, 1987). No abrupt change in dip, i.e. a shelf break, can be found in the Eocene section landward of this point (Fig. 2). Apparently during the early and middle Eocene there existed a wide, uniformly seaward dipping margin off New Jersey, and the only major defacement occurred between Sites 612 and 613, in considerable water depths at great distance from the shoreline.

A third distinction between Pleistocene and Paleogene slope defacement is the character of sediment found within the channels. Major Pleistocene features off the mid-Atlantic states are generally swept clean of thick accumulations of debris (McGregor and others, 1982), thus preserving their sharp, V-shaped cross section. Samples and visual observations from these and other nearby canyon axes, however, occasionally reveal an irregular veneer of coarse-grained, shallow-water debris and blocks of lithified sediment eroded from canyon walls (Farre, 1985). This displaced material will eventually be transported to the continental rise by turbidity currents or debris flows, and will probably be replaced by more debris that will remain in the canyon only temporarily.

Not all continental slopes are swept so cleanly. In many ancient settings, canyons have been filled with pebbly mudstone deposits that are considered diagnostic of slope canyon fill (Stanley and Unrug, 1972). This sediment is characterized by a mixture of grain sizes and materials gathered from a variety of environments, bound in a muddy matrix and deposited within base-of-slope canyons during catastrophic gravity flows. If the processes clearing Pleistocene canyons offshore New Jersey were to end, these canyons, too, might begin accumulating pebbly mudstone fill.

Only two of the four Paleogene channels offshore New Jersey contain sediment that comes at all close to matching either of these "typical" canyon fill deposits (Table 1). These are: 1) the glauconite-bearing, silty marls of basal Paleocene age in Site 605, and 2) the solitary Cretaceous limestone clast in the lower/middle Eocene channel at Site 613. This latter example is contained within a few tens of meters of obviously slumped sediment that ironically lacks any other evidence of reworked material. On the basis of geochemical character these slumps probably originated no more than a few kilometers upslope in a zone of moderately oxygen-depleted bottom water (Taraffa and others, 1987). It is obvious they were transported to Site 613 shortly after the time of their original depostion because 1) they are within the same biozone as the enclosing *in situ* sediment, and 2) they show clear evidence of soft sediment deformation.

The Role of Deep Structures

Differences between Pleistocene and Paleogene margin defacement are suggested further by noting the proximity of the latter to underlying structures and deep-seated faults. Local processes related to these structures may have contributed to the inferred slope failure.

The lower/middle Eocene "cut-and-fill-episode" is discussed more thoroughly here because it is the clearest to trace in seismic profiles, and because it extends the farthest landward of all four of the Paleogene channel systems (Fig. 11). Three categories of margin reflectors are identified: 1) acoustically laminated, parallel, but gently seaward-dipping "shelf" strata (deposited in water far deeper than the modern shelf, however); 2) gently folded, rotated and slumping reflectors forming an unstable "slope"; and 3) irregular, sometimes chaotic, discontinuous reflectors making up the channel fill deposits. The distinction between types 2 and 3 is subtle and somewhat dependent upon the orientation of the profile: rotated slumps are much more easily recognized in dip lines than in strike lines, while the reverse is true for the recognition of irregular reflectors that characterize fill deposits.

The edge of the Mesozoic carbonate margin (Fig. 2) lies beneath the "unstable slope" facies (Fig. 11). As mentioned previously, the edge of this margin prograded seaward throughout the Early Cretaceous, and was overstepped by an advancing clastic wedge in Barremian time. It is not known how long a time intervened between the end of carbonate buildup and clastic burial, but it is reasonable to assume that there was sufficient time for the margin edge to have been exposed subaerially and to have developed extensive secondary porosity.

Few strike lines are available that have both the resolution and penetration to demonstrate that faults continue down to the edge of the Mesozoic carbonate margin. Nonetheless, USGS line 93 (Fig. 12) shows that moderately deep-seated faults closely coincide with the location of this underlying structure (Fig. 11). Numerous offsets of the unconformable Cretaceous/Tertiary boundary (at this location probably lowermost Eocene resting on Maestrichtian, Fig. 2) are concentrated above the carbonate edge. All faults are steeply dipping normal offsets with no apparent preferred orientation to their slip planes. If the fundamental cause were either 1) differential compaction of post-Barremian draping facies, or 2) slope failure along seaward-dipping bedding planes, the majority of faults would be down-to-the-basin. That they are not suggests some other cause.

Lower/middle Eocene slump deposits frequently lie directly above these K/T boundary faults (Fig. 12). In general, the lower Eocene unit appears in profiles to be more disrupted and choatic than the middle Eocene. This is probably due to two

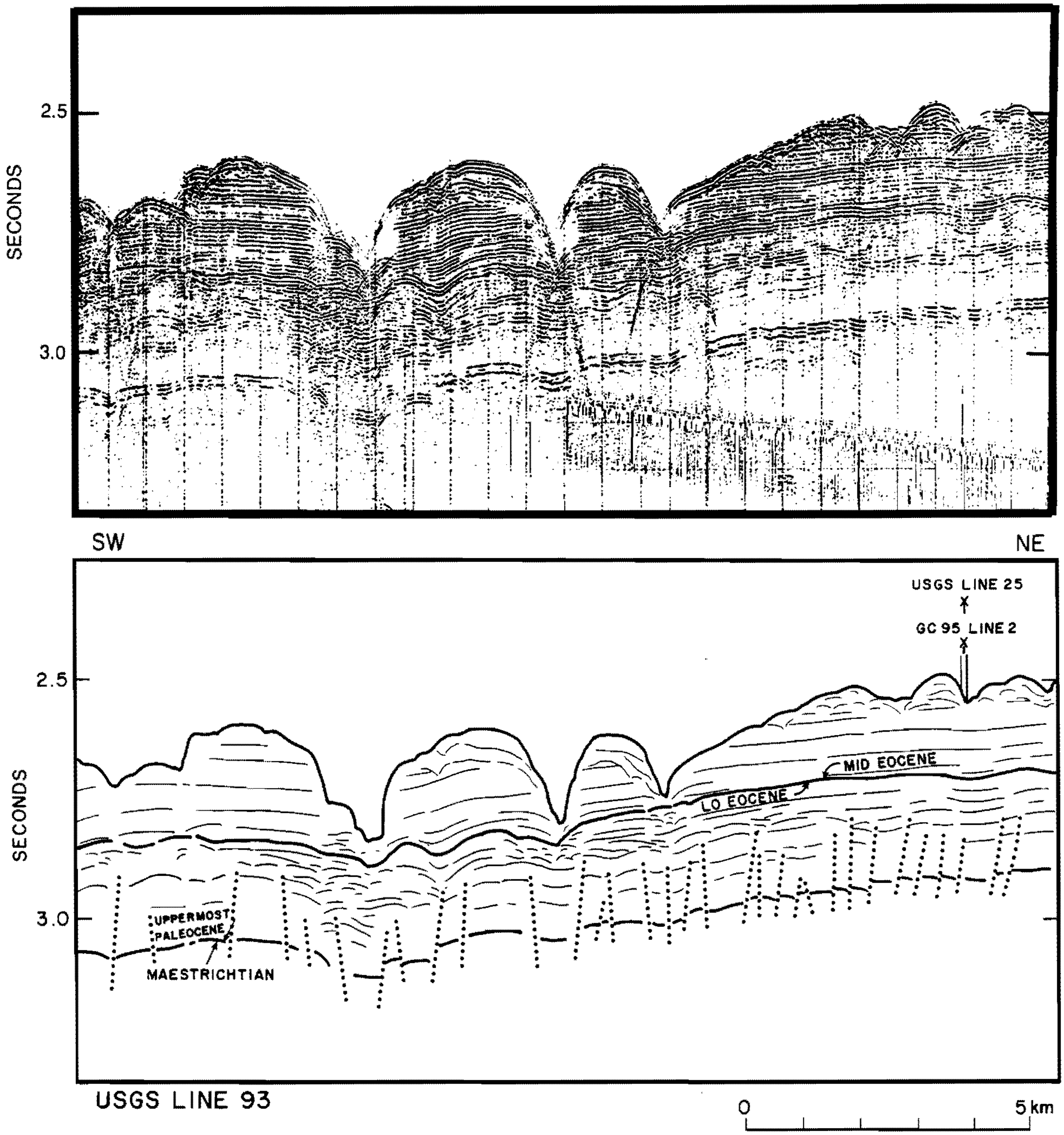

FIGURE 12 - Original and line drawing interpretation of USGS Line 93 showing numerous steeply dipping, variously oriented normal faults cutting the unconformable K/T boundary. Slumps in the overlying Paleogene strata tend to be concentrated above these faults (profile location in Fig. 11).

processes: 1) early silica diagenesis in the lower Eocene led to more brittle deformation than in the less indurated and more ductile middle Eocene section (see Wilkens and Schrieber, 1987); and 2) many (perhaps all) are growth faults that have had a long history of movement, so it follows that the lower Eocene rocks show greater offset than younger ones.

Working Hypotheses of Paleogene Slope Erosion

Hypothesis 1 - slumping triggered by deep porosity reduction

The spatial correlation between the buried Mesozoic carbonate structure, deep-seated growth faults, and overlying lower/middle Eocene slumps and channels suggests the following link. Secondary porosity developed along the seaward edge of this structure during subaerial exposure in the Early Cretaceous. Once buried by a blanket of post-Barremian clastic debris, this structure was subjected to extensive collapse. It is likely that this porosity reduction generated variously oriented growth faults that settled over long periods of time.

True dips of lower Eocene reflectors above the carbonate edge have seaward gradients of roughly 1:12. Not much of a fault scarp on seafloor was required to increase this local dip to the threshhold of sediment failure. Once started, a slump scar would have widened along-slope and eroded up-slope, similar to the way failure (by whatever cause) is defacing the lower slope off New Jersey at present (Farre and others, 1983). The mildly deformed "unstable slope" facies (Fig. 11) is evidence of these slumps. Seaward of the buried edge of the carbonate margin, where dips were steepest, slumping was especially intense and led to the "chaotic, debris-filled channel" facies. In this manner, the lower/middle Eocene cut-and-fill episode is interpreted as an arrested development of continental slope canyons: each began as a slump scar, but none ever breached the shelf break and hence none was ever provided with the sand-rich turbidity currents that erode fully-mature canyons.

Growth faults can be traced up through the entire Eocene section, and the process of deep porosity reduction, faulting and slumping must have been active at least up to Oligocene time. An important question, however, is why should the slumping process have occurred so sporadically over these tens of millions of years? Furthermore, is it circumstantial that the four Paleogene episodes described here so closely match the time of coastal onlap events?

There are many possible links between changing sea level and slumping. One of these involves changing hydrostatic pore pressures. During lowstands, seawater depth on the slope would have decreased and the hydrostatic "head" on freshwater aquifers reaching the Paleogene margin would have increased (see Hathaway and others, 1976). The result would have been increased *in situ* pore pressures, and these valuses may have reached the point of failure in Paleogene sediments that were on fairly gradual slopes.

Hypothesis 2 - very rapid canyon cutting during eustatic lowstands

The close agreement in time between some (but not all) events on the coastal onlap curve (Vail and Hardenbol, 1979) and the Paleogene cut-and-fill episodes suggests a more direct cause and effect relationship than suggested above (see also Poag and Mountain, 1987). It could be argued that the more traditional mechanism of canyon formation was responsible (Daly, 1936), and that wall collapse filled in slope canyons that were originally carved by turbidity currents. The difficulties with this model are several.

The first argument against this model is that in the upper slope areas of known stratigraphic gaps (for example, near DSDP Site 612), there is no seismic evidence that canyons cut into any of the Paleogene unconformities (Figs. 4 and 7). Of course, even the very large Pleistocene canyons such as Hudson or Wilmington have only very subtle surficial expressions on the shelf; their upper slope expressions, however, are impossible to miss. If one is to argue that the Paleogene features are seaward extensions of coastal rivers, the only explanation for no counterparts on the slope is that they were planed off by submarine currents shortly after they were formed. A link between sea-level changes and the location and intensity of near-surface currents has been proposed for this and other time periods (Pinet and Popenoe, 1982; Olssen and Wise, 1985).

A second problem with this model is the fairly short (1-2 MY) chronologic break across the Paleogene lower slope/upper rise channels. Only ice growth is thought to be able to drop sea level this fast (Pitman and Golovchenko, 1978), and there is little conclusive evidence for ice growth in pre-Oligocene time (Miller and others, 1985).

A third problem with this model for canyon formation is the scarcity of shallow water debris resting on the unconformities. Only the basal Paleocene glauconitic sands at Site 605 (van Hinte, Wise and others, Shipboard Report) provide evidence for cross-shelf sediment transport. A single Maestrichtian clast was found on the lower/middle Eocene contact at DSDP Site 612 (Poag and Low, 1987), but its significance is not clear. All other samples resting on the channeled strata described along the lower slope or upper rise -- most importantly the massive slumps in cores 613-37 to 39 -- contain locally derived, slope material.

Perhaps the only way to address these difficulties is to say that not enough is known about the various ways in which slope canyons come to be buried and preserved in the geologic record. The pebbly mudstones and shoe-string sand bodies that identify ancient slope and fan valleys (Stanley and Unrug, 1972) may be only two of several types of fill. Perhaps in the absence of a large supply of coarse-grained material, canyons collapse on their own, broaden their cross-sectional profile, and fill up with inter-canyon slope sediment. The actual composition of shelf and slope sediment may be an important controlling factor that is not yet understood; the siliceous- and carbonate-rich pelagic sediments that accumulated on the Paleogene margin are unlike the sediments found on most slopes where Pleistocene canyons have been studied.

New Patterns in the Neogene

The stacked arrangement of pre-Oligocene channels suggests a process common to each of these erosional events (Fig. 4). That subsequent features along the uppermost rise did not continue this alignment into Neogene time suggests a major change in margin processes sometime after the Eocene. Unfortunately, the 25 MY gap between these and the oldest post-Eocene strata yet sampled along the upper rise leaves an explanation for this change open to wide speculation. Two possible causes are proposed.

The first cause involves mid-Tertiary re-structuring of the margin by bottom currents. Erosion by thermohaline currents in the deep North Atlantic occurred near the Eocene/Oligocene boundary and formed the widespread reflector A^u (Tucholke and Mountain, 1979). This erosional unconformity can be traced seismically as far landward as the uppermost rise. Undercutting along the base of the slope and the Blake Escarpment by the same current that formed Reflector A^u has been proposed as a mechanism for the 5 to 25 km retreat of the margin that occurred sometime after Eocene time (Paull and Dillon, 1980; Schlee, 1981). If indeed much of the slope was removed off New Jersey (and this has yet to be fully documented), then it would not be surprising to find Paleogene patterns of slope channels do not continue into the Neogene.

A second factor that may have contributed to this changed alignment of margin features is the contrast in composition between Paleogene and Neogene sediments. Paleogene sediments from the base of the slope and upper rise are dominantly chalk or limestone, and biogenic silica has contributed further to varying degrees of diagenesis; porcellanite is now found in most samples from below the middle Eocene. It is likely that early diagenesis in the Paleogene strata contributed to the tendency of these older sediments to slump and not fail entirely.

Because of this diagentic "armoring" of the slope, channeled Paleogene topography was never completely restructured between cut-and-fill episodes, and slumps tended to concentrate in the same slope decleavities from one episode to the next. Neogene and Quaternary sediments, by contrast, are carbonate-poor mud with variable amounts of coarse-grained clastic debris. These cohesionless strata could not hold slopes as steep as those of the Paleogene, and more frequent and widespread slope failures prevent channels from concentrating in any one location for long periods of time.

Sand-Rich Miocene Rise Deposits

It is paradoxical that during the early and middle Miocene, when sedimentation rates on the rise were among the highest of the entire Cenozoic, there is no evidence in profiles from the study area that major canyons cut across the continental slope (Figs. 4 and 7). Highly reflective sequences of strata onlap the top of the Paleogene along the upper rise (Fig. 3), and just 20 km seaward of the base of the slope they are more than 500 m thick. Their similarity in seismic character to the uppermost middle Miocene sands at Sites 604, 604A and 613 proves beyond a reasonable doubt that they, too, contain thick bodies of coarse-grained sediment. This interval corresponds closely to the unit between reflectors A^u and Merlin (Mountain and Tucholke, 1985), and an isopach map of this unit shows a very large depocenter 300 km south of New Jersey, off the mouth of Chesapeake Bay. These authors proposed that this was a submarine fan depocenter whose form was partly controlled by contour currents flowing southwest along the margin, and that it later became the nucleus of the Chesapeake Drift.

There are at least three explanations for this lack of early Neogene slope canyons offshore New Jersey. First, these rise sediments may not have come from the adjacent New Jersey margin. Early and middle Miocene bottom currents could have been swift enough to transport base-of-slope fan sediment along the margin from some unrecognized sediment source north of the study area. Second, early and middle Miocene deposits may have reached the rise off New Jersey through fan valleys emanating from the known fan off Chesapeake Bay. Lastly, it must be acknowledged that margin-parallel profiles have been examined in a corridor only a few tens of kilometers wide along the slope (Fig. 3). Important early Neogene slope canyons may yet be found closer to the study area than Chesapeake Bay.

The only Miocene canyon within the study area was drilled at DSDP Site 612 (Fig. 7) and dates to 11.7 to 9.5 MY. The Eocene outcrop belt prevents the tracing of this canyon down the middle and lower slope (Fig. 2), but strike lines along the uppermost rise show that no equivalent canyon extends out onto the rise (Figs. 4, 6, and 10). The uppermost middle Miocene conglomeratic mud on the rise contains Eocene clasts, implying that some of this unit traversed the adjacent slope. How sediment by-passed this upper rise setting and contributed to the thick, coarse-grained Neogene sediment on the rest of the continental rise is a major problem yet to be solved. The answer could be that numerous channels too small to be detected seismically passed sediment across the rise; alternatively, it is possible that rise canyons are ephemeral features that are destroyed during burial.

Quaternary Processes

None of the several growth faults observed in Miocene and older sediments offshore New Jersey cut Quaternary reflectors. Either the deep seated faulting has stopped, or is too slow to develop offset within the transitory cover of Quaternary sediments. Obviously, faulting is no longer the important agent for canyon initiation that it may have been during the Paleogene.

Despite the large number of modern canyons cut into the North American margin, a majority are restricted to the slope. Few maintain more than a few tens of meters of relief seaward of the slope/rise contact, making it unlikely that they contribute significantly to central and lower rise sedimentation. Most of the large canyons that do cross the rise, however, also indent the shelf edge. Two of these canyons, Hudson and Wilmington, are within 100 km of the study area; during the Pleistocene, each passed large volumes of shallow-water debris to the lower rise and beyond. Because they are deeply entrenched across the upper rise (the Hudson canyon is as much as 600 m deep), it is likely that neither one contributed coarse material to the upper rise; only the finer-grained sediments are likely to have spilled out of the canyons as overbank deposits.

Consequently, it follows that canyons restricted to the slope are predominantly responsible for Quaternary deposition on the upper rise. Furthermore, consistent with the model of two-stage canyon development proposed by Farre and others (1983) and supported by this study, slumping appears to be an important process in this slope-to-rise transport.

This is shown by strike profiles along the uppermost rise and drilling at Site 613 (Fig. 6). The Pliocene/Pleistocene contact is unconformable at this drill site; several tens of meters of thoroughly homogenous, slumped "middle" Pleistocene mud rest on an abbreviated upper Pliocene section. Bedding-plane slides appear in seismic profiles to have stripped off the top of the Pliocene to various depths along strike. This diagnostic sequence boundary can be traced seaward on numerous profiles. The slumped interval (acoustically transparent to weakly chaotic along the uppermost rise) gradually becomes more recognizable; chaotic reflectors gain strength, and within 50 km of the base of the slope their basal contact decreases grade and these reflectors become more regular, parallel and level. Pleistocene debris flows terminated at this point, and their seaward correlatives were suspended fine-grained components deposited as turbidites. Several units, identical to this in seismic character, have been traced along the upper rise (Shor and others, 1986). Their lateral extent has yet to be determined; whether they are numerous and narrow, or few and widespread does not invalidate the conclusion that slumping has been an important process during the Quaternary.

CONCLUSIONS

1. A model of two-stage slope canyon formation (Farre and others, 1983) is supported by several episodes of Cenozoic erosion of the New Jersey margin. The first stage involves lower to middle slope sediment failure. Through headward erosion, these slump scars have the potential of reaching sources of cohesionless detrital sediment. If fed into upper slope scars, these detrital sediments can generate turbidity currents that mark the second, more active phase of canyon development. During this phase the slump scars are deeply incised, lengthened and cleared of the slump debris that accumulated during the earlier phase.

2. Paleogene hiatuses along the upper slope and coincident channeling along the lower slope were probably caused by collapse of the underlying Mesozoic carbonate margin. The episodic pattern of these events is poorly understood, but the involvement of sea-level oscillations cannot be discounted.

3. The absence of Paleogene channels landward of the middle slope and the scarcity of shallow-water debris within these channels implies that headward erosion did not reach far enough landward to tap a source of clastic sediment. In this manner, Paleogene slumping and channeling represents an arrested stage of slope canyon development.

4. Retreat of the margin during Oligocene time moved the slope tens of kilometers landward of its previous position. This removed the topography that had controlled the location of slope channels throughout the Paleogene. Sediments along the margin changed from dominantly silica-cemented carbonates to cohesionless muds between the Paleogene and the Neogene; this, too,

may have contributed to changes in style of margin erosion and deposition.

5. A delta spread across the mid-Atlantic shelf in the lower and middle Miocene, providing a large volume of clastic sediment to canyons already indenting the shelf. The sediment supply was so great that in some areas the shelf prograded out onto the slope. Any canyons that may have transported sediment across the slope during this outbuilding phase either were not in the area studied here, or were themselves eroded by late middle Miocene events.

6. A late middle Miocene canyon (11.7 to 9.5 MY) cut across the slope and directly beneath Site 612. No equivalent feature is recognized in seismic profiles along the upper rise, and it is suspected that few canyons crossed the entire continental rise during the Miocene.

7. Pliocene strata on the uppermost rise are alternating clays and glauconitic muds, prone to bedding-plane slides along especially porous and unstable sand-rich layers. Extensive lower slope failures during the Pleistocene frequently stripped off the top of this Pliocene unit. These events contributed lower slope and uppermost rise debris flows and laterally equivalent turbidites to the upper rise. Canyons that through headward erosion had reached clastic sediment sources on the shelf were the only ones able to keep their valleys clear and pass turbidites to the lower continental rise.

8. Channel-filling sediments along a continental margin can be fine-grained, pelagic sediment unsuitable as hydrocarbon resevoir rocks.

9. Slumping along the lower slope and within inter-canyon areas that is unrelated to changing global sea level can form seismic sequence boundaries. In dip sections these can be identical to patterns frequently interpreted as surfaces formed by slope by-pass during sea-level lowstands.

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MIOCENE SEA LEVEL EVENTS IN THE MARYLAND COASTAL PLAIN AND THE OFFSHORE BALTIMORE CANYON TROUGH

R.K. OLSSON, A.J. MELILLO, AND B.L. SCHREIBER

Department of Geological Sciences, Rutgers University, New Brunswick, New Jersey 08903

ABSTRACT

Biofacies analysis of benthic foraminifera and planktonic foraminiferal biostratigraphy in Miocene formations in the Maryland coastal plain subsurface are used to identify depositional sequences and to correlate them with the third-order cycles of Haq and others (1987). Four sequences are identified in the Calvert Formation and correlated with cycles TB 1.5, 2.1, 2.2, and 2.3. The Choptank Formation is upper Miocene (Zone N16) and correlated with cycle TB 3.1. The St. Marys Formation is placed in Zone N16 and in cycle TB 3.2. In downdip coastal wells an upper upper Miocene unit (Zone N17) suggests the presence of the Eastover Formation and cycle TB 3.3.

An unconformity and associated long hiatus which represents three missing sequences separates the Choptank Formation from the Calvert Formation. The missing sequences (TB 2.4, 2.5, and 2.6) occur offshore in the Baltimore Canyon Trough. Their absence in Maryland is probably due to erosion in the coastal plain during a low

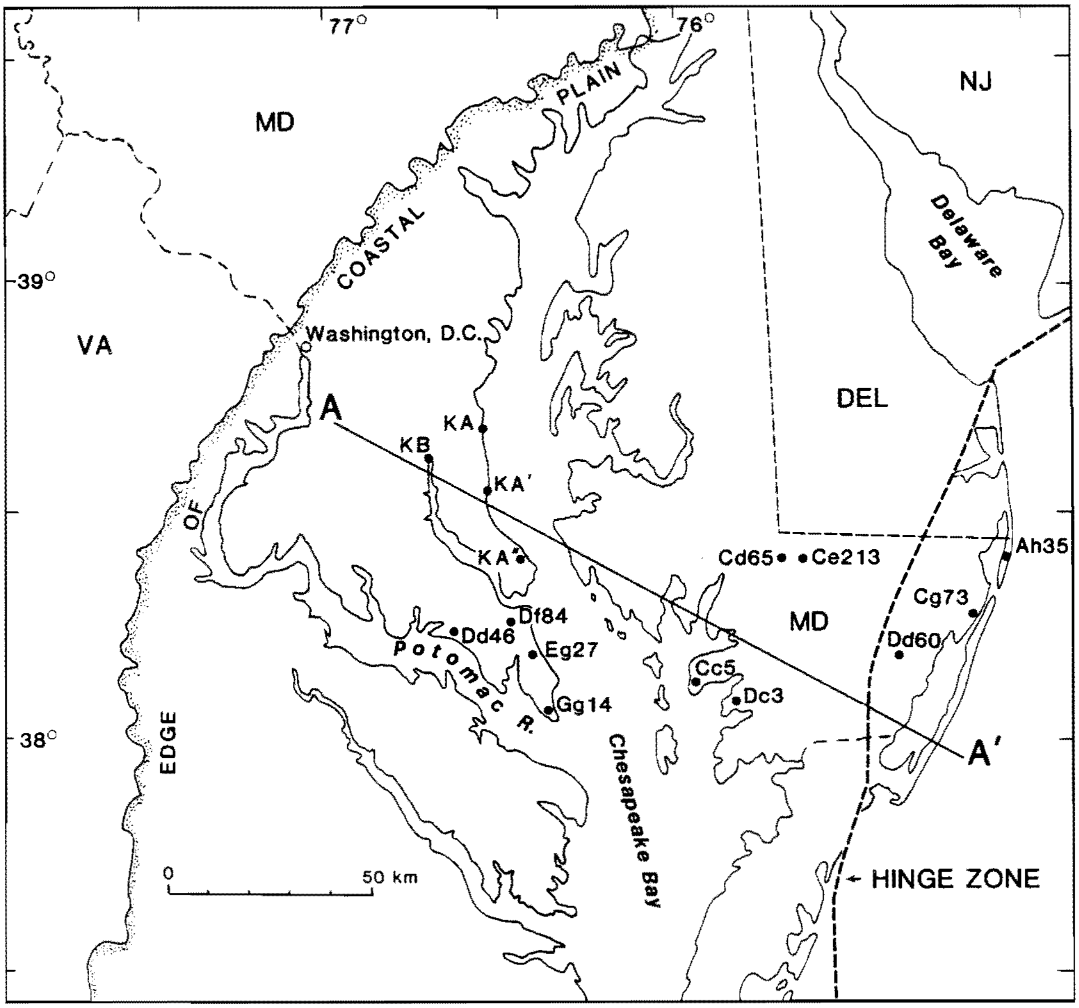

FIGURE 1. Outline map of Maryland coastal plain showing locations of wells used in this study. Locations KA, KA', KA", and KB are outcrop locations of Kidwell (1984). The cross-section shown in figure 3 is drawn along line A-A'.

stand of sea level when submarine canyons formed on the New Jersey slope at the end of the middle Miocene (10.5 Ma).

Paleoslope modeling indicates that relative sea level stood between 15 to 50 m above present sea level during deposition of transgressive deposits in the Maryland coastal plain.

INTRODUCTION

Geologist's view of stratigraphy has been significantly changed by the concept of depositional sequences brought forward by Vail and others (1977) and the publication of the EXXON global cycle chart. Revisions of the original cycle chart (Haq and others, 1987) based upon integration of outcrop stratigraphy with seismic stratigraphy have produced a sophisticated new chart showing the history and relative magnitude of numerous global sea level events. The impact of these events would appear to have a major control on deposition and erosion on continental margins. Since depositional sequences are unconformity bounded, disconformities and their associated hiatuses in continental margin stratigraphic sections provide the means for the recognition of depositional sequences. Both physical and paleontological criteria must be utilized in evaluating sequence boundaries. In this study we integrate lithology and geophysical logs with planktonic foraminiferal biostratigraphy and the distribution of benthic foraminiferal biofacies in analyzing the Miocene sequence stratigraphy of the Maryland coastal plain and the adjacent offshore Baltimore Canyon Trough An important objective of this study is to relate the sequence stratigraphy to the global cycle chart (Haq and others, 1987) and to submarine events that have been identified in Deep Sea Drilling Project (DSDP) records in order to gain an understanding of how sea level events affected the history of this margin.

The Miocene deposits of the Maryland coastal plain have received intensive study since the latter part of the nineteenth century. These studies have focused mainly on outcrop sections of the Chesapeake Group and have provided data on geologic age and environment of deposition of the Calvert, Choptank, and St. Marys Formations (Shattuck, 1904; Cooke and others, 1943; Gibson, 1962, 1982, 1983a, 1983b; Gernant, 1970; Cavallero, 1974; Abbott, 1978; Gernant and others, 1977; Kidwell, 1984). These studies recognized that these formations were deposited in generally open-marine shelf paleoenvironments. The age assignments given these Miocene formations has differed in these studies. There is also disagreement on whether these formations are separated by disconformities.

Age determinations have been based on molluscs, ostracodes, foraminifers, diatoms, and radiolarians. No standard micropaleontological biostratigraphic zonation has been applied to the Chesapeake Group although, Abbott (1978) and Andrews (1978) adapted a regional diatom zonation for correlation of the Calvert and Choptank Formations. There is general agreement that the Calvert is lower to middle Miocene (Shattuck, 1904; Cooke and others, 1943; Andrews, 1978; Gibson, 1983b) but the Choptank and St. Marys Formations have been placed alternately in the middle Miocene (Shattuck, 1904; Cooke and others, 1943; Gernant, 1970; Andrews, 1978; Gibson, 1982, 1983b), upper Miocene (Cavallero, 1974; Gibson, 1982, 1983b; this study) and in the case of the St. Marys even in the Pliocene (Cavallero, 1974).

A disconformity separating the Calvert and Choptank Formations is generally recognized (Shattuck, 1904; Gernant, 1970; Andrews, 1978, Gibson, 1982, 1983b; Kidwell, 1985; this study). Gernant (1970) first suggested that an unconformity separated the Choptank and St. Marys. Subsequent studies (Gibson, 1982, 1983b; Kidwell, 1984; this study) also place an unconformity between these formations.

We focus here on the distribution of the Calvert, Choptank, and St. Marys Formations in the Maryland coastal plain subsurface with emphasis on planktonic foraminiferal biostratigraphy and on benthic foraminiferal biofacies analysis. The objective of the study is to identify and characterize Miocene depositional sequences in the Maryland coastal plain and their extension offshore in the Baltimore Canyon Trough.

Eleven rotary wells (Fig. 1) situated in shallow updip to coastal downdip locations provided the framework for this study. In contrast to outcrop, these wells penetrated deeper-water facies which contained a greater number of planktonic foraminiferal species used for biostratigraphic age determinations. Geophysical well logs supplemented the biostratigraphic correlations and provided the control for the preparation of the downdip cross-section (Fig. 2).

Most of the samples used in this study were rotary well cuttings taken at 3.3 m (10 ft.) intervals. A small number of core samples were also used in some wells. The mixing of specimens downward is common in rotary well cuttings. Therefore, only the first downhole occurrence of a species as well as peak abundance intervals were considered in paleoenvironmental reconstructions. Biostratigraphic age determinations were based on the first downhole occurrence of diagnostic planktonic species. A previous study by Melillo (1985) in the Baltimore Canyon Trough on core samples from ASP and AMCOR wells and cuttings from the COST B3 Well provided the basis for correlation of coastal plain sequences in this offshore basin.

LITHOSTRATIGRAPHY

The Calvert Formation varies from a diatomaceous clay to a glauconitic clayey diatomaceous sand. However, there are lithologic differences within the Calvert Formation. The upper part has a higher clay content than the interval immediately below which is sandier and has a higher foraminiferal and diatom content. Below this it is more clayey with abundant microfossils. The basal part of the Calvert Formation is sandier in the Dc3 and Cc5 wells (Fig. 1). These lithologic changes are accompanied by changes in foraminiferal composition. The Calvert Formation lies disconformably on Upper Oligocene glauconitic sands. The Calvert Formation differs markedly from the overlying

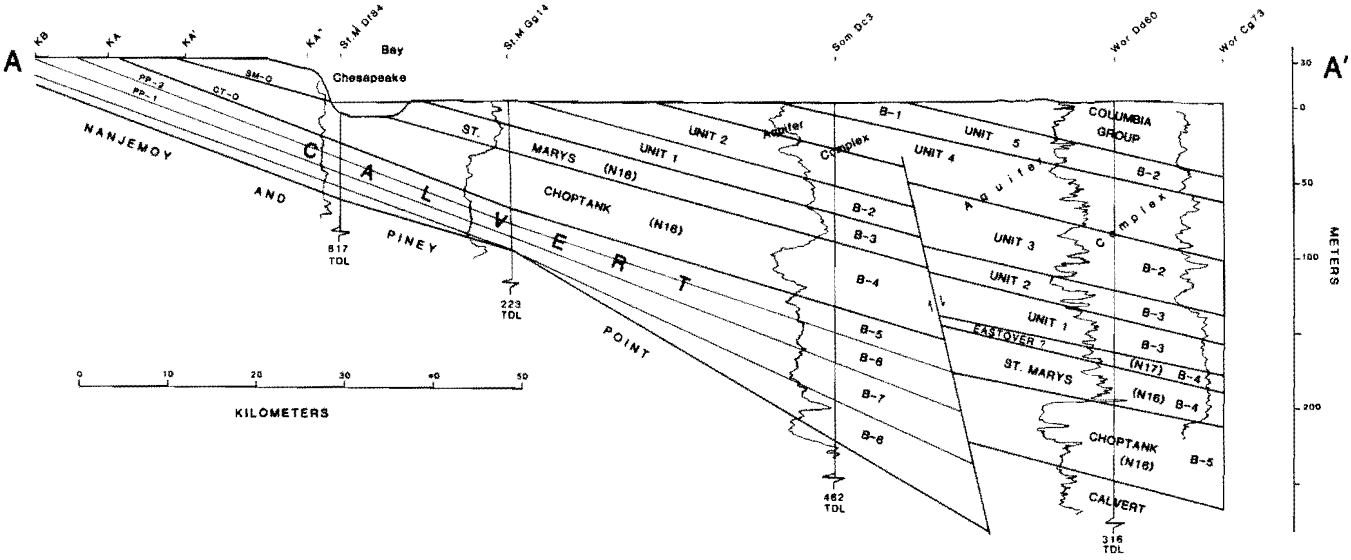

FIGURE 2. Cross-section drawn parallel to dip (line A-A' in Fig. 1) showing distribution of formations and biofacies units. Faulting is probably related to the hinge zone shown in Figure 1. Note correlation of unconformities (SM-0, CT-0, PP-2, and PP-1) identified in outcrop along in the Calvert Cliffs (Locations KA, KA' KA" in Fig. 1). B = biofacies. N = planktonic foraminiferal zone. See text for explanation of units 1-5.

Choptank Formation. The Calvert Formation is highly fossiliferous and contains abundant calcareous and siliceous microfossils which include foraminifers, ostracodes, diatoms, radiolarians, and silicoflagellates. In subsurface it is distinguished from the Choptank Formation by an increase in clay content which coincides with a sharp rightward shift on the resistivity log of coastal plain wells.

The St. Marys Formation consists of sandy clays, clayey sands and sands underlying the Yorktown-Cohansey Formations. A distinctive geophysical log signature marks the boundary which coincides with an increase and change in foraminiferal content, indicating a downsection change to a deeper inner shelf environment of deposition. The St. Marys Formation is separated from the underlying Choptank Formation in subsurface by a disconformity (Hansen, 1981). The contact between these formations is characterized in coastal plain wells by abrupt decreasing gamma ray values. The change is apparently because of less clayey sands and abundant shell material in the Choptank. The Choptank Formation is also differentiated from the St. Marys Formation on benthic foraminiferal content. Foraminifera are less abundant in the St. Marys Formation than in the Choptank Formation, and the assemblage composition differs. The Choptank Formation was deposited in somewhat deeper environments than the St. Marys Formation. Diatomaceous layers occur only in the Choptank Formation.

The Yorktown-Cohansey Formations in Maryland are a complex of alternating and integrading sands, silts, and clays that are an important aquifer system in the coastal plain. Hansen (1981) referred to this stratigraphic interval as the aquifer complex. Three aquifers are present: the Manokin, the Ocean City, and the Pocomoke (in ascending order). These sediments were deposited in inner shelf, shoreline, and bay-lagoon environments. They contain low diversity foraminiferal benthic assemblages at certain levels. Planktonic foraminifera are rare to absent with no age diagnostic species evident. Consequently, the geologic age of these formations is uncertain. The distribution of benthic foraminiferal assemblages in far downdip wells suggests that the aquifer complex is composed of four separate stratigraphic units (units 2 to 5, Fig. 2). A disconformity separates the Yorktown-Cohansey Formations from the Columbia Group above (Hansen, 1981).

BIOSTRATIGRAPHY

Biofacies analysis can be used to divide the Calvert Formation into four intervals on the basis of benthic foraminiferal assemblages. Each of the intervals has a different paleobathymetry history. In addition, planktonic foraminifera suggest each is a different geologic age as well. The four intervals and their biofacies are well represented in the Dc3 Well where the Calvert rests on the upper Oligocene and is overlain in turn by the Choptank Formation (Fig. 3) Biostratigraphic data from this well and from the Gg14 and Dd46 wells (Figs. 1,2) provide the most useful information so far on the distribution of planktonic foraminifera in the Calvert Formation.

Previously, Gibson (1967) reported Praeorbulina glomerosa glomerosa (Blow) from the upper part of the Calvert Formation in outcrops at Plum

Point, Maryland. He placed the upper Calvert in the Globigerinatella insueta Zone (= Zone N7) and correlated it with the Pungo River Formation in eastern North Carolina. Walters and Snyder (1986) also identified the G. insueta Zone in vibracores in the uppermost part of the Pungo River Formation in southern Onslow Bay, North Carolina. Praeorbulina has not been observed in the Pungo River Formation.

We have also identified the Globigerinatella insueta Zone (= Zone N7) in the upper Calvert Formation but not in the uppermost strata. In a core taken at 138 feet at the Dd46 Well (Figs. 1,2) we observe a typical G. insueta zonal assemblage which includes G. insueta Cushman and Stainforth, Globigerinoides bisphericus Todd, Globorotalia fohsi peripheroronda Blow and Banner, and Globorotalia mayeri Cushman and Ellisor among a diverse assemblage. In wells Gg14 and Dc3 Praeorbulina glomerosa curva (Blow) and P. glomerosa glomerosa occur in the uppermost part of the Calvert Formation. The co-occurrence of these subspecies places this part of the Calvert in the Praeorbulina glomerosa Zone (= Zones N8 to lower N9). The sample from 138 feet at the Dd46 well is stratigraphically below the level at Plum Point where Gibson (1967) found P. glomerosa glomerosa. We correlate the upper Calvert Formation at Plum Point with the P. glomerosa Zone (= Zones N8 to lower N9) since P. glomerosa glomerosa does not range below this zone (Bolli and Saunders, 1985).

At the Dc3 well the P. glomerosa Zone (= Zones N8 to lower N9) lies within the interval which contains benthic biofacies 5 (Fig. 3). The G. insueta Zone (= Zone N7) lies within the interval which contains biofacies 6 (Fig. 3). Planktonic foraminifera are sparse in the biofacies 7 interval so that it is unclear whether an older zone is present. Globorotalia scitula praescitula Blow is fairly common in this and higher intervals. In low latitude sections this species first occurs at the base of N5 (Bolli and Saunders, 1985), although Poore (1978) indicates that its first occurrence is close to the base of N7 in the North Atlantic. However, at DSDP Site 563 in the middle North Atlantic G. scitula praescitula first occurs in the Catapsydrax dissimilis Zone (= N5) (Melillo, 1985). Thus biofacies 7 probably lies within part of N5 to N7 (Fig. 3).

The lowermost part of the Calvert Formation in the subsurface contains a different planktonic foraminiferal assemblage than its upper parts. In contrast to the assemblages in the upper parts which can be correlated with low latitude tropical-subtropical biostratigraphy, the lowermost part contains cooler water temperature elements which correlate with the middle latitude biostratigraphy of Kennett (1973). The key species in this lowermost interval are Globorotalia zealandica Walters and Globorotalia incognita Walters. These species play a key role in the G. zealandica - Globorotalia miozea evolutionary

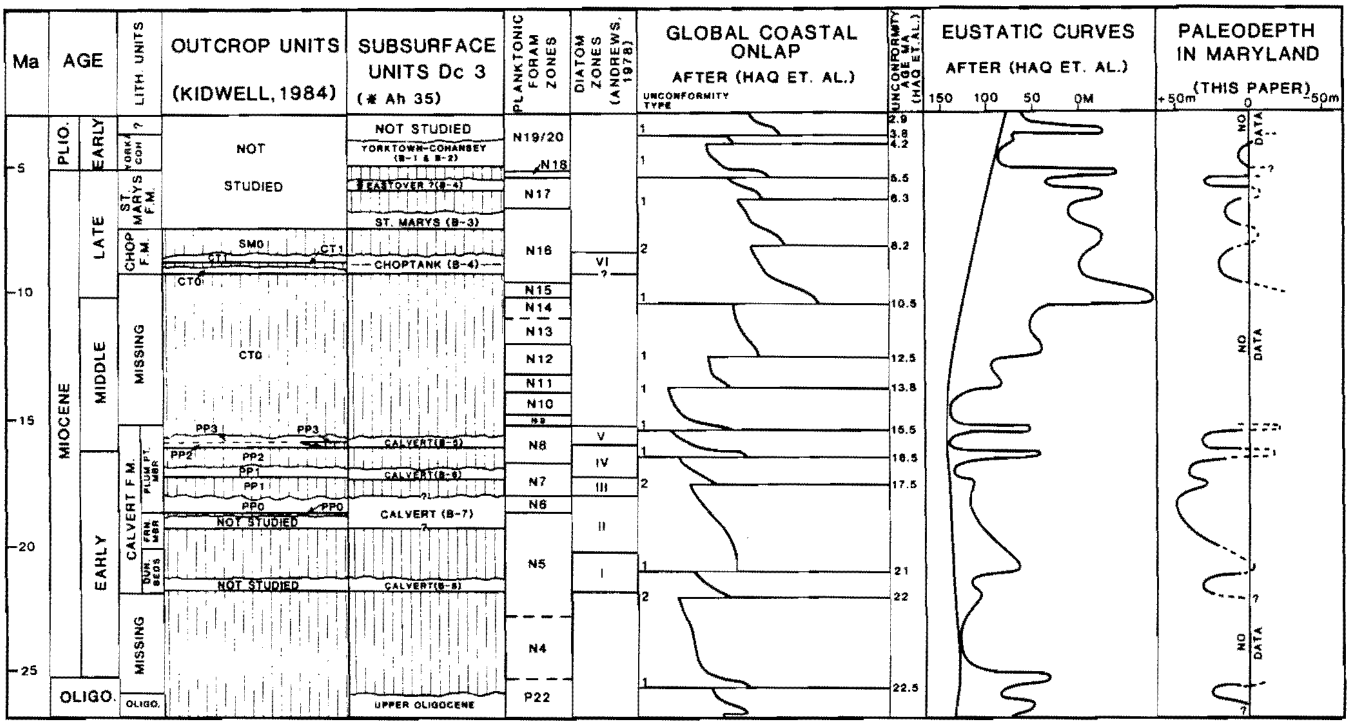

FIGURE 3. Correlation of formations in Maryland coastal plain with the third order cycles of Haq and others (1987). Correlation of outcrop units and hiatuses identified by Kidwell (1984) is also shown. Note comparison of eustatic curve of Haq and others with curve derived from paleoslope modeling in Maryland. Time scale from Haq and others.

lineage in early Miocene mid-latitude regions (Walters, 1965; Srinivasan and Kennett, 1981; Kennett and Srinivasan, 1983). This bioseries was placed in the subgenus *Globoconella* lineage by Srinivasan and Kennett (1981). The evolution of *G. incognita* to *G. miozea* through *G. zealandica* and *G. scitula praescitula* is considered by them to represent a continuum. *Globorotalia miozea* is not present in the Calvert Formation. This indicates that its evolution had not yet occurred and, thus, this provides a useful level for correlation. The co-occurrence of *G. incognita* and *G. zealandica*, and the absence of *G. miozea*, places the lowermost Calvert Formation in the middle latitude *Catapsydrax dissimilis* Zone of Srinivasan and Kennett (1981). They correlate their *C. dissimilis* Zone with the low latitude N5-N6 zones (*C. dissimilis* and *C. stainforthi* zones, respectively) and the *G. miozea* Zone with the *G. insueta* Zone (N7). The absence of *G. scitula praescitula* suggests that this stratigraphic level is below the evolutionary transition of this species from *G. zealandica*. This would correlate the lowermost Calvert with the lower half of the middle latitude *C. dissimilis* Zone and Zone N5. The lowermost Calvert is also differentiated from the upper parts of the formation because it contains benthic biofacies 8.

The Calvert in the subsurface of Maryland rests upon strata which contain the *Globigerina ciperoensis* Zone (P22) according to Olsson and others (1980). The *Globorotalia kugleri* Zone (N4) is apparently missing in the Maryland coastal plain.

The planktonic foraminifera which occur in the Choptank Formation and undip St. Marys Formation are similar although greater numbers are present in the Choptank Formation. *Neogloboquadrina acostaensis* (Blow), *Neogloboquadrina continuosa* (Blow), *Neogloboquadrina pseudopachyderma* (Cita, Premoli-Silva and Rossi), and *Globigerina apertura* Cushman are found in both formations. This association places them in the *G. acostaensis* Zone (N16) of Tortonian age. In the farthest downdip Ah35 well (Figs. 1,2) a core sample in the uppermost part of the St. Marys Formation contains *Sphaeroidinellopsis disjuncta*, Blow, *Neogloboquadrina humerosa praehumerosa* (Natori), and *Globorotalia pleisiotumida* Blow and Banner. Also, all specimens of *N. acostaensis* are dextrally coiled whereas all specimens of this species are sinistrally coiled in the St. Marys in the updip wells. This indicates that the uppermost part of the St. Marys in coastal downdip wells contain a higher stratigraphic interval. This interval can be placed in Zone N17 of Messinian age on the basis of the overlapping ranges of *G. plesiotumida* and *S. disjuncta* and the occurrence of *N. humerosa praehumerosa* (Bolli and Saunders, 1985). Further documentation of this zonal placement is indicated by the shift from sinistral to dextral cooling in *N. acostaensis* which Stainforth and others (1975) show to take place in the uppermost part (Messinian) of the upper Miocene. This unit possibly correlates with the Eastover Formation of Virginia.

The identification of Zone N17 in the upper St. Marys suggests that the poorly constrained Yorktown-Cohansey strata in Maryland are most probably Pliocene in age. These strata were deposited in inner shelf environments and lack definitive species of planktonic foraminifera. They contain biofacies 1, 2, and 3. To the south in Virginia the lower beds of the Yorktown are placed in Pliocene Zones N18 or N19 whereas the upper beds are placed in Zones N19 or N20 (Gibson 1983; Synder and others, 1983). The distribution of biofacies in the Yorktown-Cohansey Formations (Figs. 3, 4) suggests that five separate intervals are present in the subsurface of Maryland.

DISTRIBUTION OF BIOFACIES IN MARYLAND WELLS

AGE	FORMATION	UNITS	Dd46	Eg27	Gg14	Ce213	Dc3	Dd60	Cg73
PLIOCENE?	YORKTOWN-COHANSEY AQUIFER COMPLEX	1	X	X	A	1	A	2	2
		2	X	X	A	A	A	A	A
		3	X	X	X	X	A	2	2
		4	X	X	A	A	A	3	3
		5	X	X	A	2	2	2	2
LATE MIOCENE	EASTOVER?		X	X	X	X	X	3	4
	St. MARYS		X	X	3	3	3	4	4
EARLY-MID MIOCENE	CHOPTANK		X	X	4	4	4	5	5
	CALVERT		X	X	4/5	5	5	X	X
			5	5	6	X	6	X	X
			X	X	7	X	7	X	X
			X	X	X	X	8	X	X

FIGURE 4. Distribution of different types of biofacies in Maryland coastal plain wells. "__" indicates absence of foraminifera. "X" indicates section is absent or not sampled.

The Choptank Formation is separated from the Calvert Formation by a large hiatus. Zones N9 to N15 are missing in the Maryland coastal plain. Middle Miocene zones N9, N10, N11-12 have been identified in the offshore Baltimore Canyon Trough by Melillo (1985). This suggests that one or more significant erosional events effected the Maryland coastal plain during late Middle Miocene time.

BIOFACIES ANALYSIS

Biofacies analysis follows the paleoslope modeling technique established by Olsson and Nyong (1984). This includes the recognition of benthic foraminiferal biofacies by cluster analysis and factor analysis and then applying the graphic paleoslope analysis (Olsson and Nyong, 1984). Biofacies analysis of the Yorktown/Cohansey, St. Marys, Choptank and Calvert Formations identifies eight biofacies (Table 1).

Biofacies 1 consists almost exclusively of two species, *Ammonia beccarii* (Linne) and *Elphidium articulatum* (d'Orbigny). These species are

BIOFACIES 1	*Ammonia beccarii* *Elphidium articulatum*	BIOFACIES 5	*Valvulineria floridana* *Textularia agglutinans* *Florilus pizarrensis* *Spiroplectammina gracilis*
BIOFACIES 2	*Elphidium gunteri* *Buliminella elegantissima*	BIOFACIES 6	*Uvigerina peregrina* *Gyroidinoides regularis* *Sigmoilina tenuis* *Siphogenerina spinosa*
BIOFACIES 3	*Bolivina directa* *Nonionella auris* *Buccella anderseni* *Florilus incisum*	BIOFACIES 7	*Buliminella exilis* *Siphogenerina spinosa* *Stimostomella bradyi* *Bolivina fragilis*
BIOFACIES 4	*Hanzawaia concentrica* *Spiroplectammina gracilis* *Bolivina plicatella* *Bolivina multicostata*	BIOFACIES 8	*Uvigerina flinti* *Gyroidina parva* *Cassidulina subglobosa*

TABLE 1. Benthic foraminiferal species that are characteristic of each biofacies.

moderately euryhaline forms occurring in modern day inner shelf environments, especially in lagoons and estuaries (Murray, 1973). This biofacies is found in the upper part of the Yorktown-Cohansey in the subsurface. Sands and silts in the equivalent updip interval are barren of foraminifera. Thus this biofacies appears to occupy a position immediately adjacent to the shoreline during deposition of the Yorktown-Cohansey.

Biofacies 2 is characterized by *Buliminella elegantissima* (d'Orbigny) and *Elphidium gunteri* (Cole) and occurs in the Yorktown-Cohansey Formations (Fig. 4). In the uppermost part of the Yorktown-Cohansey biofacies 2 occurs just downdip of biofacies 1. Biofacies 2 is distributed for the most part in the lowermost part of Yorktown-Cohansey in subsurface except in the farthest downdip wells where it occurs in the Manokin aquifer (unit 3, Fig. 2).

Biofacies 3 occurs in the St. Marys where it is most common (Fig. 4). This biofacies is also found in the lower part of the Manokin aquifer in wells Dd60 and Cg73 (unit 2, Fig. 2). Biofacies 3 contains a more diverse assemblage than biofacies 2. Common species Biofacies 3 include *Bolivina directa* Cushman, *Nonionella auris* (d'Orbigny), *Fursenkoina fusiformis* (Cushman), *Buccella anderseni* McLean, and *Florilus incisum* (Cushman).

Biofacies 4 appears downdip of biofacies 3 (Fig. 4) which is characterized by *Hanzawaia concentrica* (Cushman), *Spiroplectammina gracilis* (von Muenster), *Bolivina plicatella* Cushman, *Textularia agglutinans* d'Orbigny, and *Bolivina multicostata* Cushman. In the upper portion of the Choptank Formation, *Cibicides lobatulus* (Jacobs and Walker) and *Valvulineria floridana* Cushman are more common than *Bolivina plicatella* and *B. multicostata*. This difference in biofacies 4 is not significant enough to warrant the recognition of two distinct biofacies in the Choptank Formation. It appears to be more related to some relatively minor change in paleoenvironmental or sedimentological conditions. It may be that the two sequences identified in outcrop exposures of the Choptank Formation by Kidwell (1984) correspond, respectively, to the lower and upper divisions of biofacies 4 in subsurface sections of this formation. In the Choptank Formation biofacies 4 is replaced downdip by biofacies 5. The shallowing upsection paleobathymetric trend exhibited by the biofacies in the Yorktown-Cohansey Formations is also evident in the St. Marys and Choptank Formations. Each of the biofacies from 1 to 5 are linked to one another by replacement along dip and also vertically in stratigraphic section.

Biofacies 5 occurs in the updip part of the middle Calvert and in the uppermost part of the downdip Calvert Formation. Diagnostic species in this biofacies include *Florilus pizarrensis* (Berry), *Uvigerina peregrina* Cushman, *Valvulineria floridana* Cushman and *Bolivina paula* Cushman and Cahill.

Biofacies analysis (biofacies 5 to 8) in the Calvert Formation is restricted to three updip wells. Wells drilled farther downdip did not penetrate the Calvert even though they were drilled to 305 m (1000 ft.). The greater depth to the Calvert is apparently due to faulting of the basement (Fig. 2). Expanded sections of the Choptank Formation in coastal wells suggests growth faulting. As a result, analysis of biofacies along dip in the Calvert is more limited than in the overlying units. Nevertheless, a distinct vertical distribution of biofacies is observed.

Biofacies 6 is dominated by *Uvigerina peregrina* and *Gyroidinoides regularis* (Phleger and Parker). *Buliminella tenuata exilis* Cushman and *Siphogenerina spinosa* Bagg characterize Biofacies 7. In the lowermost Calvert biofacies 8 is distinguished by *Uvigerina flintii* Cushman, *Gyroidina parva* Cushman and Renz, and *Cassidulina subglobosa* Brady. This biofacies was found only in wells Cc5 and Dc3.

Biofacies 5 to 7 are linked to one another in juxtaposition and indicate a deepening trend in the Calvert. They are also linked to biofacies 1 to 4 (Fig. 4). Biofacies 8 is somewhat distinct from the others and appears to be indicative of shallower paleobathymetry than biofacies 7.

PALEOSLOPE MODELING

The distribution of biofacies in the Maryland subsurface follows a pattern which is useful for paleoslope modeling (Olsson and others, 1983; Olsson and Nyong, 1984). This distribution (Figs. 4 and 5) exhibits the trends, both downdip and in vertical section, that allow the application of Walthers' Law of sedimentary facies to the biofacies. Using the paleoslope method (Olsson and Nyong, 1984) the Miocene biofacies are integrated into a single profile which approximates the paleobathymetric distribution of Miocene benthic foraminiferal species. The paleoslope model thus constructed (Fig. 6), is the basis for estimating the magnitude of sea level rise for each sequences in the Maryland Miocene section (Fig. 3).

We believe that such paleoslope models provide a more reliable constraint on the determination of paleobathymetries in a stratigraphic section than methods that compare a sample directly with recent foraminiferal distributions. Paleoslope modeling indicates that the maximum rise in sea level in Maryland during Miocene time was about 50 m above present sea level (Fig. 3). Sea level rises above present sea level varied between 15 m and 50 m during the Miocene. Our estimates of the magnitude of sea level rise are about one-third lower than the values shown on the cycle curve of Haq and others (1987). The geologic history of the coastal plain and adjacent offshore is one of gradual subsidence of a passive margin. Evidence of regional tectonics leading to significant uplift is lacking. Consequently,

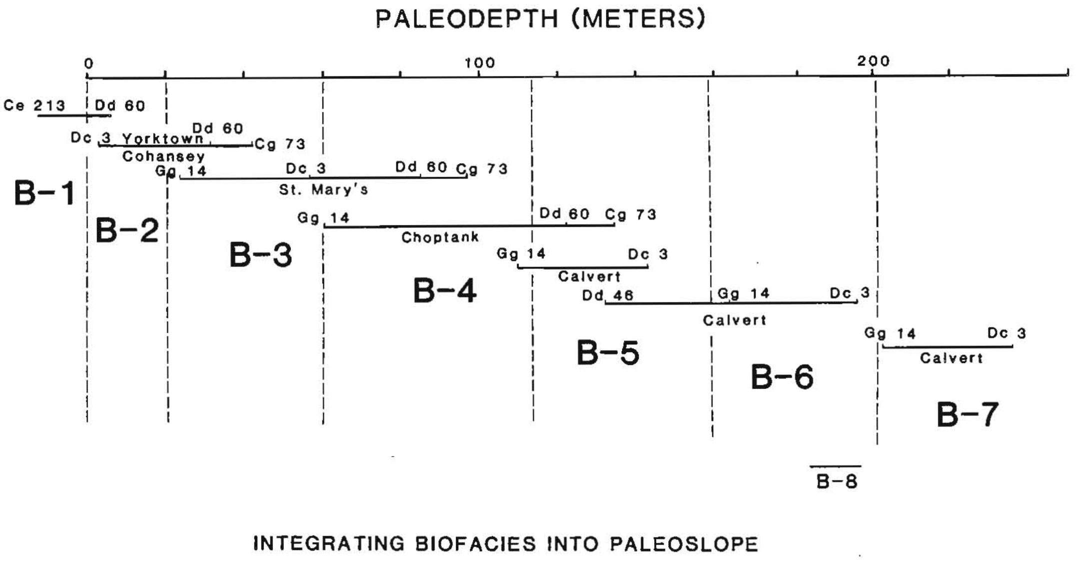

FIGURE 5. Integration of biofacies into paleoslope. The distance between wells with each biofacies is expressed as depth using a gradient of approximately 1 m per kilometer.

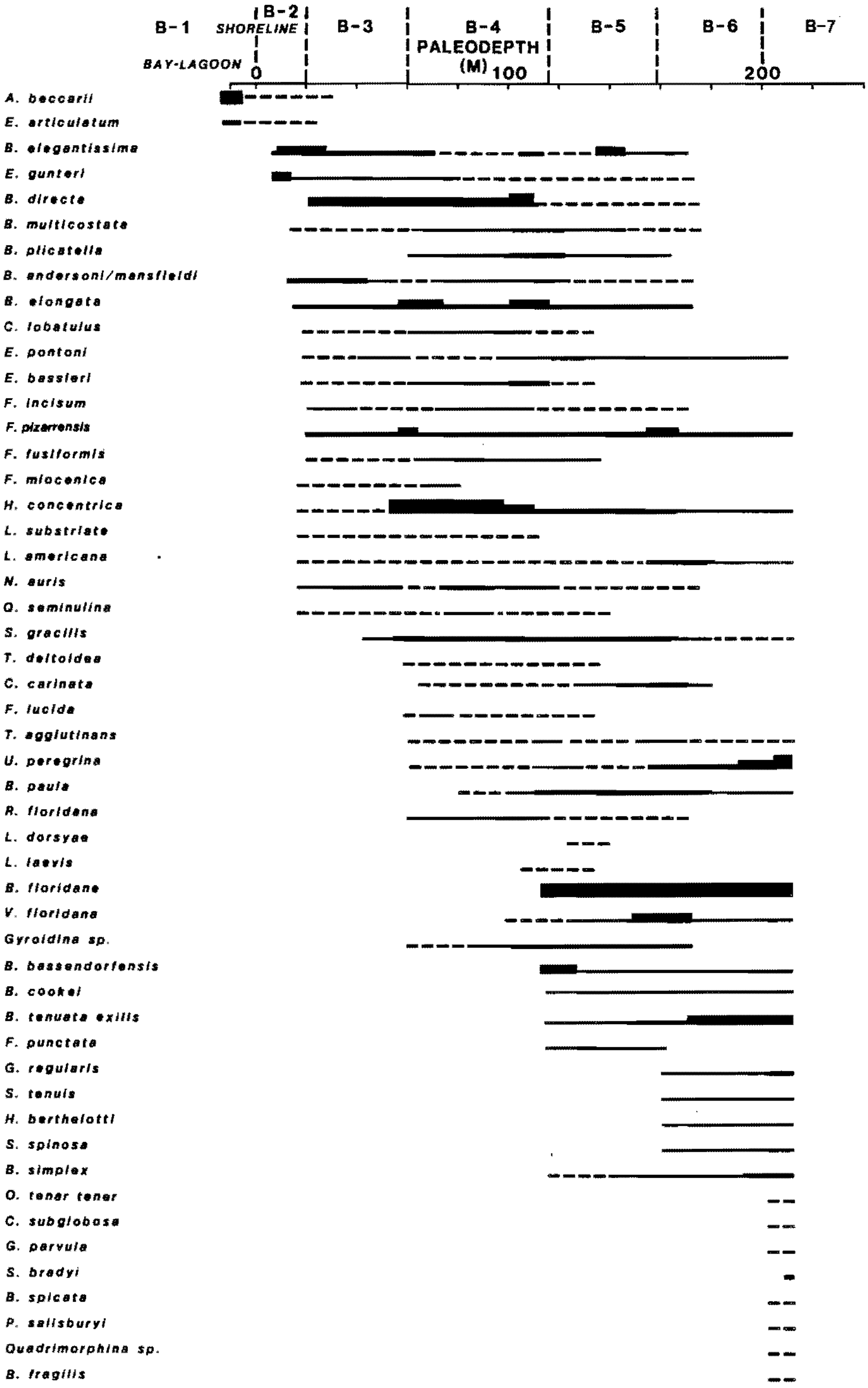

FIGURE 6. Paleoslope model derived from Figure 5 showing paleodepth distribution of most common species. Paleodepth calculations shown in Figures 3 and 7 are derived from this model.

we believe that our estimates of paleobathymetries for Miocene depositional sequences are reasonably accurate values of sea level rise events on the western Atlantic margin. On the other hand, it is not possible to estimate the magnitude of a sea level fall in the coastal plain because this part of the record is missing due to hiatuses.

DISCONFORMITIES AND DEPOSITIONAL SEQUENCES IN THE MARYLAND COASTAL PLAIN

Numerous disconformities are now recognized in outcrop of the Miocene strata in Maryland. Disconformities separate the Calvert, Choptank, St. Marys and Yorktown-Cohansey Formations (Shattuck, 1904; Gernant, 1970; Andrews, 1978; Gibson, 1982; Kidwell, 1984) in outcrop. In addition, Kidwell has recognized four disconformities within the Calvert Formation and one within the Choptank Formation (Fig. 3). In the subsurface hiatuses are also recognized between the time of deposition of the St. Marys, Choptank, and Calvert Formations (Hansen, 1981). Also, hiatuses are present between the time of deposition of the Yorktown-Cohansey (aquifer complex) Formations and the Columbia Group (Hansen, 1981) and between the Calvert Formation and the Upper Oligocene. Weathering at the top of the St. Marys Formation noted in wells in this study suggests the presence of a disconformity. Furthermore, a Tortonian equivalent for the St. Marys in updip wells and the identification of a Messinian age in the uppermost part of the St. Marys (Eastover Formation?) in coastal wells supports the presence of a hiatus in updip wells between the St. Marys and the Yorktown-Cohansey.

Our studies suggest that the distribution of biofacies in the Maryland coastal plain subsurface (Fig. 4) may indicate that hiatuses are present in the Calvert Formation because of the abrupt change between biofacies and the differences in paleobathymetry histories. In addition, in the Calvert Formation the change in biofacies is accompanied by a shift in the E-log signatures which suggests a boundary surface separates each biofacies. The planktonic foraminiferal biostratigraphy is consistent with this interpretation. In addition to the hiatuses which we place within the time of deposition of the Calvert Formation, we also interpret the Messinian equivalent of the uppermost part of the St. Marys in coastal wells as indicative of a separate unit (Eastover?) separated by a disconformity from the older St. Marys Formation below and the Yorktown-Cohansey Formations above (Fig. 3).

An extensive study of outcrops of the Calvert and Choptank Formations in Maryland and Virginia by Kidwell (1984) revealed the presence of burrowed disconformable contacts within and between these formations. She divided the Calvert into four depositional sequences, each of which is separated by a disconformity (Fig. 3). Similarly she divided the Choptank Formation into two sequences. Using the diatom biostratigraphy established by Abbott (1978, 1982), Kidwell correlated the outcrop sequences with the seismic sequence stratigraphy of Vail and others (1977).

The construction of a cross-section parallel to dip using key wells and data from Kidwell (1984) has enabled us to correlate the biofacies intervals in the St. Marys, Choptank, and Calvert Formations with the hiatuses recognized in outcrops by Kidwell (Figs. 2, 3). These correlations show that the major hiatuses established by Kidwell are traced to the boundaries between biofacies. Within the Calvert, the PP-1 hiatus correlates with the biofacies 6 and 7 boundary, and the PP-2 hiatus correlates with the biofacies 5 and 6 boundary. We correlate the disconformity placed within the Fairhaven Member of the Calvert (ie. between the Dunkirk beds and the Fairhaven beds) by Abbott (1978, 1982) and Andrews (1978) with the biofacies 7 and 8 boundary. In addition, the CT-0 disconformity and the SM-0 disconformity correspond, respectively, with the Choptank-Calvert and the St. Marys-Choptank boundaries in the subsurface (Figs. 2, 3).

We correlate the Miocene sequences we have identified using biostratigraphy, paleoecology, lithostratigraphy, and E-logs with the cycle chart of Haq and others (1987) (Fig. 3). In the right-hand column of Figure 3 is our estimate of the paleobathymetry of each cycle based on paleoslope modeling. Our analysis of paleobathymetries indicates that maximum rise of sea level during the Miocene was about 50 m above present sea level. Sea level elevations fluctuated from about 15 m to 50 m (Fig. 3). Lowstands of sea level can not be measured in the coastal plain.

Our identification of depositional sequences compares favorably with the Miocene sequence stratigraphy interpretations of Greenlee and Vail (1987) in the Baltimore Canyon Trough. They identify four sequences in the early Miocene. The upper three sequences correlate chronostratigraphically with the lower three sequences of the Calvert Formation. The lowermost sequence identified in the Baltimore Canyon is represented by a hiatus in the Maryland coastal plain. Of the four sequences identified in the middle Miocene in the Baltimore Canyon only the lowermost is present in Maryland in the uppermost sequence of the Calvert. The upper three are absent in Maryland. All three sequences of the late Miocene are identified in the Baltimore Canyon and the Maryland coastal plain. We correlate the early Pliocene sequence identified by Greenlee and Vail with the Yorktown-Cohansey.

Biostratigraphic documentation of the above sequences in the Baltimore Canyon is scarce but, nevertheless, Melillo (1985) has identified lower, middle, and upper Miocene zones in AMCOR, ASP, and COST B3 wells. Miocene biostratigraphy in the Baltimore Canyon like in the Maryland coastal plain shows the influence of middle and low latitude waters. In the lower Miocene Melillo has identified the Globorotalia kugleri (N4) Zone which occurs in the lowermost Miocene TB1.4 depositional sequence (Haq and others, 1987). As noted above, this sequence is missing in the Maryland coastal plain. Other zones identified from oldest to youngest are the middle latitude Catapsydrax dissimilis Zone, with Globorotalia zealandica, (= the low latitude C. dissimilis and C. stainforthi Zones) and the middle latitude Globorotalia miozea Zone (= the low latitude Globigerinatella insueta Zone). The oldest of these zones correlates with the lowermost Calvert interval (biofacies B-8) which also contains the middle latitude C. dissimilis Zone (with G. zealandica). The Calvert interval with biofacies B-7 correlates with the C. dissimilis and C. stainforthi low latitude zones and the Calvert interval with biofacies B-6 correlates with the G. insueta and G. miozea zones. Thus the lower Miocene sequence interpretation of Greenlee and Vail is supported by Melillo's biostratigraphic analysis in Baltimore Canyon and our analysis of the Calvert Formation.

Melillo (1985) has noted the presence of the Globorotalia fohsi peripheroronda, G. fohsi, fohsi, and G. fohsi robusta zones in the middle Miocene. These zones occur in the middle two sequences of the middle Miocene and support the interpretation of Greenlee and Vail (1987). The Praeorbulina glomerosa and Globorotalia siakensis zones were not noted by Melillo. In Maryland the uppermost Calvert interval (B-5) contains the P. glomerosa Zone. Three of the four sequences identified by Greenlee and Vail are also identified by our analysis and by Melillo.

In the Baltimore Canyon wells Melillo (1985) has identified the upper Miocene middle latitude Neogloboquadrina continuosa and Globorotalia conomiozea Zones. We correlate the Choptank and the St. Marys Formations with the N. continuosa Zone which is broadly equivalent to the low latitude Globorotalia acostaensis Zone. The uppermost St. Marys (Eastover?) in coastal wells in Maryland we correlate with the G. conomiozea Zone. Our correlations agree with the upper Miocene sequences identified by Greenlee and Vail (1987) in the Baltimore Canyon.

Within the Yorktown-Cohansey Formations or aquifer complex (Hansen, 1981) in the far downdip subsurface, the distribution of aquifers and biofacies suggests that several depositional sequences are represented. Five informal units are recognized (Fig. 2). Unit 1 is the lower confining bed of the Manokin aquifer. Although Hansen included this unit in the St. Marys Formation, there is a distinct difference in biofacies from the St. Marys Formation. We place this unit in the Yorktown-Cohansey. The fact that unit 1 rests disconformably, respectively, on Zones N16 updip and N17 downdip shows that it is unconformably separated from the St. Marys Formation. It is most probably Pliocene in age, but this is yet to be demonstrated using biostratigraphy.

Units 2 and 3 are contained within the Manokin aquifer. They differ in that unit 3 contains the shallow inner shelf biofacies 2 which is dominated by the species Elphidium gunteri and Buliminella elegantissima in contrast to deeper water biofacies 3 in unit 2. Foraminifera are absent in unit 4 which corresponds to the Ocean City aquifer. Unit 4 consists of sand with silty clay layers and carbonaceous material. It most probably represents a beach complex.

Unit 5, the Pocomoke aquifer, contains biofacies 1 in the updip position and biofacies 2 downdip. A confining bed separates the Pocomoke from the Columbia Group (Hansen, 1981).

The distribution of these biofacies and aquifers suggest that several depositional sequences are present. Without biostratigraphic constraint they can not be compared with the cycle chart of Haq and others (1987).

THE MIDDLE MIOCENE HIATUS

The middle Miocene hiatus (Fig. 3) is the largest hiatus in the Maryland coastal plain Miocene section, representing a geologic time gap of approximately 5-6 Ma. In contrast, one sequence is missing in the lower Miocene and none are missing in the Upper Miocene. An event or series of events must be invoked to explain the large middle Miocene hiatus. We suggest that a marked lowering of sea level at the end of the middle Miocene as indicated on the cycle chart of Haq and others (1987) was the event that led to erosion of most of the middle Miocene sediments in the Maryland coastal plain. Miller and others (1987) have correlated a buried submarine canyon identified at DSDP Site 612 on the New Jersey slope with an uppermost middle Miocene hiatus in the slope and elsewhere in DSDP Site 548 on the Irish margin. They suggest that this submarine erosion event corresponds to a glacio-eustatic lowstand indicated by oxygen isotope data.

As shown on the cycle chart (Haq and others, 1987) this Miocene lowstand is marked by a second-order cycle boundary. The middle Miocene lowstand may have been of sufficient magnitude to expose and erode away middle Miocene sediments from the coastal plain. Only further out in the Baltimore Canyon were these sediments preserved (Melillo, 1985; Greenlee and Vail, 1987).

CONCLUSIONS

Four depositional sequences are recognized in the Calvert Formation in the subsurface of the Maryland coastal plain (Fig. 7). Each of these sequences are recognized on the basis of biostratigraphy and benthic foraminiferal biofacies composition which indicates that the sequences were deposited under different paleobathymetries (Fig. 7). Construction of a cross section parallel to dip shows that the sequences recognized in outcrop by Kidwell (1984) can be correlated with the subsurface sequences (Fig. 2). Kidwell's study pointed out a number of stratigraphic discontinuities within the Calvert. She interpreted them mostly as surfaces which were related to transgression and correlated three of the surfaces with interregional unconformities shown on the cycle chart of Vail and Hardenbol (1979). These correlations need reexamination in view of the revised cycle chart (Haq and others, 1987) and the biostratigraphic data generated in this study. We correlate (Fig. 3) the Calvert sequences with the third order cycles TB1.5, 2.1, 2.2, and 2.3, respectively and the hiatuses that separate the sequences with unconformity ages of 21 Ma, 17.5 Ma, 16.5 Ma, and 15.5 Ma, respectively. We also interpret the Calvert sequences as transgressive deposits in view of the uniformity of foraminiferal distribution in each biofacies and uniform lithology. Regressive or highstand deposits may be present in outcrops as noted by Kidwell (1984).

The Choptank and St. Marys formations are

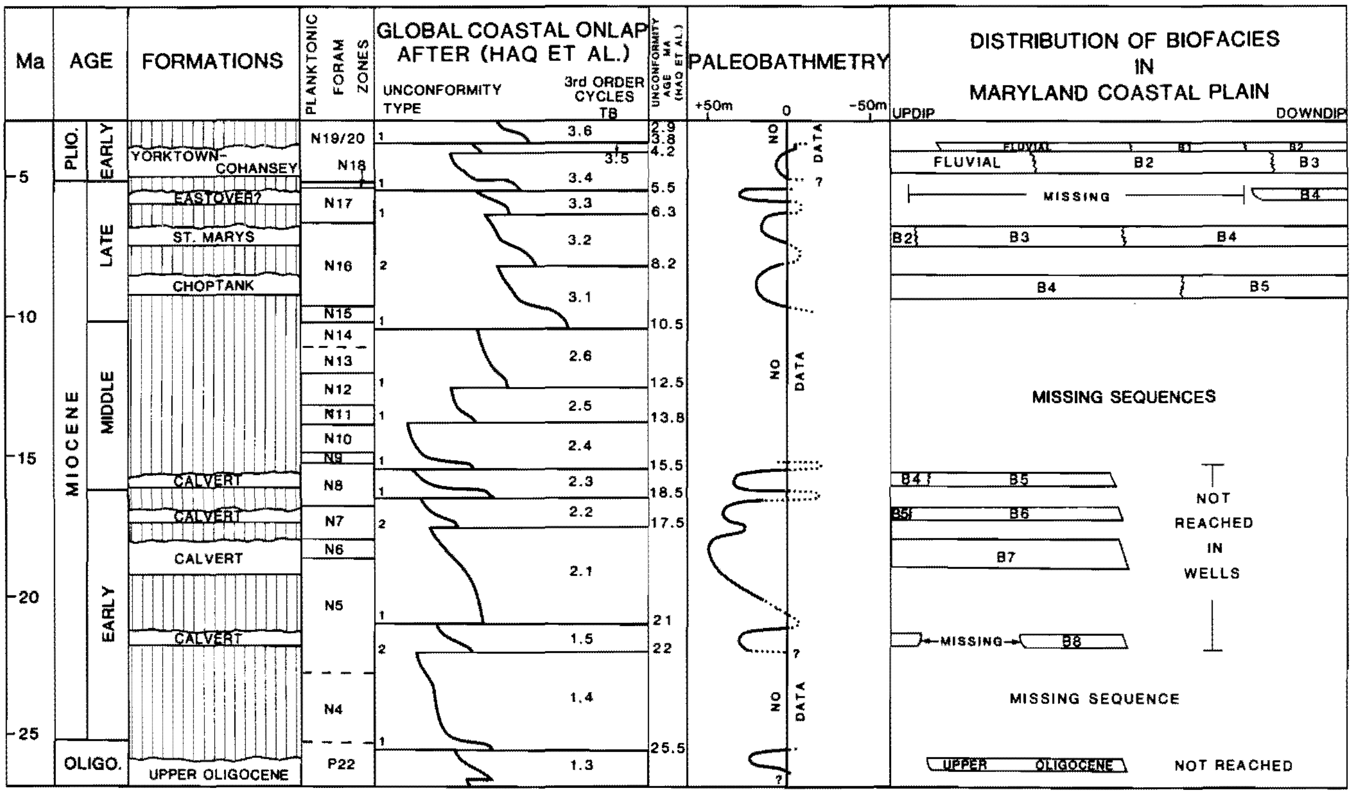

FIGURE 7. Distribution of biofacies in the Maryland coastal plain and their correlation with the third order cycles of Haq and others (1987).

placed in the upper Miocene. The Choptank and lower St. Marys are placed in Zone N16 and the upper (downdip) St. Marys (Eastover?) in Zone N17. The Choptank is correlated with the third order cycle TB 3.1 and the St. Marys is correlated with cycles TB 3.2 and 3.3, respectively. A large hiatus of approximately 5.5 Ma separates the Choptank Formation from the Calvert Formation. This hiatus was caused by erosion of the coastal plain during the canyon cutting event at 10.5 Ma (Miller and others, 1987). Kidwell notes topographic relief in outcrop of as much as 14 m on the unconformable surface. The hiatuses separating the Choptank and St. Marys sequences are correlated with unconformity ages of 8.2 Ma, 6.3 Ma, and 5.5 Ma, respectively.

Paleoslope modeling of the Miocene sequences estimates sea level elevations of transgressive deposits to range from 15 m to 50 m above present sea level (Fig. 3, 7).

The overlying Yorktown/Cohansey (Pliocene?) section in the coastal subsurface contains five units which are separated by biofacies differences and aquifer characteristics. Lack of biostratigraphic data prevents correlation of these units with the cycle chart of Haq and others (1987).

This study shows that biofacies analysis is a useful tool for recognition of depositional sequences in the Miocene, and Pliocene? as well, where eustatic changes occurred on a frequency as high as 1 Ma.

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UPPER PALEOCENE TO MIDDLE EOCENE DEPOSITIONAL SEQUENCES AND HIATUSES IN THE NEW JERSEY ATLANTIC MARGIN

R.K.[1] OLSSON AND S.W. WISE, JR.[2]

ABSTRACT

Biofacies analysis of benthic foraminifera and planktonic foraminiferal and calcareous nannofossil biostratigraphy in upper Paleocene to middle Eocene formations in the New Jersey coastal plain are used to identify depositional sequences and to correlate them with the third-order cycles of Haq and others (1987). Two sequences separated by hiatuses are identified in the upper Paleocene and correlated with cycles TA2.1 and 2.3.

Three sequences are recognized in the lower Eocene and correlated with cycles TA2.4-2.5, 2.7-2.8, and 3.1 Hiatuses separating these sequences in the lower Eocene correlate with type 1 unconformities on the Haq and others (1987) cycle chart. The hiatus corresponding to the type 1 unconformity between cycles TA2.9 and 3.1 on the cycle chart is also recognized in DSDP sites 605 and 613.

Three sequences are delineated in the middle Eocene. They correlate respectively with cycles TA3.3, 3.4, and 3.5-3.6 and are separated by hiatuses.

Paleoslope modeling indicates that relative sea level stood between 55 m and 120 m above present sea level during deposition of transgressive deposits in the New Jersey coastal plain. The shoreline would have shifted about 70 km during this change.

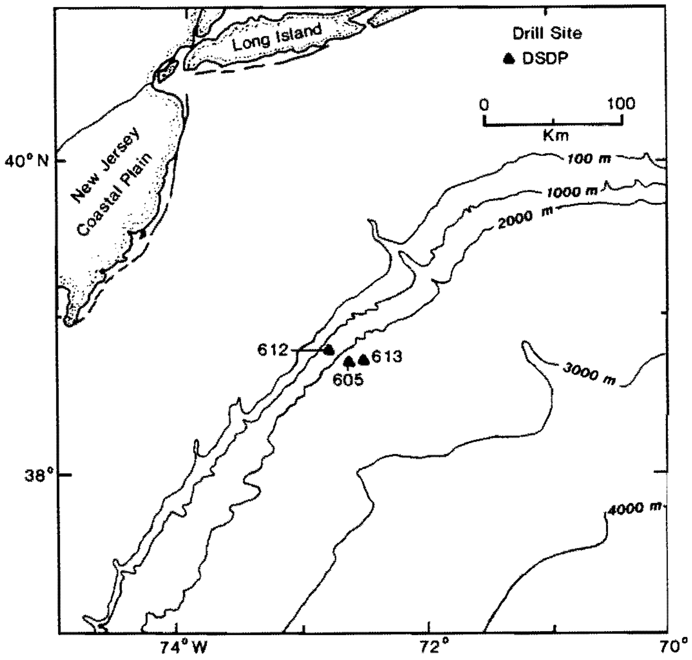

FIGURE 1. Outline map of the New Jersey Atlantic margin showing locations of DSDP drill sites and coastal plain.

INTRODUCTION

Deep Sea Drilling on the New Jersey slope at Sites 605, 612, and 613 penetrated Paleocene and Eocene sections (Fig. 1). The objective of this drilling was to provide data on the depositional history of the margin by linking data from DSDP holes with outcrops, existing wells on land, and wells along the continental shelf (Poag, Watts, and others, 1987). Olsson and Wise (1987) correlated Site 605 with New Jersey coastal plain formations using planktonic foraminifera and calcareous nannofossils. A strong seismic reflector at Site 605 which is associated with a lower Paleocene unconformity was correlated with a prominent unconformity in the coastal plain, and an upper Paleocene hiatus at Site 605 was also correlated with a similar hiatus in the coastal plain. In this study, emphasis is placed on benthic foraminiferal biofacies as an aid to characterizing the upper Paleocene to middle Eocene depositional sequences in the New Jersey coastal plain. In addition, hiatuses recognized here in the coastal plain are compared with data on hiatuses at DSDP Sites 605, 612, and 613 (Fig. 1).

Deposition during the late Paleocene to middle Eocene in the New Jersey coastal plain occurred in inner to outer shelf and upper slope environments. Lithologic differences between units are often subtle in subsurface where facies changes have blurred lithologic boundaries. It is now clear that numerous sea-level events have controlled deposition by shifting lithofacies and biofacies back and forth along the margin during rises and falls of sea level. In the New Jersey coastal plain, lithologically similar strata occur in the lower and middle Eocene, so that paleontological criteria must be employed to separate such strata on the basis of age. Benthic foraminifera and their implied paleobathymetries based upon paleoslope modeling techniques (Olsson and Nyong, 1984) are used to differentiate depositional sequences.

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[1] Department of Geological Sciences, Rutgers University, New Brunswick, NJ 08903
[2] Department of Geology, Florida State University, Tallahassee, FL 32306

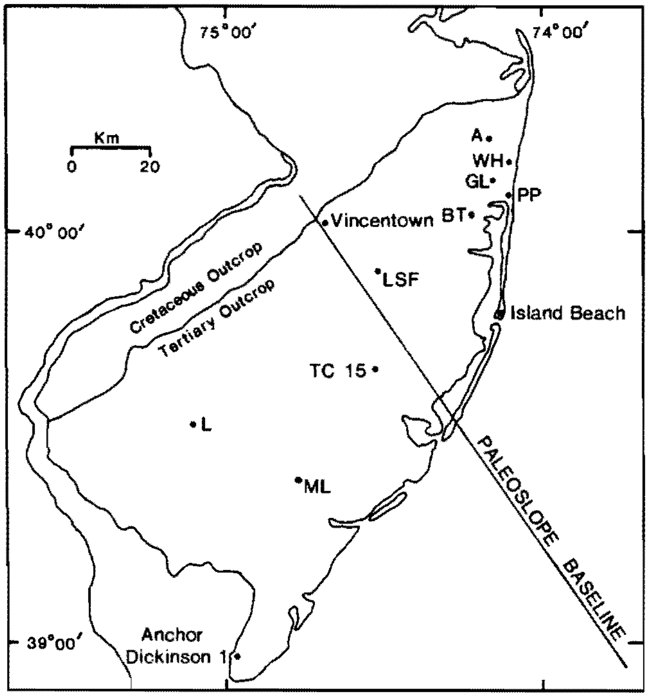

FIGURE 2. Outline map of New Jersey coastal plain showing locations of wells used in this study.

The results of this study are based on data generated from outcrops, core samples from seven wells and cuttings from two wells in the coastal plain, and core samples from DSDP Hole 605 (Figs. 1 and 2). Analysis of outcrops incorporates the data from the Paleocene work of Olsson (1960) and Youssefnia (1978). The planktonic foraminiferal biostratigraphy for the Paleocene is based on split spoon samples from three wells (Glendola, Leggette, and Whitesville Wells) and cable tool samples from the Point Pleasant Well (this study). Paleocene planktonic foraminifera have been previously reported from the Whitesville Well (Olsson, 1969). The Eocene planktonic foraminiferal biostratigraphy in the coastal plain is developed principally in two wells (Leggette and Point Pleasant Wells) where the most complete set of samples were available for study. This data were supplemented by observations from more limited split spoon core samples from the Allaire, Brick Township, and Lebanon State Forest Wells and, cuttings from three wells (Anchor Dickinson 1, Island Beach, and Transco 15). In addition, 22 core samples from DSDP Hole 605 taken in the uppermost part of the Lower Eocene were analyzed for planktonic foraminiferal biostratigraphy. The zonal scheme given in Toumarkine and Luterbacher (1985) was used in this study. It is recognized that certain ambiquities and uncertainties occur in applying zonal criteria in the lower Eocene.

The calcareous nannofossil biostratigraphy for the Paleocene to middle Eocene in the coastal plain was developed in four wells (Glendola, Leggette, Point Pleasant, and Whitesville Wells) and reported on by Jiang and Wise (1987).

The benthic foraminiferal biofacies analysis and paleoslope modeling was conducted on core samples from the seven wells noted above. Previous benthic foraminiferal studies on portion of four of these wells by Charletta (1980) and Enright (1969a) were evaluated and incorporated in this analysis. Data from the three wells with cutting samples (Charletta, 1980) were also evaluated and incorporated in the conclusions of this study.

The sampling interval in coastal plain wells with core samples range between 5 and 10 feet, with some more widely spaced samples. Intervals for cutting samples range between 10 and 20 feet. This interval of sampling appears adequate to document the major stratigraphic sequences in the upper Paleocene to middle Eocene. Nevertheless, the coastal plain section is condensed and thin so that it is not possible in this paper to fully assess the magnitude of hiatuses and to determine whether some sequences are truly absent or are absent due to sampling gaps.

COASTAL PLAIN FORMATIONS

The Hornerstown Formation in outcrop consists of nearly pure deposits of glauconite (Fig. 3). In the subsurface the Hornerstown Formation becomes more clayey and in places in the far downdip it is replaced by a gray clay. The Hornerstown Formation was deposited during two cycles of sea level change (Olsson and Wise, 1987). A prominent lower Paleocene hiatus and associated strong seismic reflector at Site 605 on the New Jersey slope has been correlated by Olsson and Wise (1987) with a lower Paleocene hiatus in the Hornerstown Formation.

The Vincentown Formation (Fig. 3) is a calcareous sand in outcrop. In the subsurface, it becomes more fine-grained and glauconitic and is replaced by glauconitic quartzose silts and glauconitic clays. Increased silt content in the upper Paleocene at Site 605 correlates closely with the Vincentown Formation (Olsson and Wise, 1987).

In the subsurface, a gray clay, silt, and fine sand unit separates the Vincentown and Manasquan Formation . This unit was first noted by Olsson and Wise (1987) as an uppermost Paleocene unit distinct from the Vincentown Formation below and the Manasquan Formation above. Previously, this unit had been logged as beds belonging to one or the other of these formations. The age and benthic foraminiferal biofacies separates this unit from these formations as well.

The Manasquan Formation (Fig. 3) contain two lithofacies which have been given member status: a lower Farmingdale Member and an upper Deal Member (Enright,1969b). The Farmingdale is a slightly clayey glauconitic medium to coarse quartzose sand. Glauconite is abundant and in places nearly pure glauconite beds are present. This lithofacies occurs in outcrop and in the updip subsurface. It thins and disappears in the downdip subsurface.

The Deal Member was originally described as the Ash Marl by Cook (1868) because of its distinctive ash color which varies from grayish-green to yellowish-gray to greenish-yellow. Updip and in outcrop, the Deal Member is a

FORMATION		LITHOLOGY	
		OUTCROP	SUBSURFACE
SHARK RIVER	TOMS RIVER	ABSENT	GLAUCONITIC, FINE TO MEDIUM SAND
	SQUANKUM	ARGILLACEOUS GLAUCONITIC SAND	SILTY & SANDY CLAY (DEAL)
MANASQUAN	DEAL	SLIGHTLY GLAUCONITIC CLAYEY FINE SAND & SILT (ASH MARL)	
	FARMINGDALE	GLAUCONITIC SAND TO GREENSAND	
UNNAMED		ABSENT	CLAY, SILTS & FINE SAND
VINCENTOWN		CALCAREOUS SAND	GLAUCONITIC SILTS & CLAYS
HORNERSTOWN		GREENSAND	GLAUCONTIC CLAY

LITHOFACIES DISTRIBUTION

INNER SHELF TO UPPER SLOPE

SANDS ——— GREENSANDS & GLAUCONITE RICH SANDS & SILTS ——— CLAYS

FIGURE 3. Lithology of coastal plain formations in outcrop and in the subsurface. Lower part of figure shows bathymetric distribution of lithofacies.

slightly glauconitic, clayey, fine quartz sand to clayey sandy silt unit. Downdip it varies from a slightly sandy, clayey silt to a silty clay. The Deal Member replaces the Farmingdale Member as it thins in the subsurface. In turn, it also replaces in the subsurface the lower facies of the overlying Shark River Formation. The Deal Member, thus, becomes the dominant Eocene lithology in subsurface and compares closely with Eocene clays encountered in DSDP Sites 605, 612 and 613.

The Shark River Formation (Fig. 3) also contains two lithofacies which have been given member status: a lower Squankum Member and an upper Toms River Member (Enright, 1969b). The Squankum Member is an argillaceous, glauconitic sand which grades rapidly in subsurface into the Deal Member. The Toms River Member is a micaceous, slightly clayey and glauconitic fine to medium sand. It occurs only in the subsurface.

The distribution ot the four Eocene lithofacies in the New Jersey coastal plain suggests that they were deposited in adjacent environments, with the sandier deposits accumulating nearer to the shoreline and the clay deposits forming farthest from the shoreline. Glauconite rich sediments were deposited in intermediate positions between sand and clay. Dominance of the Deal Member lithofacies in the subsurface indicates that deep paleobathymetric depths prevailed continuously here during the early to middle Eocene. This can be misleading because paleoecologic and biostratigraphic data indicate several abrupt changes in paleobathymetry and several hiatuses. Thus, although the lithologic changes that occur might be interpreted as facies transgressing time, in fact, these lithofacies were characteristic of Eocene continental margin environments. The juxtaposition of these lithofacies in the subsurface is due to sea-level events which shifted deposition of each lithofacies to its corresponding bathymetric position over the surface of a previous depositional sequence. Biostratigraphic and foraminiferal biofacies analysis suggest that these lithofacies are transgressive deposits from six separate cycles of deposition. They each have different paleobathymetric histories. Deposition during the early and middle Eocene in New Jersey occurred in middle shelf to upper slope environments.

FORAMINIFERAL BIOFACIES

The two Paleocene depositional sequences recognized in the New Jersey coastal plain contain strikingly different benthic foraminiferal assemblages (Fig. 4). The lower sequence, which includes the upper Hornerstown and Vincentown Formations and their subsurface equivalents, contains a diverse assemblage with many well-known, typical Paleocene species. Although complete biofacies analysis is the subject of another study, some of the general characteristics are pointed out here (Fig. 4). The most common and widespread species in the subsurface include *Cibicidoides alleni* (Plummer), *Cibicides marylandicus* Shifflett, *Bolivinopsis emmendorferi* (Jennings), and *Spiroplectammina wilcoxensis* (Schwager). In the most downdip wells *Bulimina pseudocacumenta* Olsson, *Epistominella minuta* (Olsson), *Gavelinella danica* (Brotzen), and *Tappanina selmenensis* (Cushman) become more abundant. These assemblages are indicative of Paleocene shelf environments and have been termed "Midway-type fauna" by Berggren and Aubert (1975). In outcrop sections of New Jersey and Maryland, an inner shelf assemblage consisting of *Anomalinoides umboniferus* (Schwager), *Globulina gibba* d'Orbigny, *Cibicidoides alleni*, and *Pararotalia perclara* (Loeblich and Tappan) has been delineated by Youssefnia (1978). Preliminary studies indicate that up to four biofacies can be recognized in the upper Paleocene shelf environments.

The uppermost Paleocene depositional sequence contains few benthic foraminifera (Fig. 4). The most persistant species is *Spiroplectammina spectabilis* (Grzybowski). In the Leggette Well, *Pulsiphonina prima* (Plummer) and *Tappanina selmensis* are the most common species in a small-sized, low-diversity assemblage. In the Point Pleasant Well the sequence is barren of foraminifera and coccoliths. Coccoliths are also absent and planktonic foraminifera are sparse in this sequence in the Glendola Well. *Spiroplectammina spectabilis* appears to be the only in-place benthic species in this well and in the Whitesville Well. In the Whitesville Well, coccoliths indicate a CP8 and CP9 Zone for the sequence (Jiang and Wise, 1987). In the Leggette

TYPICAL PALEOCENE FORAMINIFERAL ASSEMBLAGES OF COASTAL PLAIN

NEARSHORE		*Anomalinoides umboniferus* *Cibicidoides alleni* *Globulina gibba* *Pararotalia perclara*
SHELF " MIDWAY-TYPE "	SHALLOW	*Bolivinopsis emmendorferi* *Cibicides alleni* *Cibicides marylandicus*
	DEEP	*Bulimina pseudocacumenata* *Epistominella minuta* *Tappanina selmensis*
OXYGEN MINIMUM		*Spiroplectammina spectabilis*

FIGURE 4. Typical Paleocene foraminiferal assemblages in the New Jersey and Maryland coastal plains.

Well, planktonic foraminifera are abundant and well-preserved in the sequence. *Morozovella velascoensis* (Cushman) and *Morozovella subbotinae* (Morozova) among others indicate Zone P6a. Coccoliths correlate the interval with Zone CP8.

The limited benthic assemblage and the absence of foraminifera in one well suggests that poor environmental conditions existed during deposition of this uppermost Paleocene depositional sequence. The dominance of the agglutinated species *S. spectablis* in two wells and the poor preservation of foraminifera in three wells may indicate that oxygen minimum conditions were present during deposition of the sequence. The diverse planktonic foraminiferal assemblage which lacks fully developed adult forms in the Leggette Well indicates middle to outer shelf depths.

Benthic foraminifera are abundant and diverse in the lower and middle Eocene strata of New Jersey. They are representative of shelf and upper slope environments and occur in five recognizable biofacies distributed in several depositional sequences (Fig. 5). Biofacies 1 is associated with fine to medium sand lithofacies and is characterized by dominance of the benthic species *Epistominella minuta* (Olsson), *Hanzawaia mauricensis* (Howe and Roberts), *Pararotalia inconspicua* (Howe), and *Uvigerina elongata* Cole. The percentage of planktonic foraminifera present in assemblages of this biofacies ranges from less than one percent to about 7 percent. This biofacies indicates an inner shelf environment where bathymetry was on the order of 50 m. Biofacies indicative of shallower paleodepths are absent in lower and middle Eocene strata in New Jersey.

Biofacies 2 (Fig. 5) occurs in a very fine sand and silt lithofacies. Representative species include *Alabamina mississippiensis* Todd, *Gyroidinoides octameratus* (Cushman and Hanna),

DISTRIBUTION OF EOCENE BIOFACIES IN NEW JERSEY

DEPOSITIONAL SEQUENCE AFTER HAQ ET AL., 1987	AL	LSF	BT	PP	L	TC 15	AD	ZONES FORAMS & NANNOS
TA 3.5-3.6	X	Biofacies 1 P% < 1	Biofacies 1 P% < 1	Biofacies 1 P% < 1	Biofacies 1 P% < 1	X	X	P 14? CP 14?
TA 3.4	X	Biofacies 1 P% < 1	Biofacies 1 P% 1-5	Biofacies 1 P% 4-6	Biofacies 2 P% 20	Biofacies 2 P% 10	X	P 12 CP 13-14
TA 3.3	X	Biofacies 4 P% 20	Biofacies 4 P% 14-33	Biofacies 4 P% 58	Biofacies 4 P% 70	Biofacies 4 P% 50-60	X	P 11 CP 13
TA 3.1	X	X	X	Biofacies 1 P% 7	Biofacies 2 P% 26	X	Biofacies 3 P% 50	P 9 CP 12A
TA 2.7-2.8	Biofacies 3 P% 40-50	Biofacies 4 P% 20-70	Biofacies 4 P% 83	Biofacies 4 P% 60-71	Biofacies 4 P% 70	Biofacies 4 P% 60-70	Biofacies 5 P% 75	P 8 CP 10-11
TA 2.4-2.5	X	X	Biofacies 4 P% 46-65	Biofacies 4 P% 42-60	Biofacies 4 P% 74-65	Biofacies P% 80	Biofacies 5 P% 90	P 8B CP 9

BIOFACIES 1	*Epistominella minuta* *Hanzawaia mauricensis* *Pararotalia inconspicua* *Uvigerina elongata*
BIOFACIES 2	*Alabamina mississippiensis* *Gyroidinoides octacameratus* *Hanzawaia mauricensis*
BIOFACIES 3	*Bulimina whitei* *Kolesnikovella elongata* *Turrilina robertsi*
BIOFACIES 4	*Pyramidina subrotundata* *Siphonina claibornensis* *Turrilina robertsi* *Cibicidoides sps.*
BIOFACIES 5	*Anomalinoides spissiformis* *Bulimina sp.* *Trifarina wilcoxensis*

P% = % PLANKTONIC FORAMINIFERA
X = NO DATA

FIGURE 5. Distribution of benthic foraminiferal biofacies in New Jersey coastal plain wells.

as well as *H. mauricensis*. Biofacies 2 is found downdip of Biofacies 1, with which it has many species in common. The percentage of planktonic foraminifera averages 10 percent.

Biofacies 3 (Fig. 5) is identified by the common occurrence of the benthic species *Bulimina whitei* Martin, *Kolesnikovella elongata* (Halkyard), and *Turrilina robertsi* (Howe and Ellis). Forty to fifty percent of the assemblages of this biofacies consists of planktonic foraminifera. It occurs in the farthest updip position in the Allaire Well and also in the farthest downdip well, the Anchor Dickinson Well (Fig. 5). The lithology associated with this biofacies is very fine sand, silt and clay.

The most extensive biofacies found in the lower to middle Eocene in the coastal plain is biofacies 4 (Fig. 5). This biofacies is characterized by large-sized species of the the genus *Cibicidoides* and the species *Pyramidina subrotundata* (Cushman and Thomas), *Siphonina claibornensis* Cushman, as well as *T. robertsi*. The percentage of planktonic foraminifera varies from 20 to 80 percent. The dominant lithology is silty clay and clay. Glauconite is widespread and in places very abundant.

Biofacies 5 occurs farthest downdip from biofacies 4 (Fig. 5). The most abundant species in this biofacies include *Anomalinoides spissiformis* Cushman and Stainforth, *Trifarina wilcoxensis* (Cushman and Ponton), and *Bulimina* sp. In addition, species such as *Gavelinella capitata* (Gumbel), *Gavelinella micra* (Burmudez), and *Nuttalides truempyi* (Nuttall), which are common in Eocene bathyal and abyssal environments (Tjalsma and Lohmann, 1983), are present in small numbers. The percentage of planktonic foraminifera varies from 75 to 90 percent. Clay predominates with this biofacies.

DEPOSITIONAL SEQUENCES

The distribution of the upper Paleocene and Eocene biofacies in the New Jersey coastal plain shows that abrupt changes occur between biofacies in the wells used in this study (Figs. 4,5). In addition, biostratigraphic criteria show or suggest that hiatuses separate most of the biofacies. We use the vertical changes of biofacies and the hiatuses that separate them to identify depositional sequences. The distribution of biofacies in the coastal plain (Figs. 4,5) indicates the relative paleobathymetric history of each depositional sequence and is a useful way in which to identify a sequence.

The upper Paleocene sequence (Fig. 6) is separated from the lower Paleocene sequence by a prominent unconformity which can be traced to

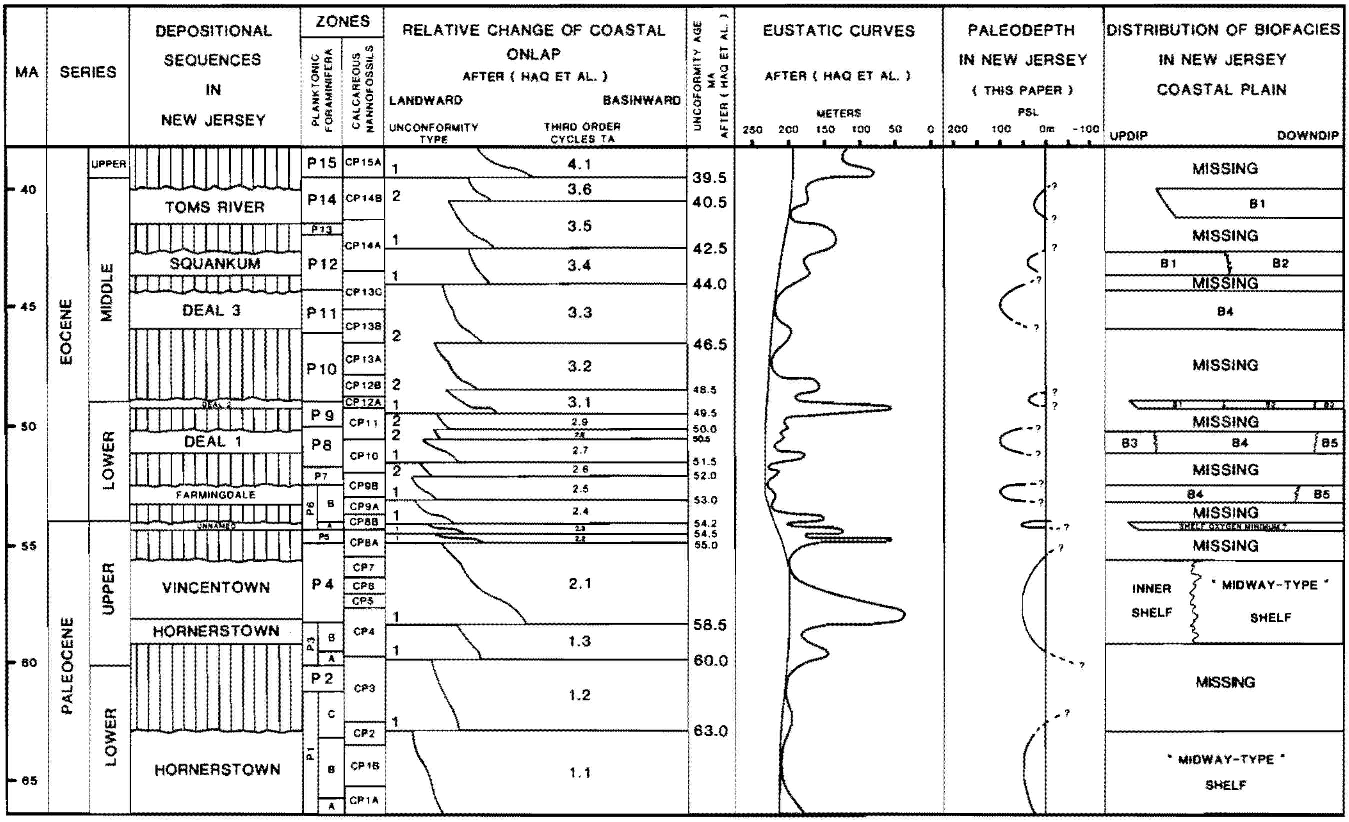

FIGURE 6. Distribution of biofacies in the New Jersey coastal plain and their correlation with the third order cycles of Haq and others (1987). Note comparison of eustatic curve of Haq and others with curve derived from paleoslope modeling in New Jersey. Time scale from Haq and others.

DSDP Site 605, where it is associated with a strong seismic reflector (Olsson and Wise, 1987). This upper Paleocene sequence, which in New Jersey contains the upper part of the Hornerstown Formation and the Vincentown Formation, was correlated by Olsson and Wise (1987) with the third-order cycle TP2.2 of Vail and others (1977). The underlying lower Paleocene sequence was correlated with third-order cycle TP1. The hiatus was correlated with the top of TP1. Cycle TP2.1 is missing in the coastal plain. In the revised sea-level cycle chart (Haq and others, 1987), a different terminology is used and the number of sequences designated in the Paleocene have increased. In this paper we attempt to correlate with the revised sequence stratigraphy (Fig. 6).

The lower Paleocene unconformity correlated to Site 605 is a type 1 unconformity (subaerial and submarine in extent). Two type 1 unconformities are shown on the revised cycle chart in the middle Paleocene (Fig. 6). We correlate the upper Paleocene sequence (Hornerstown-Vincentown) with the third-order cycle TA2.1. The sequence in New Jersey contains planktonic foraminifera zones P3 and P4 and coccolith zones CP4 to CP7. Thus, the strong seismic reflector at Site 605 apparently correlates with the type 1 unconformity shown at the top of TA1.2 or TA1.3 on the revised cycle chart (Fig. 6). We cannot differentiate sequence TA1.3 and its associated type 1 unconformity. There are no abrupt biofacies changes evident within the Hornerstown and Vincentown Formations which might suggest a boundary between depositional sequences. We conclude that TA1.3 may be miscorrelated on the cycle chart and that it is possibly a lower Paleocene event. If so, the unconformity at Site 605 may be coalesced unconformities of cycles TA1.2 and 1.3.

The uppermost Paleocene depositional sequence rests unconformably on the Vincentown Formation (Fig. 6). A hiatus which encompasses possibly part of P4 and all of P5 is suggested on planktonic foraminiferal criteria. The sharp contrast between the benthic foraminiferal assemblages in this sequence and those in the underlying Paleocene and the overlying Eocene sequences suggest that it is bounded by unconformities. A hiatus within the lower Eocene cannot be resolved via planktonic foraminiferal nor coccolith biostratigraphy, however. We correlate the uppermost Paleocene sequence with the third-order cycle TA2.2 or TA2.3 of the revised cycle chart and suggest that the sequence represented is probably TA2.3. Two third-order cycles (TA2.2 and 2.3) of 1.0 and 1.5 Ma duration are shown (Fig. 6) in the uppermost Paleocene on the revised cycle chart.

Type 1 unconformities separate these cycles at their bases and at their tops. At Site 605, a hiatus is suggested between planktonic foraminiferal Zones P4 and P5 (Saint-Marc, 1987). This correlates with the type 1 unconformity at the top of TA2.1 (Vincentown Formation) in the coastal plain. Zone P5 at Site 605 is very thin in comparison to Zone P6a, which is thicker and well developed. Third-order cycle TA2.3, thus, appears to be the one represented in the New Jersey margin. Saint-Marc (1987) noted a very high percentage of planktonic foraminifera (95-99%) in the interval of this sequence. He suggested either that a low rate of terrigeneous-hemipelagic sedimentation or that a higher rate of productivity could explain the high planktonic abundance. A higher rate of productivity, as well as slow sedimentation may well have led to oxygen minimum conditions in the latest Paleocene shelf environments of the coastal plain. These conditions are, indeed, suggested by the coastal plain benthic foraminifera in this sequence. Cycle TA2.2 cannot be identified in the coastal plain and is poorly represented at Site 605. Exposure of the coastal plain at the end of this cycle may have removed any sediments that were deposited during its transgressive and highstand phases. Validation of this cycle in the New Jersey margin will require further study.

Six depositional sequences are recognized in the Eocene section examined in this study, with three in the lower and three in the middle Eocene (Fig. 6). The lowermost sequence correlates with planktonic foraminiferal Zone P6b. In the Leggette Well (Fig. 2) the top of this sequence contains Morozovella aequa (Cushman and Renz), Morozovella formosa gracilis (Bolli) and Morozovella subbotinae in a biofacies 3 benthic assemblage (Fig. 7). Jiang and Wise (1987) correlate the interval of this sequence with nannofossil Zone CP9a. In the Point Pleasant Well, the same foraminiferal species appear at the top of this

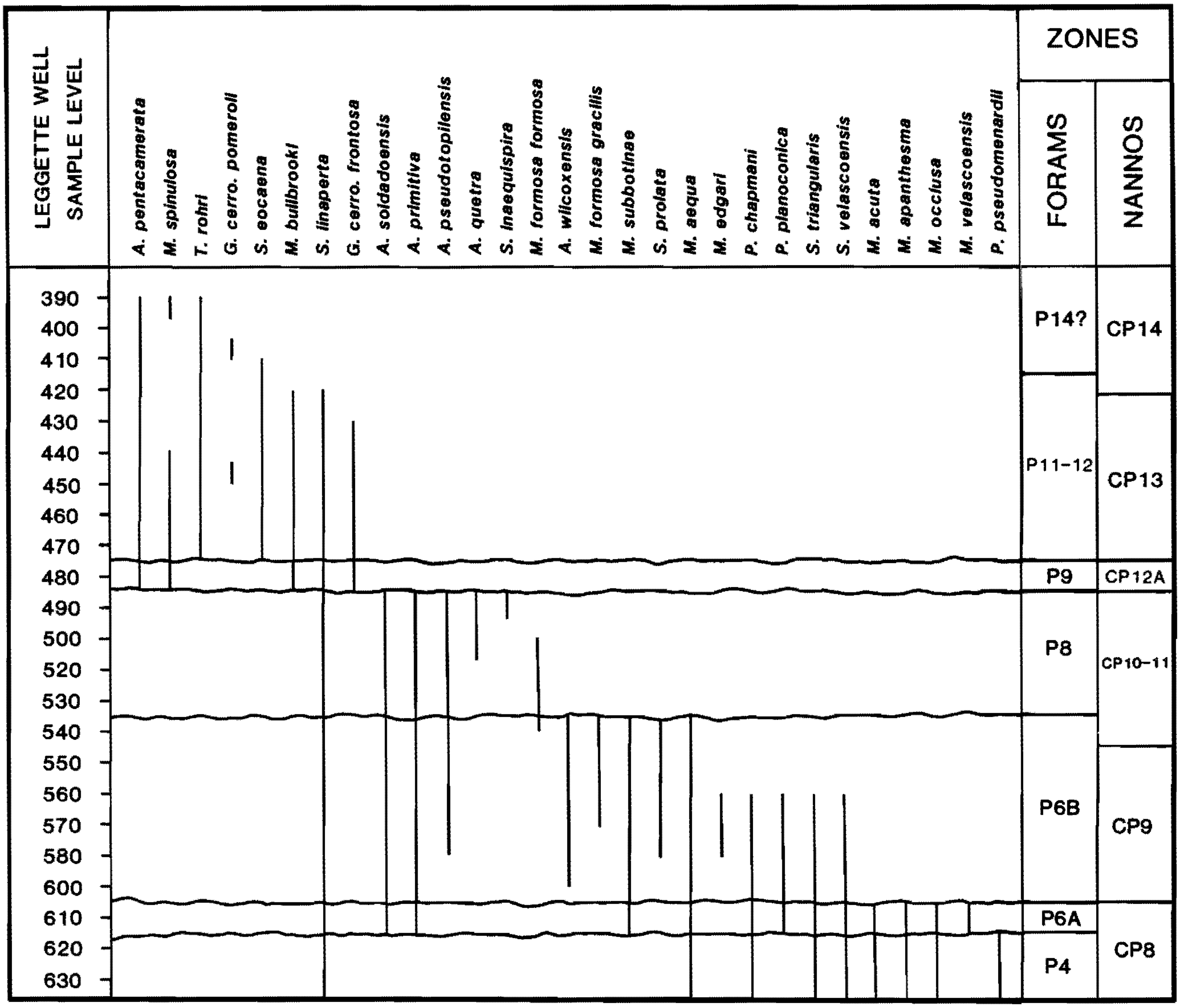

FIGURE 7. Distribution of upper Paleocene to middle Eocene planktonic foraminiferal species in the Leggette Well.

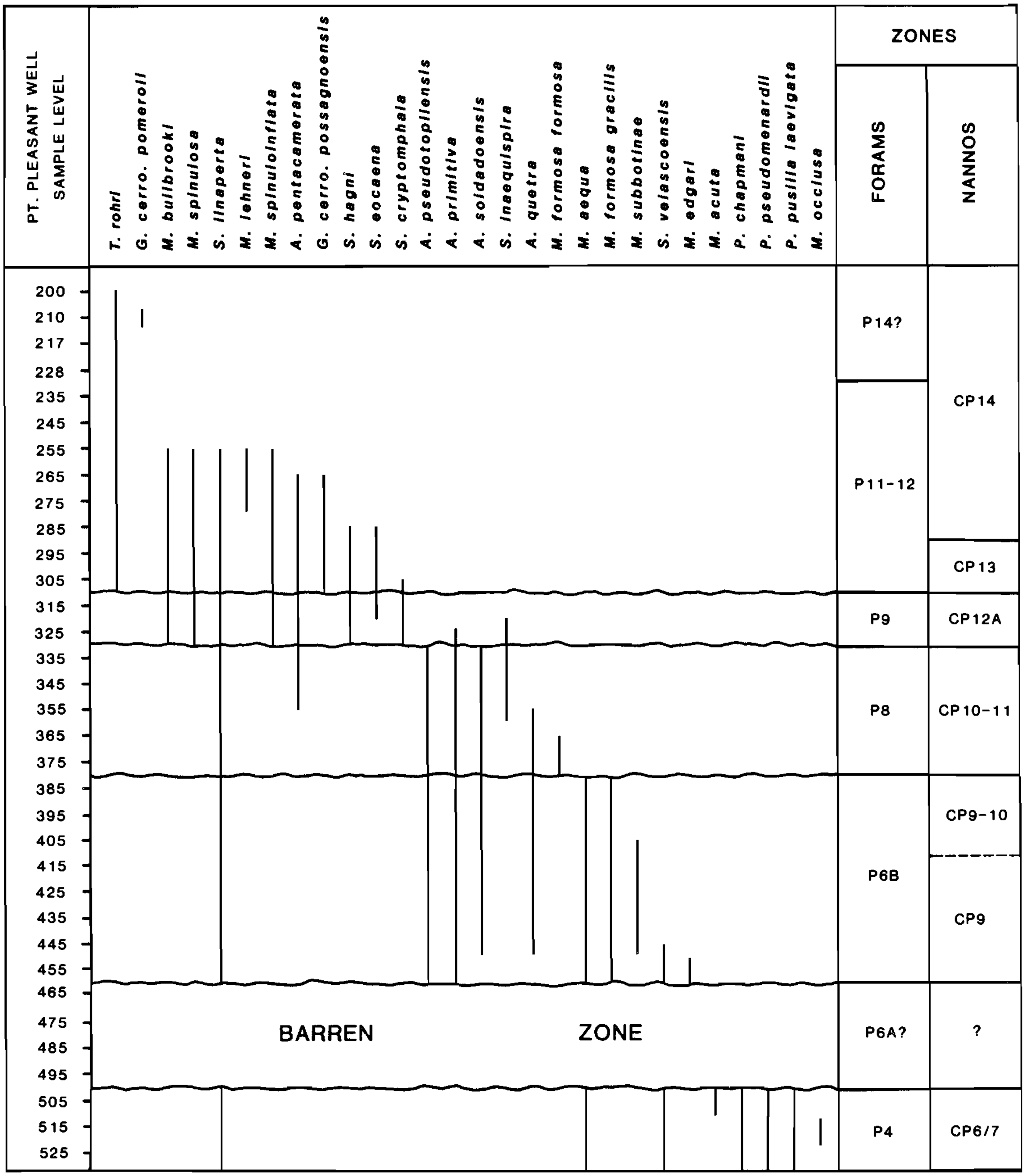

FIGURE 8. Distribution of upper Paleocene to middle Eocene planktonic foraminiferal species in the Point Pleasant Well.

sequence (Fig. 8). Nannofossil correlation is less precise than in the Leggette Well but suggests Zone CP9. Benthic foraminiferal biofacies 4 occurs in this sequence in the Point Pleasant Well. The lithology of this sequence varies from a glauconitic, very fine sand to a clay and greensand (Fig. 3). Formal names applied to these lithologies include the Farmingdale, Manasquan, and Deal Members. In general, this sequence is more glauconitic than the sequences above and below.

The overlying sequence is characterized lithologically as light ash gray to gray-white glauconitic clay and silt (Fig. 3). Foraminifera and siliceous microfossils (radiolaria, diatoms, sponge spicules) are often very abundant. The formal name Deal Member is applied here (Fig. 6). The sequence interval is correlated with Zones P8 and CP10-11. Benthic biofacies 3 and 4 are associated with this sequence. In places, siliceous microfossils are very abundant and lenses and layers of porcellanite are sometimes encountered. This sequence appears to be separated from the underlying sequence by an unconformity. The appearance of M. aequa in the top of the underlying sequence suggests that the topmost part of P6 is missing (Figs. 7,8). Jiang and Wise (1987) point out that nannofossil Zone CP9b is possibly missing in the Leggette Well. In addition, foraminiferal Zone P7 cannot be clearly delineated in the wells studied herein. The first appearance of Morozovella formosa gracilis downhole occurs with the first appearance of M. aequa, which suggests that Zone P7 may be missing (Figs. 7,8). We therefore show a hiatus separating the two sequences (Fig. 6).

At the top of the lower Eocene, a thin sequence (about 10 feet) of glauconitic, gray-white sandy clay to light ash and greenish-gray clayey greensand (Deal Member) contains biofacies 1 and 2 (Figs. 3,5). Nannofossils and planktonic foraminifera place this sequence in Zone CP12a and P9, respectively. The abrupt change in paleobathymetry from deep to shallow may indicate an unconformity. If so, this unconformity correlates with a hiatus in the lower Eocene at Site 605 (Fig. 9). Applegate and Wise (1987) noted the relatively short interval occupied by Martini nannofossil Zone NP13 and suggested that an undetected hiatus occurs within or at the boundary of Zone NP13. An examination of the planktonic foraminifera across this interval shows that this is indeed the case (Fig. 9). In sample 605-32-2, 20-21 cm and above, an assemblage which contains Morozovella aragonensis (Nuttall), M. spinulosa (Cushman), M. bullbrooki (Bolli), M. spinuloinflata (Bandy), Subbotina eocaena (Guembel), and S. hagni (Gohrbrandt) suggests an upper Zone P9. Morozovella formosa formosa (Bolli) appears in the next sample below with Acarinina soldadoensis (Bronnimann) and A. pseudotopilensis (Subbotina) indicating Zone P8. Since M. formosa formosa goes extinct within Zone P8 (Toumarkine and Luterbacher, 1985), the upper part of the zone is probably missing. A single occurrence of Planorotalites palmerae (Cushman and Barmudez) along with several specimens of Turborotalia griffinae (Blow) in this sample is regarded as evidence of downhole contamination or mixing of P9 species at the unconformity. Out-of-place species do not occur in the next sample below (Fig. 9). A similar unconformity is also present near Site 605 at Site 613, where Valentine (1987) records the absence of the middle part of Zone CP11 and NP13 (Fig. 9). He estimates a hiatus with a duration of about 1.1 Ma.

In the lower part of the middle Eocene, the Deal Member lithology persists as light ash-gray to gray-white glauconitic clay (Fig. 3). Abundant foraminifera in this sequence belong to biofacies 4 (Fig. 5). Planktonic foraminifera suggest Zones P11-P12. Nannofossil biostratigraphy (Jiang and Wise, 1987) places this sequence within CP13 and CP14 (NP15 to NP17) (Fig. 6). Consequently, a hiatus probably separates this sequence from the lower Eocene. A hiatus of short duration is noted at DSDP Site 612 (Miller and Katz, 1987; Poag and Low, 1987) between the lower and middle Eocene (Fig. 9). Miller and Katz (1985) noted the first appearance of numerous benthic foraminiferal species, while Poag and Low (1985) recorded a major change across this boundary from a Bulimina dominated biofacies below to a Stilostomella - Cibicides - Cassidulina biofacies. In the coastal plain, a shallow to deeper depth change occurs with the replacement of biofacies 1 and 2 in the lower Eocene by biofacies 4 in the middle Eocene.

Overlying this deep-water facies, in sharp contrast, is a sequence which contains biofacies 1 and 2 (Fig. 5). Lithologically, this sequence consists of light ash-gray, glauconitic, very fine sand, silt, and clay (Fig. 3). This has been termed the Squankum Member of the Shark River Formation (Enright, 1969b). Nannofossil (Jiang and Wise, 1987) and planktonic foraminifera correlate this sequence with Zones CP13 to

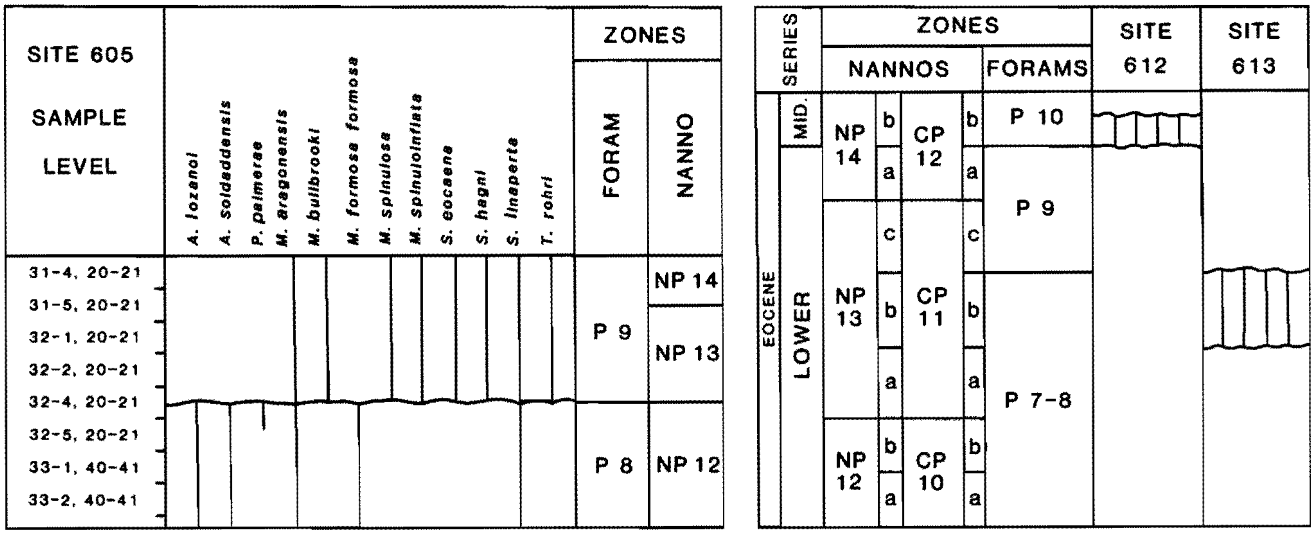

FIGURE 9. Distribution upper lower Eocene planktonic foraminiferal species in DSDP Site 605. Comparison is made to Sites 612 and 613 (Miller and Katz, 1985; Poag and Low, 1985; Valentine, 1985).

CP14 and P11 to P12, respectively (Fig. 6). We place a hiatus between this sequence and the one below on the basis of their sharp contrast in paleobathymetric histories. We are unable to biostratigraphically resolve a hiatus with microfossils due to the lack of diagnostic marker species.

Gray, glauconitic silty and clayey fine, medium and coarse sand constitutes the third sequence in the middle Eocene (Fig. 3). The lithology has been termed the Toms River Member of the Shark River Formation (Enright, 1969b). A shallow inner shelf biofacies B1 occurs in this sequence. Planktonic microfossils are sparse, so that biostratigraphic resolution is limited. Nannofossils suggest Zone CP14. The presence of the foraminiferal species Globorotalia cerroazulensis pomeroli (Tourmarkine and Bolli) without G. cerroazulensis frontosa (Subbotina) (Figs. 7,8) may indicate Zones P13 or P14.

CORRELATION OF THE NEW JERSEY EOCENE DEPOSITIONAL SEQUENCES WITH THE CYCLE CHART

As is evident from the above discussion, the lower and middle Eocene depositional sequences are difficult to separate on lithologic criteria alone. Benthic foraminiferal biofacies analysis reveals separate paleobathymetric histories, aiding in the recognition of depositional sequences. Detailed biostratigraphy using nannofossils and planktonic foraminifera is necessary to assess hiatal boundaries. This has been most successful in the lower Eocene and less so in the middle Eocene. Our correlations of the New Jersey Eocene sequences with the revised cycle chart (Haq and others, 1987) is therefore based upon integrating planktonic biostratigraphy with benthic foraminiferal sequence studies (Fig. 6).

The lowermost Eocene sequence is termed the Farmingdale sequence. It apparently correlates with third-order cycles TA2.4 and 2.5. The type 1 unconformity shown on the cycle chart at the top of TA2.6 is apparently expressed in the coastal plain by a hiatus which spans part or all of Zones P7 and CP9b. The sequences identified in the Deal Member lithology are termed Deal sequences 1, 2, and 3. Deal sequence 1 is correlated with third-order cycle TA 2.7 and 2.8. The type 1 unconformity at the top of TA2.9 correlates with a hiatus which encompasses the upper part of Zone P9 in the coastal plain and at DSDP Site 605, thus corroborating the suggestion by Applegate and Wise (1987) that part of Zone NP13 is missing.

Deal sequence 2 apparently is correlated with third-order cycle TA3.1 and Deal sequence 3 with TA3.3 (Fig. 6). The hiatus drawn between Deal sequences 2 and 3 would correlate largely with TA3.2. This hiatus corresponds to a type 2 unconformity on the cycle chart. A recently drilled well downdip from the Leggette Well at Mays Landing, New Jersey (Fig. 2), is reported as containing lowermost middle Eocene strata (Poore and Bybell, 1987) in a Deal Member lithology. This strata is either missing or too thin to be recognized in the wells used in this study, given the constraints of the sampling interval. Thus, it is possible that Deal sequence 3 also includes third-order cycle TA3.2

The Squankum sequence and the Toms River sequence are correlated with third-order cycles TA3.4 and 3.5 to 3.6, respectively (Fig. 6). The unconformity at the top of the Toms River sequence corresponds to the type 1 unconformity at the top of TA3.6.

PALEOBATHYMETRY

The distribution of biofacies in the Eocene follows a pattern which is useful for paleoslope modeling using the techniques of Olsson and Nyong (1984). Paleoslope modeling of the upper Paleocene is not possible in the current study because too few sections are available in the data base. Nevertheless, as previously discussed, several biofacies are recognized and are useful for estimates of paleobathymetry, especially when compared to the Eocene paleoslope model.

The distribution of wells used in establishing the paleoslope baseline is shown in figure 2. Some of the data derived in this analysis is based upon paleoecologic studies of New Jersey Eocene foraminifera by Enright (1969a) and Charletta (1980). Figures 5 and 10 show the distribution of biofacies along the paleoslope baseline, and figure 11 is the derived paleoslope model using the graphic technique of Olsson and Nyong (1984). Estimates of sea level change in the coastal plain are derived from the paleoslope model and are referenced to present day sea level. These estimates are the basis for the paleodepth curve shown in figure 6. Paleoslope estimates of sea-level change in the New Jersey coastal plain indicate that maximum highstands of sea level in the Eocene were about 120 m above present sea level (Fig. 12). Lowstands during deposition in the coastal plain stood about 55 m above present sea level. Sea-level change from lowstand to highstand was about 65 m. The shoreline would have shifted around 70 km during this change. Maximum lowstands of sea level during sea-level falls can not be derived if sea level retreated from the coastal plain as suggested by the hiatuses representing type 1 unconformities. The estimates derived here are for third-order cycle highstands of sea level.

The paleoslope estimates of sea-level highstands in the coastal plain during the late Cretaceous (Olsson and Nyong, 1984), the Miocene (Olsson and others, 1987), and in this study of the Paleocene and Eocene are much less than the values given on the Exxon cycle charts by Vail and others (1977) and Haq and others (1987). The contrast in paleobathymetric values derived from paleoslope models and from the cycle charts is between paleoecologic data and the seismic reflection data (onlap-offlap patterns) used to derive some (e.g., the 2nd order cycle boundaries) sea-level values in the cycle chart. The distribution of foraminiferal assemblages within depositional sequences deposited on a tectonically slowly subsiding passive margin should accurately reflect rise and fall in sea level. Nevertheless, paleoslope studies must be integrated with seismic reflection studies in order to resolve these discrepancies in sea-level estimates.

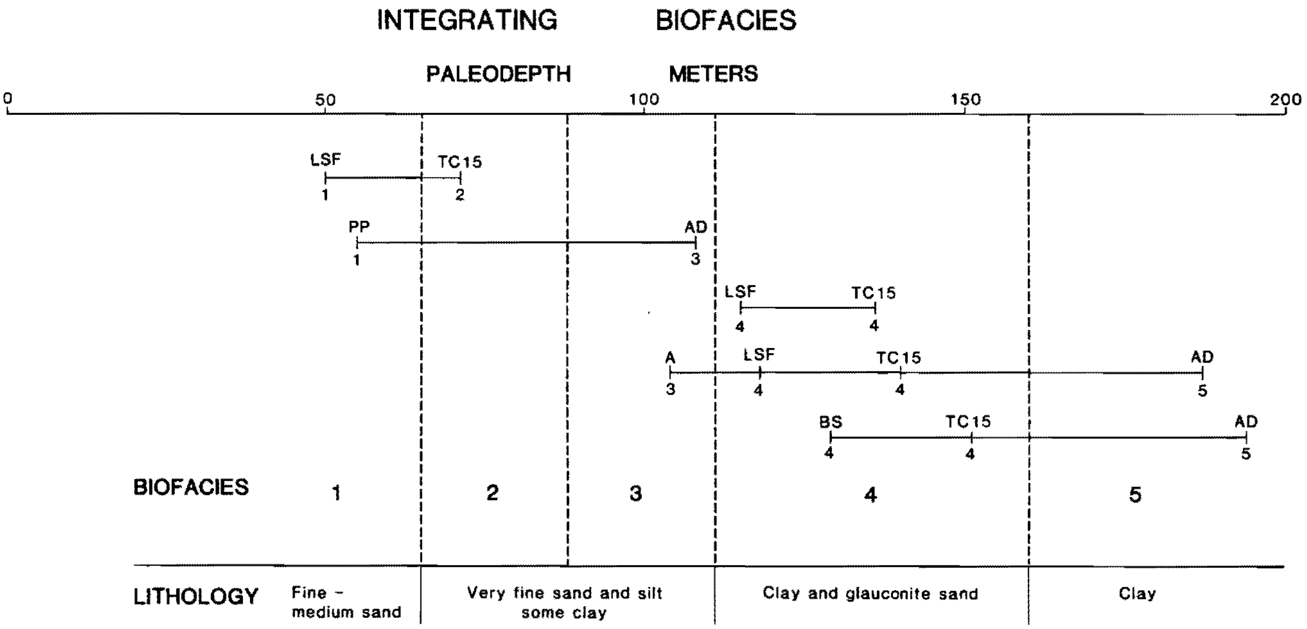

FIGURE 10. Integration of biofacies into paleoslope. The distance between wells with each biofacies is expressed in depth using a gradient of approximately 1 m per kilometer.

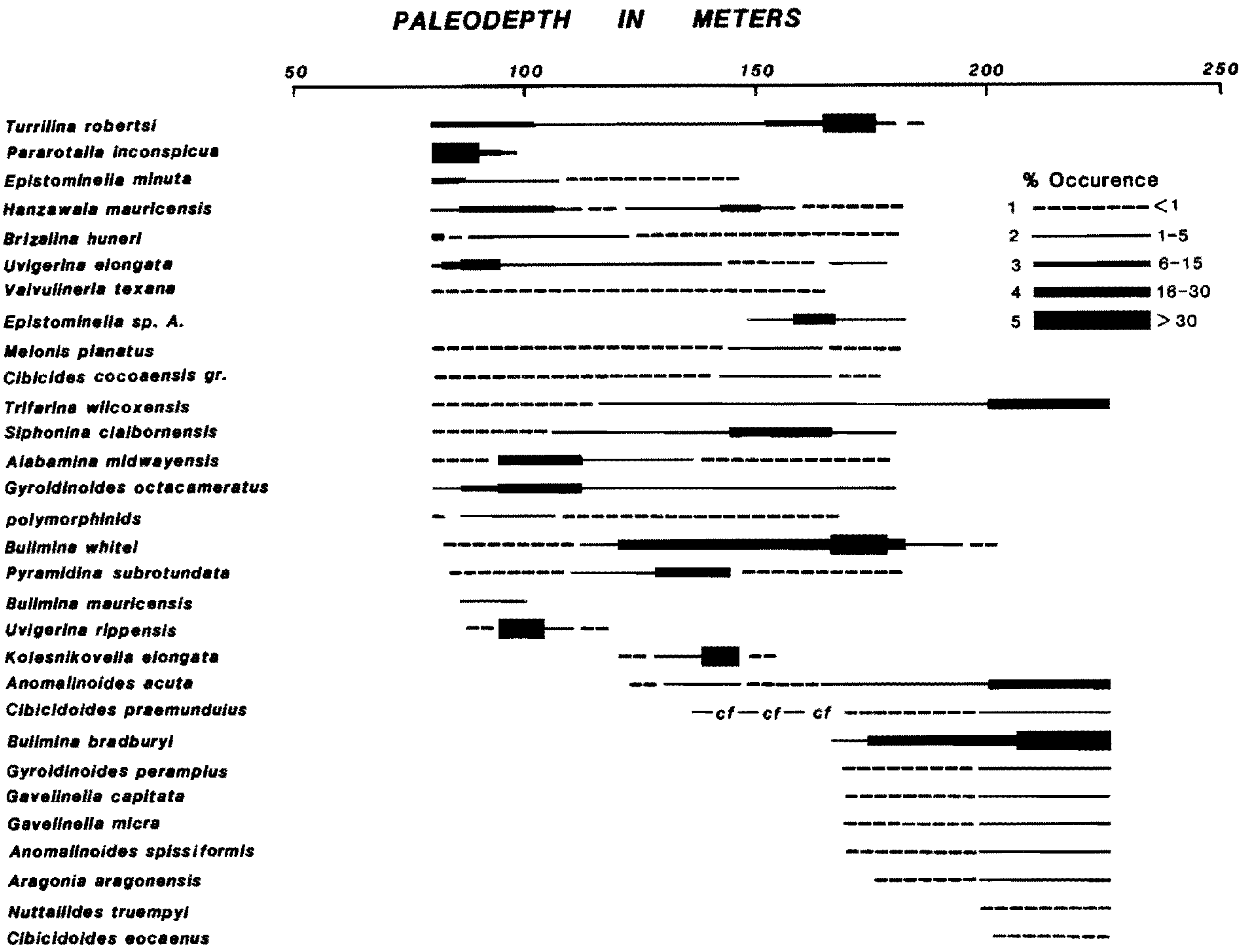

FIGURE 11. Paleoslope model derived from figure 10 showing paleodepth distribution of most common species. Paleodepth calculations shown in figures 6 and 12 are derived from this model.

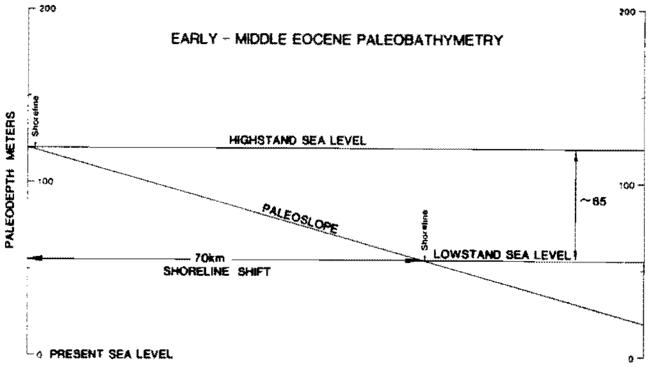

FIGURE 12. Estimate of early and middle Eocene paleobathymetry in the New Jersey coastal plain. Figure compares the lowest and highest stand of sea level in the coastal plain.

SUMMARY

Foraminiferal biofacies analysis, foraminiferal and nannofossil biostratigraphy, and lithofacies changes in the upper Paleocene to middle Eocene in the coastal plain of New Jersey leads to the recognition of eight depositional sequences. Foraminiferal paleoslope analysis shows that each of the sequences have different paleobathymetric histories (Fig. 6). The upper Paleocene sequences are correlated with sequences TA2.1 and TA2.3 on the cycle chart of Haq and others (1987). Two hiatuses are recognized in the Paleocene, a prominent hiatus which separates sequence TA2.1 from the lower Paleocene sequence TA1.1. This hiatus correlates with a similar hiatus at Site 605 where it coincides with a strongly reflective seismic reflection horizon (Olsson and Wise, 1987; Wise and van Hinte, 1987). The hiatus corresponds with the type 1 unconformity that separates sequences TA1.2 and TA1.3 on the cycle chart as recalibrated by Haq and others (1987) and sequences TP1 and TP2.1 on the cycle chart as calibrated by Vail and others (1980). On the revised cycle chart, a type 1 unconformity that separates sequences TA1.3 and TA2.1 occurs during the most profound sea-level fall of the Paleocene. A hiatus corresponding to this type 1 unconformity cannot be documented in the coastal plain and at Site 605 on biostratigraphic criteria (Olsson and Wise, 1987; Wise and van Hinte, 1987). Sequence TA1.3 on the cycle chart is possibly miscorrelated and may fall below the Hornerstown and Vincentown sequence. If so, the hiatus associated with the prominent reflector at Site 605 might be due to coalesced unconformities. An alternate explanation would place the profound sea level fall below sequence TA1.3 and replace the type 1 unconformity at the top of TA1.3 with a type 2 unconformity. The Hornerstown Formation might then correlate with sequence TA1.3 and the Vincentown Formation with sequence TA2.1.

The hiatus which separates the uppermost Paleocene sequence and the Vincentown Formation is also noted at Site 605 (Saint-Marc, 1987). Two type 1 unconformities correspond to this hiatus on the cycle chart of Haq and others (1987), so that this may be another case of coalesced unconformities in the Paleocene.

In the lower Eocene three hiatuses correspond with type 1 unconformities on the Haq and others (1987) cycle chart. They occur between sequences TA2.3 and TA2.4, between sequences TA2.6 and TA2.7, and between sequences TA2.9 and TA3.1. The hiatus between sequences TA2.9 and TA3.1 is also present at Site 605 and at Site 613 (Valentine, 1987). Elsewhere, this hiatus which occurs in Zone P9 has been identified in Libya (Barr and Berggren, 1981), in northwest Europe (Aubry, 1985), in Egypt (Abul-Nasr and Thunell, 1987), and in the California Coast Range (Berggren and Aubert, 1983). Four hiatuses are recognized in the middle Eocene, but only one of these, that at the top of sequence TA3.6, is associated with a type 1 unconformity. The other three, which separate sequences TA3.1 and TA3.2, TA3.3 and TA3.4, and TA3.4 and TA3.5, correspond with type 2 unconformities. The hiatus between TA3.1. and TA3.2 is also noted at Site 612 (Miller and Katz, 1987; Poag and Low, 1987).

Foraminiferal biofacies analysis and paleoslope analysis have proven successful in recognizing sea-level cycles because foraminifera respond readily to changes in sea level. The results of this study indicate that type 1 unconformities can be recognized in continental margins such as the New Jersey coastal plain and that some of them extend as far as the New Jersey slope. Several hiatuses apparently correspond with type 2 unconformities. All of the third-order cycles shown on the Haq and others (1987) cycle chart could not be identified. This may in part, be due to the discontinuous samples that were available for this study. Apparently, in at least two cases, cycles are not present because they were, most-likely, destroyed by erosion. Continuous sampling would facilitate a test to determine whether or not all the third-order cycles of Haq and others (1987), can indeed, be delineated.

ACKNOWLEDGEMENTS

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REGIONAL UNCONFORMITIES CORED ON THE NEW JERSEY CONTINENTAL SLOPE

C. WYLIE POAG AND DORIS LOW

U.S. Geological Survey, Woods Hole, Massachusetts 02543

ABSTRACT

On the basis of lithologic, foraminiferal, seismostratigraphic and downhole logging characteristics, we identified seven distinctive erosional unconformities at the contacts of the principal depositional sequences at Site 612 on the New Jersey Continental Slope (water depth 1404 m). These unconformities are present at the Campanian-Maestrichtian, lower Eocene-middle Eocene, middle Eocene-upper Eocene, upper Eocene-lower Oligocene, lower Oligocene-upper Miocene, Tortonian-Messinian, and upper Pliocene-upper Pleistocene contacts. The presence of coarse sand or redeposited intraclasts above six of the unconformities suggests downslope transport from the adjacent shelf by means of sediment gravity flows, which contributed in part to the erosion. Changes in the benthic foraminiferal assemblages across all but the Campanian-Maestrichtian contact indicate that significant changes in the seafloor environment, such as temperature and dissolved oxygen content, took place during the hiatuses.

Each identified unconformity can be traced widely on seismic reflection profiles and most have been identified from wells and outcrops on the coastal plain and other offshore basins of the U.S. Atlantic margin. Furthermore, their stratigraphic positions and equivalence to similar unconformities on the Goban Spur, in West Africa, New Zealand, Australia, and the Western Interior of the U.S., suggest that most contacts are correlative with the global unconformities and sea-level falls of the Vail depositional model.

INTRODUCTION

Publication of the Vail depositional model in which sequences are bounded by unconformities of interregional or even global extent (Vail and others, 1977; 1984), has focused renewed attention on gaps in the stratigraphic record (e.g., Schlee, 1984; Poag and others, 1985; Poag and Low, 1987). Continuously cored boreholes drilled along multichannel seismic reflection profiles can provide the essential documentation, lacking in Vail's papers, of relationships between seismic reflection, geophysical logging, lithologic, and faunal characteristics of such unconformities.

The initial shipboard identification of major unconformities at Site 612 on the New Jersey Slope (Fig. 1) indicated that they generally correspond with global unconformities and low sea levels of the Vail model, and that several are equivalent to unconformities documented on the Goban Spur (Poag and others, 1985). Also, they could be traced updip to the continental shelf, where they form part of the stratigraphic framework discussed by Poag and Schlee (1984) and Poag (1985a) (Fig. 2). This paper provides further documentation for the geological and geophysical characteristics of these unconformities and their enclosing strata, especially emphasizing their genesis, implications of paleoenvironmental change, and regional relationships. An analysis of sedimentological characteristics, derived from thin sections at several of these same unconformities, is presented by Cousin and Thein (1987).

METHODS OF FAUNAL ANALYSIS

Poag and others (1985) took two to four samples of 20 cc volume at irregular intervals above and below the unconformities at Site 548. We have used a more systematic approach to sampling at Site 612, attempting thereby to identify the "normal" (background) lithic and faunal characteristics of strata above and below the unconformable contacts and their associated disturbed zones. We attempted to maintain a 3-cm interval between each 20 cc sample, and to take seven samples above and seven below each contact. High-density sampling for other studies, however, prevented strict adherence to this plan. We also sampled as close to the contact (above and below) as possible. Microfossil samples were prepared by disaggregation in a warm Calgon solution (a commercial water softener), wet sieving on a 74 μm screen, and oven drying at 70°F.

A census of benthic foraminiferal assemblages was carried out on approximately 300 specimens per sample, derived from aliquots of the >74 μm size fraction, to determine the relative abundance of genera and species (see Poag and Low, 1987, Tables 1-7). The results may be compared with those of Miller and Katz (1987) and Katz and Miller (1987), who analyzed the >150 μm size fraction in some of the same core sections. We find that an assessment of the original community structure is more accurate when the small species (which are often predominant) are included in analyses. In many cases, those specimens that dominate the coarse sieve fractions are actually only a small portion (numbers of individuals) of the total assemblage. Sedimentological composition of the >74 μm fraction was also noted, and provides complementary information to shipboard interpretations.

We also examined the planktonic foraminiferal assemblages to refine the biozonations of Miller (1987) and Hart (1987), which were based on more

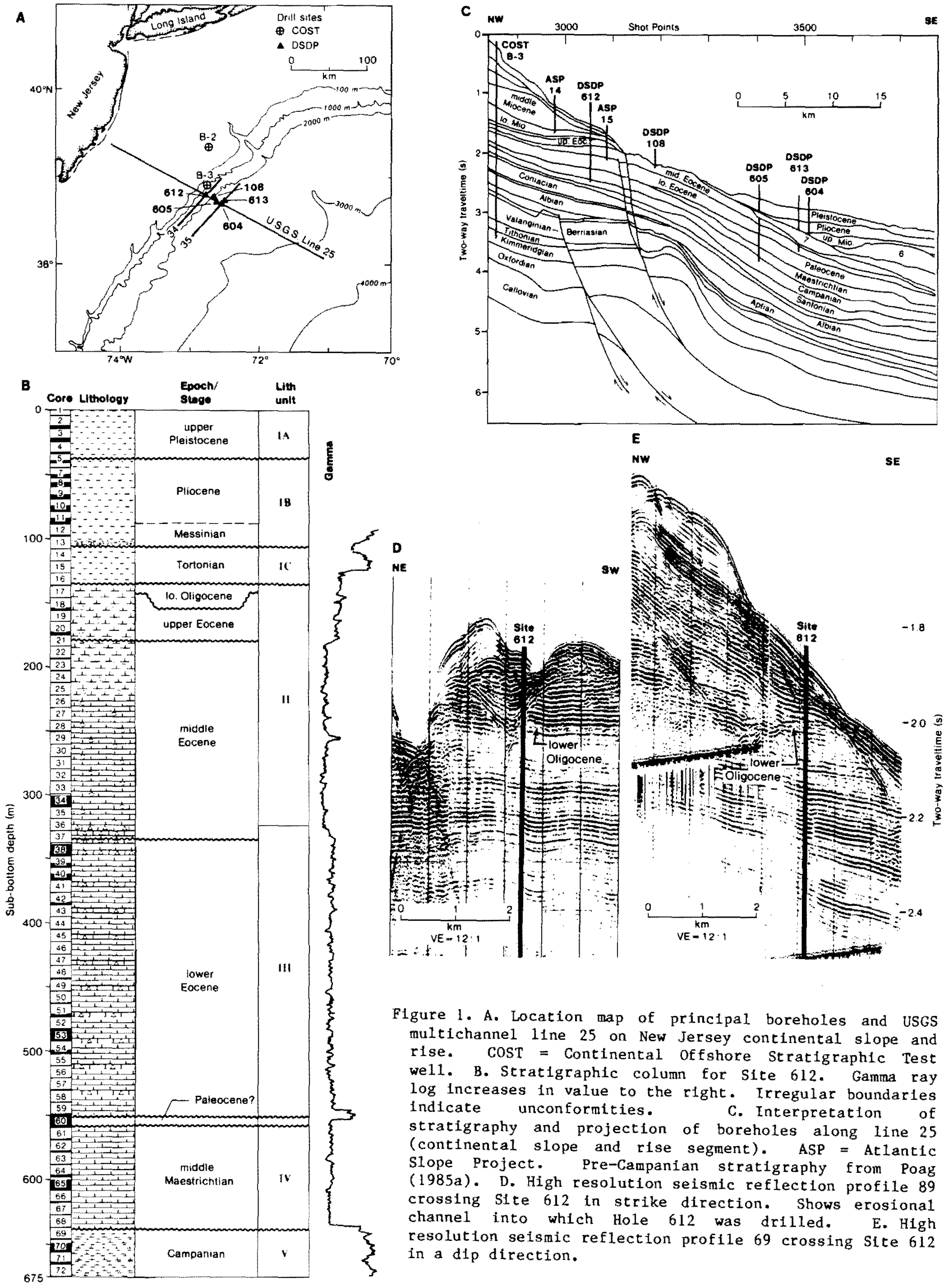

Figure 1. A. Location map of principal boreholes and USGS multichannel line 25 on New Jersey continental slope and rise. COST = Continental Offshore Stratigraphic Test well. B. Stratigraphic column for Site 612. Gamma ray log increases in value to the right. Irregular boundaries indicate unconformities. C. Interpretation of stratigraphy and projection of boreholes along line 25 (continental slope and rise segment). ASP = Atlantic Slope Project. Pre-Campanian stratigraphy from Poag (1985a). D. High resolution seismic reflection profile 89 crossing Site 612 in strike direction. Shows erosional channel into which Hole 612 was drilled. E. High resolution seismic reflection profile 69 crossing Site 612 in a dip direction.

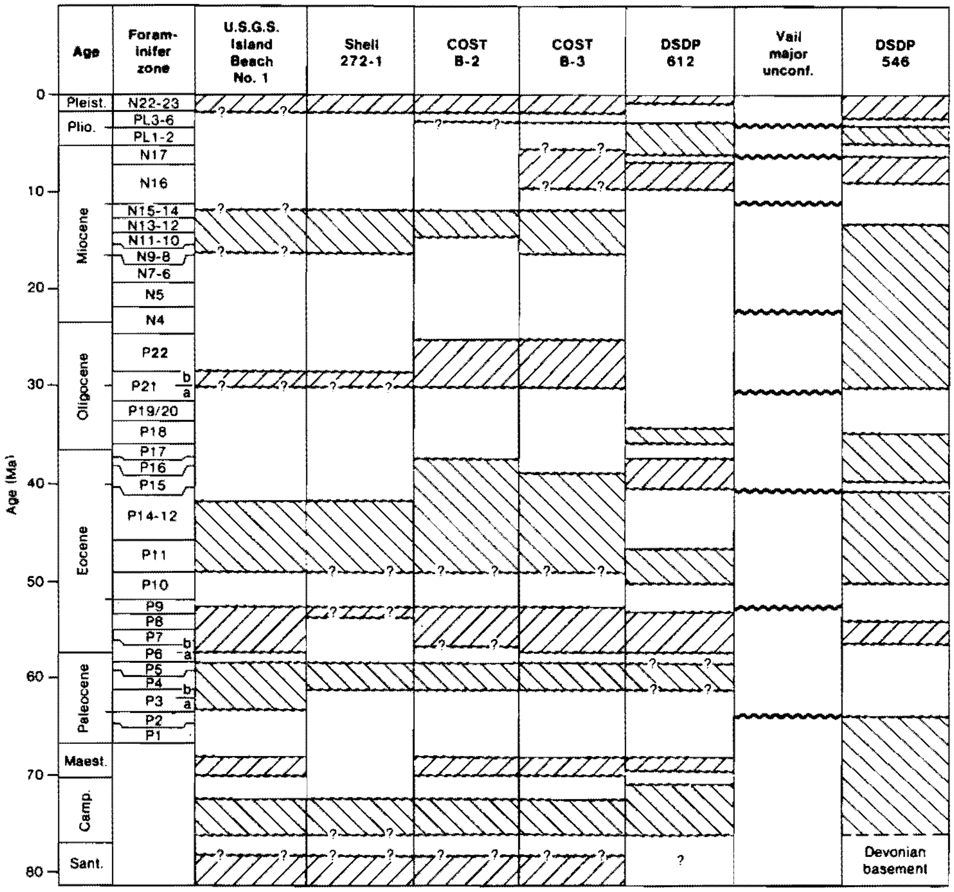

Figure 2. Chart showing relationships of global unconformities (wiggly horizontal lines) of the Vail depositional model to depositional sequences and stratigraphic gaps of the New Jersey continental shelf and slope and the continental slope of Ireland (DSDP 548). Data from Poag (1985a), Poag and others (1985), and Poag and Low (1985).

widely spaced samples and coarser size-fractions.

Our presentation proceeds chronostratigraphically from oldest to youngest levels in the cores, to emphasize the historical development of the New Jersey margin.

CAMPANIAN-MAESTRICHTIAN CONTACT

Lithologic Characteristics

The unconformable contact between Campanian and Maestrichtian strata is at approximately 8 cm in 612-69-3, where a concave scour surface separates dark gray to black, fissile, finely glauconitic, pyritic, laminated shale and chalk (below) from lighter gray, coarsely glauconitic (scattered, dark, distinctly larger grains) pyritic, marly, foraminiferal-nannofossil chalk (above) (Fig. 3; Poag and Low, 1987). Horizontal burrows, extending to at least 10 cm below the contact, are filled with the lighter gray, coarser, glauconitic sediment from above the contact.

The >74 µm sieved residues of samples between 53 cm and 18 cm, are planktonic foraminiferal oozes, containing small amounts of *Inoceramus* fragments, fish skeletal debris, ostracode valves, pyrite, and glauconite. Several indurated burrow casts in sample 612-69-3, 15-18 cm contain glauconite in abundance; this was apparently derived from above the contact, as glauconite is rare elsewhere in the residue.

A marked lithologic change just below the unconformity (612-69-3, 14-10 cm), is manifest by intense pyritization: euhedral aggregates as long as 2 cm are present along with many indurated burrow casts, large fragments (not prisms) of *Inoceramus*, benthic foraminifers, ostracodes, fish skeletal debris, and glauconite grains.

Just above the unconformable contact (612-69-3, 7-4 cm), the sediment also is indurated, but in this interval the cementing agent is calcite. The sample did not completely disaggregate, and microfossils, though abundant, are secondarily calcified. Glauconite grains also are abundant.

The uppermost sample examined (612-69-2, 131-127 cm) disaggregated completely, yielding an assemblage of abundant planktonic foraminifers, but they are poorly preserved due to secondary calcification. Glauconite is common, and burrow casts are filled with euhedral pyrite. A few *Inoceramus* prisms, ostracodes, and fragments of fish skeletons also are present.

Biochronology

The Campanian planktonic foraminiferal assem-

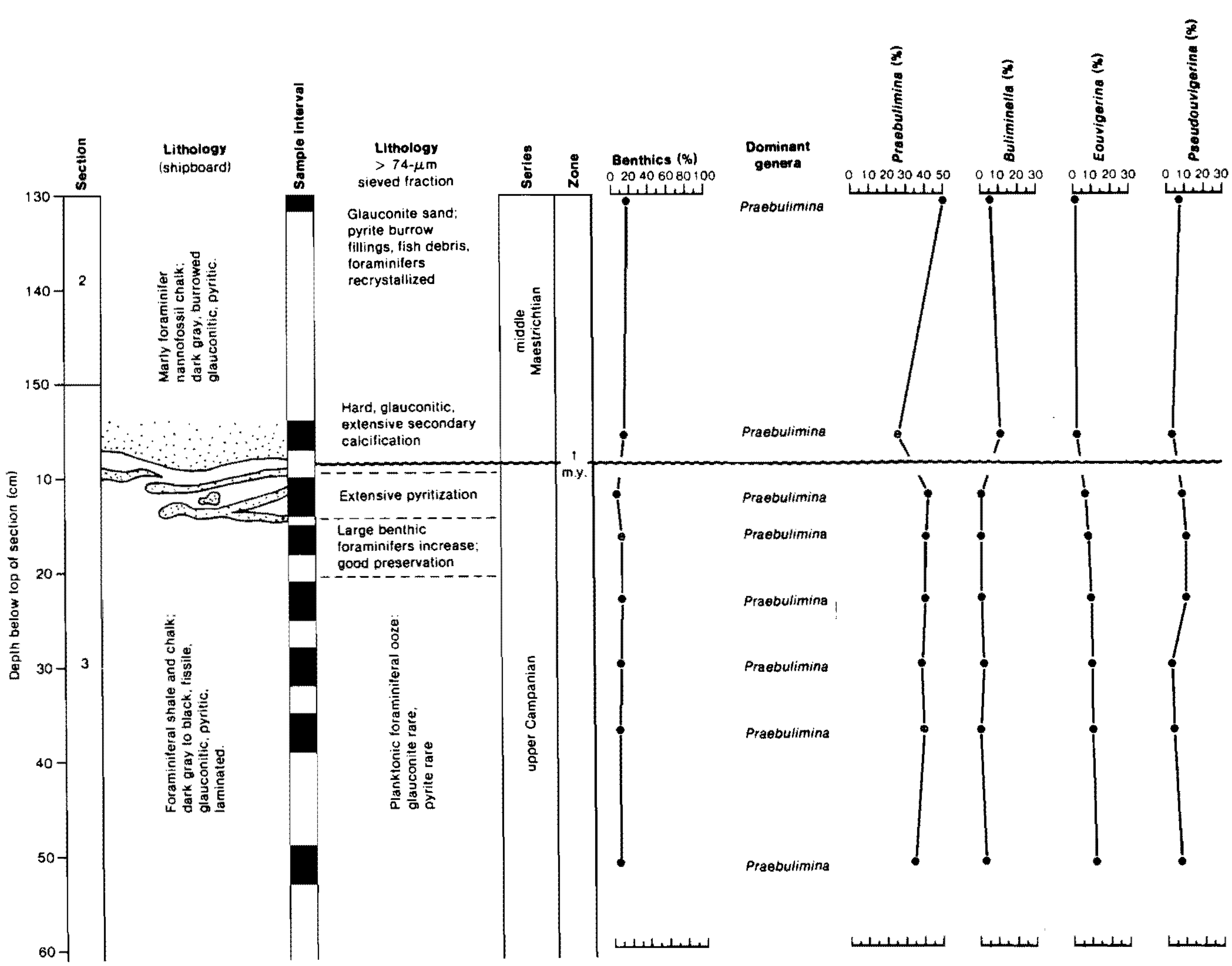

Figure 3. Chart showing lithic and benthic foraminiferal characteristics across the Campanian-Maestrichtian contact at Site 612. Sample intervals indicated by heavy black rectangles in column 4. Biochronology based mainly on planktonic foraminifers, supplemented by nannofossils. Length of hiatus represented by unconformity is indicated at the contact, within the column labeled "zone".

blage immediately below the unconformity (612-69-3, 14-10 cm) is abundant and diverse, containing, among other species, Rosita fornicata, R. patelliformis, Globotruncanita stuartiformis, Globotruncana ventricosa, G. orientalis, G. linneiana, G. arca, Archaeoglobigerina cretacea, and Globigerinelloides multispina. Many of these taxa span the upper Campanian and lower Maestrichtian elsewhere; others are restricted to the Maestrichtian in indigenous assemblages, and have been brought downward as burrow-fill. However, the presence of Kyphopyxa christneri, a typical Campanian form for this region (Poag, 1980), along with the absence of several diagnostic Maestrichtian species that are present above the contact, defines the age of this sample as Campanian.

Just above the unconformity (612-69-3, 7-4 cm), the foraminiferal assemblage is small, containing poorly preserved specimens of many of the same planktonic species noted below the unconformity. The presence of Kyphopyxa christneri and a specimen of Marginotruncana sp. indicates redeposition of older forms. Specimens of Gansserina gansseri appear to be indigenous, however, indicating a middle Maestrichtian age for this sample. A hiatus estimated to be at least 1 my is represented by the unconformity.

The highest sample examined in the Cretaceous section at Site 612-69-1 (124-120 cm) is also of middle Maestrichtian age, but contains a more complete and well-preserved suite of planktonic foraminifers, including Rosita contusa, Globotruncanella havanensis, G. petaloidea, and Globotruncana aegyptiaca.

Geophysical Characteristics

The Campanian-Maestrichtian contact is marked by significant shifts in both the sonic (interval transit time) and gamma-ray curves recorded by downhole logging (Fig. 4). The higher gamma-ray values in the Campanian section reflect a larger component of terrigenous clay as compared with the Maestrichtian section.

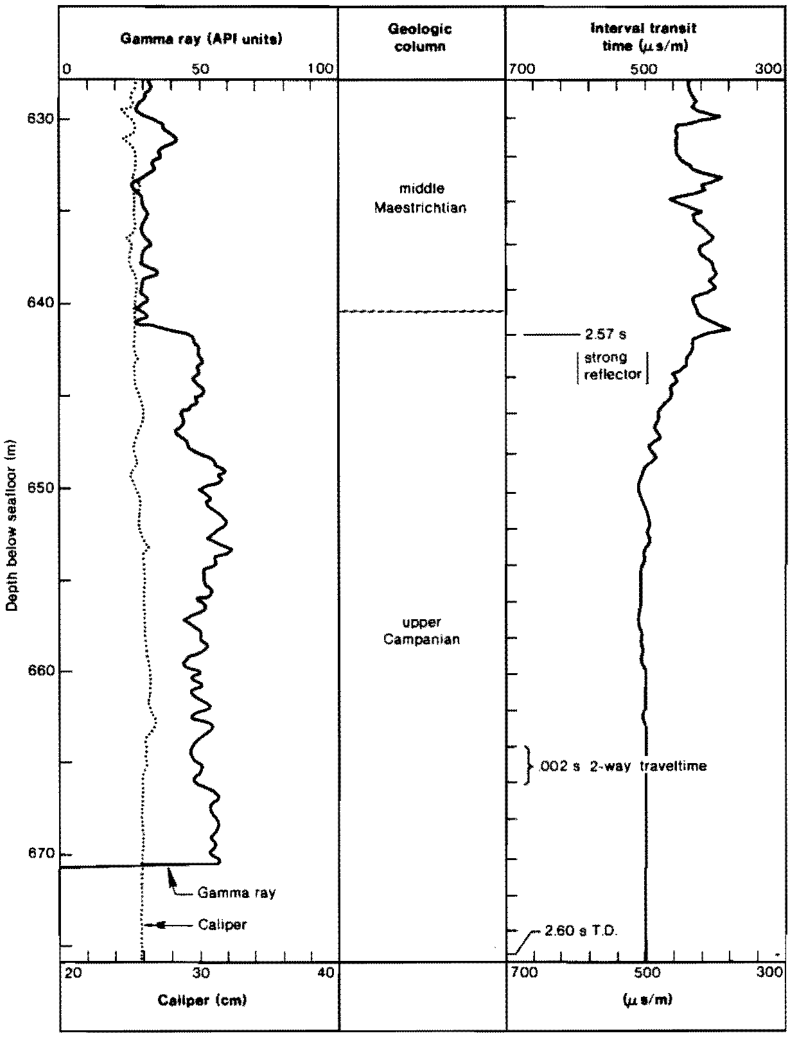

Figure 4. Logging characteristics across the Campanian-Maestrichtian contact.

Sonic velocities (reciprocal of interval transit time), on the other hand, are greater in the Maestrichtian section. The major inflection of the sonic log at 2.57 sec corresponds to a strong seismic reflector at this position on Line 25. The reflector can be traced updip to the shelf, and downdip to the rise, although its amplitude is diminished beneath the middle Eocene submarine outcrop belt (Fig. 6).

Benthic Foraminiferal Characteristics

The benthic foraminiferal assemblages undergo moderate change across the Campanian-Maestrichtian contact (Fig. 3). The percentage of benthic foraminifers increases from 5-9 below the contact to 11-13 above. *Praebulimina* dominates every sample, ranging from 36% to 42% below the contact and from 27.5% to 48.7% above. The lowest value (2.8%) is associated with the indurated layer just above the contact. At the same place, *Buliminella* increases from 0 to 1.2% (below) to 4.5 to 11.9% (above), and the highest value for this genus is immediately above the contact. A reciprocal trend is present in *Eouvigerina* and *Pseudouvigerina*, whose values decrease from 7.1 to 12.4% and 2.8 to 10.6% respectively, below the contact, to 0.9 to 1.3% and 3.2 to 6.5% above. The number of benthic genera remains nearly constant (22-29 below; 21-27 above). Additional samples in the Maestrichtian interval are needed to strengthen these conclusions.

Regional Relationships

The Campanian-Maestrichtian unconformity can be traced widely on seismic reflection profiles beneath the shelf, slope, and rise of the middle Atlantic margin of the U.S. and has been documented on the shelf and upper slope in the COST B-2 and COST B-3 wells and at the shoreline in the USGS Island Beach No. 1 well (Poag, 1985a). On the coastal plain, it has been recorded by by Owens and Gohn (1985) and, indirectly, by Olsson and Nyong (1984). This unconformity falls within the upper part of supercycle Kb of the Vail model (Vail and Mitchum, 1979; Fig. 2), an interval whose details have just recently been published by Haq and others (1987). It is a persistent phenomenon on the U.S. Atlantic margin (Poag and Schlee, 1984), and elsewhere, it has been documented in the western interior of the U.S.

(Weimer, 1984) and from central West Africa (Seiglie and Baker, 1984).

LOWER EOCENE (YPRESIAN) - MIDDLE EOCENE (LUTETIAN) CONTACT

Lithologic Characteristics

The contact between the Ypresian and Lutetian stages at Site 612 is more complex than the Campanian-Maestrichtian contact. Instead of a sharply defined scour surface, burrowing across the contact has mixed two biosiliceous chalk units (Fig. 5; Poag and Low, 1987). Below the contact (which is at ~81 cm in core 612-37-3), the dominant lithology is dark, yellowish brown chalk. It is penetrated by numerous horizontal burrows filled with light greenish gray, coarsely glauconitic chalk derived from above the contact. The lighter chalk above the contact is thinly laminated, frequently burrowed, and contains occasional layers of glauconite, as well as scattered coarser grains of glauconite. In addition, there are hard clasts of redeposited, dark gray, Upper Cretaceous chalk at several places in the light gray unit, indicating that this is part of a debris flow deposit. A 3-cm-long clast is present between 78 and 75 cm; another was sampled at 40-36 cm.

In the >74 μm sieved fraction, the lithology below the unconformity is varied. The lower two samples (612-37-3, 135-132 cm and 128-126 cm) contain hard fragments of tan and white chalk that did not disaggregate completely during preparation. These fragments have a vuggy surface texture caused by hollow impressions of foraminiferal tests. The free foraminiferal specimens are poorly to moderately preserved. Fish skeletal debris is also a common component of this residue.

The dominant lithology in the next samples (612-37-3, 124-105 cm) is tan porcellanitic chalk. The sediment hardly disaggregated at all and the few foraminiferal specimens released are barely recognizable as foraminifers.

Between 612-37-3, 100 cm and the unconformity (~81 cm), all three samples disaggregated completely, yielding a well-preserved planktonic foraminiferal ooze. Additional elements of the residue include echinoid spines, fish skeletal debris, glauconite grains, and euhedral pyrite.

The section above the unconformity is even more variable than that below (Fig. 5). The first sample (612-37-3, 79-75 cm) was taken from

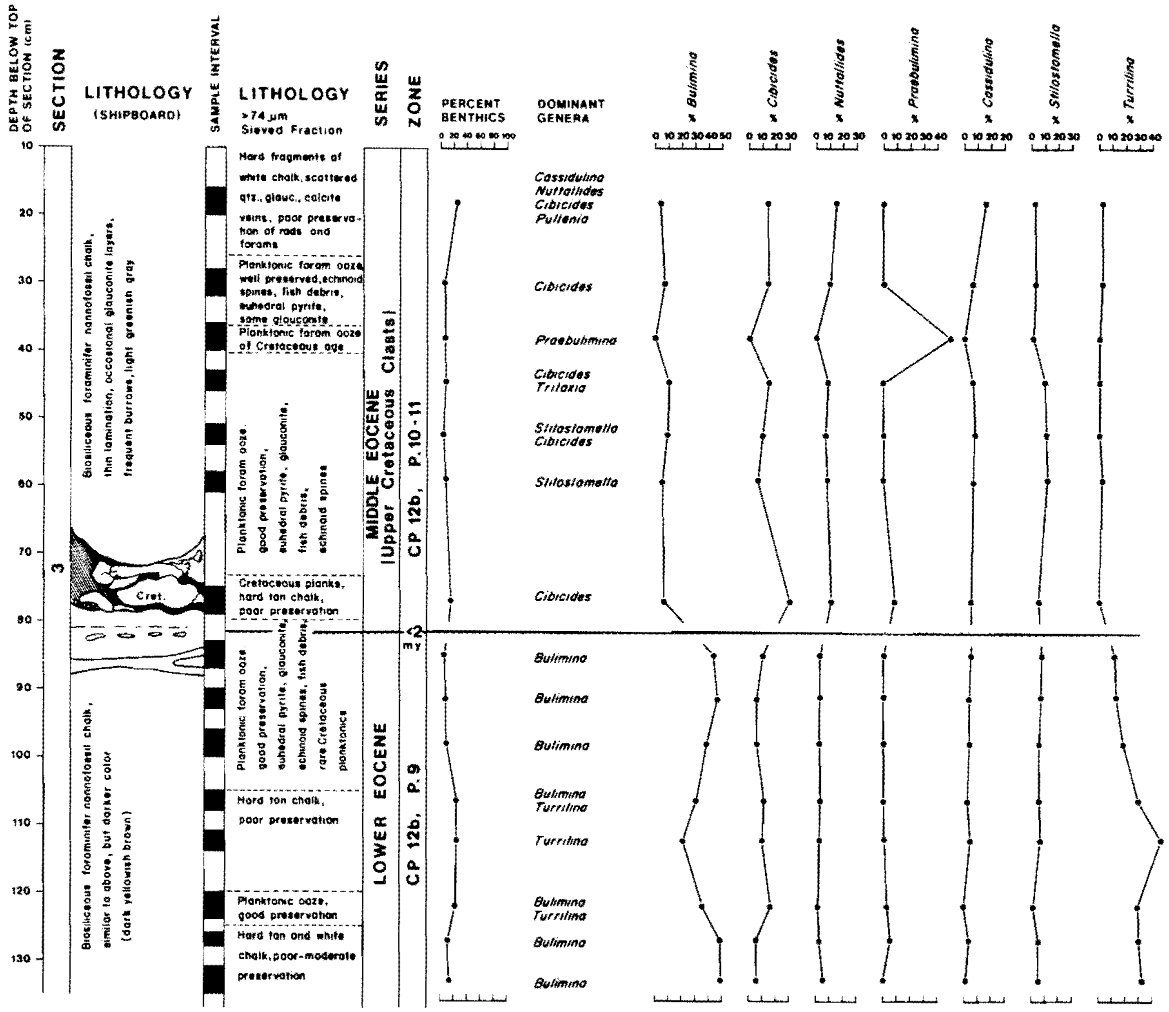

Figure 5. Chart showing lithic and benthic foraminiferal characteristics across the lower Eocene - middle Eocene contact at Site 612. See explanation of figure 4 for additional information.

the hard Upper Cretaceous intraclast, composed of tan chalk that resisted disaggregation. Preservation of planktonic foraminifers is poor, though numerous hollow impressions of them create vuggy surfaces on the chalk fragments.

Softer sediment in the three samples from 612-37-3, 61 cm to 43 cm disaggregated completely, leaving an excellently preserved planktonic foraminiferal ooze, along with echinoid spines, fish skeletal debris, glauconite grains, and euhedral pyrite.

The sample at 612-37-3, 40-36 cm also disaggregated, but the resultant planktonic foraminiferal ooze is entirely of Late Cretaceous age, indicating derivation from a redeposited clast. Along with the excellently preserved foraminifers, are echinoid spines, fish skeletal debris, mica flakes, and euhedral pyrite.

Another soft sediment sample at 612-37-3, 31-28 cm, contained white chalk, which disaggregated almost completely, yielding a moderately-well-preserved planktonic foraminiferal ooze of middle Eocene age. This residue includes echinoid spines, fish skeletal debris, a few ostracodes, some glauconite grains, and euhedral pyrite. The youngest sample examined (612-37-3, 20-16 cm) was another firm white chalk that only partly disaggregated, yielding large chalky chips and a few grains of glauconite and quartz. The surfaces of the chips are vuggy where planktonic foraminifers were imbedded. A few free specimens of radiolarians and planktonic foraminifers are present, but poorly preserved, accompanied by fish skeletal debris and fragments of calcitic vein fillings.

Biochronology

A sample below the unconformity at 612-37-3, 100-96 cm, contains a well preserved in situ assemblage of lower Eocene (Ypresian) planktonic foraminifers, including Morozovella caucasica, M. aragonensis, Acarinina matthewsae, A. broedermanni, A. pentacamerata, Subbotina crociaperta, and Muricoglobigerina senni, which indicate Zone P. 9 (Blow, 1979). A few specimens of redeposited Upper Cretaceous planktonic species are also present here. A similar association of in situ and redeposited forms is present in the other samples below the unconformity.

In sample 612-37-3, 61-58 cm, above the unconformity, the planktonic foraminiferal assemblage contains Morozovella aragonensis, Subbotina frontosa, and Muricoglobigerina soldadoensis; at 46-43 cm and 31-28 cm, there are, in addition, Globigerinoides(?) higginsi, Acarinina bullbrooki, and Truncorotaloides quetra. These assemblages indicate that the indigenous taxa represent zones P. 10-11 of the middle Eocene (Lutetian; Blow, 1979), but that specimens from Zone P. 8 or older (T. quetra; lower Eocene) have been incorporated by redeposition. The length of the hiatus represented by the unconformity can not be directly determined from the foraminifers because only part of a zone (undetermined amount) is missing (this is also true for the nannofossil record; Fig. 5; see Valentine, 1987). Miller and Katz (1987) used the change in sedimentation rate at this contact to estimate a 1-5 my hiatus (most of hiatus appears to represent basal Zone P. 10). However, the nannofossil zone CP 12b, which encompasses the unconformity, is only 2 my in total length (Berggren and others, 1985), so the hiatus must be less than 2 my.

Geophysical characteristics

The sonic log records a sharp upward increase in velocity at the Ypresian-Lutetian contact (Fig. 6; at 2.26 sec on the integrated sonic log). It correlates with a strong seismic reflector at this depth on Line 25 (Poag, 1987). An upward velocity decrease at 2.25 sec is associated with the top of a zone of silica diagenesis (porcellanite "front"). The impedance contrast causing the high-amplitude seismic reflector on line 25 is probably a result of the close proximity of the diagenetic front and the unconformity.

The gamma-ray log also indicates a significant decrease in clay content above the unconformity, as the curve deflects from consistently higher values below the contact, to lower values above.

Benthic Foraminiferal Characteristics

Distinctive changes in the benthic foraminiferal assemblages across the Ypresian-Lutetian contact are expressed most clearly by the upward reductions of Bulimina and Turrilina at the contact. Below the contact, Bulimina is predominant or co-predominant in seven of eight samples (Fig. 5; Table 2 of Poag and Low, 1987). Above the contact Bulimina is consistently less than 10.5% (range = 3.8 to 10.2%; mean = 5.9%) and is no longer predominant. Turrilina is predominant or co-predominant in three samples below the contact (range = 10.8 to 40%; mean = 24.4%), declines upward from a peak at 612-37-3, 114-111 cm, to 10.9% just below the contact, and essentially disappears above it (present in 3 samples, ranging from 1.9% to 3.5%). Nuttallides, on the other hand, increases above the contact (1.6 to 4.1% below; 7.9 to 14.6% above).

Among the predominant genera of in situ samples above the contact are Stilostomella, Cibicides, Cassidulina, Tritaxia, Nuttallides, and Pullenia. Among the redeposited Cretaceous assemblages above the contact, Cibicides, and, especially, Praebulimina are predominant (Praebulimina = 49.8% at 612-37-3, 40-36 cm). The Praebulimina values are similar to those in the Campanian-Maestrichtain section described above, from which the clast probably was derived. Benthic percentage varies little throughout the section.

In a separate study, Miller and Katz (1987) also demonstrated a major change in the benthic foraminiferal assemblage across this contact, manifested by the abrupt appearance above it of

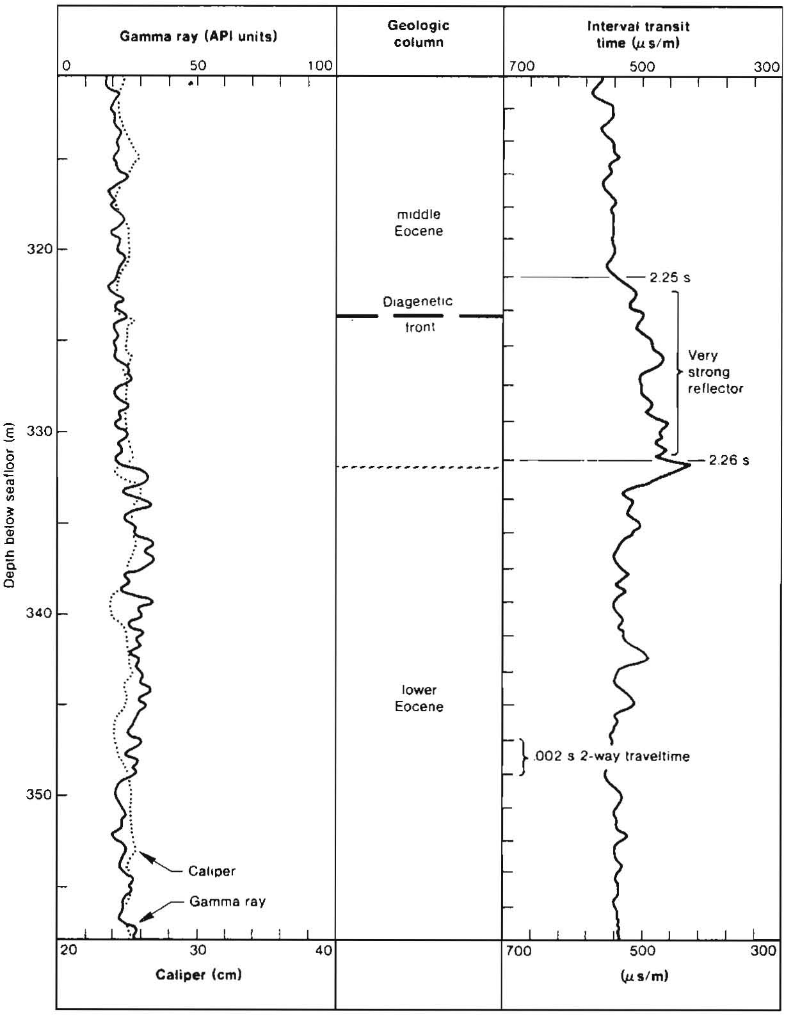

Figure 6. Downhole logging characteristics across the lower Eocene - middle Eocene contact.

10 species.

Regional Relationships

The Ypresian-Lutetian contact is widely recognizable on seismic reflection profiles crossing the New Jersey margin and has been documented as far shoreward as the USGS Island Beach No. 1 well (Poag, 1985a). Ward (1984) and Ward and Strickland (1985) also noted it as a widespread unconformity among outcropping formations of the coastal plain, from South Carolina to Maryland. It also crosses the sites of the COST B-2 and B-3 wells, where the accompanying silicification impairs foraminiferal preservation and obscures the biozonation (Poag, 1985a). The contact also was cored downdip at Sites 605 and 613, and can be traced on seismic profiles well seaward under the continental rise (Poag, 1985b, 1987). Popenoe (1985) also recorded this unconformity on the continental slope of Georgia, where it is exhibited on seismic reflection profiles as an erosional swath cut by the paleo-Gulf Stream.

Thorne and Watts (1984) suggested that the seismic reflector associated with this contact at Site 612 might be due to the porcellanite alone, and thus could be a diachronous reflector. However, at both sites 612 and 613, the porcellanite front is too close to the unconformity (2-8 m) to be recognized as a separate reflector on our seismic profiles. The fact that the reflector retains its position at the Ypresian-Lutetian contact for hundreds of kilometers from the shoreline to the continental rise argues against significant diachronism.

The Ypresian-Lutetian contact on the New Jersey margin corresponds to the contact between supersequences Ta and Tb of the Vail depositional model (Vail and Mitchum, 1979). The bounding major global unconformity is placed by these authors within Zone P. 9, and the concurrent sea-level drop is estimated to have taken place at 49.5 Ma. On the Berggren time scale (Berggren and others, 1985) this would be 52.6 Ma. An equivalent Ypresian-Lutetian unconformity (assuming a dating accuracy of no greater than ~1 my) was documented by Poag and others (1985) at Sites 548 and 549 on the Goban Spur (continental slope of Ireland), where it is associated with turbidite deposition. The Ypresian-Lutetian unconformity also is one of the most thoroughly documented supersequence boundaries outside the western North Atlantic. It has been recorded in Libya (Barr and Berggren, 1981), northwestern Europe (Aubry, 1985), California (Berggren and Aubert, 1983), Australia (Steele, 1976; McGowran, 1979; Quilty, 1980), and New Zealand (Loutit and Kennett, 1981a, 1981b). Such widespread erosion of equivalent stratigraphic sections suggests a common cause, such as the change in sea level postulated for the Vail model.

MIDDLE EOCENE (LUTETIAN) - UPPER EOCENE (PRIABONIAN) CONTACT

Lithologic Characteristics

The contact between the Lutetian and Priabonian sections is marked by an irregular scour surface at 612-21-5, 115 cm, which separates medium gray, biosiliceous, sparsely burrowed, nannofossil ooze (below), from dark greenish gray, glauconitic, quartz sand (Fig. 7). The 1.5-cm-thick sand is overlain by ~17 cm of dark brownish gray, biosiliceous, nannofossil ooze, which contains distorted inclusions of lighter gray sediment. Above this is a thick interval of light brownish gray, mottled, biosiliceous, nannofossil ooze.

The >74 μm sieved fraction of the six samples below the unconformity contains a microfossil ooze of chiefly radiolarians and planktonic foraminifers. Diatoms, sponge spicules, and fish skeletal debris are also present, along with a few shards of glass, which Thein (1987) attributes to a bolide impact. No sample was taken in the sand immediately above the contact, where Thein (1987) and Cousin and Thein (1987) identified microtektites in a matrix of angular quartz, feldspar, glauconite, and clay clasts; they interpret this layer as a reworked impact breccia. At 612-21-5, 109-105 cm, in the dark brownish gray ooze, a very small residue is dominated by radiolarians and euhedral pyrite; foraminifers are sparse. At 612-21-5, 101-97 cm, a residue from the top of the dark ooze contains hard lumps of tan clay and abundant framboidal pyrite, radiolarians, diatoms, and planktonic foraminifers. Echinoid spines, sponge spicules, and fish skeletal debris are less abundant. A few glauconite and quartz grains also are present. The tan clay is partly soluble in dilute HCl, but leaves a residue of thin brown flakes.

From 612-21-5, 94 cm to 65 cm, residues from 4 samples of the lighter brownish gray ooze

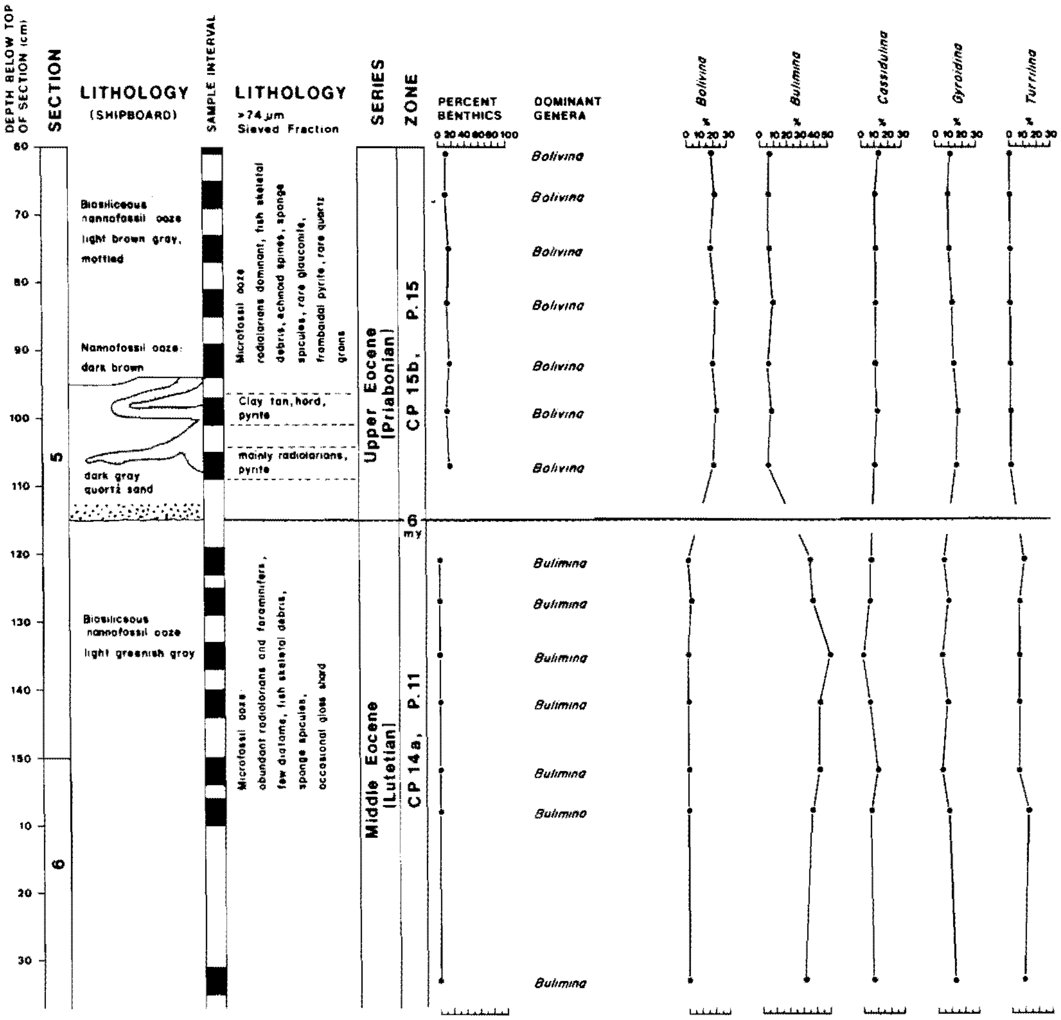

Figure 7. Chart showing the lithic and benthic foraminiferal characteristics across the middle Eocene - upper Eocene contact at Site 612.

(Fig. 7) have approximately the same constituents as below, lacking only the tan clay lumps. The uppermost sample (612-21-5, 61-57 cm) is similar, but diatoms are notably scarce.

Biochronology

A diverse assemblage of well-preserved planktonic foraminifers from below the unconformity includes: *Hantkenina mexicana dumblei*, *Globigerinatheka index*, *Morozovella lehneri*, *M. spinulosa*, *Acarinina densa*, *A. broedermanni*, *A. matthewsae*, *Subbotina frontosa*, *Globigerinoides(?) higginsi*, and *Hastigerina(?) bolivariana*. This assemblage is diagnostic of Zone P. 11 (Blow, 1979), but includes a few younger forms (*Hantkenina alabamensis*, *Truncorotaloides topilensis*, *T. rohri*) that probably came from the burrows.

The first diagnostic assemblage from above the unconformity, at the top of the dark ooze (612-21-5, 101-97 cm), is sparse, but contains *Globigerina cerroazulensis cerroazulensis*, *G. linaperta*, *Dentoglobigerina galavisi*, *Globorotaloides suteri(?)*, and *Globigerinatheka semiinvoluta*. This assemblage belongs to Zone P.15. Redeposition of middle Eocene specimens into this upper Eocene section is manifest by the presence of *Morozovella lehneri*. These assemblages indicate that the hiatus represented by the Lutetian-Priabonian unconformity is approximately 6 my.

Geophysical Characteristics

The gamma-ray curve displays a distinct upward increase in value across the Lutetian-Priabonian contact, coincident with a moderate decrease in sonic velocity at 2.09 sec on the integrated sonic log (Fig. 8). This depth corresponds to the position of a strong unconformable reflector that crosses Site 612 along Line 25, and can be easily traced updip and downdip from the borehole (Poag, 1987).

Benthic Foraminiferal Characteristics

Striking changes in the benthic foraminiferal assemblages take place across the Lutetian-Priabonian contact at Site 612 (Fig. 8; Table 3 of Poag and Low, 1987). The benthic percentage increases upward from 4-7% below the contact to 12-17% above it. At the same level, *Bulimina*, which is predominant below the contact (30.5-52.1%), is replaced by *Bolivina* (18.1-21.9%) above the contact. Ancillary changes include the near absence of *Turrilina* above the contact, although it is persistent (4.8-11.0%) below. *Cassidulina* and *Gyroidina*, on the other hand, are consistently more abundant above the contact.

The numbers of benthic genera do not vary significantly across the contact, and are similar to those near the Ypresian-Lutetian unconformity.

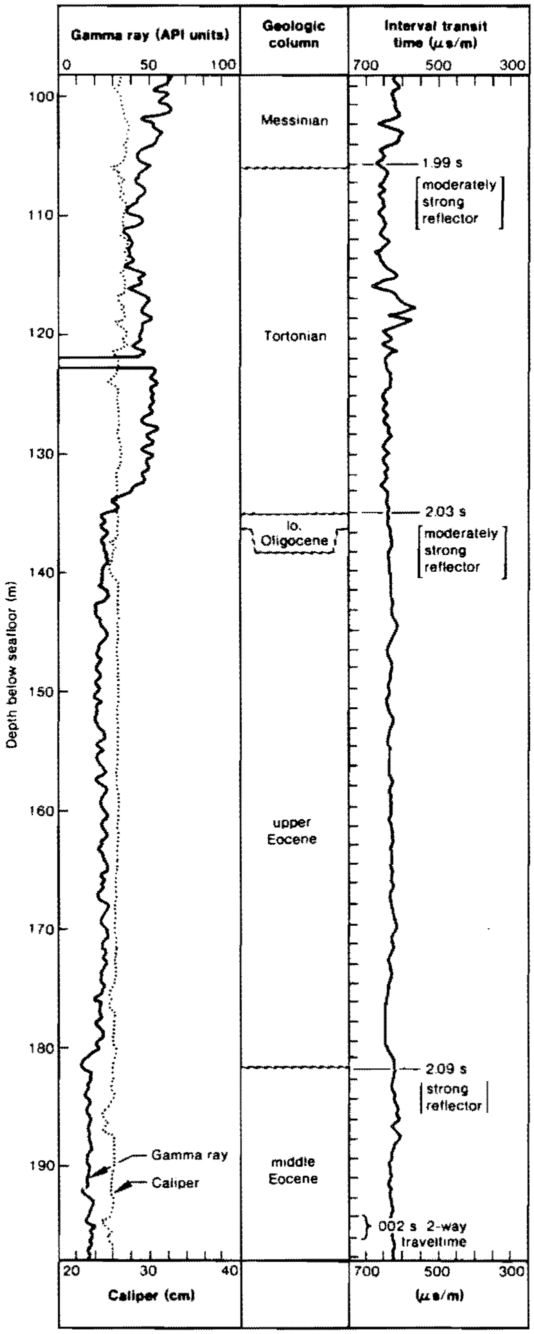

Figure 8. Downhole logging characteristics across the middle Eocene - upper Eocene, upper Eocene - lower Oligocene, lower Oligocene - upper Miocene, and Tortonian - Messinian contacts at Site 612.

Regional Relationships

The seismic expression of this contact is widely traceable throughout the New Jersey margin (Poag, 1985a, 1985b, 1987). It has been documented in several shelf wells, and is well known across the adjacent coastal plain (Ward, 1984; Ward and Strickland, 1985). It is particularly notable basinward from Site 612, where the unconformity is exposed at the seafloor as a broad submarine outcrop of middle Eocene chalk (Hollister, Ewing and others, 1972;

Robb and others, 1983; Poag, 1985b; Hampson and Robb, 1984; Farre, 1985; Farre and Ryan, 1985). Farther seaward it is onlapped by Tertiary and Quaternary sequences of the continental rise (Tucholke and Mountain, 1979; Mountain and Tucholke, 1985; Poag, 1985b; Poag and Mountain, 1987). Popenoe (1985) noted that this unconformity is expressed on seismic profiles off North Carolina as a seaward erosional swath of the paleo-Gulf Stream.

The Lutetian-Priabonian contact at Site 612 corresponds to the contact between supersequences Tb and Tc of the Vail depositional model (Vail and Mitchum, 1979), which is placed between planktonic foraminiferal zones P. 14 and P. 15 (Fig. 7). It coincides with a major sea-level fall and global unconformity, which Vail and Mitchum date at 40 Ma. Berggren and others (1985) would place the P. 14-P. 15 contact at 40.2-41.2 Ma. Snyder and Waters (1985) examined this stratigraphic interval on the Goban Spur, but did not report an unconformity associated with the P. 14-P. 15 boundary. Poag and Low (1987) reexamined the zonal boundary interval at Site 548 and found a distinct erosional contact at 548A-18-2, 25 cm, between the sample assigned by Snyder and Waters to P. 14 (548A-18-2, 65-68 cm; having abundant *Truncorotalia topilensis*) and a sample at (548A-18-1, 110-112 cm (P. 15; having no *T. topilensis*). An unconformity at this contact also has been recorded from New Zealand (Loutit and Kennett, 1981a, 1981b) and from Australia (Steele, 1976; McGowran, 1979; Quilty, 1980).

UPPER EOCENE (PRIABONIAN) - LOWER OLIGOCENE (RUPELIAN) CONTACT

Lithologic Characteristics

Priabonian strata consist of grayish, yellow green, homogeneous, biosiliceous, foraminiferal, nannofossil ooze, which extends upward from core 612-17-1 to the top of the core catcher of 612-16 (Fig. 9; Poag and Low, 1987). Section 7 and basal section 6 of 612-16 comprise distinctly lighter-gray, biosiliceous, foraminiferal, nannofossil ooze. There is no transition between the two lithologies, and no contact was recovered; it apparently was lost between 612-16 core catcher and the base of 612-16-7.

The washed residues of the >74 µm size

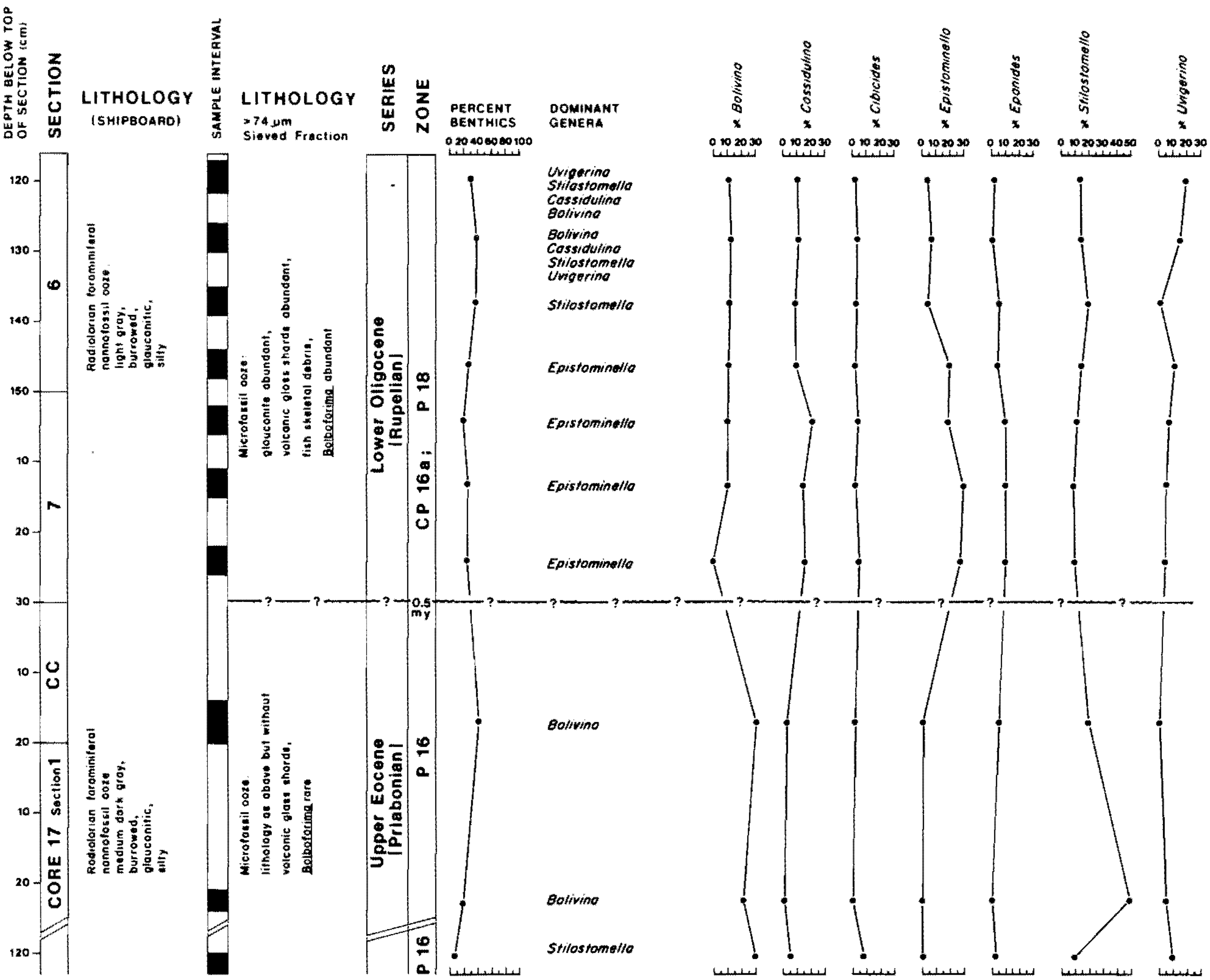

Figure 9. Chart showing lithic and benthic foraminiferal characteristics across the upper Eocene - lower Oligocene contact at Site 612.

fraction are more diagnostic than shipboard descriptions and photographs for separating these sequences (Fig. 9). The lighter gray Oligocene ooze contains abundant volcanic glass shards, which are completely absent from the grayish, yellow green ooze below the section 612-16-7.

Biochronology

The foraminiferal assemblage of core 612-16-7 is a typical early Oligocene planktonic association as described in more detail in the following section. On the other hand, the core catcher of 612-16 and section 1 of 612-17 contain Hantkenina alabamensis, a diagnostic late Eocene (Zone P. 16-lower P. 17) marker.

Geophysical Characteristics

The Oligocene interval is so thin that its upper and lower contacts cannot be distinguished as separate reflectors on Line 25 (see discussion of Oligocene-Miocene contact below). Neither can a significant gamma ray change be observed (Fig. 8). However, the interval is clearly distinguishable on high resolution profiles 89 and 69, at the intersection of which Site 612 was drilled (Fig. 1).

Benthic Foraminiferal Characteristics

The benthic foraminiferal assemblage changes markedly between the two lithologies as shown on Figure 9 (Table 4 of Poag and Low, 1987). Epistominella is consistently predominant in the lower Rupelian section, whereas Bolivina predominates in the upper Priabonian. Bolivina is consistently more abundant below than in any samples above the core catcher of 612-16. Conversely, Cassidulina and Epistominella are important in the Rupelian, but sparse in the Priabonian. Benthic percentage is higher in the Rupelian.

Regional Relationships

This contact is not well documented anywhere on the New Jersey margin. At ASP 15 it was not

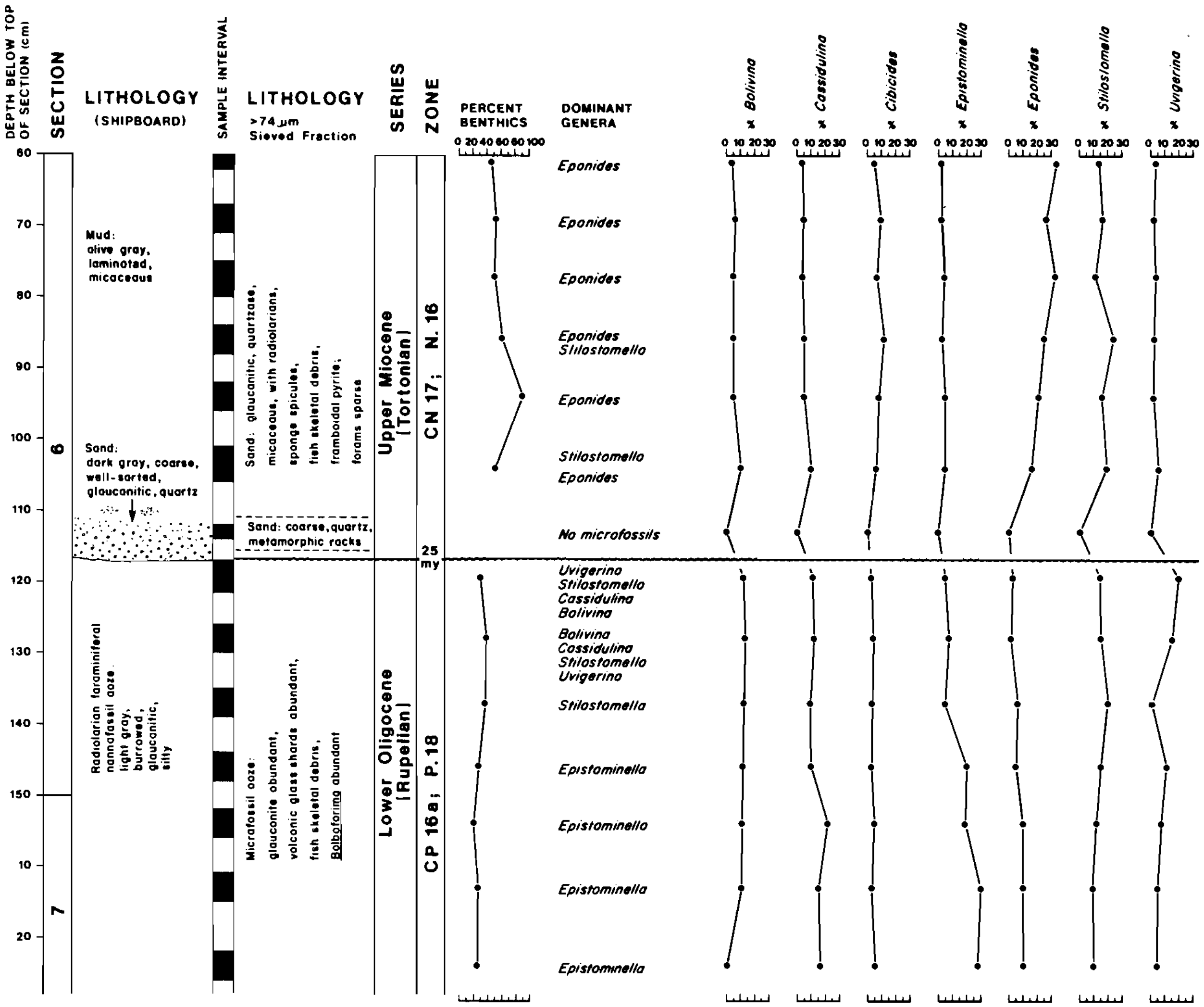

Figure 10. Chart showing the lithic and benthic foraminiferal characteristics across the lower Oligocene - upper Miocene contact at Site 612.

recovered, although lower Oligocene and upper Eocene strata were penetrated. Elsewhere, across the shelf and on the coastal plain, lower Oligocene strata were removed at most locations by late Oligocene erosion (Olsson and others, 1980; Poag, 1985a, 1987). However, an equivalent upper Eocene-lower Oligocene unconformity has been recorded in the Southeast Georgia embayment at ASP Site 5 and in the COST GE-1 well, (Poag and Hall, 1978; Popenoe, 1985).

LOWER OLIGOCENE (RUPELIAN) - UPPER MIOCENE (TORTONIAN) CONTACT

Lithologic Characteristics

The contact between the Rupelian and Tortonian sections at Site 612 is marked by a sharply defined scour surface at 612-16-6, 116 cm, which separates light gray microfossil ooze below, from dark gray, well-sorted, coarse, quartzose sand (5 cm thick) above (Fig. 10; Poag and Low, 1987). The coarse sand grades upward into a 1-cm interval of finer sand, which is overlain by olive gray, faintly laminated mud. We interpret the sand-to-mud section to be a turbidite.

In the >74 μm sieved fraction, the light gray ooze below the unconformity contains a rich microfossil assemblage of chiefly planktonic foraminifers, radiolarians, and *Bolboforma* (Poag and Karowe, 1987), accompanied by abundant glauconite grains and shards of volcanic glass. Fish skeletal debris and sponge spicules are accessory elements.

The residue of the sand at 612-16-6, 114-112 cm is barren of microfossils, consisting of coarse to fine quartz grains, fine glauconite grains, and metamorphic rock fragments. The residues of the six samples from the overlying mud contain fine quartz sand, mica flakes, glauconite grains, lignitic plant fragments, framboidal pyrite, radiolarians, sponge spicules, and fish skeletal debris; foraminifers are scarce to moderately abundant.

Biochronology

The rich assemblage of well-preserved planktonic foraminifers below the unconformable contact contains *Globigerina ampliapertura*, *G. gortanii*, *G. increbescens*, *G. tapuriensis*, *G. praeturritilina*, *G. praebulloides*, *G. gemma*, *Dentoglobigerina galavisi*, *D. praesepis*, *Globorotaloides suteri*, and *Catapsydrax unicava*, an assemblage assignable to lower Oligocene (Rupelian) Zone P. 18. Redeposited middle Eocene planktonics, such as *Truncorotaloides rohri*, are common in these assemblages, attesting to exposures of middle Eocene strata updip. Other distinctive elements of this assemblage are large numbers of *Bolboforma irregularis* (Poag and Karowe, 1987).

Above the barren turbidite sand that marks the unconformity, sparse planktonic foraminiferal assemblages contain *Globorotalia juanai*, *G. menardii*, *G. scitula*, *Globigerinoides obliquus*, *Globigerina decoraperta*, *G. nepenthes*, *G. bulloides*, *Neogloboquadrina continuosa*, *N. acostaensis*, *Orbulina universa*, *Globoquadrina dehiscens*, *Sphaeroidinellopsis seminulina*, and *Dentoglobigerina altispira*. This assemblage is assignable to upper Miocene Zone N. 16 of the Tortonian Stage. Thus, a hiatus of approximately 25 my is represented by the unconformable Rupelian-Tortonian contact.

The presence of a few redeposited specimens of *Globorotalia siakensis*, *G. praefohsi*, *G. acrostoma*, *Catapsydrax unicavus*, and *C. dissimilis*, indicate that middle Miocene and lower Miocene to upper Oligocene strata were exposed updip.

Geophysical Characteristics

On the downhole geophysical log, gamma-ray values increase markedly above the unconformity (Fig. 8), reflecting the abundance of terrigenous clay in the Tortonian mud. At the same level (2.03 sec on the integrated sonic log), a slight decrease in sonic velocity correlates with a moderately strong seismic reflector on line 25 (Poag, 1987).

Benthic Foraminiferal Characteristics

The most striking change in the benthic foraminiferal assemblages in this section is their absence (along with all other microfossils) in the coarse sand immediately above the contact (Fig. 10; Table 4 of Poag and Low, 1987). This absence is followed by a marked increase in the percentage of benthic foraminifers (44-93%; mean = 56%) in the overlying mud as compared to the underlying Oligocene assemblages (18-35%; mean = 27%). This relative increase may reflect, in part, diagenetic alteration of the Tortonian assemblages, whose specimens are commonly pyritized and broken. No obviously displaced shallow-water species are represented in the Tortonian assemblages, although the sediments undoubtedly were derived, in part, from updip sources.

In the lowest four samples examined (612-16-7, 26 cm to 612-16-6, 135 cm) *Epistominella* is the predominant benthic genus (20-31%). *Stilostomella* becomes the dominant genus (19%) at 612-16-6, 148-144 cm, but it is quickly replaced at 612-16-6, 130-126 cm, by a more equable assemblage dominated by four genera (*Bolivina*, *Cassidulina*, *Stilostomella*, and *Uvigerina*) which persists to the unconformable contact. Above the nonfossiliferous sand, *Eponides* predominates in most samples (16-33%), but *Stilostomella* is copredominant in two.

Several other distinct abundance changes take place among individual genera. *Epistominella* and *Eponides*, for example, have reciprocal trends. *Epistominella* is most abundant in the lowest two samples (28-31%), declines progressively upward to the contact (4.3%), and remains sparse above the sand (0.3-4.8%). *Eponides*, on the other hand, is sparse (1.0-9.5%) below the contact, increases to 16% just above the sand, and reaches more than 25% in the highest three samples analyzed. The trend of *Cassidulina* is similar to that of *Epistominella* although at lower values; *Cibicides* is similar to *Eponides* (also at lower values). The relative abundance of *Stilostomella* is rather constant, except for its absence in the sand.

The two oxygen-sensitive genera, *Bolivina* and *Uvigerina* have parallel trends, being generally rather sparse, but increasing between 612-16-6, 139 cm and the contact. This suggests that oxygen levels may have decreased slightly in the early Oligocene prior to the erosion.

The number of benthic genera ranges from 21 to 26 below the contact (mean 22.8) and from 24 to 28 above the contact (mean 25.6). There probably is little meaningful difference in these values.

The long hiatus (25 my) separating the Rupelian and Tortonian sections at Site 612 encompassed a well-documented period of significant evolutionary change among benthic foraminifers (Douglas, 1979; Douglas and Woodruff, 1981; Woodruff and Douglas, 1981; Berggren and Schnitker, 1982) as the Neogene-modern fauna replaced that of the Paleogene. Some of the assemblage changes documented here, therefore, may be accounted for, in part, by evolution, and not by paleoenvironmental shifts alone.

Regional Relationships

The Rupelian-Tortonian contact at Site 612 juxtaposes supersequences Tc and Te of the Vail depositional model (Vail and Mitchum, 1979). Supersequence Td and most of Te are not represented. This part of the stratigraphic column at Site 612 is, however, not representative of the late Paleogene - early Neogene depositional regimes of the New Jersey slope, because upper Oligocene, lower Miocene, and middle Miocene strata (supersequences Td and Te) are well represented in surrounding sections (e.g., COST B-3 well; 300 m thickness; P. 21b-N. 14). These Oligocene and Miocene units can also be traced widely on seismic reflection profiles (Hampson and Robb, 1985; Farre, 1985; Poag, 1985a, 1987). The seismic reflections that bound each of these supersequences coalesce into the single reflector that marks the channel thalweg at Site 612. However at ASP 15, only 6 km along strike to the southwest, 40 m of lower Miocene and middle Miocene (N. 13-14) strata are present, and at ASP 14 these units are even thicker (~250 m; N. 6 - N. 12[?]). Thus the major channel cutting at Site 612 appears to have taken place during the early Tortonian (post N. 14, as documented at ASP 15 and COST B-3; pre-N. 16, as documented at Site 612), when the seaward edge of the middle Miocene delta was truncated as a result of lowered sea level (Poag, 1985a, 1987). The channeling presumably represents part of a global period of erosion, which, according to Vail and Mitchum (1979), coincided with the N. 15-16 boundary at 9.8 Ma; Berggren and others (1985) date this boundary at 11.1 Ma. Prior to this however, a major period of erosion in the late Oligocene removed most of the lower Oligocene strata from the coastal plain, shelf, and slope (Poag and Schlee, 1984; Poag, 1985a; Ward and Strickland, 1985). This mid-Oligocene period of erosion has also been documented on the Goban Spur (Poag and others, 1985), off North Carolina (Popenoe, 1985), and in Australia (McGowran, 1979; Quilty, 1980). Recently published oxygen-isotope analyses show that a significant increase in the global ice-volume was associated with this unconformity (Keigwin and Keller, 1984; Miller and others, 1985).

TORTONIAN-MESSINIAN CONTACT

Lithologic Characteristics

The contact between Tortonian and Messinian strata at Site 612 is manifest by a fractured scour surface at 13-6, 93 cm, which separates two distinctly different muddy lithologies (Fig. 11; Poag and Low, 1987). Below the contact is a light olive gray, micaceous, homogeneous mud; above is a dark olive gray, micaceous mud, containing layers and scattered patches of glauconite sand. A 2-cm bone fragment, surrounded by glauconite sand, is present at 612-13-6, 83-81 cm. The contact is disturbed by a diagonal fracture, 5 cm deep, filled with dark, glauconitic mud derived from the younger section. There is no obvious mixing of sediments across the contact by burrowing.

In the >74 μm sieved fraction, the seven samples below the unconformity (Fig. 11) contain chiefly fine, micaceous, quartz sand and a subordinate amount of fine, dark green, glauconite sand. Lignitized plant fragments and burrow casts made of framboidal pyrite are common. Radiolarians and planktonic foraminifers are abundant throughout the section except in the two samples closest to the unconformity, where the foraminifers are rare and fragmented.

Above the unconformity, dark green, medium to coarse glauconite sand is the dominant lithic component in the >74 μm fraction. A few poorly sorted grains of quartz also are present. Microfossils are dominantly radiolarians, as foraminifers are sparse.

Biochronology

The planktonic foraminiferal assemblages below the Tortonian-Messinian contact are moderately rich and diverse. The lowest sample at 612-13-6, 149-145 cm contains *Sphaeroidinellopsis subdehiscens*, *S. seminulina*, *Globigerina apertura*, *G. bulloides*, *G. quinqueloba*, *G. praebulloides*, *Globigerinita glutinata*, *G. uvula*, *Orbulina universa*, *Globigerinoides mitra*, *G. obliquus*, *Globorotalia merotumida*, *G. scitula*, *G. plesiotumida*, and *Neogloboquadrina pachyderma*. This and a similar, but less diverse, assemblage at 612-13-6, 118-114 cm, may be assigned to Zones N. 16-17 of the Tortonian Stage (as defined by Kennett and Srinivasan, 1983). In addition to these foraminifers, two radiolarian species, *Didymocyrtis laticonus* and *Diartus pettersoni*,

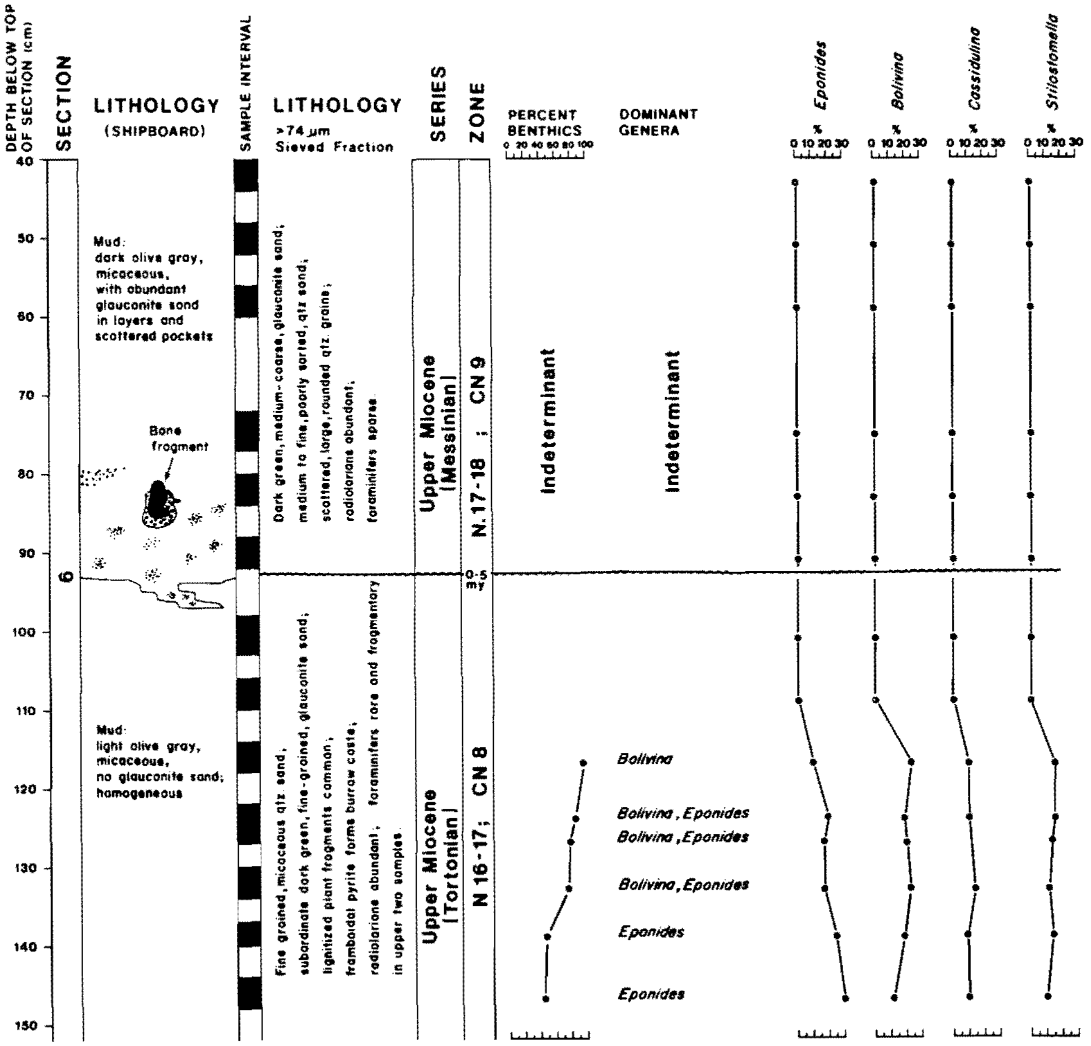

Figure 11. Chart showing lithic and benthic foraminiferal characteristics across the Tortonian - Messinian contact at Site 612.

which do not extend stratigraphically higher than middle Zone N. 16 (Theyer and others, 1978; Sanfilippo and others, 1985; Berggren and others, 1985; Palmer, 1987), are consistently present below the unconformity.

The six samples immediately above the unconformity (612-13-6, 92-40 cm) are difficult to date with planktonic foraminifers, which are scarce. However, upper Miocene radiolarians (Diartus hughesi and Didymocyrtis antepenultima, which do not occur lower than upper Zone N. 16, are consistently present in these samples. The nearest sample above the contact having a datable planktonic foraminiferal assemblage is at 612-13-5, 124-120 cm (119 cm above the contact). The presence here of Sphaeroidinellopsis subdehiscens, S. seminulina, Neogloboquadrina humerosa, N. acostaensis, Globigerina bulloides, G. nepenthes, G. incisa, G. falconensis, Globigerinoides conglobatus, G. extremus, Globorotalia cultrata, G. scitula, and Globigerinella obesa, indicates that this sample represents Zone N. 17-18 of the latest Miocene (Messinian Stage). These combined data indicate, therefore, that a short hiatus of perhaps 0.5 my is represented by the unconformable Tortonian-Messinian contact.

Geophysical Characteristics

On the downhole geophysical log, the Tortonian-Messinian contact is marked by higher gamma-ray values in the Messinian section (Fig. 8), probably reflecting the larger quantity of glauconite. Sonic velocity also is greater in the Messinian section, and the velocity change across the unconformity at ~1.99 sec (integrated sonic log) correlates with a moderately strong seismic reflector at the same depth on line 25 (Poag, 1987).

Benthic Foraminiferal Characteristics

The major microfaunal change associated with the Tortonian-Messinian contact is the dominance of radiolarians and virtual absence of foraminifers between 612-13-6, 110 cm (below the contact) and 612-13-6, 40 cm (above the contact). However, below the contact, several interesting trends can be observed among the benthic foraminifers. The percentage of benthic specimens is high throughout, increasing progressively upward from 44% to 92% (Fig. 11;

Table 5 of Poag and Low, 1987). These high values suggest either that planktonic foraminifers have been preferentially dissolved, or that the original assemblage has been altered by the downslope transport of shallower-water forms. The presence of *Nonionella* and *Buccella* does suggest some downslope redeposition, but these forms are only minor components of the assemblage.

Among the predominant genera, *Eponides* is the principal form in the lower two samples (26-30%). It declines upward, becoming copredominant with *Bolivina* in the next three samples, finally to become subordinate to *Bolivina*. The other two common genera, *Cassidulina* and *Stilostomella*, display no significant stratigraphic trends, although *Cassidulina* disappears temporarily at 612-13-6, 127-125 cm.

The number of benthic genera varies only slightly throughout the section (range = 23-29; mean 24.5) and is similar to values cited from lower in the section.

Regional Relationships

The Tortonian-Messinian contact has not been documented previously on the New Jersey shelf and slope. Samples were not collected this high in the section at the COST B-2 and B-3 wells, nor at other commercial well sites. Intermittent coring coupled with long erosional hiatuses prevented its recognition in the shallower boreholes, such as ASP and AMCOR. However, a pronounced seismic reflector, previously interpreted as the Miocene-Pliocene contact (Poag, 1985a, 1985b) now appears to be more correctly equated with the Tortonian-Messinian unconformity. The unconformity can be traced widely across the New Jersey margin on seismic reflection profiles, and has been described from outcrops and boreholes on the Atlantic Coastal Plain, where it separates the Tortonian Eastover Formation from the Pliocene Yorktown Formation (Ward and Blackwelder, 1980; Ward, 1984; Ward and Strickland, 1985). An equivalent unconformity is known from the nearshore region of North Carolina (Riggs, 1984), from the deeper Georgia-North Carolina margin (Popenoe, 1985), and from coastal plain localities in northern and southern Florida (Adams and others, 1977). Further evidence of widespread erosion during this period comes from the Mediterranean region (Hsü, 1977; Adams and others, 1977), the continental slope of Ireland (Poag and Low, 1985), New Zealand (Loutit and Kennett, 1981a, 1981b), Australia (McGowran, 1979; Quilty, 1980), and Puerto Rico (Seiglie and Moussa, 1984).

The Tortonian-Messinian unconformity correlates with a sea-level fall and major global unconformity in the middle of supersequence Tf of the Vail depositional model, which Vail and Mitchum (1979) place at 6.6 Ma, in the middle of foraminiferal Zone N. 17. Berggren and others (1985) place this level at 6.4 Ma. The sea-level fall is especially notable for having contributed to the closing of the Straits of Gibraltar and the ensuing evaporation of Mediterranean waters (Messinian "salinity crisis"; Hsü and others, 1977), which profoundly affected global paleoclimates and oceanic circulation patterns in the late Miocene (e.g., Poag and Low, 1985).

UPPER PLIOCENE - UPPER PLEISTOCENE CONTACT

Lithologic Characteristics

The unconformable contact of upper Pliocene and upper Pleistocene strata was recovered in core 612-5-3, at 39 cm (Figs. 12; Poag and Low, 1987). Below a concave scour surface, the sediment is a homogeneous dark gray mud, containing scattered layers, clumps, and burrow casts of glauconite sand. Immediately above the contact is a 4-cm zone of coarse, dark green to black, glauconite sand, mixed with chunks of the underlying mud. Above the sand, separated by a sharp contact, is a lighter gray, finely laminated mud.

The >74 µm sieved fraction reveals a much more complicated succession of lithologies (Fig. 12). The lower section (612-5-3, 124-48 cm) consists of slightly micaceous, quartzose, glauconite sand. Small aggregates of gypsum (no euhedral cyrstals) are rare and foraminifers are rare to common. The sample immediately below the contact (612-5-3, 44-40 cm) is quite different; quartz sand is dominant, gypsum aggregates are abundant, and foraminfers are rare. Above the contact (612-5-3, 39-35 cm), the residue is quartzose glauconite sand, with only sparse gypsum aggregates and rare foraminifers. The next higher sample (612-5-3, 30-26 cm) contains chiefly quartz sand, glauconite, mica, and gypsum; foraminifers are rare. The youngest section (612-5-3, 24-0 cm and 612-5-2, 150-120 cm) is principally micaceous quartz sand, frequently containing burrow casts of framboidal pyrite. Glauconite is rare, and foraminifers range from few to common.

Biochronology

This contact was originally dated on the basis of closely spaced nannofossil samples (Poag, Watts and others, 1987). Planktonic foraminiferal assemblages are sporadically and incompletely represented in many of the samples in this section, but a few moderately rich assemblages corroborate the late Pliocene and late Pleistocene dates derived from nannofossils. At 612-5-3, 124-120 cm, an assemblage containing *Globigerinoides ruber*, *G. conglobatus*, *G. trilobus*, *Globigerina inflata*, *G. incisa*, *G. bulloides*, *Orbulina universa*, *Globigerina glutinata*, *Neogloboquadrina pachyderma*, *Globorotalia crassaformis*, *G. scitula*, *Sphaeroidinella dehiscens*, and *Sphaeroidinellopsis seminulina* is present. The co-occurrence of *S. seminulina* and *G. inflata* indicate that this assemblage belongs to Zone PL. 5 of the late Pliocene. A similar assemblage at 612-5-3, 78-74 cm contains, in

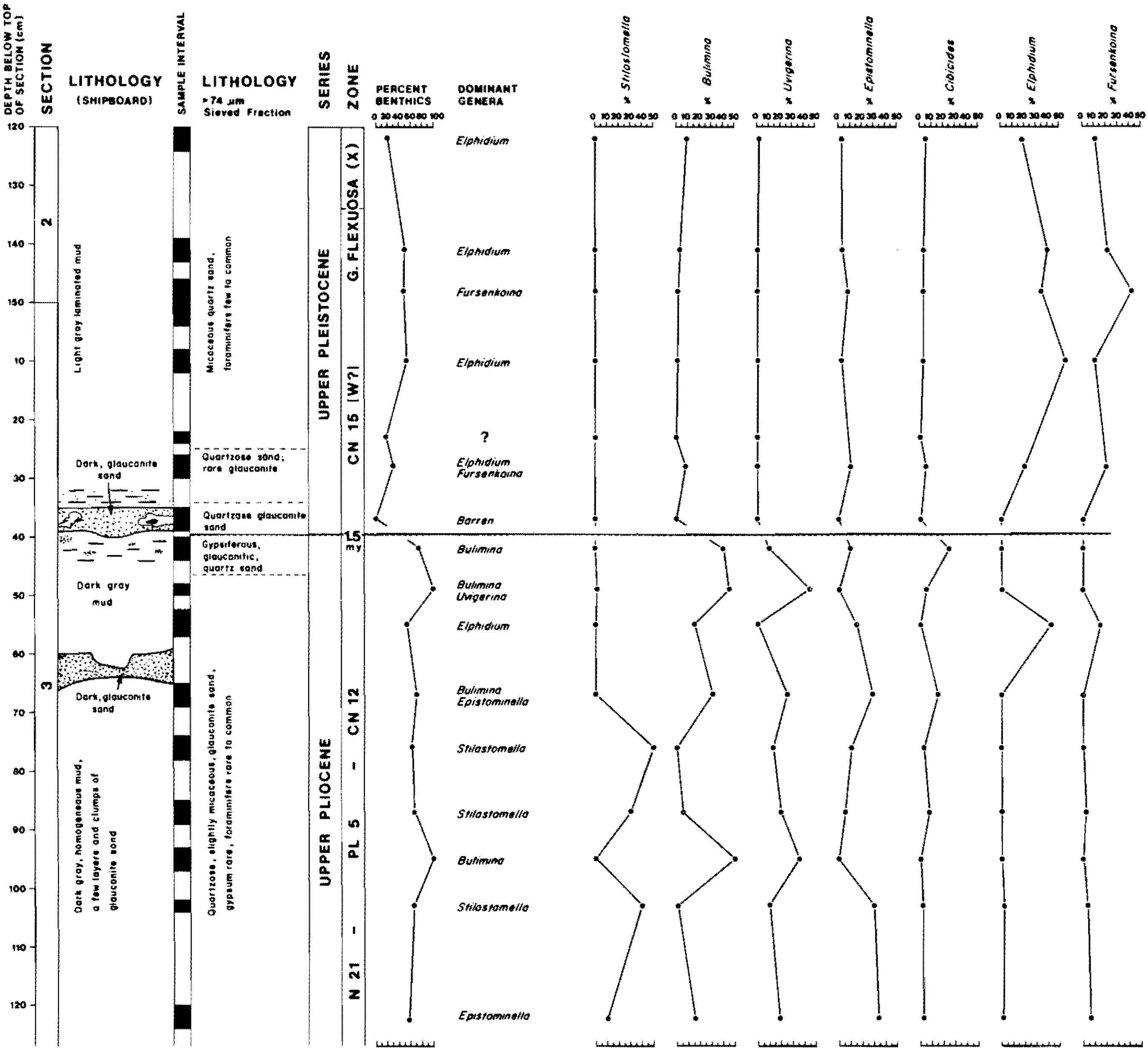

Figure 12. Chart showing the lithic and benthic foraminiferal characteristics across the upper Pliocene - upper Pleistocene contact at Site 612.

addition, Globorotalia miocenica, corroborating the assignment to Zone PL. 5, which spans 3-2.1 Ma.

Above the contact, the most complete planktonic assemblage is present at 612-5-2, 124-120 cm; it includes Globigerinoides ruber (pink form), G. conglobatus, Globigerina inflata, G. bulloides, G. rubescens, Globorotalia flexuosa, G. hirsuta, G. truncatulinoides, Neogloboquadrina dutertrei, Pulleniatina obliquiloculata, and Globigerinella aequilateralis. This is a typical late Pleistocene assemblage. The most notable element is the abundance of Globorotalia flexuosa, and its predominance among the globorotaliids. These characteristics are generally diagnostic of the Globorotalia flexuosa Zone (Kennett and Huddelstun, 1972; Poag and Valentine, 1976), which is equivalent to the X Zone of Ericson and Wollin (1968) and stage 5 of the oxygen-isotope chronology (Berggren and others, 1980). The interval between the G. flexuosa Zone and the unconformity contains a glacial planktonic assemblage dominated by G. inflata and N. pachyderma and probably belongs to isotopic Stage 6. Oxygen-isotope stage 5 spans the interval of 0.09-0.12 Ma. The nannofossil evidence places this Pleistocene section at no older than 0.44 Ma. These data indicate, then, that a hiatus of approximately 1.5-2 my separates the upper Pliocene from the upper Pleistocene section at Site 612.

Geophysical Characteristics

The upper Pliocene - upper Pleistocene contact at Site 612 is too close to the seafloor to have been logged. The end of the drill pipe had to remain within the hole below the top of the Miocene section during logging. However, the contact forms a distinct reflector on high-resolution seismic lines and can be traced widely across the New Jersey margin (Poag, 1987).

Benthic Foraminiferal Characteristics

The upper Pliocene-upper Pleistocene contact at Site 612 is marked by several important changes in the benthic foraminiferal assemblages (Fig. 12; Table 6 of Poag and Low, 1987). Benthic percentage is particularly high below the contact, (54-99%; mean 75%) and drops significantly above it (21-54%; mean 38%), even though the number of displaced shallow-water specimens (Elphidium, Fursenkoina, Quinqueloculina, is higher above it. Several genera are predominant below the contact. Bulimina and Stilostomella are most consistent, but share predominance in two samples with Uvigerina and Epistominella. Elphidium, a displaced shallow-water form, is predominant at 612-5-3, 57-53 cm).

The glauconite sand immediately above the contact is barren of microfossils, but they are common in several of the higher samples. Elphidium and Fursenkoina, both displaced shallow-water forms, predominate in most samples above the contact.

In addition to the relative abundance of Bulimina, Uvigerina, and Epistominella below the contact, Cibicides also is most abundant there.

Regional Relationships

Site 612 is the only firm documentation of a fully marine Pleistocene-Pliocene contact yet established in the New Jersey shelf and slope region. Elsewhere, intermittent coring (ASP sites), the presence of unconformities (AMCOR sites), and lack of shallow-depth sampling (commercial and COST wells) prevent analysis of the contact. A distinct seismic discontinuity separates inferred Pliocene from Pleistocene strata over most of the shelf and slope (Poag, 1985a, 1987) indicating its regional significance. An equivalent seismic unconformity is also present beneath the continental rise, where it was documented at Sites 604 and 613 (Poag, 1985b, 1987).

The upper Pliocene - upper Pleistocene contact at Site 612 corresponds approximately to the contact between supersequences Q and Tf of the Vail model. Vail and Mitchum (1979) assigned the Q-Tf boundary to the Pliocene-Pleistocene boundary, which they dated at 2.8 Ma. This stratigraphic interpretation is peculiar to the Exxon group (e.g., Lamb and Beard, 1972; Stainforth and others, 1975) and has little support elsewhere (e.g., Poag and Valentine, 1976; Haq and others, 1977; Berggren and others, 1980). A variety of recent studies show that the base of the Pleistocene is approximately 1.6 Ma (Berggren and others, 1980). The boundary between supersequences Q and Tf at 2.8 Ma falls within Zone Pl.5 of Berggren and Van Couvering (1974; Berggren and others, 1980), which is an upper Pliocene level.

Poag and Low (1985) noted an unconformity within middle to upper Pliocene strata at Site 548 on the Goban Spur, which also appears to represent the Q-Tf contact. They originally estimated that the Goban Spur contact was at approximately the 3.8 Ma level, but the imprecision of both nannofossil and foraminiferal zonation at Site 548 could easily accomodate a date of 2.8 Ma.

SUMMARY AND CONCLUSIONS

Marked changes in lithologic, microfaunal, seismic reflection, and downhole logging characteristics give evidence that seven (and perhaps eight) major sequence boundaries at Site 612 are erosional unconformities. The presence of sand layers or exotic lithoclasts immediately above the scour surface at six of the seven contacts indicates that deposition was reinitiated by a sediment gravity flow, which must have eroded at least a portion of the missing sediment. These relationships are similar to those noted for several equivalent supersequence boundaries on the Goban Spur (Poag and Low, 1985; Poag and others, 1985). The source of sand at the Miocene and younger unconformities of Site 612 could have been almost anywhere on the adjacent shelf and upper slope, as the entire sedimentary section there is chiefly terrigenous detritus (Poag, 1985a). However it is more difficult to explain a quartz sand layer at the middle Eocene - upper Eocene contact, in the midst of a thick carbonate (chalk) section. Presumably this sand was emplaced during a rapid sea-level fall, which brought a siliciclastic shoreline much nearer to Site 612 than during the rest of the Paleogene.

The persistent presence of gravity-flow deposits at the unconformable supersequence boundaries implies a common cause of sediment displacement. Most of the unconformities can be traced onto the adjacent shelf and even onto the coastal plain, where the stratigraphic evidence strongly implicates sea-level fall as that common agent (Fig. 13; Poag and Schlee, 1984; Ward and Strickland, 1985; Poag, 1987; Poag and Ward, 1987). Poag and others (1985) reached a similar conclusion regarding equivalent supersequence boundaries on the Irish continental slope, as did Loutit and Kennett (1981a, 1981b) for New Zealand, and Steele (1976), McGowran (1979) and Quilty (1980) for Australia. The process of slope erosion envisioned by Poag and others (1985) was depression and elevation of a turbulent water-mass boundary across the continental slope in unison with falling and rising sea level. This was based on the suggestion of Sarnthein and others (1982) that internal waves and turbulence, caused by density differences at water-mass boundaries, could cause significant erosion where a boundary intersects the seafloor. Stanley and others (1983) discussed similar relationships regarding water-mass boundaries on the New Jersey margin. They showed that the mudline on the modern New Jersey margin can range from 200 to 1000 m depending on several variables. Beneath the shelf water-mass (shoreline to the shelfbreak), erosion takes place continually from the interplay of storms, fronts, tides, and internal waves. The upper few hundred to 1000 m below the intersection of the shelf and slope water-masses, is a transitional zone of periodic resuspension induced by surface waves, tidal currents, wind-stress currents, internal waves (Southard and Stanley, 1976), and shear forces between major water masses and oceanic fronts (Ruzecki and Welch, 1977; Karl and others, 1983; Pietrafesa, 1983). This alternation of deposition and resuspension triggers sediment flow along the middle and lower slope. A falling sea level would depress the transition zone, causing resuspension and sediment flow even farther down the slope, affecting sites such as 612.

Deep flowing boundary currents, such as the Gulf Stream and the Western Boundary Undercurrent, also are effective agents for eroding the continental slope and rise (e.g., Tucholke and Mountain, 1979; Pinet and Popenoe, 1982, 1985a, 1985b; Popenoe, 1985; Ledbetter and Balsam, 1985). Geographic and bathymetric shifts of such currents, coincident with sea-level changes, could be of particular importance in forming major unconformities. Such shifts were demonstrated by Ledbetter and Balsam (1985) on the New Jersey margin, where the Western Boundary Undercurrent accelerated, moved shoreward by ~150 km, and shoaled by ~1000 m (relative to its modern velocity and position) during the last glacial. These data do not apply directly to depths as shallow as Site 612, but a current-swept middle Eocene outcrop belt is only 3 km downdip from 612; it is reasonable to assume that this erosional swath also temporarily shifted updip during glacio-eustatic sea level falls.

At most of the contacts studied, definitive changes in the benthic foraminiferal assemblages and other sedimentary constituents indicate that the seafloor environment changed during the hiatus. Although our analyses do not always point directly to paleobathymetric changes, they do suggest that water mass properties, such as dissolved oxygen content, temperature, and nutrient content, changed considerably, implying that related large scale tectonic or paleoclimatic shifts triggered paleoceanographic responses. The growing body of field data, such as we have presented here (including biostratigraphy, lithostratigraphy, sedimentology, downhole logging, and seismostratigraphy), when

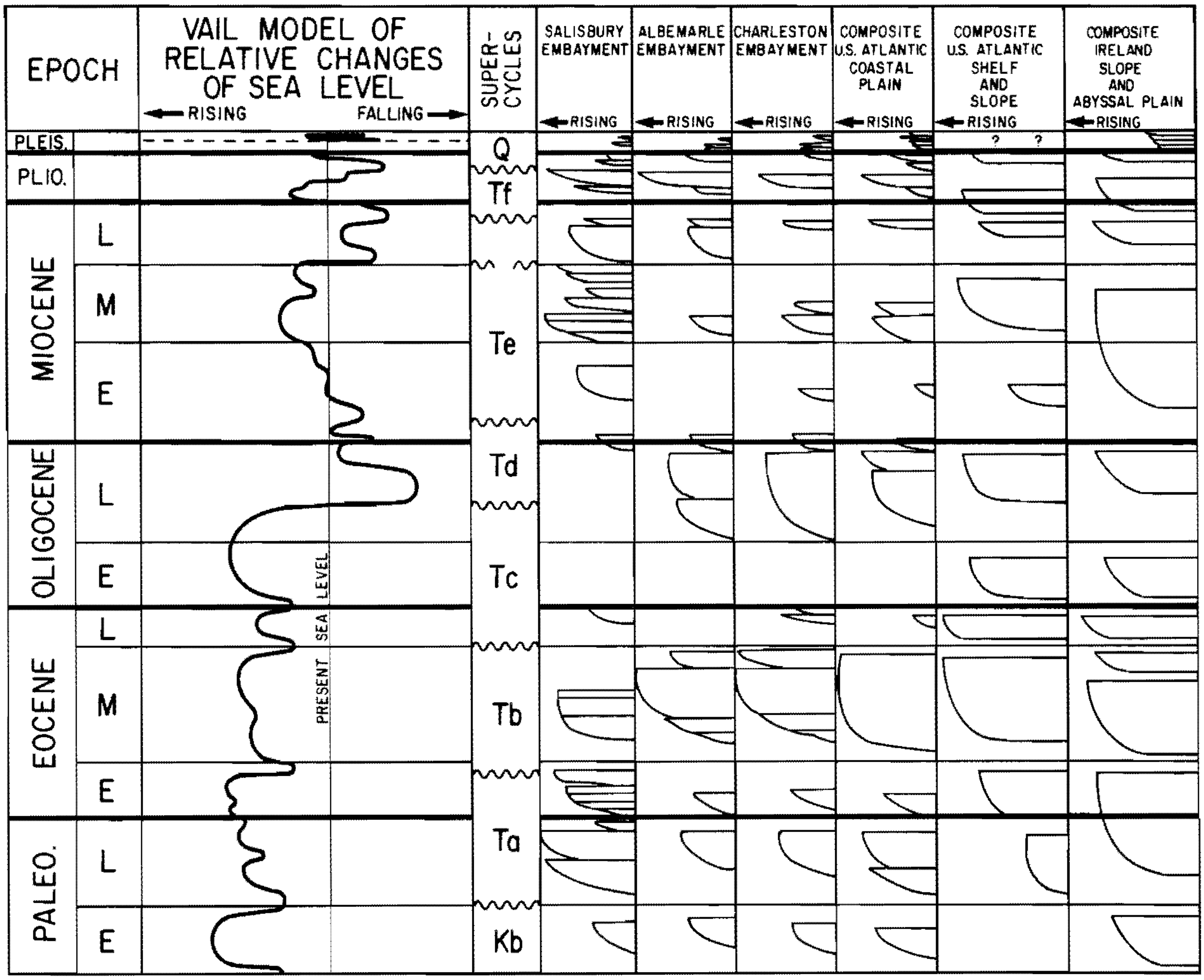

Figure 13. Relationships of the Cenozoic depositional sequences and stratigraphic gaps of the U.S. Atlantic margin and the Irish margin to the sea-level curve, supercycles and global unconformities of the Vail depositional model. (From Poag and Ward, 1987).

integrated with stable isotope analyses, demonstrate a clear link during much of the Cenozoic between widespread shelf and slope (and even abyssal) erosion, increased global ice volumes, cooler global climate, and lowered sea levels (Vail and others, 1977; Frakes, 1979; Miller and Fairbanks, 1983; Keigwin and Keller, 1984; Poag and Schlee, 1984; Poag, 1985a; Poag and others, 1985; Poag and Low, 1985; Miller and others, 1985; Aubry, 1985; Woodruff and Savin, 1985; Keller and Barron 198X). Some authors have interpreted the oxygen-isotope record as an indication that significant global ice-volumes were present well into the Late Cretaceous (Matthews and Poore, 1980; Matthews, 1984). Thus there seems to be no requirement that tectonism played a role in short-cycle sea-level change (at least in post-Eocene time), as advocated by Pitman (1978), Watts (1982), Watts and Thorne (1984) and Thorne and Watts, 1984). On the other hand, there is no a priori reason to assume that a single mechanism was responsible for each supercycle. Tectonism undoubtedly was a modifying factor during some intervals (Klitgord and Schouten, 1986), amplifying or damping the effects of ice-volume change, and was, perhaps, the principal cause of pre-Oligocene supercycles.

Our results substantiate the applicability of the Vail supercycle framework model to North Atlantic continental margins, confirming that six of the eight Cenozoic global unconformities are present in this region. We also provide further evidence of a stratigraphic gap between the Campanian and Maestrichtian stages as suggested by Haq and others (1987).

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LATE PALEOZOIC SEA LEVELS AND DEPOSITIONAL SEQUENCES

CHARLES A. ROSS[1] AND JUNE R. P. ROSS[2]

1 Chevron U.S.A., Inc., P. O. BOX 1635, Houston, TX 77251
2 Department of Biology, Western Washington University, Bellingham, WA 98225

ABSTRACT

Cyclic sea level charts for the Lower Carboniferous (Mississippian), Middle and Upper Carboniferous (Pennsylvanian), and Permian show considerable variability in the duration and magnitude of third-order depositional sequences, and also in the position of general sea level as represented by second-order sea level. Transgressive and highstand system tracts are numerous on the cratonic shelves of the late Paleozoic continents. Shelf margin wedges are less well represented except at times of general lower sea levels. Most low stand wedges and all low stand fan systems are structurally deformed and make up many of the accretionary wedges and displaced terranes that lie structurally emplaced against the former Paleozoic margins of the cratons.

More than seventy named third-order depositional sequences (mesothems) seem well defined in Carboniferous and Permian rocks. They may be grouped into six named second-order supercycles which in turn are parts of the Kaskaskia and Absaroka megacycles (or Sloss sequences).

Most third-order sequences, wherever possible, are named for the marine limestone formation(s) or member(s) that represents the highstand facies of that particular sequence. It is also the name bearer of the associated sea level rise and fall. The second-order sequences are named for areas where the general relationships between the second-order sequences are well shown as in the Upper Mississippi River Valley, in southeastern Arizona and southwestern New Mexico, and in western Texas.

Although glaciation appears to be the cause of the relatively short term sea-level changes associated with these sequences, other longer term causes also are suspected in order to explain some of the phenomena. These longer term causes may relate to timing and rates of plate motions, orogenic events, and mid-oceanic ridge construction.

INTRODUCTION

During the later part of the Paleozoic Era, major sea-level fluctuations having about 1 to 3 million years duration (third-order cycles) are inferred from study of depositional environments and stratigraphic relations in the rock record in many parts of the world. The authors' initial studies on these changes in sea level and their paleogeographic distribution (Ross, 1979; Ross and Ross, 1979, 1981a, 1981b, 1985a, 1985b) are elaborated on in this paper with charts in a similar format to that used for Mesozoic and Cenozoic sea-level cyclic fluctuations by Haq, Hardenbol, and Vail (1987 and this volume).

The third-order cycles of sea-level changes are global in extent, and not relative, local sea-level changes (Vail and Mitchum, 1977). They may be grouped together into larger, second-order cycles by major events that partially determined broader patterns of late Paleozoic deposition. These are comparable to major events that determine Mesozoic and Cenozoic depositional patterns.

In the late Paleozoic the events associated with second-order cycle patterns appear related to tectonic events and changes in paleogeography, such as the various steps in joining together of Euramerica and Gondwana late in Early Carboniferous. Similar types of events for the Mesozoic and Cenozoic, for example, would be the steps in the opening of the North and South Atlantic oceans and the Gulf of Mexico or the various steps of the Himalayan orogeny and the closing of the Tethys. Second-order cycles are subdivisions of first-order cycles (or Sloss sequences, Sloss, 1963, 1964) which are, at least in part, the culmination of a series of trends seen in second order-cycles, such as the final step in joining Gondwana and Euramerica into Lesser Pangaea. In addition to naming and describing these late Paleozoic second-order cycles, we also name and describe the third-order cycles that they contain.

COMPARISON WITH MESOZOIC/CENOZOIC CYCLES

In late Paleozoic strata, it is possible to identify interregional unconformities and to correlate these from one region to another with biostratigraphic evidence. Mitchum and others (1977) defined a depositional sequence as "a stratigraphic unit composed of a relatively conformable succession of genetically related strata and bounded at its top and base by unconformities or their correlative conformities". With the type of detailed physical and biostratigraphic criteria available for late Paleozoic strata, unconformities have been consistently usable and traceable, however, the identification of correlative conformities in deeper basins has been difficult (or impossible) to establish and correlate. Therefore, the late

Paleozoic cycles discussed in this paper are best known and most easily studied on the more stable shelves of cratons.

With certain differences, sequence-stratigraphy concepts and terminology (Fig. 1) used by Haq and others (1987 and this volume) are applied to late Paleozoic cyclic sequences. The most obvious difference is the lack of continuity between shelf sediment systems and ocean floor fan systems. Also, most cratonic margins of Paleozoic age do not occupy their former Paleozoic geographic positions (Ross and Ross, 1981b, 1983). All the Paleozoic cratons and ocean floor fans have been carried by spreading sea-floor plates to their present geographic positions. In that process, most Paleozoic margins of the cratons were structurally deformed to various degrees and have younger margins made up of complexly folded and faulted accreted terranes composed of microcratons and fans and margin deposits (Monger and Ross, 1971). In some cases, their reconstruction is not possible because the sedimentary deposits have been destroyed in subduction zones. Further, because of their complex internal structure, accreted terranes are not suitable for use in detailed seismic analysis aimed at reconstructing depositional patterns. Such reconstruction is possible for the relatively undeformed Cretaceous and Cenozoic depositional systems in many parts of the world. Eventually it may be possible to identify strata in these accreted terranes as being the same age as relatively undeformed strata on the craton and, in that way, bring together different parts of individual sequences. However, it is not possible at this time to identify undeformed late Paleozoic lowstand system wedges and oceanic fans on seismic profiles.

Another difference lies in the depositional setting. Most of the rock record on which the Cenozoic sea-level curves and the sequence-stratigraphy concepts have been developed are in clastic-rich depositional settings on passive cratonic margins, such as the Gulf of Mexico. Paleozoic sequence-stratigraphy, on the other hand, is based mainly on cratonic shallow water carbonate-rich or mixed carbonate-clastic successions with numerous calcareous fossils. In these Paleozoic stratigraphic sections, detailed fossil zonations and time correlations are well established. Seismic interpretations and seismic correlation for these rocks are much less useful for developing sequence-stratigraphic interpretations than in cratonic margin and continental slope settings of the Cretaceous and Cenozoic. Thus, those principles and concepts of sequence-stratigraphy that deal with transgressive depositional system tracts and highstand depositional system tracts are the ones most readily applicable to Paleozoic strata. Lowstand depositional fan system tracts, lowstand depositional wedge system tracts, and shelf margin depositional wedge system tracts are less commonly preserved or are less easily identified in Paleozoic rocks and are usually tectonically deformed.

On the other hand, many features of sequence-stratigraphy (Fig. 1) are well known in Paleozoic successions, particularly in shallow intracratonic basins and their margins. Both types of sequence boundaries are well represented. The first type shows extensive erosion and valley and canyon cutting on the shelves during very low sea-level stands below the cratonic shelf margin (type 1 boundaries or type 1 unconformities of the Exxon group). The second type of sequence boundary is shown by weathering, non-deposition, minor solution (in carbonates), and hard ground development during times when sea-level stands dropped to positions at or just above the shelf margin edge (type 2 boundaries or type 2 unconformities of the Exxon group).

TRANSGRESSIVE AND HIGHSTAND SYSTEM TRACTS

The stable parts of Paleozoic cratonic shelves include a great variety of thin, widespread rock units. Most are fluvial to shallow subtidal deposits of mixed carbonate and clastic sediments that are part of transgressive and highstand systems tracts. Because the topography of these shelves had extremely low relief (nearly flat), a small increase or decrease in sea level resulted in great lateral displacement of the shoreline and the lithofacies and biofacies. Thus, the actual area available for retaining additional sediments because of a small rise in sea level was very large. These wide shallow shelf areas also led to warm, very shallow, and geographically extensive areas which encouraged carbonate-producing faunal communities to subdivide the environment into many, clearly defined and bounded, specialized communities. These communities shifted back and forth across and laterally along the shelves giving rise to widespread traceable limestones that are only slightly diachronous at different places.

Because of the depositional features of late Paleozoic shelves, during sea-level highstands, the small amount of clastic sediment that was available was retained mainly in the shoreline clastic facies and relatively little of it by-passed the shallow subtidal carbonate facies. Those parts of the outer shelves that became too deep for carbonate production as sea level rose were starved of sediments and have thin, phosphate-rich dark shales forming condensed sections (Heckel, 1986). This is particularly typical of many Pennsylvanian cycles and indicates that vertical growth of the carbonate communities, for whatever reason, commonly was unable to keep up with the rate of sea level rise. The resultant deposits demonstrate the contrasting lithologies of these cyclical repetition of facies that accompanied each sea-level rise and fall; eg., cyclothems (Weller, 1930; Wanless and Shepard, 1936) and megacyclothems (Moore, 1958). In intracratonic basins on the stable parts of the cratons, repeated condensed sections are commonly stacked on one another and may represent several depositional sequences.

These mainly passive intracratonic shelves and margins in the Mississippian and Permian

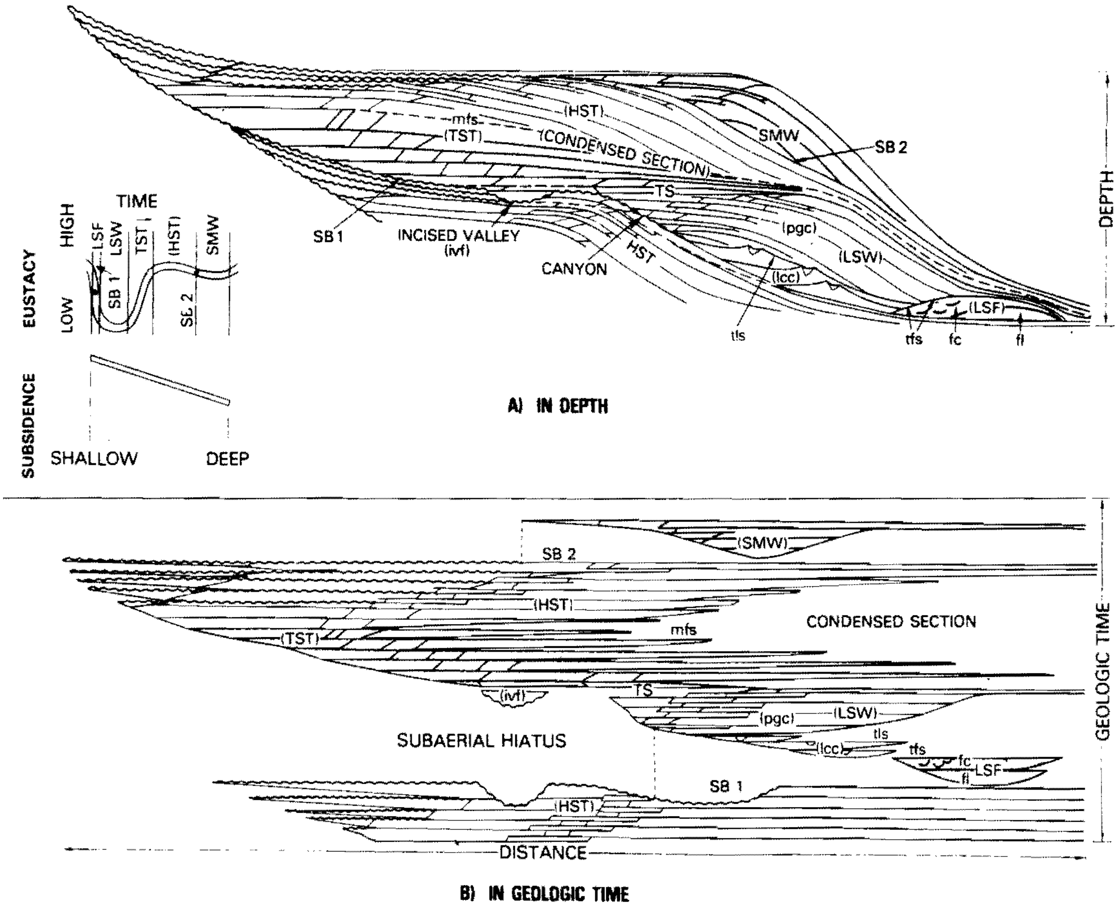

FIGURE 1. Idealized model of the concepts of depositional sequences. A, Cross-section showing stratigraphic relations. B, Same cross-section redrawn to show time relations. Surfaces are: SB = sequence boundaries (SB 1 = type 1 sequence boundary, SB 2 = type 2 sequence boundary), DLS = downlap surfaces (mfs = maximum flooding surface, tfs = top fan surface, tls = top leveed channel surface), and TS = trangressive surface. System tracts (or genetic stratigraphic units) are: HST = highstand systems tract, TST = transgressive systems tract, LSW = lowstand wedge systems tract (ivf = incised valley fill, pgc = prograding complex, and lcc = leveed channel complex), and SMW = shelf margin wedge systems tract. (After Haq and others, 1987.)

commonly were dominated by carbonate transgressive and highstand system tracts perhaps because they were times of higher carbonate production or were times of slower rates of sea-level change (see discussion by Kendall and Schlager, 1982). Examples would be the Illinois Basin (Lineback, 1981) and the Tobosa Basin (western Texas and eastern New Mexico) during the Tournaisian and the early to middle Visean (Lane and De Keyser, 1980) and the Delaware Basin during the Leonardian and Guadalupian. The maximum flooding surfaces (mfs) at the end of the construction of the transgressive systemstract and just prior to construction of highstand systems tract are not well represented by condensed sections on these shelves, however, the condensed sections are stacked in the adjacent shallow intracratonic basins.

SHELF MARGIN AND LOW STAND WEDGES

Shelf-margin wedge system tracts and lowstand wedge system tracts are difficult to differentiate in late Paleozoic rocks mainly because of later structural disturbances along the former cratonic margins. The best candidates for shelf-margin wedge system tracts are in rocks of the Morrowan and Atokan Series along the southern and western edges of the North American craton. In eastern Oklahoma and northern Arkansas, the type areas for the Morrowan and Atokan Series, the lithologic facies and restricted geographical distribution of these rocks along the edge of the Paleozoic craton seem to fit a carbonate/clastic shelf-margin wedge setting and represent deposition during a series of generally low sea-level highstands (Figs. 2, 3). The lower (latest Chesterian to Atokan) parts of the Ely Limestone in northeastern and eastern Nevada probably are typical of a carbonate shelf-margin wedge system tract, and were deposited during the same cycles of low sea-level stands.

During the late Paleozoic, parts of the Euramerican craton were not stable. For example, in Oklahoma, Texas, New Mexico, Colorado, and Utah, and in the greater Donetz basin of Ukrainia, long linear high angle to vertical reverse faults created a system of structural basins and blocks. These included the Paradox and Oquirrh basins and the Ancestral Rocky Mountains. Thick wedges of commonly arkosic sediment poured into the basins from the adjacent uplifts. Along the western margin of the Euramerica craton, the Antler-Sonoma orogeny formed complex uplifts, imbricated thrust sheets, and adjacent foredeep basins. The Appalachian orogeny along the southeastern margin produced extensive alluvial plains and terrestrial clastics. Much of that material was redeposited during sea-level lowstands as turbidite fans in the Ouachita-Marathon trough during the late Mississippian and Pennsylvanian.

Identifying lowstand wedge systems and lowstand fan systems tracts in late Paleozoic rocks is difficult because most of these sediments later became parts of the deformed rocks within the Hercynian-Appalachian-Ouachita-Marathon and Antler-Sonoma orogenic belts. The source for much of the clastic sediment in those orogenic belts was from the advancing northern edge of Gondwana (or from oceanic arcs and microcontinents in PaleoPanthalassa in the case of the Antler-Sonoma belt) rather than from the passive margins of Euramerica. Along the southern margin, most of these clastics are coarse turbidites, including megaclastic turbidites, and it is extremely difficult to separate the effects of tectonic oversteepening of mobile shelf edges and other tectonic events along the active margin of Gondwana from the effects of eustacy events. These turbidite clastics were thrust over the southern margins of Europe and North America (Euramerica) in many places and have obscured and deformed any late Paleozoic lowstand wedges and lowstand fans that were deposited there.

In summary, it is possible to recognize the various parts of the sequence-stratigraphic depositional model within late Paleozoic strata. On the cratonic shelves and shelf margins, the highstand, transgressive, and shelf margin wedge systems seem clearly identifiable. The lowstand wedge and lowstand fan systems are more difficult to identify and may be mostly concealed by Hercynian orogenic structures.

LATE PALEOZOIC SEA-LEVEL CHARTS

Although we closely follow the format of the Mesozoic and Cenozoic charts prepared by Haq and others (1987 and this volume), several changes were dictated by the different kind of available evidence. First, we include no data on magnetostratigraphy. Most late Paleozoic paleomagnetic studies have been concerned with the important problem of finding reliable pole positions in order to construct apparent polar wandering curves. Most of the base data are from red beds or volcanics. Although these are adequately located stratigraphically and geographically for polar wandering studies, they generally are not adequately dated faunally or radiometrically to add any precision to the sea-level charts. We do know that magnetic polarity reversed repeatedly during the late Paleozoic, just as it did in the Mesozoic and Cenozoic, and that the Kiaman interval of predominantly reversed poles extended from about the beginning of the Late Carboniferous to near the end of the Permian. Much of the paleomagnetic data for polar wandering studies includes averaging both normal and reversed pole positions and this suggests that when studied in sufficient detail a magnetic polarity scale eventually may be feasible for the late Paleozoic.

Time Scale

The absolute age scale (time in millions of years) is based mainly on Harland and others (1982). The Decade of North America geology scale (Palmer 1983) follows the Harland and others' scale for this part of the Paleozoic. The COSUNA chart (Salvador, 1985) does not give enough information to know how its absolute time scale was constructed or, in some cases, the details of the international correlation of several of the boundaries used to form the chart.

A more important point about all three age scales, however, is that the published estimated errors on all of these Paleozoic dates is at least ±6 million years and most are ±10 to ±12 million years or about 10 times the duration of individual sequences as estimated by Ramsbottom (1979) and by Ross and Ross (1985b). Because the absolute age scale for late Paleozoic eustatic events is only a rough estimate of the timing of these events, we have not tried to place absolute ages on the maximum flooding surfaces (mfs) or on unconformities and correlative conformity surfaces between sequences. Such ages in the late Paleozoic are not accurate enough to carry significance.

Biostratigraphy

The biostratigraphy column uses a different set of zonal faunas than used for the Mesozoic and Cenozoic. Planktonic (or nektonic) faunas include conodonts which are widely used correlation guides for the Lower Carboniferous and lower part of the Middle Carboniferous, and to a lesser extent for parts of the Permian. Also, nektonic ammonoid zones are included where possible. However, the most exhaustive zonal faunas are the shallow-water benthic foraminifers which are widespread in the carbonate facies of the transgressive, highstand, and shelf margin systems tracts. Bryozoans, brachiopods, corals, and blastoids are also useful zonal fossils in many of the shelf carbonate facies. Other fossil groups may be useful, particularly in other parts of the Paleozoic column, such as the planktonic graptolites and acritarchs in Ordovician, Silurian, and Devonian rocks. A more detailed discussion of the late Paleozoic faunal zonation that we show in Figures 2, 3, and 4 is presented in the accompanying faunal analysis by Ross and Ross (this volume).

Although different faunas are used for these

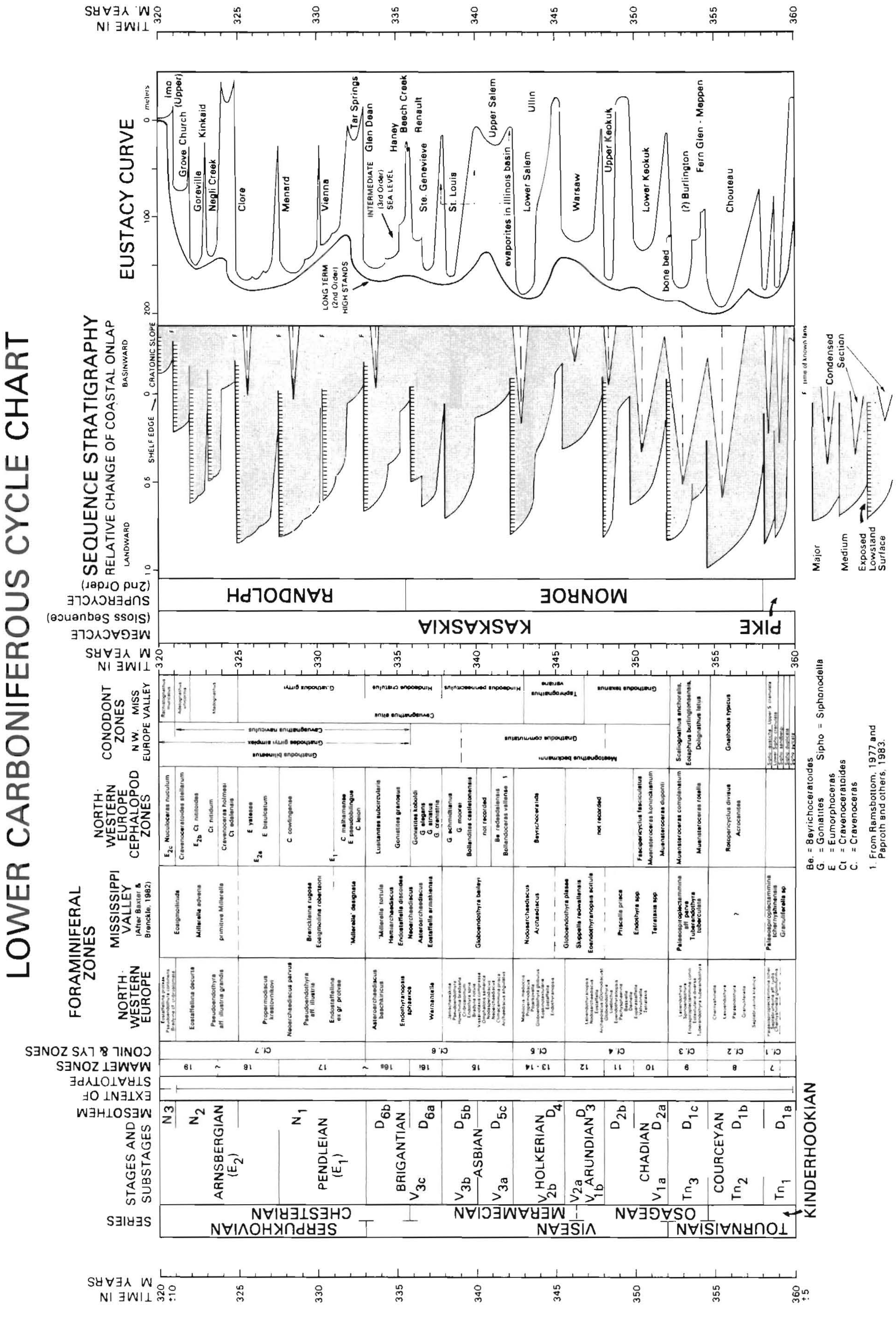

FIGURE 2. Lower Carboniferous (Mississippian) cycle chart plotted on a linear time scale. Subdivision of the Carboniferous System follows Bouroz and others (1979). Names applied to the third-order depositional sequences are taken from the highstand limestone in most cycles. See Ross and Ross, this volume, for discussion of biostratigraphy. (Modified from Ross and Ross, 1985b.)

late Paleozoic zonations, the type of zonations are similar to those used in the Mesozoic and Cenozoic charts. Most are a combination of overlapping range zones of genera and species. Some are species range zones. Others are based on first occurrences, and still others on last occurrences. In several faunas, genera have different stratigraphic ranges in different parts of the world because of paleogeographic provinciality (Ross and Ross, 1981, 1982, 1985a). This is comparable to differences in stratigraphic ranges of Cenozoic plankton because of latitudinal (or temperature) gradients or because of living in different ocean basins.

SEQUENCE NOMENCLATURE

The International Subcommission on Stratigraphic Classification (Salvador, 1987, p. 236) has recently expressed its decision that names of unconformity-bounded stratigraphic units..."require a separate and distinctive nomenclature" and..."should be formed of the name of an appropriate local geographic feature at or near the location where the unit is well developed, combined with a term that properly indicates the kind of stratigraphic unit." The subcommission further recommended these units be called "synthems" and, when needed and useful, be subdivided into "subsynthems" or combined into "supersynthems." The subcommission also mentioned "miosynthem" which they defined as "a relatively small, minor synthem within a larger synthem, but not a component of a hierarchy of unconformity-bounded units"....

Chang (1975) first used and clearly established "synthem" for a unit of approximately "System" level in size. This established a "synthem" as being of the same general magnitude as a "Sloss sequence" or a "Megacycle" as commonly used on sequence stratigraphy charts (Haq and others, 1987, this volume; Ross and Ross, this paper). Subsynthems, as subdivisions of synthems, are at the 'supercycle' or 'Second-order' level as used on the cycle charts. It is possible Chang's (1975) "interthem" should be used for this size unit. Unconformity-bounded units of a smaller size were referred to as "stratigraphic sequences" or "third-order cycles" (Vail and others, 1977; Mitchum and others, 1977; Haq, and others, 1987, and this volume), and as "mesothems" (Ramsbottom, 1977). These units are closely equivalent in size and concept to Moore's (1958) "megacyclothems". It is at this third-order level, based on current understanding, that unconformity-bounded units have become useful as worldwide stratigraphic tools and it is on this level that they are recognizable based on biostratigraphic information. These are the operational units for studying eustatic sea-level fluctuations.

Smaller unconformity-bounded units include "cyclothems" (Weller, 1930; Wanless and Shepard, 1936; Beerbower, 1964; Wanless, 1964) which have been referred to the level of fourth-order cycles. Heckel (1986) has demonstrated that numerous cycles of this level of magnitude are common within the Middle and Upper Pennsylvanian "megacycles" of Kansas. Brown (1969a,b, 1979) and Boardman and Malinky (1985) showed similar data for the Upper Pennsylvanian and Lower Permian of northcentral Texas, and Busch and Rollins (1984) used these types of units (and also smaller units) in Middle and Upper Pennsylvanian beds in Pennsylvania.

Smaller, or fifth-order, depositional cycles are recognized by Goodwin and Anderson (1985) which they refer to as punctuated aggradational cycles (or PAC's). These very small cycles commonly are contained within one depositional bed.

Names of Sequences

The sequence stratigraphy column names the large or first-order cycles after the Sloss (1963) sequences. These include parts of the Kaskaskia and Absaroka sequences.

The Mississippian (Lower Carboniferous) part of the Sloss Kaskaskia sequence includes the upper part of the Pike and all of Monroe and Randolph second-order cycles. The lowest, the Pike second-order sequence, is named for Pike County, Illinois. This is the type area of the Kinderhookian Series (Mississippian). This area also shows the relationships to the underlying predominantly dark shaly beds of the Upper Devonian which form the lower part of this second-order sequence.

The Monroe second-order sequence, is named for Monroe County, Illinois, where the largely shelf successions of the Chouteau though Ste. Genevieve carbonates are well exposed and where many of the formations in this part of the Mississippian have their type sections.

The Randolph second-order sequence is named for Randolph County, Illinois, where most of the formations from the Tar Springs through Grove Church interval of the Chesterian (Upper Mississippian) are exposed and named. This second-order sequence is a grouping of third-order sequences of mixed lithologies of sandstones, mostly dark shales, and limestones.

The lower part of the Sloss Absaroka sequence (Morrowan and Atokan strata) forms a second-order sequence called here the Pedregosa second-order sequence. It is based on nearly complete exposures of strata of this second-order sequence in and around Pedregosa basin in southwestern New Mexico and southeastern Arizona (Ross, 1973). These beds are well exposed in many basin ranges in that area and include shelf carbonates with well defined unconformities that pass into shallow carbonate basin facies with little break in deposition.

The succeeding second-order sequence includes beds of Desmoinesian, Missourian, Virgilian, and Bursum age. They also are well exposed in successions in the basin ranges in southeastern Arizona and southwestern New Mexico (Ross, 1973), and we apply the name Tombstone

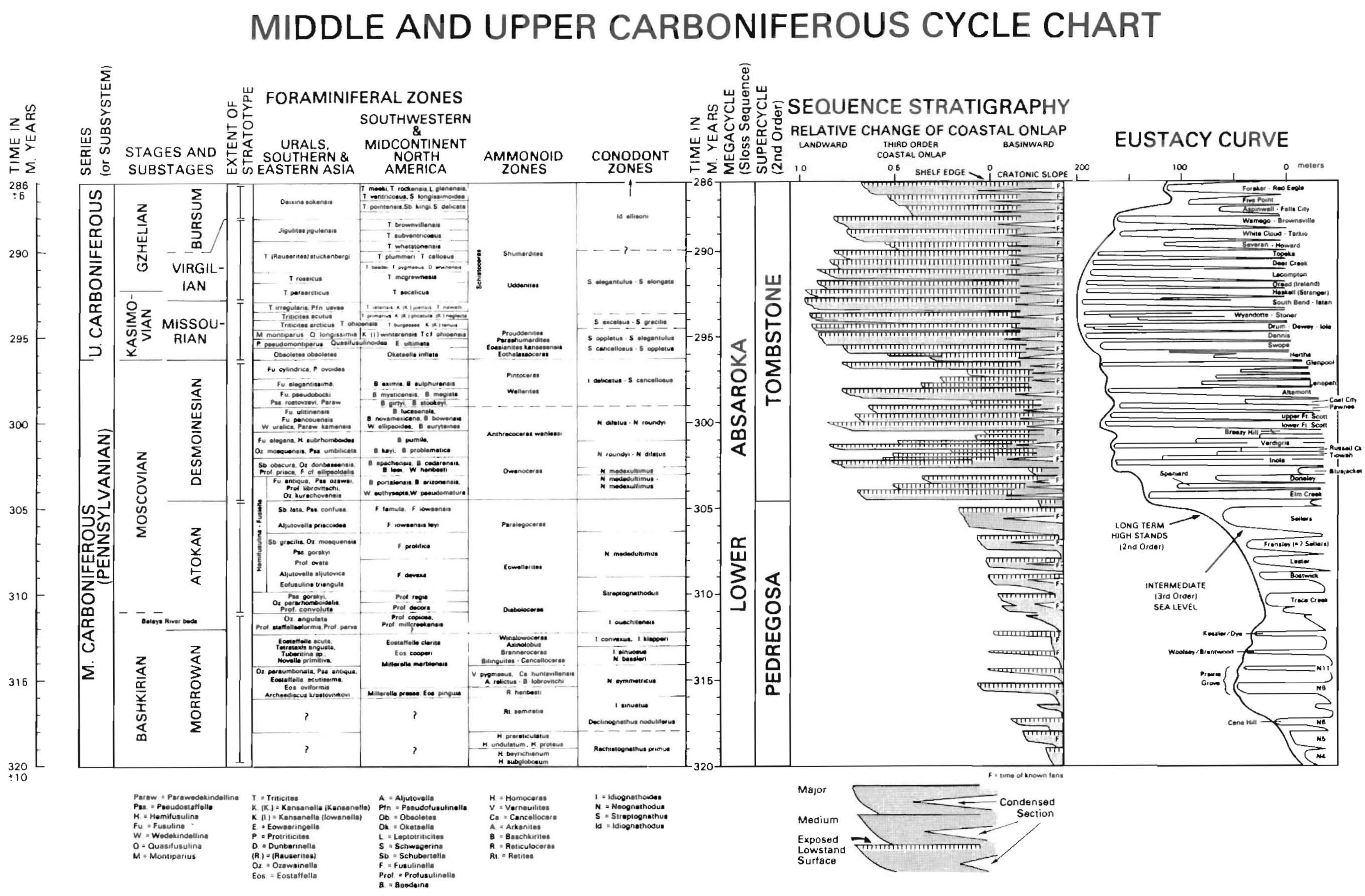

FIGURE 3. Middle Carboniferous (Lower and Middle Pennsylvanian) and Upper Carboniferous cycle chart comparing North American series and Russian Platform stages and substages with cycles of sea-level fluctuations and depositional sequences (See Fig. 2 for further explanation.) (Modified from Ross and Ross, 1985b.)

second-order sequence, named for Tombstone, Arizona, to these beds. It is possible that future studies will show that the lower Wolfcampian Neal Ranch should be included in this second-order sequence (rather than in the Transpecos second-order sequence) because Neal Ranch age rocks lie on top of the Tombstone and share many depositional features with it.

The Permian System is treated here as a large second-order cycle, although its lower one or two third-order cycles of the Wolfcamp (Neal Ranch) at its base are transitional with the Tombstone second-order cycle below, and its upper four third-order cycles (Djulfian) are transitional with nonmarine beds typical of the upper Absaroka (Triassic) nearly everywhere except in the Tethys area. Because this second-order cycle is well exposed and well studied in western Texas, we are calling it the Transpecos second-order cycle.

Third order Sequences and Sea Levels

The column showing a saw-tooth pattern is the third-order representation of relative changes in coastal onlap. Symbols on that curve show those sea-level changes that are considered to have fallen below the edge of stable cratonic margins and which were accompanied by the erosion of stream and river channels (type 1 unconformities). Other sea-level changes are considered to have fallen only to the outer portion of a cratonic shelf (type 2 unconformities). Prominent condensed sections are shown as unshaded wedges on the chart. Because these condensed sections typically separate the transgressive system tracts (below) from the highstand system tracts (above), we have not used a separate column for designating the system tracts. Only in the Pedregosa second-order cycle sequence are third-order shelf margin system tracts commonly preserved and these seem to be associated with river channels, coals, underclays, and terrestrial clastics higher on the shelf, suggesting a merging of type 1 and type 2 unconformities. As mentioned earlier, the present lack of precise absolute age assignments does not permit designation of specific ages for sequence boundaries and downlap surfaces so these ages are not included.

The eustacy curve column shows long term (second-order cycle) trends in sea-level change and shorter term (third-order) trends. Not all of the many smaller sea-level fluctuations which occur within the third-order fluctuation are shown. Some of these fourth-order trends have been documented in detail for parts of the Middle and Late Carboniferous by Heckel (1986). Suffice it to say, the third-order eustacy curve is in its self not a smooth sea-level curve, but internally these sea-level changes had minor fluctuations which rose and fell with hesitation, irregularly, and were associated with considerable lateral shifting of depositional environments both perpendicular and parallel to the shore.

Naming third-order depositional sequences (or cycles) is made more complicated because of the large number of sequences of this size which are known (60 to 70). Because the identification of each of these is based on interpretation of the strata and their fossils, it seems appropriate not to introduce a totally new and different set of stratigraphic place names lest the result be an unintelligible mire of names. The nomenclature that we have used (Ross and Ross, 1985) is based on naming each of the third-order sequences after the marine highstand limestone (or limestones in some cases) which commonly has a fossil fauna that aids in interregional correlation.

This has the advantages of using names that currently have an existing geographic basis, an established stratigraphic usage, and, often, a well known fauna associated with the name. In this system of nomenclature many depositional sequence names are taken from named members and beds. Their use as depositional sequence names can be clearly indicated by use of the words "depositional sequence" or "sequence" after the geographic part of the name. For some, an appropriate limestone name is not available and other well-known rock names have been used instead. Because of the distribution of available space on the charts, we show these names in the eustatic curve column, however, they may be used for both the sequence and the sea-level cycle because the concepts of third-order depositional sequences imply changes in sea level (either local or eustatic) during the deposition of a sequence. This makes use of the same name for the eustatic sea-level cycle as for the depositional sequence.

Thus, it is feasible to discuss the Menard Limestone as a lithologically identified rock unit, the Menard (third-order) depositional sequence as the name bearer for an unconformity-bounded unit with the Menard Limestone as its typical marine highstand limestone unit, or the Menard sea-level cycle.

This use contrasts with that of the geological Survey of Great Britian (George and others, 1976; Ramsbottom, 1973, 1977, 1979; Ramsbottom and others, 1978) which used a series of letters and numbers to identify individual third-order depositional sequences (mesothems).

CAUSES OF SEA-LEVEL CHANGES

A number of causal mechanisms have been proposed to explain sea-level changes, however, none of these by themselves completely explains the phenomena as we presently understand them in the late Paleozoic. The most difficult problems are those that deal with the apparent differences in duration and magnitude of the sea-level changes such as those of the Tournaisian-Visean compared to those of the Missourian-Virgilian. Several authors (Wanless and Shepard, 1936; Crowell, 1978; Frakes, 1979; Heckel, 1986; and others) have presented good arguments that cyclothem (or fourth-order) magnitude sea-level changes are likely associated with repeated glaciation in Gondwana and have durations that are similar to the long 0.4 million year

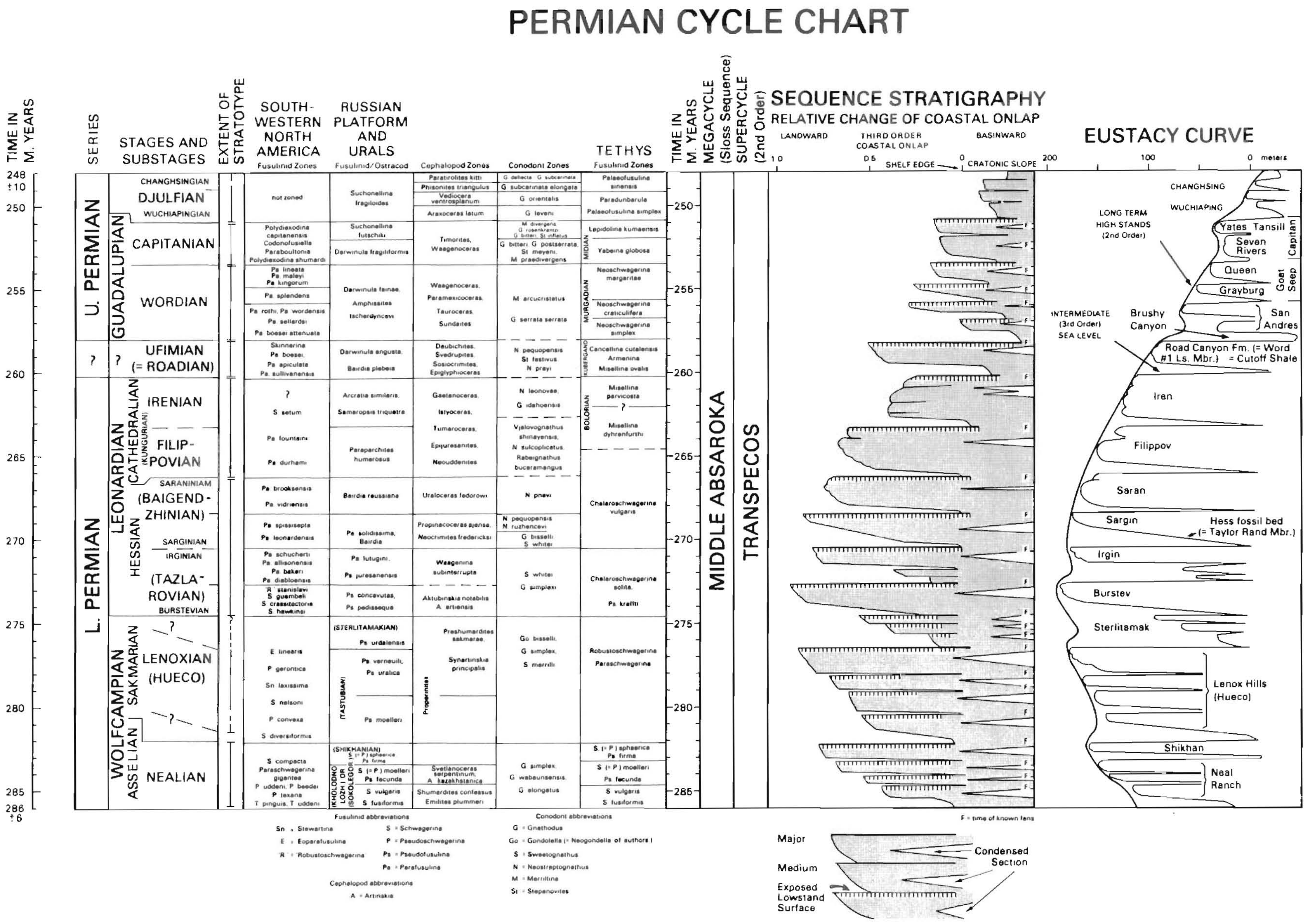

FIGURE 4. Permian cycle chart comparing North American and eastern Russian Platform - southern Urals series and stages and substages with cycles of sea-level fluctuations and depositional sequences (see Fig. 2 for further explanations). (Modified from Ross and Ross, 1985b.) (Djulfian also = Dzhulfian)

Milankovitch eccentricity cycle used by some to explain Pleistocene glaciation (Veevers and Powell, 1987).

In addition to the Milankovitch eccentricity cycles, other causes for glaciations and other causes for cyclic desposition have been proposed including changes in the geographic position of land areas, mountain building, climatic rainfall cycles, flow of ocean surface currents, and general climatic cooling. Each of these could and probably did make important contributions to late Paleozoic climates and temperature distributions.

On the other hand, when viewed on the level of third-order (mesothemic) cycles, or the even larger second-order (subsynthemic) cycles, it appears that other causes are superimposed on the smaller cycles. This suggests that the lack of a clear cut distinction between the different orders of cycles is the result of multiple causes which act in combination to determine a particular sea level at any one point in time. The shifts in patterns of magnitude, duration, and also in temperature (as indicated by the production of carbonate deposits) of these changes in eustacy strongly suggest that the relative importance of one causal mechanism relative to the importance of one or more other causal mechanisms probably changed with time.

Tectonic events, although associated with local sea level through uplift and downwarp, also involve segments of thrusting along orogenic belts which tend to add rock material to continental margins and remove it from ocean basins. This results in changes in the volume and shape of ocean basins and causes lowering of sea level. Sediments eroded from these uplifts gradually fill the basins and cause sea level to rise again. In this way, local tectonics may contribute to the fall and rise of eustacy.

The motion of crustal plates, in addition to being the driving force for orogenies, also is associated with sea-floor spreading and the rate of growth or subsidence of mid-oceanic ridges. Rapid rates mean hot, topographically high ridges, diminished ocean basin volume, and high sea levels. Donovan and Jones (1979) estimated that sea level could change by as much as 300m by this means, but at rates of about 1 cm/1000 years (or only 10m/m.y.). Pitman and Golovchenko (1983), who studied passive continental margins, summed up the problem of causal mechanisms stating that glacial fluctuation is, at present, the only known mechanism that could exceed sea level change rates of 1 cm/1000 years, but that other, still unknown, mechanisms may exist.

SUMMARY

Late Paleozoic depositional sequences include particularly good examples of transgressive and high stand system tracts and their bounding unconformities. Shelf margin wedges are only locally well preserved and formed during times of generally lower sea level, such as in the Morrowan.

Third-order depositional sequences include more than seventy Carboniferous and Permian cycles which may be grouped into six larger or second-order sequences. These six in turn are parts of the larger Kaskaskia and Absaroka sequences of Sloss.

The upper Tournaisian and lower and middle Visean sequences (Ramsbottom, 1973; Mamet, 1974; Conil and Lys, 1977; Armstrong and others, 1979, 1980; Paproth and others, 1983) have predominately carbonate transgressive and highstand systems tracts with basinward progradation of the carbonate facies in which the carbonate banks had steep outer slopes facing starved or semi-starved intracratonic basin. Isolated carbonate mounds and coalescing bioherms and mounds are common. Several carbonate producing communities were present and each formed distinctive depositional facies (for example, Tubiphytes mounds, bryozoan bioherms).

The amount of sea-level change in any one third-order cycle was commonly large, perhaps as much as 100 to 200 meters. This seems particularly evident in cycles from the upper part of the Lower Carboniferous (Chesterian) until well into the lower part of the Lower Permian (Wolfcampian) (Sinitsyn, 1975; Willman and others, 1975; Yablokov, 1975). Almost all these depositional sequences have at their base unconformities that display channelling and erosion on the shelves (type 1 unconformities).

A larger second-order cyclical sea-level trend is clearly superimposed on these depositional sequence cycles. This trend is shown by a general lowering of sea-level (Ross and Ross, 1981) around which third-order cycles are clustered in the lower part (Morrowan and Atokan) of the Middle Carboniferous. Only a few of those third order cycles reached the top of cratonic margin.

The upper part (Desmoinesian) of the Middle Carboniferous and the Upper Carboniferous are characterized by large, internally complex third-order fluctuations in sea-level with relatively short periodicity (Moore and others, 1951; Harrison and others, 1979). These sea levels reached high onto the shelves and fell to or below the shelf margin in relatively short periods of times (1 to 1.5 million years).

Although the lower Wolfcampian sequences are transitional with those of the Upper Carboniferous, by Leonardian time the duration of these sequences gradually became longer and the magnitude of the sea-level fluctuations decreased.

During the Leonardian and Guadalupian, the transgressive and highstand systems tracts again are locally dominated by carbonate facies. These are well developed and well exposed around the western and southern edges of the intracratonic Delaware Basin in west Texas and southeastern New Mexico. Although they include some carbonate transgressive systems tracts, most carbonate producing communities easily kept up with the rate of sea-level rise and the over-all depositional relations are highstand systems tract geometry with shallow, steep platform edges maintained throughout most rises of sea level. Carbonate shelf margin system tracts developed quickly when sea level did reach its high point in the cycle because most of the "accommodation volume" had been filled by transgressive carbonates as sea level rose. At the end of Guadalupian time, general (second-order) sea level had dropped to near the cratonic margin. The magnitude of the fluctuation also continued to decrease so that seas did not flood the shelves again during the Paleozoic.

The uppermost Permian (Djulfian) is represented by nonmarine or evaporite beds in most intracratonic basins (Kotljar and Stephanov, 1984), including the Delaware basin area, but is well defined with fossil zones along the cratonic margins of the Paleotethys (Noé, 1987), particularly in South China (Sheng, 1963; Chao, 1963). There, the faunas suggest very warm shelf waters and relatively low general sea levels at or near the shelf margins.

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BIOSTRATIGRAPHIC ZONATION OF LATE PALEOZOIC DEPOSITIONAL SEQUENCES

CHARLES A. ROSS[1] AND JUNE R. P. ROSS[2]

1 Chevron U.S.A., Inc., P. O. Box 1635, Houston, TX 77251
2 Department of Biology, Western Washington University, Bellingham, WA 98225

ABSTRACT

Correlation of more than seventy third-order depositional sequences in Carboniferous and Permian strata uses assemblage zones of warm water benthic and nektonic shelf faunas. These include calcareous foraminifers, bryozoans, conodonts, and ammonoids and represent tropical, subtropical, and warm temperate water faunas from carbonate shelves and adjacent cratonic basins.

After the Early Carboniferous faunal zonation was highly provincial and worldwide correlations of depositional sequences are based on interpretations of evolutionary lineages and depositional patterns in each province and in identifying times of limited dispersals between provinces.

Associated with these faunal zones there were times of expanded or reduced faunal diversities and temperature related latitudinal expansion or reduction of faunas. Higher sea levels during warmer times enhanced faunal evolution and diversity and lower sea levels during cooler times dampened evolution and diversity and resulted in many species becoming extinct.

After the Early Carboniferous, provincial faunal zonations and evolutionary lineages developed as a consequence of the formation of Lesser Pangaea. Further isolation of these provinces resulted in faunal realms in the middle Early Permian after the formation of Greater Pangaea.

INTRODUCTION

Establishing the timing of depositional events is critical in analyzing late Paleozoic sea-level fluctuations. Fortunately, a number of fossil groups have been studied in great detail so that most late Paleozoic stratigraphic successions around the world can be correlated with considerable precision. We present in this paper a brief summary of four fossil zonations that are commonly used for correlating late Paleozoic strata and try to evaluate the precision that is available. Two zonations are based on foraminifers and bryozoans which are marine benthic shelf organisms that showed preferences for different depositional environments. The other two zonations are based on cephalopods and conodonts which were nektonic and formed part of the ecosystem in the water column above these marine shelves. The empty shells of cephalopods may float and be carried by currents great distances beyond their habitat. Conodonts are elements from an organism about which we know virtually nothing of its life habits.

The following discussion is centered around the faunal zone columns on the Carboniferous and Permian cycle charts in the accompanying article on 'Late Paleozoic sea levels and depositional sequences' by Ross and Ross (this volume).

PALEOGEOGRAPHIC CHANGES

In order to understand and use late Paleozoic fossil zonations, it is important to keep in mind that changes in paleogeography, in climate and climatic fluctuations, and in ocean currents had great influences on the dispersal, evolution, and extinctions of these faunas (Ross and Ross, 1985a).

The assemblage of Lesser Pangaea at the end of the Early Carboniferous and the assemblage of Greater Pangaea in the middle of the Early Permian increasingly disrupted the tropical, subtropical, and warm temperate marine shelf faunas, giving rise to biogeographical provinces and finally to realms. These geographic changes took place over 100 million years in a series of steps (Figs. 1 to 4) that brought together first Gondwana and Euramerica and later added Angara to form the extremely large landmass Pangaea. Changes in land mass distribution also changed ocean basin shape and size, ocean current dynamics, temperatures, and ultimately changed climates and the distribution of climatic belts. Associated with these geographic changes were tectonic and orogenic events and a northward transport of Pangaea so that its northern marine shelf was moved into cooler and cooler waters. Cool ocean currents were redirected toward the equator along the western marine shelves of Pangaea and consistently westward directed warm equatorial currents flowed along the eastern marine shelves of Pangaea and fostered great faunal diversification in the Tethys region.

Gondwana was the largest of the cratonic blocks and was so large that some of its shelves were commonly within warm temperate and subtropical regions at the same time that other parts were at the southern pole. Euramerica was considerably smaller and was situated across the equator extending into the northern part of the tropical area. The ocean basin and its sediments between Euramerica and Gondwana were in the process of being deformed into the Hercynian-Appalachian-Ouachita-Marathon orogenic

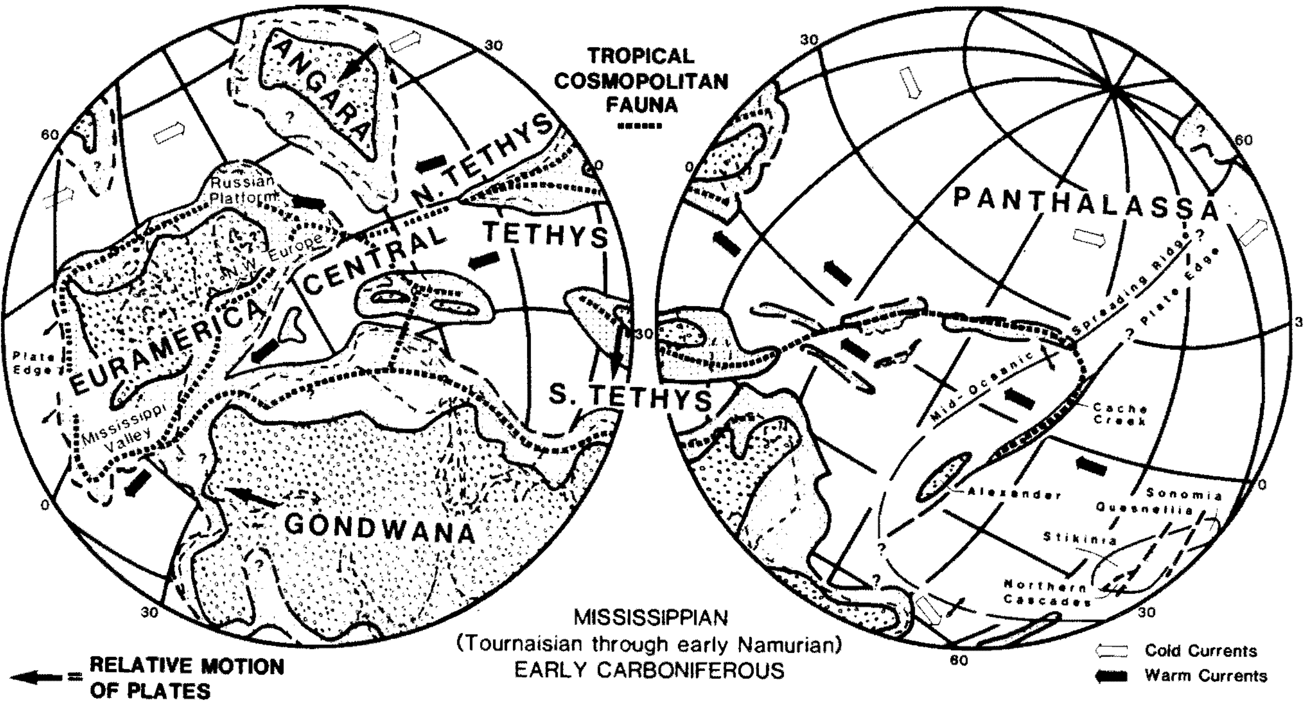

FIGURE 1. Paleogeographic map for the Tournaisian through early Namurian showing relative motions of plates (large black arrows), direction of ocean surface currents (small black arrows = warm currents; small open arrows = cold currents), and shallow shelf areas (stipple). (Revised from Ross and Ross, 1981.)

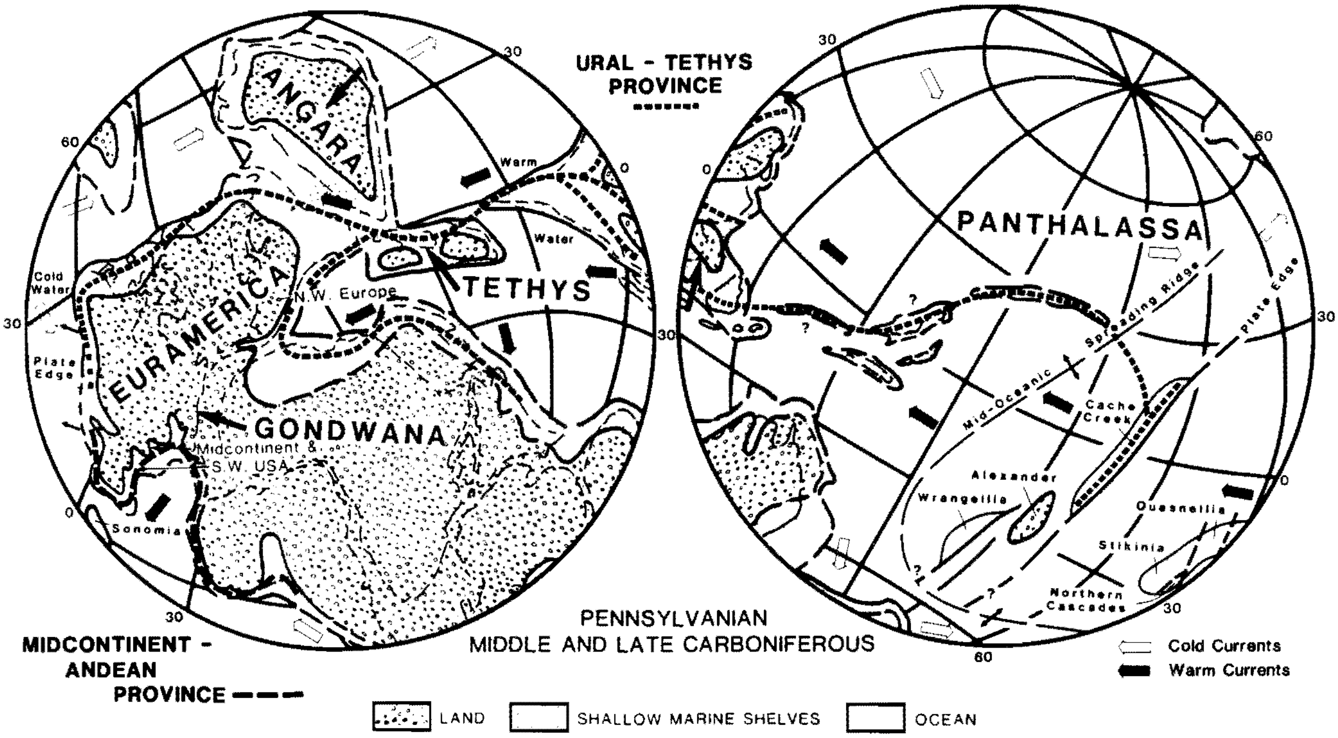

FIGURE 2. Paleogeographic map for the Middle and Upper Carboniferous. Gondwana rotated clockwise against Hercynian - Appalachian - Ouachita - Marathon orogenic belt during this time. (Revised from Ross and Ross, 1981.)

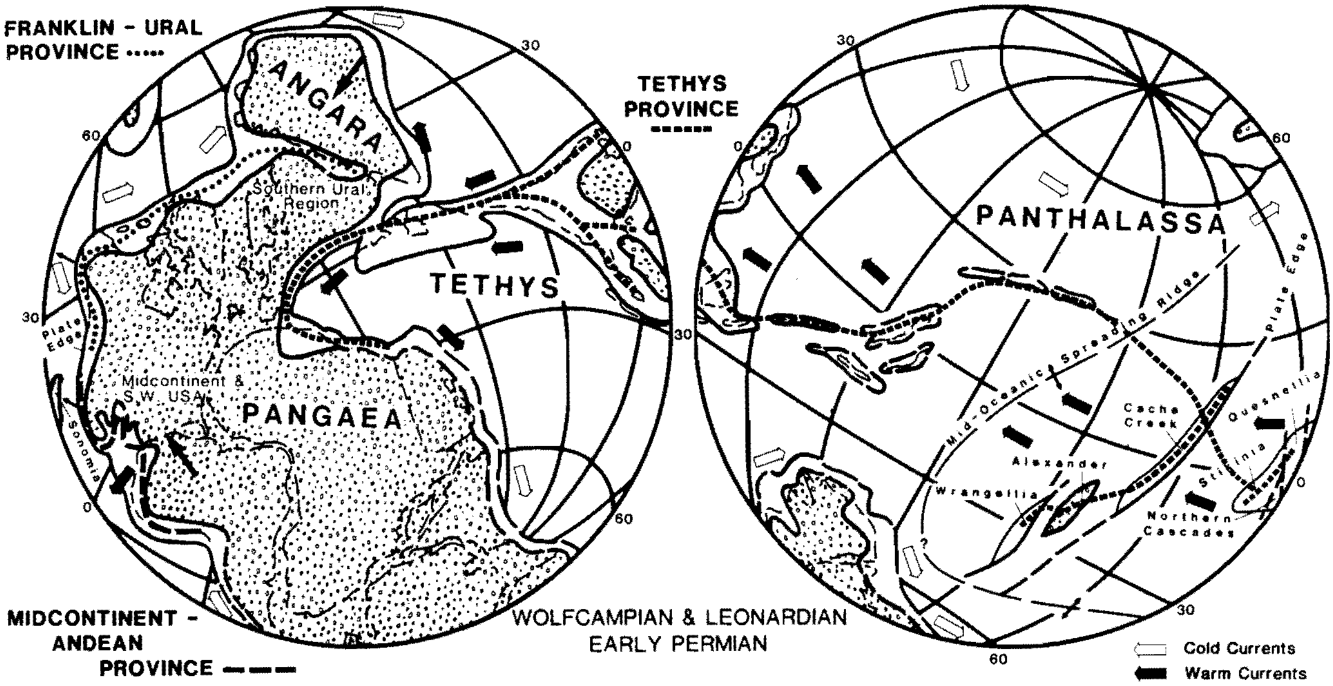

FIGURE 3. Paleogeographic map for the Early Permian shows the separation of the Ural region and Russian Platform from the PaleoTethys which dates from the Middle Leonardian. (Revised from Ross and Ross, 1981.)

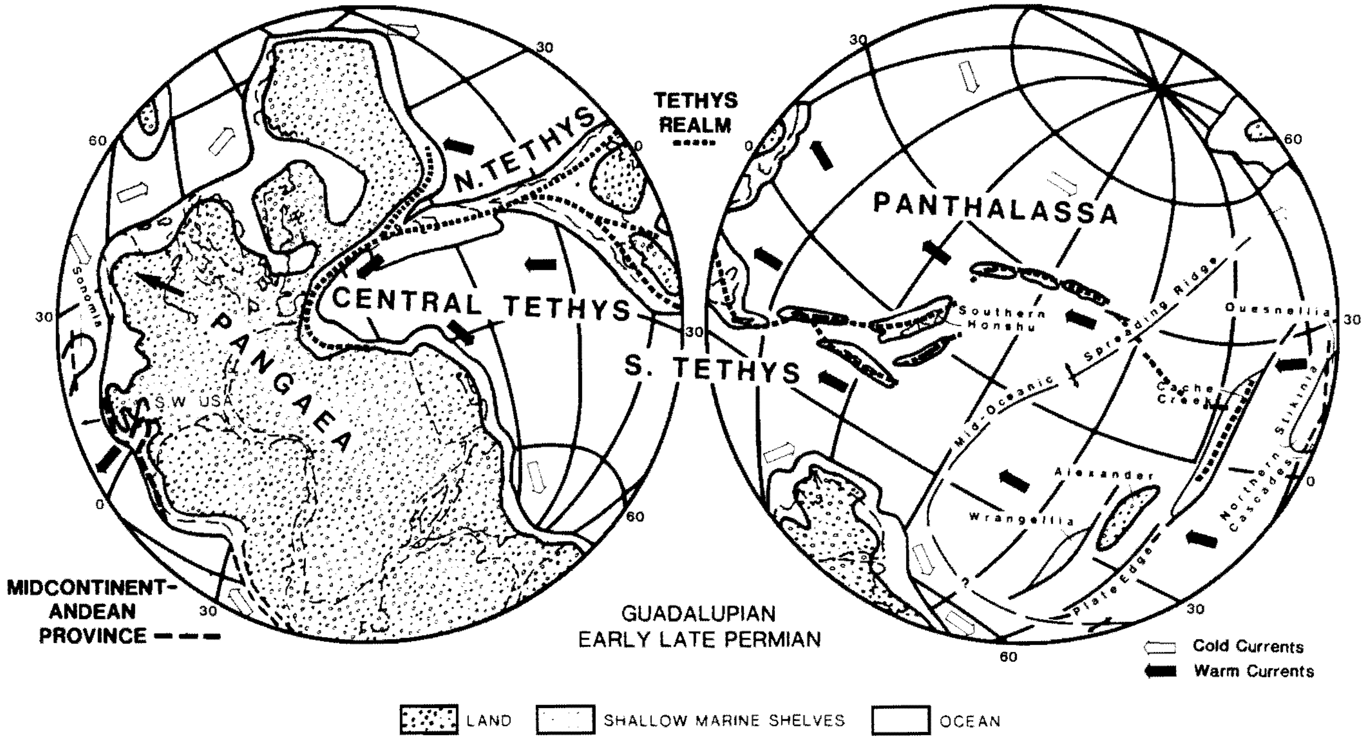

FIGURE 4. Paleogeographic map for the Guadalupian shows the position of Greater Pangaea extending across all the latitudinal climatic belts. (Revised from Ross and Ross, 1981.)

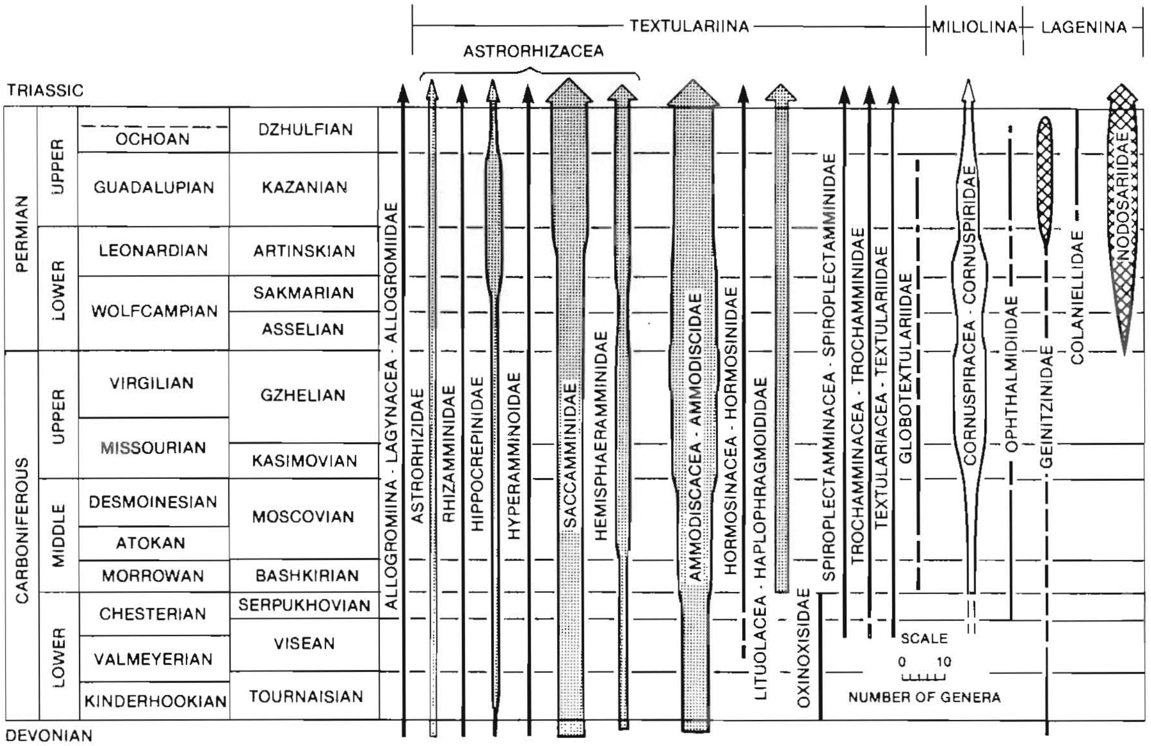

FIGURE 5. Stratigraphic ranges of Paleozoic families of Foraminiferida (data modified from Loeblich and Tappan, 1984).

belt and only the Acadian orogenic phase had been completed by the Early Carboniferous. In the middle portion of the belt, for example in Spain, central France, and the Maritime provinces of eastern Canada, several Armorican and Avalonian microcratons had been accreted to one or the other of the larger cratons during Devonian time.

Angara lay to the north and east of Euramerica in warm to cold temperate to cold latitudes. The Ural ocean basin separated Angara from eastern Euramerica during the Carboniferous.

The geographic position and makeup of the fourth major cratonic block, China, is less well known. It is made of a dozen significant cratons. South China and North China were separate in the Devonian. The Carboniferous faunas and floras are strongly indicative of a warm temperate to subtropical position and, for parts of South China, a tropical position during the Carboniferous and Permian. Western China appears formed of additional large stable blocks. Much about the paleogeographic history of the area remains to be discovered.

Each of these four major cratonic blocks have large areas of relatively undisturbed shelf deposits of late Paleozoic age which include thick carbonates, glacial-marine deposits, phosphatic sandstones, black shales, coals, and fluvial deltas, and data on these, in addition to fossil assemblages, provide the basis to reconstruct the paleogeographic relationships of these cratons with some confidence.

Ocean basins of the late Paleozoic are less known and less understood than the cratons. Active mid-ocean spreading centers during the Mesozoic and Cenozoic caused the accretion of late Paleozoic sea-floor sediments (and also some younger sediments) as structurally deformed margins to the Paleozoic cratons. The stratigraphy and fossils of these accreted margins are less thoroughly studied than those of the craton, and it has been recognized only since the early 1970's that these margins include rocks that have been transported long distances. The world ocean basin, PaleoPanthalassa, had a large western region, PaleoTethys, which contained numerous small island arcs and small to medium size cratons. Our reconstruction of central and eastern PaleoPanthalassa also suggests several oceanic plates (with island arcs and trenches) bounded by subduction zones, similar to those now in the western Pacific.

FAUNAL PROVINCES

Late in the Early Carboniferous (late Visean through Namurian B) Gondwana and Euramerica joined along the Hercynian-Appalachian-Marathon orogenic belt to form Lesser Pangaea. This divided the tropical marine shelf along the southern margin of Euramerica and eliminated the marine connection that had joined the shelf faunas of western Euramerica with those of the eastern PaleoTethys. Because parts of the coast of Gondwana remained near the south pole and in cold water, the two tropical shelf faunas became isolated, one on either side of Lesser Pangaea. Dispersals between them were infrequent along the only available warm temperate route on the northern (Franklinian) shelf of Euramerica.

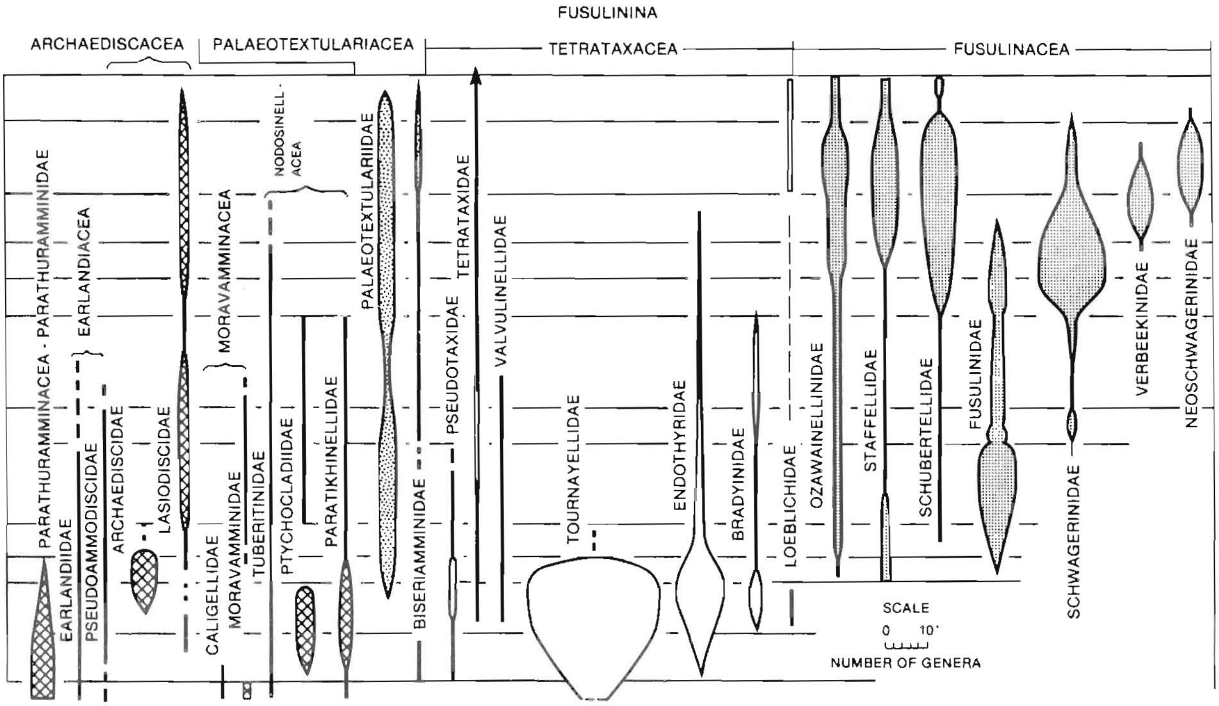

FIGURE 5. (Continued).

Equatorial warm water currents were diverted northward along the Russian Platform and southward along eastern Gondwana into much cooler regions and resulted in increased precipitation and a general climatic cooling. Glaciation in Gondwana resumed during the Early Carboniferous (Namurian) and continued with fluctuations until near the end of Early Permian.

During the Middle and Late Carboniferous and the earliest Early Permian, Lesser Pangaea continued on a northerly track and the north coast of Euramerica was gradually displaced to higher and cooler temperate latitudes. Shelf faunas occasionally dispersed by that route during times of higher sea-levels (which were also times of warmer water temperatures).

Another major geographic change resulted when Lesser Pangaea joined with Angara along the Ural orogenic belt in the middle Early Permian to form Greater Pangaea. Angara was being displaced from the north or northeast and its coasts extended well into cold temperate or boreal waters. By eliminating the currents of warm water from the Tethys area that had been flowing northward through the Ural marine connection, the northern shelf of Euramerica lost its physical connection to the Tethys and became much cooler. Its outpost populations of warm-adapted, shelf shelly faunas diminished greatly after this event.

The marine shelves of Greater Pangaea extended into both the cold northern and southern seas and effectively completed the isolation of the tropical marine shelf faunas on either side of Pangaea. Each fauna evolved independently with only occasional species dispersal by island "hopping" across Paleo-Panthalassa.

The assemblage of Greater Pangaea may have affected world climate by redirecting ocean surface currents into more latitudinally compartmentalized flow. Climates, in general, appear to have gradually warmed during the later Early Permian, suggesting a connection with paleogeographic changes.

FORAMINIFERAL ZONATION

The ranges and abundance of genera in late Paleozoic families of foraminifers are shown in Figure 5. Most of the Textulariina, particularly the Astrorhizacea are made up of conservative, long ranging genera and species. Many of these were common in the outer parts of shelves, on the slopes and in basins in dark fine clastic sediments. Except for the Kinderhookian and pre-Carboniferous strata, the astrorhizaceans are seldom used in correlation, and even in those rocks conodonts are more generally used. The first appearances of the Spiroplectamminidae and Trochamminidae contribute to Lower Carboniferous zonations. The Globotextulariidae is a minor family in the Middle and Upper Carboniferous and extends into the Upper Permian.

Among the calcareous Foraminifera, the first appearances of several genera in the Cornuspiridae are important in outer shelf (or "deeper-water") limestones in the middle and later part of the Lower Carboniferous and into the Lower Permian. Genera and species were long-ranging so that first appearances are used.

In the Lagenina, three families make their first appearance in the late Paleozoic. The Geinitzinidae and Colaniellidae are restricted to Upper Paleozoic beds and the Nodosariidae, which begin in the Early Permian, extend into the Mesozoic and Cenozoic. These are warm water families and the Nodosariidae, although common in darker gray basinal limestones, may have been transported there from other habitats.

The Fusulinina superfamilies Archaediscacea, Palaeotextulariacea, Tetrataxacea, and Fusulinacea had the most rapid diversification and have been the basis for most foraminiferal zonal schemes. These superfamilies were adapted to warm (tropical to subtropical), shallow water, carbonate environments and commonly were prolific. Some may have been host to photosymbionts. For Lower Carboniferous zones, genera and families of the Archaediscidea, Tetrataxacea, and primitive Fusulinacea are most useful.

Foraminiferal zonation of Mississippian carbonates was initially developed in the 1950's and 1960's for the Russian Platform, Ural, and Donetz Basin areas and later extended to other parts of the Soviet Union and has been summarized by Aisenverg (1964), Aisenverg and others (1968), Bogush and Juferev (1966), Brazhnikova and others (1967), Brazhnikova and Vdovenko (1973), Einor (1973), Ganelina (1956), Grozdilova (1966), Grozdilova and Lebedeva (1961), Lipina (1964, 1973), Lipina and Reitlinger (1971), Malakhova and Chuvaschov (1973), Vdovenko (1961) and many others. This scheme is based on identification of characteristic assemblages of species which form biozones.

In northwestern Europe the zonation of calcareous foraminifers in the type areas of the Tournaisian and Visean, was worked out by Conil and Lys (1964, 1973 and in Paproth and others, 1983), Mamet (see summary Mamet, 1977), and others. Mamet (1976) and Mamet and Skipp (1971) carried this zonation into the Cordillera region of western Canada and the United States (see also Skipp, 1969; Sando and others, 1969). Because in the Mississippian these areas were parts of the same large tropical faunal province as the Russian Platform, Donetz Basin, and South China, the Tournaisian and Visean foraminiferal zones are widespread and readily recognized. Mamet's early emphasis was on abundances (acme zones), sudden appearances, and evolutionary bursts of species in these shallow water, carbonate faunas. These differences in abundances may represent ecologically widespread features in northwestern Europe, however, other paleontologists in other parts of the world, particularly in the type area of the Mississippian (in the Upper Mississippi Valley) have found difficulty in using these zones as they were originally defined.

Mamet (1977) in summarizing his zonal scheme pointed out that the PaleoTethys during the Tournaisian and Visean had greater diversity than the nonTethys areas and that the foraminifers show paleolatitudinal diversity gradients. Although the key index fossils bridge these latitudinal gradients, the dispersal of many of the genera, particularly among the Bradyinidae and Palaeotextulariidae, to different parts of the PaleoTethys and to other shelf areas was in a heterochronic series of steps. Mamet (1977) suggested that island arcs in PaleoPanthalassa (that now form accreted terranes of the western Cordillera of North America), which have a mixture of species normally endemic to either the PaleoTethys or cratonic North America, served to integrate "regionally distinct biostratigraphic systems." Because phylogenetic faunal sequences are most complete in the PaleoTethys, Mamet (1977) believed most lineages originated there and later dispersed to other areas in an irregular pattern of abrupt first appearances without obvious endemic ancestors.

Baxter and others (1979), Baxter and Brenckle (1982), Brenckle and others (1982) have identified many of the foraminiferal assemblages from the region of the type Mississippian Subsystem. The Gilmore City Limestone and Humbolt Oolite of northcentral Iowa contains middle and late Tournaisian assemblages similar to those of the North American Cordilleran area and are not comparable to those of the Mississippi Valley sections. Scattered lower Visean foraminifers, such as *Priscella* gp. *prisca* and primitive *Tetrataxis*, occur in the Lower Keokuk Limestone. The upper part of the Keokuk also includes *Eoendothyranopsis*, primitive *Globoendothyra* gp. *tomiliensis* and the algae *Koninckopora tenuiramosa* and are suggestive middle Visean age.

Late middle Visean foraminifers appear in the Salem Limestone and include *Archaediscus*, *Nodosarchaediscus*, *Globoendothyra baileyi*, and continuations of primitive forms of *Eoendothyranopsis* and *Globoendothyra*.

The lower part of the St. Louis Limestone (below the breccia beds) has *Eoendothyranopsis* gp. *ermakiensis*, rare *Eostaffella* and complex-walled *Septabrunsiina*. First occurrences of *Archaediscus angulatus*, calcitornellids and calcivertellids are in the upper St. Louis. Baxter and others (1979) consider this an assemblage from the lower and middle part of the upper Visean. Ste. Genevieve foraminifers include the stellate archaediscids *Neoarchaediscus* and *Asteroarchaediscus* and *Endostaffella discoidea*. *Hemiarchaediscus*? is added to this assemblage in the lower Chesterian. This assemblage is of late, late Visean age (V_{3c}). *Endostaffella discoidea* disappears in the Glen Dean Limestone and its place in the upper Glen Dean is taken by primitive millerellid-like *"Millerella" tortula* and *"M." designata* and by *Eostaffella*.

The Menard Limestone sees the introduction of eosigmoilinids represented by the first appearance of *Brenckleina rugosa* and *Eosigmoilina robertsoni*. The highest Chesterian foraminiferal assemblage is from the Kinkaid Limestone and contains *"Millerella" cooperi* and small, primitive true *Millerella*.

Baxter and Brenckle (1982) place the Visean-Namurian boundary at the base of the

Menard Limestone, a position that most others would consider too high. We have retained a more traditional correlation and placed it one cycle lower just above the Glen Dean Limestone at the base of the Cypress Sandstone. Rich (1980, 1986) has established a useful foraminiferal zonation for Chesterian strata in the Black Warrior basin of northern Georgia and Alabama.

As with other shallow shelf faunas, foraminifers were greatly reduced in diversity by the end of the Chesterian and there was a brief time when only the Archaediscidae were abundant. Although a number of genera and most families did survive into the Middle Carboniferous, their importance in carbonate-producing communities was replaced by new genera and families in the Fusulinacea.

The Middle Carboniferous to Late Permian foraminiferal zonations are based on a succession of genera and species of fusulinaceans (Dalmatskaya and others, 1961; Douglass, 1977; Dunbar and Skinner, 1937; Kalmykova, 1967; Kanmera and others, 1976; Kotljar and Stephanov, 1984; Leven, 1967; Leven andShcherbovich, 1978; Ozawa, 1970; Ross 1967; Rozovskaya, 1975; Sheng, 1963; Skinner and Wilde, 1965; Thompson, 1964; Toriyama, 1967; and many others). A few of the lowest Middle Carboniferous zonal fossils are species that had ancestors in the upper Visean and lower Namurian (e.g., *Eostaffella*, *Millerella*, and *Pseudoendothyra*). The Morrowan foraminifers lack forms that were elongated along the axis of coiling.

The overlying Atokan had the beginnings of several very successful lineages. Species of *Profusulinella* followed rapidly by species of *Fusulinella* formed the nucleus of some shallow shelf, ecologically stable communities. These two genera persisted through the Middle Carboniferous in the Russian Platform and PaleoTethys. However, in Midcontinent North America, these same two genera had short stratigraphic ranges that only just overlapped and *Fusulinella* gave rise to *Beedeina* and *Wedekindellina* with only minor range overlap in the lower part of the Desmoinesian. In addition, on the Russian Platform *Fusulinella* gave rise to *Fusulina* (s.s.) which briefly was common there, but which was virtually unknown in the North American Midcontinent. The Russian Platform also has a number of genera which are known from the PaleoTethys, but not from the Midcontinent, for example *Eofusulina*, *Verella*, *Hemifusulina*, and *Aljutovella*.

The highest two or three depositional sequence in the Middle Carboniferous show a decrease in fusulinacean diversity and the highest Middle Carboniferous sequence in the Midcontinent appears to lack fusulinaceans. In the PaleoTethys and the Russian Platform, these cycles had a few new genera that appearred briefly, such as *Putrella*, *Pseudotriticites*, and early species of *Quasifusulinoides* and were forerunners of Late Carboniferous fusulinaceans.

The nearly complete replacement of Middle Carboniferous fusulinacean genera by new genera in the Late Carboniferous is well documented by Rozovskaya (1975) for the Soviet Union and by Thompson (1957) and Thompson and others (1956) for the Midcontinent area of North America. Some of these new fusulinids may be widespread, but this is largely concealed by the use of different generic names in the Midcontinent of North America and in the Russian platform. For example, the lineage at the base of the Upper Carboniferous of the Russian Platform and Ural region called *Fusulinella* by Soviet paleontologists is probably represented by *Eowaeringella* in the Midcontinent area; *Protriticites* from lower Upper Carboniferous beds of the Russian Platform and Ural region is remarkably similar to *Kansanella* *(Iowanella)* from the same stratigraphic interval in the Midcontinent. Also from the lower beds of the Upper Carboniferous on the Russian Platform, *Quasifusulinoides* has the same characteristics as *Fusulina* *fallensis* from the lower part of the Missourian of the Midcontinent region. Slighty higher *Obsoletes* *obsoletes* from the Russian Platform and Ural region is transitional to the genus *Triticites* in the same way that the lowest occurrences of *Triticites* *ohioensis* in the Midcontinent area are transitional with well developed species of *Triticites*.

In the middle part of the Upper Carboniferous (Upper Missourian - Lower Virgilian) in most parts of the world, well developed zones are based on similar stages of species evolution within the genus *Triticites*. Although different faunal provinces existed, the endemic lineages were not as dominant in the fusulinacean communities as those having more cosmopolitan phylogenies and more frequent dispersals.

In the middle and upper Virgilian and Bursum, this situation changed, faunas became much more endemic, and the endemic species dominated most fusulinacean communities in both provinces. Genera such as *Daixina*, *Rauserites*, *Rugosofusulina*, *Quasifusulina*, and *Fusulinella* were common and widespread from the Franklinian shelf (northern Canada) through the Ural region and into the PaleoTethys. In the Midcontinent region, *Dunbarinella*, *Leptotriticites* and several endemic species lineages of large thick-walled *Triticites* dominated. Three representatives of Franklinian shelf and Ural region faunas appear briefly in the Midcontinent and southwestern North American regions; first, *Waeringella* in the middle Virgilian, and then *Pseudofusulinella* and *Rugosofusulina* in uppermost Virgilian, Bursum and early Wolfcampian strata. The first two were dispersed from the Franklinian shelf and derived from the long lineages of *Fusulinella* that were common there during this time. *Rugosofusulina* dispersed from the Russian Platform.

Foraminiferal zonation of the Lower Permian is complicated by a rapid increase in the number of genera, a continuation of provincial distributions, and a reduction in the dispersal events between geographic distant areas. Two periods of rapid fusulinacean evolution in the Lower Permian

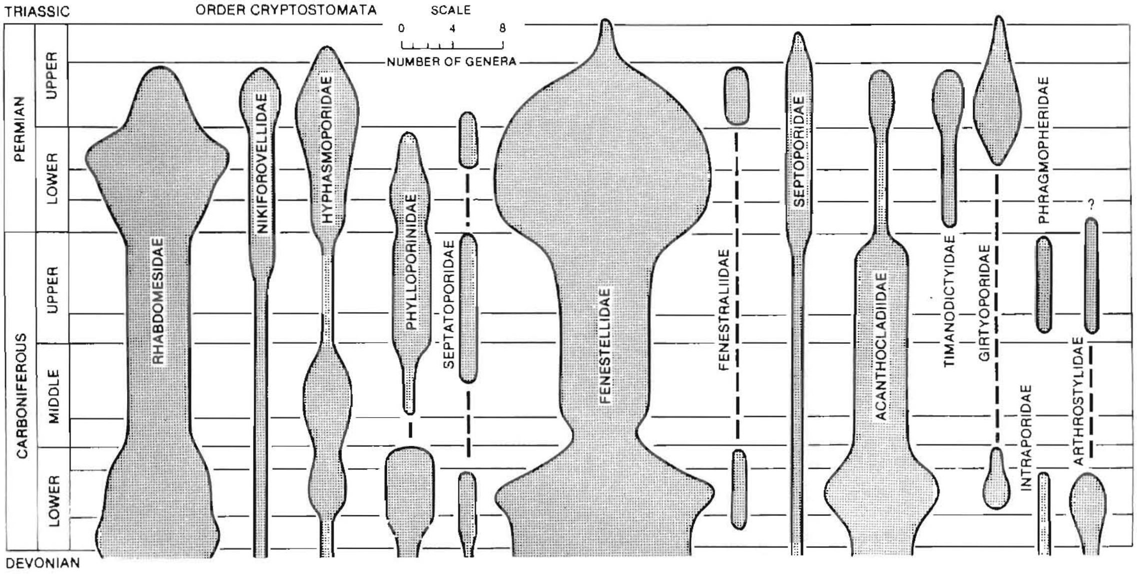

FIGURE 6A. Stratigraphic ranges of Paleozoic families of bryozoans in the order Cryptostomata (after Ross, 1981).

aided zonation. The earliest involved the Schwagerinidae which gave rise to such diverse genera as the subcylindrical Eoparafusulina and Monodiexodina and the subspherical (possibly planktonic or nektonic genera) Pseudoschwagerina, Paraschwagerina, Occidentoschwagerina, Sphaeroschwagerina, Zellia, and Robustoschwagerina, as well as a great number of other genera including Pseudofusulina and Parafusulina. Species of Schwagerinidae form most of the zonal fossils for the Lower Permian. Although some of these may be associated with reef flanks, most lived in less agitated shallow water.

About the middle of the Early Permian, the PaleoTethys region became faunally isolated from the Ural region and a second period of rapid evolution occurred, this time mainly in the Staffellidae and its descendants, the Verbeekinidae and Neoschwagerinidae. These formed a rapid succession of specific and generic zones in the PaleoTethys along with rapid evolution of endemic species in genera of Schwagerinidae. The Neoschwagerinidae were associated with wave agitated reefal features. The Verbeekinidae and Schwagerinidae lived in less strongly agitated environments. Species of Staffellidae were usually in the shallow shelf shallow lagoonal facies. These facies-related PaleoTethyan faunas are made even more difficult to study because later tectonic events commonly have structurally disturbed the facies and stratigraphic relationships.

Upper Permian fusulinacean zones are divided into those of Guadalupian age and those that are younger (Djulfian or latest Permian). Guadalupian zones are based on ranges of species of Parafusulina, Skinnerina, and Polydiexodina in the Midcontinent and southwestern North America realm and on ranges of genera and species of the Neoschwagerinidae, Verbeekinidae, Staffellidae, and Schubertellidae in the PaleoTethyan faunal realm. Ozawa (1970) devised a particularly useful species lineage zonation for some of the Neoschwagerinidae in southeast Asia starting with Misellina minor and leading to Lepidolina kumaensis. Near the end of Guadalupian, most large fusulinaceans became extinct. Lepidolina kumaensis was apparently the last survivor and occurs in strata considered earliest Djulfian by Japanese geologists.

Only the Staffellidae and relatively small Schubertellidae and Ozawainellidae survived through the Djulfian during which time a number of new genera evolved. At the species level, the Djulfian may be subdivided into zones using Paleofusulina and Codonofusiella. These fusulinid faunas are part of the PaleoTethyan faunal province and are not well distributed even within that realm. Other foraminifers, including species of Colaniella, Lasiodiscus, Abadehella, Pachyphloia, Nodosaria, and Paraglobivalvulina, are used to supplement this zonation in parts of the PaleoTethyan realm (Ishii and others, 1975; Okimura and others, 1985). None of the fusulinaceans range into Triassic strata. The cause of the extinction of most fusulinid families near the end of the Guadalupian and their final extinction at the end of the Djulfian remains unknown.

BRYOZOAN ZONATION

The stratigraphic distribution of bryozoan families for the late Paleozoic is shown in Figure 6 A, B. The extinction and evolution of genera are illustrated by the width of the outline of the family ranges. At the level of families, bryozoans demonstrate a large diversity of forms in the Tournaisian and a slightly larger diversity in the early and middle Visean (Ross, 1981a, b; 1984). This speciation was followed by a rapid, progressive decline in diversity late in the Visean through the Serpukhovian and into the Bashkirian. The Moscovian shows only a slight

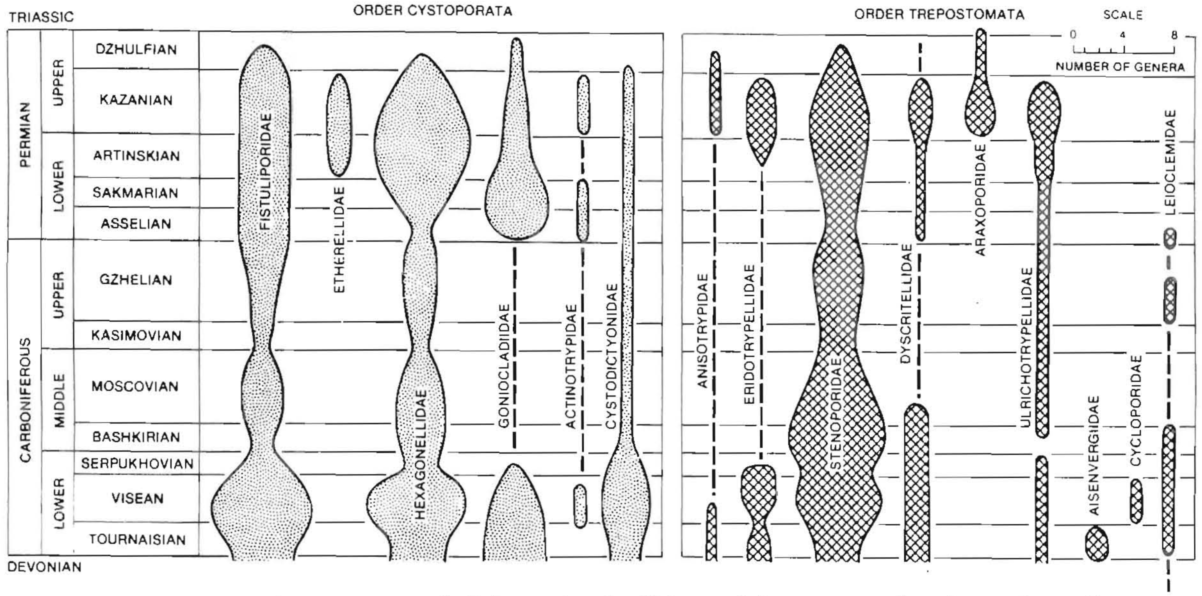

FIGURE 6B. Stratigraphic ranges of Paleozoic families of bryozoans in the orders Cystoporata and Trepostomata (after Ross, 1981).

increase in diversity and near the end of that stage shows another decline in several families. The Upper Carboniferous had a fairly stable, if somewhat low, diversity.

Beginning in the Asselian, new genera were gradually added to a number of families, particular to the Hexagonellidae, Goniocladiiae, Rhabdomesidae, Hyphasmoporidae, and Fenestellidae (Ross, 1978, 1979). A few families having only one or two genera in the Lower Carboniferous, such as the Actinotrypidae, Anisotrypidae, Eridotrypellidae, Fenestraliidae, Girtyoporidae, and others, are not known from strata of Middle or Upper Carboniferous ages and presumably survived in paleogeographic refuges that are not known at present. Genera of these families start to reappear in the Asselian and this repopulation is completed by the beginning of the Guadalupian (= Kazanian). Accompanying these Early Permian reoccurrences are a few new and distinctive families, the Timanodictyidae, Etherellidae, and Araxoporidae.

The Guadalupian (Kazanian) bryozoans show a change in faunal dominance as more and more genera of Rhabdomesidae and Fenestellidae became extinct. All the other bryozoan families declined in diversity and abundance during the Guadalupian and only nine families survived into the lower part of the latest Permian (Djulfian) and of those only four into the upper part. Only a few Djulfian genera survived into the early Triassic before becoming extinct.

These patterns of family level diversity are of interest because they reflect change that occurred at the same times as those seen in the second-order cycle sea-level curve and also many of the changes in diversity seen in other groups of marine organisms.

Studies of the stratigraphic ranges and paleogeographic distribution of individual genera (Ross, 1978, 1979, 1981, 1984 and Ross and Ross, 1981) illustrate that genera commonly have significantly different stratigraphic ranges in different faunal provinces and that the times of dispersals from one area to another are more frequent than just at the beginning or end of a particular stage. For example, there were three or four times of dispersal during the Visean and these may relate to four sea-level highstands (see Lower Carboniferous Cycle Chart, this volume). The Middle and Late Carboniferous data show similar frequent dispersal patterns. In the Permian, there are three obvious times of changes in paleogeographic dispersals, one within the lower part of the Artinskian, another at the end of the Artinskian (or perhaps the Ufimian) and the third at the end of the Guadalupian. These few episodes of dispersals suggest that Permian bryozoan were less influenced by third-order sea-level changes in their dispersals than their Carboniferous ancestors.

Detailed studies of bryozoan species distributions in closely controlled stratigraphic successions aid in identifying different depositional sequences. Trizna (1958) in her study of Lower Carboniferous bryozoans of the Kuznets Basin (Fig. 7) found six assemblages which differed in species associations and in species abundances. Although Trizna's data does not have detailed bed by bed distributions, we have reconstructed such a distribution using stratigraphic and faunal data from Selyatitsky and others (1975). The Lower Carboniferous of the Kuznets Basin includes many sandstones, evaporites, tuffs, and dolostones. There also are widespread fossiliferous Tournaisian limestones and, in the northwestern part of the basin, some fossiliferous Visean limestones. As with the Moscow basin, the succession has numerous unconformities that separate non-fossiliferous and some fossiliferous sequences. Using reported foraminiferal data, we have assigned the bryozoan assemblages to the Russian

Platform and northwest Europe depositional sequences.

CONODONTS

Conodonts occur in a great variety of lithologies that represent many contrasting depositional environments and indicate that they were nektonic (Seddon and Sweet, 1971). Shallow water deposits typically have less diversity of conodont elements than deeper water deposits, and in some parts of the Paleozoic succession, an approximate depth zonation of conodont-bearing organisms is possible (Merrill, 1972, 1975).

Conodont zonations for the late Paleozoic, as with most other biostratigraphic zonations, have regional differences which probably relate to environmental and geographical separation. Conodonts show these regional differences less distinctly than some other faunal groups. They are widely used for correlating parts of the Carboniferous and Permian where they are extremely valuable.

The conodont zones of the latest Devonian and earliest Mississippian (Kinderhookian) are known in remarkable detail. In western North America, Sandberg (in Sando, 1985a) has worked out a scheme based on the first occurrences of species of Siphonodella, Gnathodus, Scaliognathodus and Dolignathodus from the later part of the Devonian to about the middle of the Visean. The remainder of the Lower Carboniferous is based on assemblage zones utilizing a number species of Cavusgnathus and Hindeodus, Taphrognathus varians, Gnathodus girtyi, Adetognathus unicornis and Rachistognathus muricatus. Although many of these assemblages are known from Europe, they are not represented well in the Mississippian type area where a generally similar set of assemblage zones using different species and some different genera is used (Collinson and others, 1971).

The Kinderhookian part of the conodont zonation is particularly detailed in that it adds three zones below the first widely traceable foraminiferal zone. The middle Visean through lower Namurian part of the conodont zonation becomes less detailed and the Mississippi Valley and Arkansas sections have more provincial faunas than earlier ones.

Early Pennsylvanian Morrowan zones again show a pattern of assemblage zones of short duration. Lane and others (1971) and Dunn (1974) and Lane and Straka (1974) generally agree on the ranges of most genera and species, however, they use quite different zonal units for the upper part of the Morrowan. As with the detailed zones near the base of the Tournaisian, these conodont zones are considerably more detailed for the Morrowan than are foraminiferal zones.

Atokan, Desmoinesian, and Upper Carboniferous conodont zones have been reviewed by Merrill (1972, 1975). These assemblage zones have species with considerable overlap in species morphologies and require large numbers of specimens to be usable. Permian conodonts are more diverse. Clark and Behnken (1971, 1979) and Behnken (1975) studied their ranges in some detail from both the Great Basin area of the western United States and from west Texas. Kozur (1978) examined conodont zones in the Permian of Europe. Sweet (1970) determined a number of upper Permian zones below the base of the Triassic in the PaleoTethys area. Wang and Wang (1981) were able to apply part of these two sets of ranges to a study of Chinese Permian conodonts and established preliminary zones. The Permian conodont zonation shown on the cycle chart is a compilation of this data. It is not complete and many zonal boundaries are likely to be moved as additional occurrences fill in the ranges.

The Nealian (= Neal Ranch age beds) at the base of the Lower Permian has a conodont assemblage that contains many Late Carboniferous holdovers. The Lenox Hills age beds (Lenoxian) has the additional species Sweetognathus merrilli. In the Leonardian, a number of new species appear in fairly rapid succession to form the basis for a preliminary zonation. Neostreptognathus, Gnathodus, and Merrillina contribute most of the guide species to those zones that are younger than Leonardian.

CEPHALOPODS

Ammonoid cephalopods were one of the first groups to be used for a detailed subdivision of the Carboniferous (see summaries by Paproth and others, 1983; Ramsbottom and Saunders, 1984; Miller and Furnish, 1958; Ruzhentsev, 1960, 1962; Saunders and others, 1979) and Permian (see summaries by Chao, 1965; Furnish, 1973; Miller and Furnish, 1940; Smith, 1929). They show very rapid evolutionary changes that are placed in well defined lineages. Their nektonic habitats were apparently depth partitioned because deeper water lithofacies tend to have increasingly diverse faunas. In contrast to the conodonts, which also show evidence of depth stratification, empty shells of ammonoids floated well and were commonly distributed great distances from their actual habitat range. Because of the shell's propensity to float after the death of the animal, many empty shells were deposited as wind and current flotsam on beaches in death assemblages representing mixed communities (or depth) faunas. In addition, ammonoids are a relatively common fossil (usually compressed) in black shales in environments of slow deposition. They are also known in some "deeper" water carbonate debris and turbidite beds, but as scattered individuals.

Ammonoids show less provincialism than some of the benthic carbonate shelf faunas, probably as a result of their nektonic habitats by which some genera occupied deeper, more widely distributed cooler temperature water masses. Those ammonoids that display the greatest provinciality, such as the Perrinidea, likely were adapted to warm surface waters and had temperature restrictions on their dispersals. Distribution

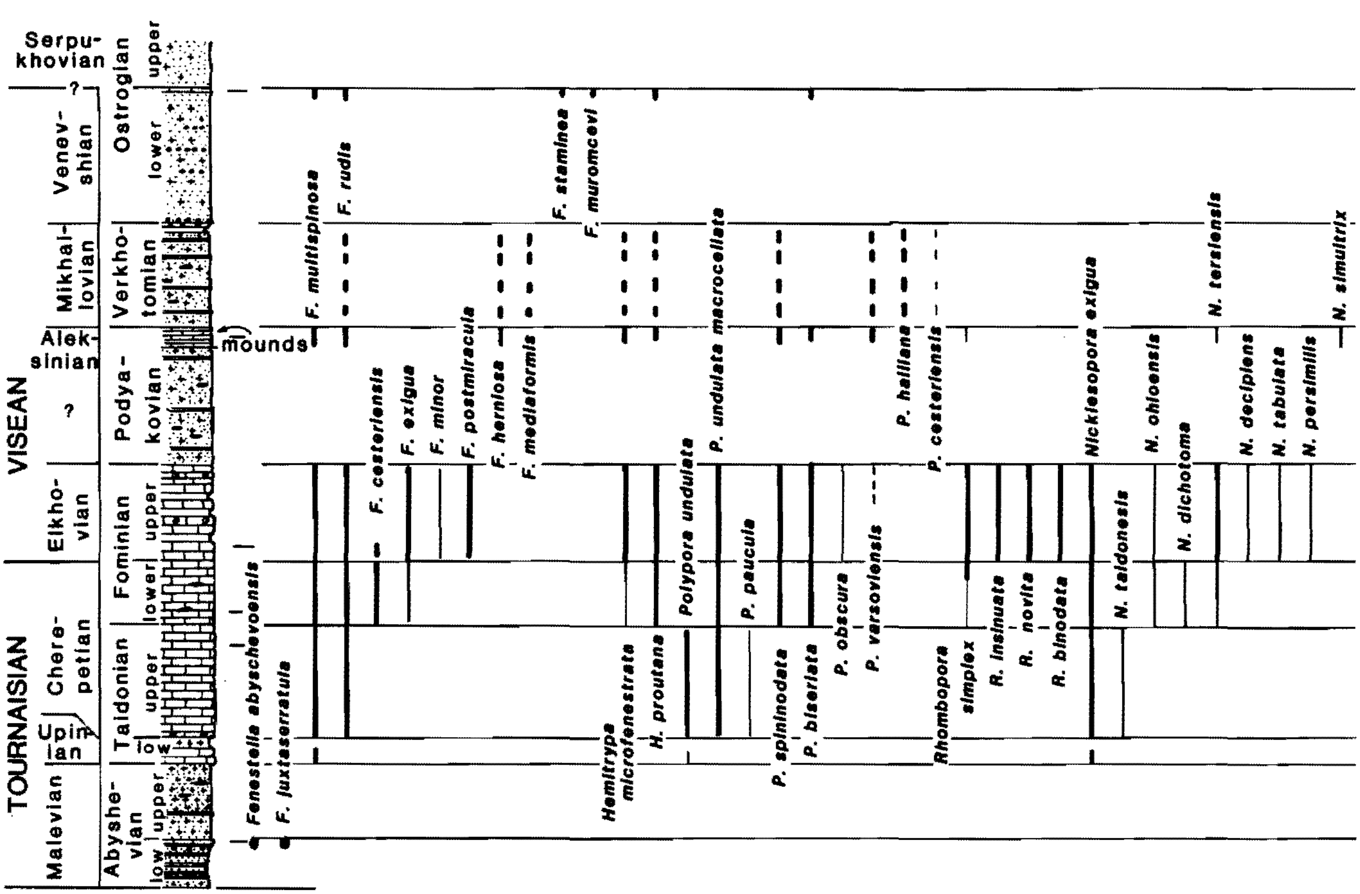

FIGURE 7. Distribution of some bryozoan species and genera in the Lower Carboniferous of the Kuznets Basin, U.S.S.R., (data from Trizna, 1958, and Selyatitsky and others, 1975).

of genera in other families, such as Rectuloceras, suggest random or fortuitous dispersals were common at certain times in the Carboniferous and Permian.

As in most other faunal groups, ammonoids show expansions and reductions in their diversities and geographical ranges. The most obvious is the one at the base of the mid-Carboniferous boundary (Saunders and Ramsbottom, 1986) which may be the result of the effects of greatly lowered ocean temperatures or the result of diminished food supply because ammonoids were predators relatively high in the ecosystem food pyramid.

Ammonoid zonations were the basis for subdividing the Belgium Tournaisian, Visean, and Silesian (Namurian and Westphalian) strata into zones (Paproth and others, 1983), based on species of Muensteroceras, Beyrichoceratoides, Goniatites and related genera. In the lower Namurian, species of Eumorphoceras, Cravenoceras, Cravenocertoides, and Nuculoceras form the typical zonal scheme. The species zones of Homoceras in the Chokierian and Alportian form the basal Middle Carboniferous zones and are not widely distributed outside of northwestern Europe. These zones are equivalent to the lower part of the Morrowan. Upper Morrowan zones include species assemblages of Retites, Recticuloceras, Hudsonoceras, Verneultites, Baschkirites, Banneroceras, Gastrioceras, and others.

In North America, Atokan, Desmoinesian, and late Carboniferous cephalopods (Böse, 1919; Unklesbay, 1954; Miller and Furnish, 1958) are more common than in northwestern Europe where a generally impoverished fauna is associated with a few thin marine bands. The Midcontinent cephalopods are common in Missourian strata and occur as high as the lower part of the Wabaunsee of the Virgilian. They are not known in younger strata in that area because of unfavorable facies. Strata of Middle and Upper Carboniferous ages on the Russian Platform contain some genera, but few species, in common with those of the Midcontinent and southwestern North America (Ruzhentsev, 1960, 1965).

Permian cephalopod zones were summarized by Furnish (1973). Although Furnish attempted to define or redefine time-stratigraphic stages to be the direct equivalent of cephalopod zones in that article, his discussion of the actual cephalopod zones showed that twelve well-defined assemblage zones can be recognized based on the distribution and stratigraphic ranges of genera in thirty-one families. Furnish (1973) also showed that these families are not evenly distributed geographically. Asselian, Tastubian, Sterlitamakian, and Aktastinian (Wolfcampian to middle Leonardian on the cycle charts used here) have more widely distributed assemblages than do higher Leonardian and Upper Permian cephalopod assemblages. Only in the middle part of the Leonardian and again in the lower part of the Guadalupian were Paleo-Tethyan and southwestern North America cephalopod families closely associated into biogeographical units.

The Asselian saw the introduction of new genera and families, such as the Metalegoceratidae, Paragastriceratidae, and Popanoceratidae. In southwestern North America the provincial Perrinitidae first appear either just below or in lower Wolfcampian strata (Furnish, 1973). *Properrinites* and *Akmilleri* appear in the Upper Wolfcampian Lenox Hills Formation.

Lower Leonardian (Tazlarovian) cephalopods include *Metalegoceras*, *Eothinites*, and lower Baigendzhinian cephalopods have *Paragastrioceras* and *Uraloceras* in eastern Europe. In southwestern North America species of *Medlicottia*, *Metalegoceras*, *Popanoceras*, and *Metaperrinites* are widespread in lower Leonardian beds.

The Cathedralian (upper Leonardian) (Ross, 1986) contains distinctive species of *Medlicottia*, *Eumedlicottia*, *Pseudohalorites*, *Neocrimites*, *Almites*, and *Perrinites*, many of which also appear in Coahuila (Mexico), western Guatemala, Timor, arctic Canada, South China, Pamir, and Darvas.

The Roadian contains species of *Eumedlicottia*, *Perrinites*, *Glassoceras*, and others. One locality also contains *Texoceras*, *Peritrochia*, and *Paraceltites*. These assemblages are known mainly in western and northern North America.

The Wordian has a large number of species and genera of cephalopods, particularly common are species of *Agathiceras*, *Popanoceras*, *Stacheoceras*, *Waagenoceras*, *Pseudogastrioceras*, and others. Similar species are known from Sicily and Timor.

The upper Guadalupian (Capitanian) is characterized by several species of *Timorites*. The highest of these assemblages is known only in abundance from Timor (Furnish's, 1973, 'Amarassian Stage') where species of *Strigogoniatites*, *Epadrianites*, *Stacheoceras*, *Timorites*, primitive *Cyclolobus*, *Hyattoceras*, *Sundaites*, *Syrdenites*, *Episageceras*, and *Xenodiscus* are reported.

The highest Permian Series, the Djulfian includes three cephalopod zones (Furnish, 1973). The lower one has *Araxoceras*, *Vescotoceras*, *Prototoceras*, *Pseudogastrioceras* and *Cyclolobus*. The succeeding zone has *Vedioceras* and species of *Cyclolobus*, *Dzhulfoceras*, and longer ranging genera. The youngest assemblage has a diversity of heavily ribbed xenodiscids (Chao, 1965) as well as longer ranging forms. *Phisonites triangulus* is present in the lower part of this zone and *Paratirolites kitti* in the upper part and separate the zone into two subzones.

BRACHIOPODS, BLASTOIDS, AND CORALS

Three groups, the brachiopods, blastoids, and corals, have been used for many provincial correlations and, to a lesser extent, for interprovincial correlations. Of these, the brachiopods have abundant provincial faunas which show some dispersals between provinces during the Carboniferous. As with most benthic groups, brachiopods show tropical cosmopolitan distributions during the Tournaisian and early and middle Visean. By the late Visean and early Namurian, decreases in diversities and geographical restrictions becomes apparent. Middle and Upper Carboniferous and Early Permian distributions are very provincial with only a few common genera between even tropical provinces. The Ural-Franklin province is united by having a common brachiopod fauna at this time. During the Guadalupian, southwestern North America and the PaleoTethys had quite different brachiopod faunas. Relatively little consideration has been given to the dispersal history of brachiopod genera or the timing of dispersals. Grunt and Dmitriev (1973) examined some aspects of Permian dispersals in the Soviet Union, however, the overall subject remains not well studied or understood.

Corals also are a group that have good provincial zonation (Federowski, 1981). In the Lower Carboniferous Vaughan (1915) and Hill (1948; 1957) described a good coral zonation for the Tournaisian and Visean (Dinatian) of northwestern Europe. Sando (1985a) and Sando and Bamber (1984) have detailed coral zonation for rocks of similar age in the western margin of the North American craton during the Mississippian. The Middle and Upper Pennsylvanian coral zonation in North America has been summarized by Sando (1985b).

The Lower Permian corals were divided by Minato and Kato (1965a, b; 1971) into two coral provinces, one dominated by waagenophyllid corals, the other by durhaminid corals. Stevens (1982, 1983) and Wilson (1982) have extended studies of colonial Early Permian corals to various parts of western North America. Hill (1958) used coral distributions to help examine Sakmarian geography.

Blastoids in the Chesterian of the type region of the Mississippian show a remarkable species succession (Waters and others, 1985) which seems to be a useful provincial zonation. This group was not widely distributed outside of that province during the Mississippian and was not again abundant until the Permian, and then only in Timor.

DISCUSSION

From the preceding discussions of different fossil groups, it is possible to generalize many of the ecological and environmental conditions at different times during the late Paleozoic.

Changes in the geographic configuration of cratons during the late Paleozoic were sequential. Each major step in this sequence resulted in changes in ocean surface currents, in their temperatures, and in their directions. These, in turn, resulted in changes in world climates as shown by expansion and reduction in glaciation, particularly in Gondwana, in reduction and expansion of carbonate production, fluctuations in sea level, and in the dispersal, extinction, and evolutionary patters of warm water shelf faunas.

The rapid evolution (and subsequent zonation) of the shelf faunas is associated with depositional sequences and may be related to physical changes in the environment, such as temperature, and to the repeated flooding of shallow shelves which encouraged community diversification and specialization. Certainly the adaptive opportunities were greatly increased at these times of sea-level highstands. The subsequent lowering of sea level and restriction of shelf areas may have caused ecosystem disorder and the extinction of some species.

The Tournaisian and most of the Visean were relatively warm and during sea-level highstands had many diverse carbonate-producing communities. Sea level fluctuations were of low magnitude with relatively long frequences. In most of the Tournaisian and Visean, general sea levels were high and the shift of shorelines remained on the shelves. Dispersals were very common with nearly cosmopolitan faunas having latitudinal gradients.

Late in the Visean and in the Serpukhovian (Chesterian), world temperatures cooled rapidly and remained cool during the Bashkirian. Shelf faunal diversity became very low as a result of many extinctions. Surviving genera and families commonly contained only a few species. Carbonate production was generally minor except for a relative narrow equatorial belt. Sea level was generally low. Sea-level fluctuations, however, were of considerable magnitude and of relatively short frequency. Dispersal of benthic shelf faunas was poor. Yasamanov's (1981) studies of Ca/Mg (Fig. 8) suggested the decline in diversity in the later part of the Lower Carboniferous was related to a lowering of surface water temperatures and the gradual increase in diversity during the Early Permian was related to a gradual warming trend.

The Moscovian was slightly warmer than the Bashkirian and, although there was minor faunal diversification, many of the surviving Lower Carboniferous genera became extinct by the end of this stage. Conservative, low diversity shelf communities were the rule, faunal dispersals were irregular and probably fortuitous. Although general sea level rose, sea-level fluctuations continued to be of large magnitudes and of short frequencies.

Late Carboniferous and earliest Permian were times of gradual warming, few extinctions and modest diversification. Dispersals were only slightly more common than during the Moscovian. The low diversity carbonate mud-bank and mound communities of the Late Carboniferous gradually expanded into somewhat more complex biohermal communities during the earliest Permian. Sea-level fluctuations continued to have short frequencies and large magnitudes.

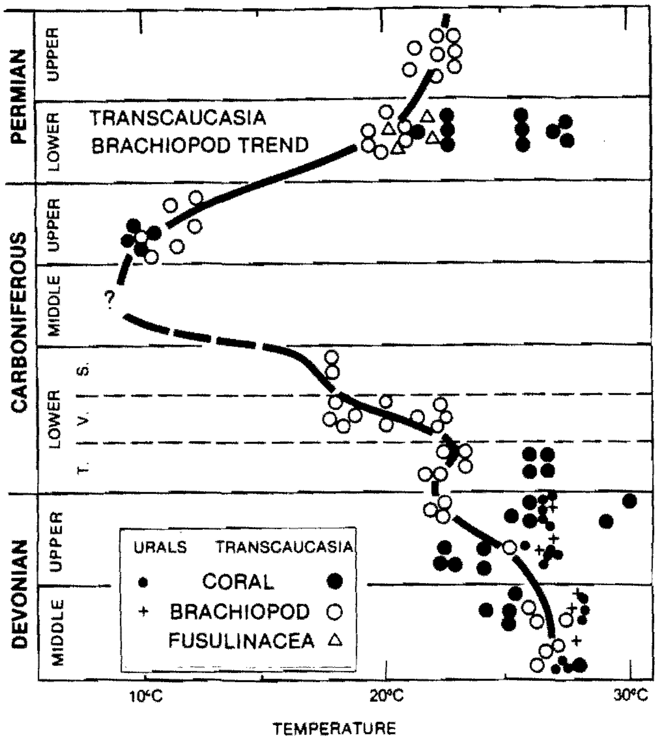

FIGURE 8. A temperature curve for the Devonian through Permian derived from Ca/Mg ratios in corals, brachiopods, and fusulinids (redrawn from Yasamanov, 1981).

Later Early Permian (Leonardian) was warm, perhaps as warm as the Tournaisian, and the shelf carbonate faunas show marked diversification. Reef-forming communities gradually evolved independently on both tropical shores of Pangaea. This pattern continued into Guadalupian. Dispersals were extremely rare across Paleo-Panthalassa giving rise to strongly provincial faunas, which were further emphasized in the Guadalupian by increased faunal diversity, particularly in the PaleoTethys. During the later part of the Guadalupian, extinctions again became increasingly common. Although many families had a few surviving genera, they were composed of only a few species. Sea-level fluctuations in the Leonardian and Guadalupian became longer in duration and less in magnitude.

The latest Permian (Djulfian) saw a burst of diversity in the Tethyan faunal realm, which produced some distinctive and briefly successful lineages. These, and the few remaining survivors of the Guadalupian, suffered extensive extinctions before the end of the Permian. The stratigraphic records suggests four rapid sea-level fluctuations of relatively small magnitude which were superimposed on a general lowering of sea level. The shelf faunas in the Tethys include genera and species that may have been adapted to warm, perhaps very warm water, however, in many other parts of the world, it is difficult to find any faunas or strata that can be identified as being of latest Permian (Djulfian) age.

CONCLUSIONS

Late Paleozoic sea-level fluctuations on the scale of 1 to 3 million years are identified worldwide by specific and generic range zones of many invertebrate groups. Although geographic provinciality was common in the Carboniferous and Early Permian, dispersals of some species and genera took place infrequently.

The resulting fossil zone assemblages are provincial species and genera having independent evolutions and stratigraphic ranges in different provinces combined and mixed with more cosmopolitan (or at least more widely dispersed) species and genera which tie the correlations between different provincial zones together. This type of zonation is dependent upon infrequent dispersals of a relatively small number of species during usually brief times that were favorable for the dispersals. These were apparently warmer times having high sea levels.

Changes in the configuration of continents during the Carboniferous to form Lesser Pangaea and in the middle Early Permian to form Greater Pangaea changed the pattern of oceanic surface currents and progressively isolated the tropical shelves on either side of the supercontinent. Although the Tethys portion of the huge single world ocean, Paleo-Panthalassa, contained many microplates, dispersals of warm water faunas eastward across the main part of the ocean was difficult because of a westward flow of equatorial currents (Ross and Ross, 1981).

The zonation of the Carboniferous and Permian into about seventy warm-water shelf faunal zones is possible and permits the identification of individual third-order sea-level fluctuations worldwide. In part, these sea-level fluctuations themselves may be one of the main contributing causes of the rapid evolution that aids in the zonation. In their role as stratigraphic markers, the recognition of these depositional sequences along with their faunas offers a different and useful approach by which to add more precision to the correlation of late Paleozoic strata.

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SEA LEVEL CHANGES AND CENOZOIC LARGE FORMINIFERA WITH INNER STRUCTURES

GEORGE A. SEIGLIE
Chevron U.S.A Inc., P. O. Box 1635, Houston, Texas 77251

ABSTRACT

Sea-level changes affect organisms living on the shelves, especially the large calcareous foraminifera with photosynthetic symbiotic algae. The evolutionary patterns of Cenozoic shallow water large foraminiferal genera correlate with sea level fluctuations.

Most newly evolved genera of larger foraminifera first appeared during sea-level highstands. The number of new genera appearing in each highstand is approximately proportional to the duration of the highstand, except for the Early Oligocene. This may be a consequence of the high diversification of species of Lepidocyclina during the Oligocene, which probably occupied a diversity of ecological niches that otherwise would have been occupied by new genera. The greatest number of new genera appeared during the Middle Eocene which had the longest Cenozoic sea-level highstand. The major extinctions of large foraminifera possibly occurred during major sea-level lowstands.

INTRODUCTION

Most of the foraminiferal genera used in this study may be considered "larger foraminifera". This is, however, a controversial expression. Glaessner (1945) and Pokorny (1958) listed as "larger foraminifera": fusulinids, some Orbitolinidae and Peneroplidae (Soritidae), Alveolinidae, Camerinidae (=Nummulitidae), and the orbitoidal foraminifera. Haynes (1981) added the families Miscellanidae, Chapmaninidae, and some genera of Rotaliidae. All these foraminifera lived in shallow water and have in common large size, calcareous tests. Ross (1974) defined large foraminifera as those foraminifera that are greater than 3mm^3, independent of the nature of the wall and the living habitat. This definition is the closest to the expression "larger foraminifera". Saint-Marc (1977) included as larger foraminifera Cretaceous genera with agglutinated walls and inner structures. For the purpose of this paper foraminifera irrespective of whether they are shallow or deeper-water dwellers, which are large or relatively large, and have complex inner structures, which necessitate thin section examination are regarded as large foraminifera.

It has been postulated that during sea-level highstands oceanic waters are rich in nutrients and hence, the phytoplankton and planktonic foraminifera are abundant, while during sea-level lowstands the oceans are relatively improverished and the planktonic foraminifera scarce (Lidz, 1974). Although different ecological factors are involved, the biology and habitat of shallow-water large foraminifera were also affected by sea-level changes.

Shallow-water large foraminifera live on tropical and subtropical shelves. The shelves are the marine areas most strongly affected by sea-level changes. Even minor world wide sea level changes achieve significance in a shelfal regime. In addition, the rate of sedimentation is higher on the shelves during high sea-level stands. Relationships between large foraminifera have been already indicated by Adams (1983) and Ross and Ross (1985).

Calcareous shallow-water large foraminifera have been associated since their appearance with microscopic symbiotic algae (zooxanthellae) and evolved with them, as suggested by Ross (1974), Lee et al. (1979), and Hallock (1985). The relationships between symbiotic algae and large foraminifera have been studied for many of the species of living genera, as for example: Röttger (1972) for Heterostegina, Ross (1972) for Marginopora, and Hanson and Burchardt (1977) for Amphistegina. The calcareous large foraminifera necessarily live in shallow water because their associated symbiotic algae require light to live. The warm water allows them to build their calcareous tests in a relatively short time (Ross, 1972), and the clastic sediments, must be a minimum or absent.

Their environment must be stable, as they may require up to two years for their reproduction (Ross, 1972; Hallock, 1984). In contrast, the smaller foraminifera that live in unstable environments, like Ammonia beccarii, reproduce every 3 or 4 weeks (Bradshaw, 1961). The required environmental stability would be strongly affected by sea-level changes which in turn would severely modify the character and/or composition of large foraminifera.

To study the effects of the sea-level changes on the large foraminiferal assemblages, which is the purpose of this paper, I have used genera rather than species. The reason for this is that the utilization of species is impracticable because the required information in terms of species is not frequently available (Newell, 1982). I prefer to use the genera in the large foraminifera rather than family because the number of families of shallow water large foraminifera is also too small to have statistical value.

One hundred and eight genera of shallow-water large foraminifera have been considered for this study. They are listed by family and subfamily in the appendix. Fifteen genera were not included in the charts because their stratigraphic ranges are not fully documented; another four genera were omitted because they are only known for the Recent. The relationships of the deep water large foraminifera and the sea-level changes are explained at the end of this paper for comparison.

The absolute ages used in this study are taken from Berggren et al. (1985), and the eustatic sea-level changes are compiled according to Vail and Hardenbol (1979), and Keller and Barron (1983).

TAXONOMY

The genera of shallow-water large foraminifera utilized in this study are included in three suborders of Foraminiferida: Textulariina, Miliolina and Rotaliina. Their classification is shown in the appendix with the list of shallow-water large foraminifera.

Six genera of shallow-water Textulariina were considered for this report, but two of them, Liebusella and Textulariella also occur in deeper, bathyal waters. The taxonomic positions of the other four genera are not well defined at present and they are only referred to as Textulariina.

The shallow-water large foraminiferal genera of the suborder Miliolina are placed in 9 families and 3 superfamilies. The superfamilies follow the usage of Loeblich and Tappan (1984). The families in which the genera were included follow de Castro (1971), Ghose (1972), Hamaoui and Fourcade (1973), Henson (1950), Robinson (1974a, 1974b), and Seiglie et al. (1977).

The large-foraminiferal genera of the suborder Rotaliina are classified in 17 families and 6 superfamilies. The superfamilies are according to Loeblich and Tappan (1984). The genera are included in the families in agreement with Caudri (1972), Deloffre and Hamaoui (1973), Glaessner and Wade (1959), and Loeblich and Tappan (1964). The family "Rotaliidae" is placed in quotation marks, because Lévy et al. (1986) showed that Rotalia, the type genus of the family, should be included in the family Discorbidae based on its inner structure.

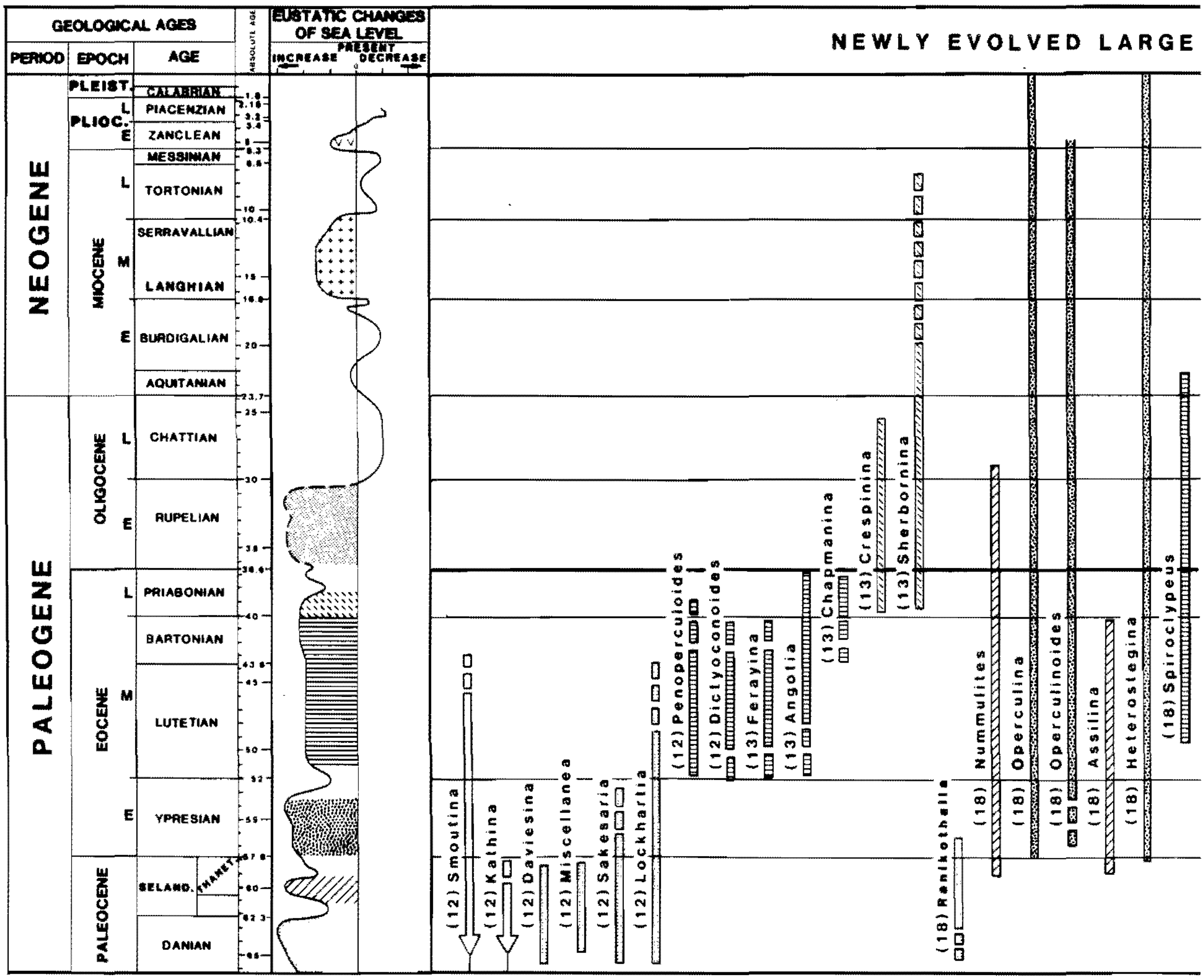


Figure 1a. Newly evolved large foraminifera

INFORMATION ON BIOSTRATIGRAPHY

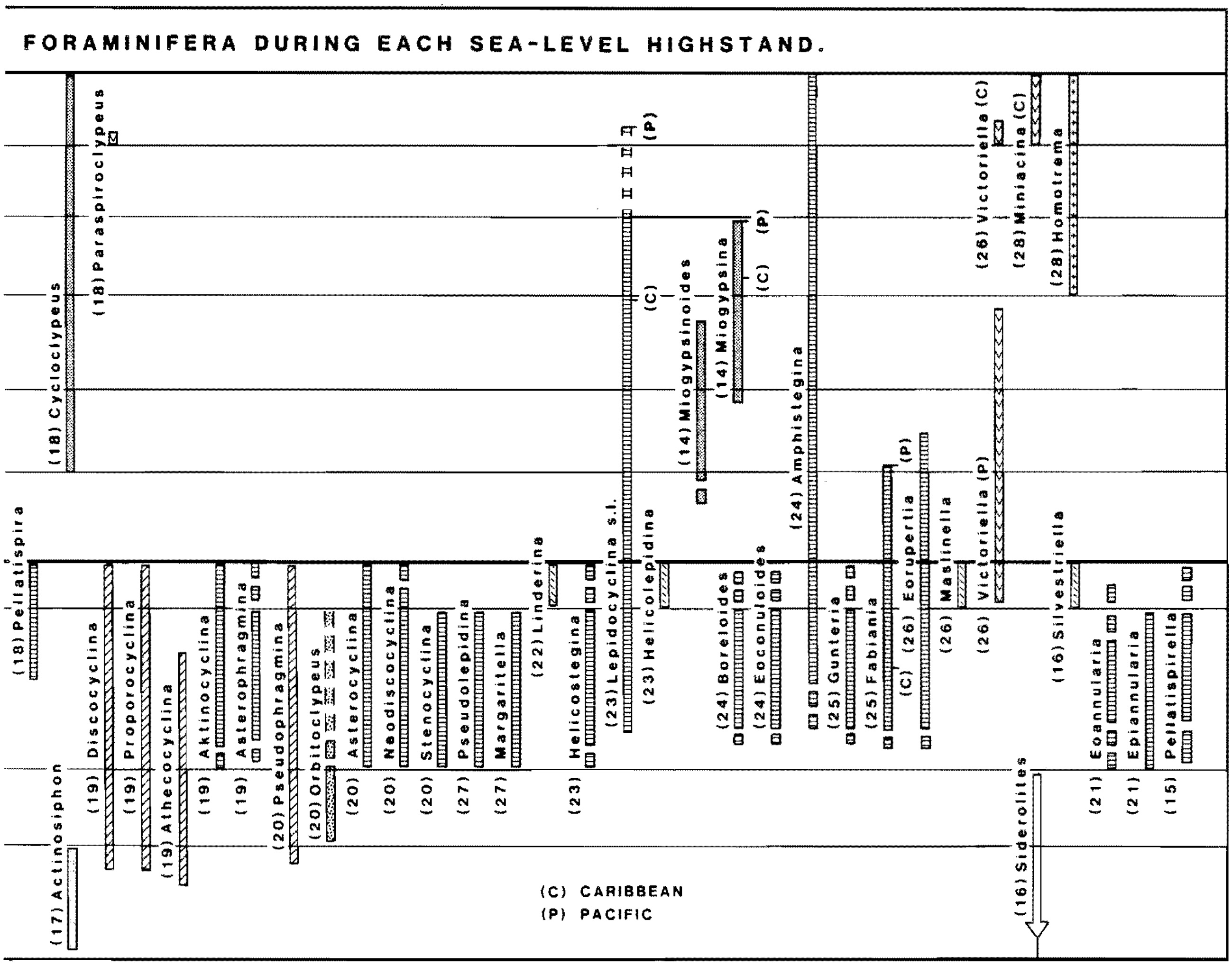
The biostratigraphic distribution of the shallow-water large foraminiferal genera is shown in Figures 1a and 1b. The stratigraphic range of these genera, which live mostly in inner and middle neritic carbonate environments, cannot be accurately correlated with the planktonic foraminiferal zones. In addition, different authors have differing opinions about their stratigraphic ranges.

The stratigraphic distribution shown in the charts is a compilation of the ranges of the genera of Miliolina given by Adams (1968, 1970, 1984), Adams and Belford (1979), Beckmann et al. (1981), de Castro (1971) Cole (1965), Ferrer et al. (1973), Frost and Langenheim (1974), Hamaoui and Fourcade (1973), Henson (1950) and Seiglie et al. (1977).

The stratigraphic ranges of Rotaliina were compiled from Adams (1984) Berggren and van Couvering (1974), Beckmann et al. (1981), Blondeau et al. (1974), Blow (1979), Bombita and Popescu (1977), Brun et al. (1982), Butterlin (1984), Butterlin and Monod (1969), Caudri (1972, 1974, 1975), Cole (1957, 1958a, 1958b, 1958c, 1969), Cole and Bermudez (1944), Cole and Applin (1964), Eames et al. (1967), Frost and Langenheim (1974), Glaessner and Wade (1959), Kugler and Caudri (1975), Rahagi (1984), and Schaub (1963). In addition, one species of Eorupertia has been found by the writer in Oligocene reefs of southwestern Puerto Rico, and Carpenteria bulloides, described from Puerto Rico by Galloway and Heminway (1941), is actually a Caribbean, Early Pliocene species of Victoriella or a closely related genus.

SHALLOW-WATER LARGE FORAMINIFERA AND SEA-LEVEL CHANGES

It is logical to assume that shallow-water large foraminifera evolved and were more abundant during sea-level highstands. At such times, ocean waters were warmer because of the absence of glaciations, and tropical shelves were extensively submerged providing



during each sea-level highstand; Rotaliina.

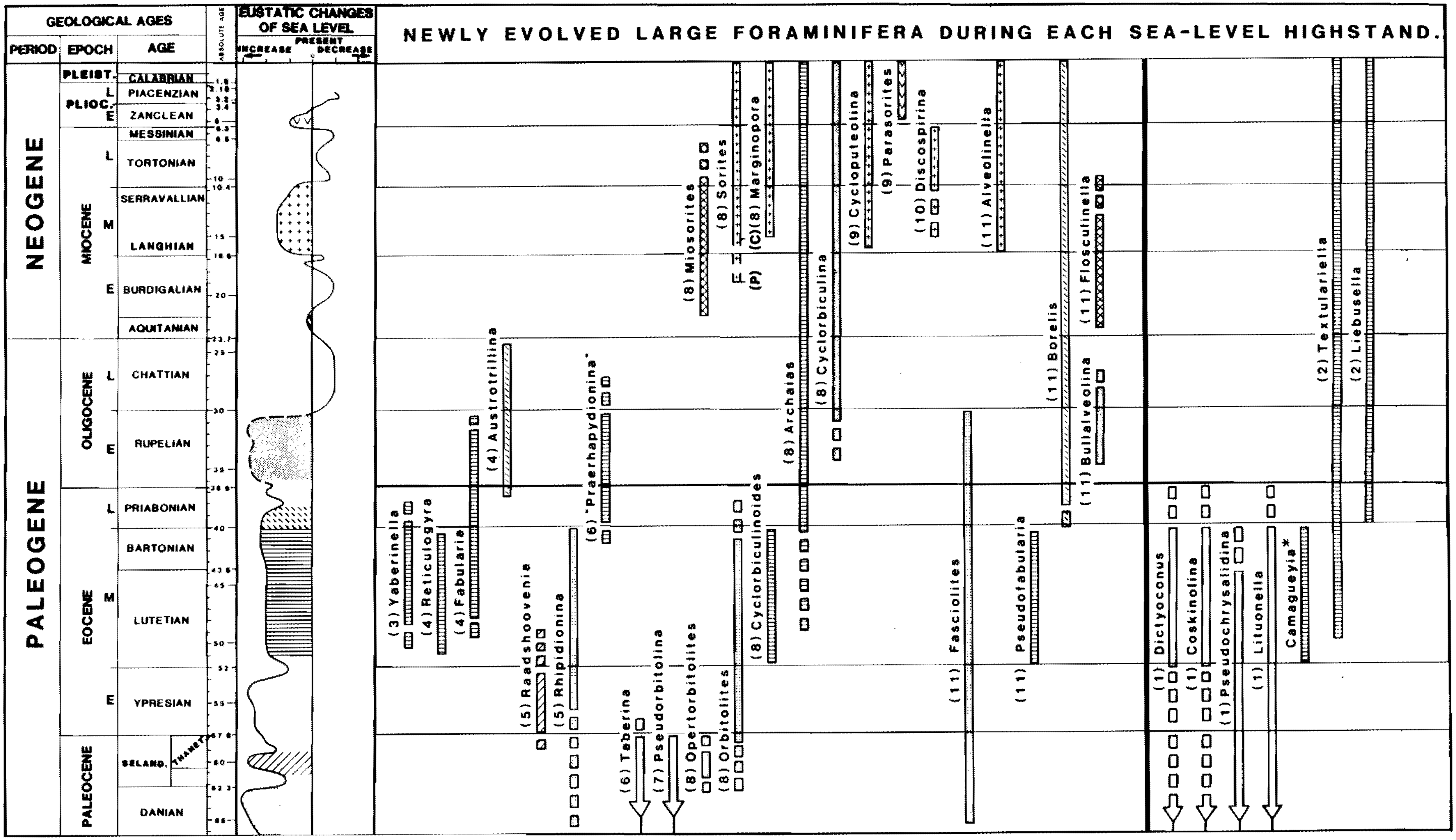


Figure 1b. Newly evolved large foraminifera during each sea-level highstand; Miliolina and Textularina.
*Possibly not agglutinated.

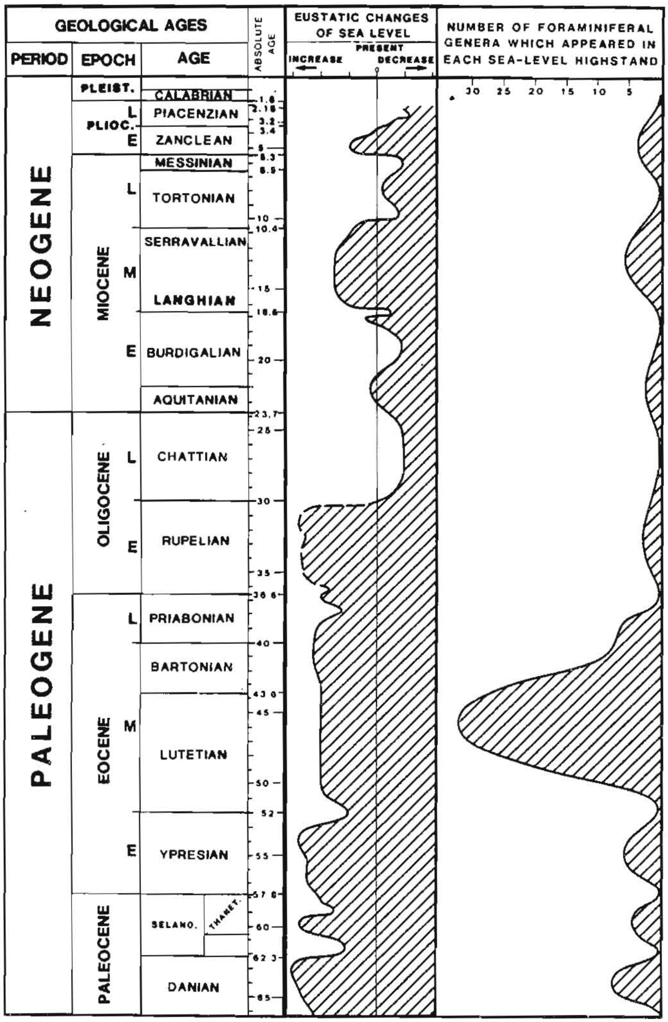


Figure 2. Sea-level highstands and number of new evolved genera during each highstand.

the foraminifera with varied environments and new areas for colonization, and enriched nutrients. Although the ranges of the large foraminifera cannot be precisely correlated with the planktonic foraminiferal zones, it is clear that most of the genera of shallow-water large foraminifera appeared during the major sea-level highstands (Figs. 1a, 1b).

Haq et al. (1987) have recently developed a new analysis of the Cenozoic sea-level changes that is more detailed than the generalized curve used in this study. The most significant events for this study are the major sea-level highstands, allowing evolution of large foraminiferal genera and the major sea-level lowstands, which are associated with the major extinctions. Minor low sea-level fluctuations probably have not had serious effect on the extinctions and have little consequence on the conclusions of this study.

The geographic position where these events occurred is also significant. For example, a sea-level highstand on a continental shelf having its border in 250m of water and a high rate of sedimentation will provide new areas for colonization, but the shelf will have a higher rate of sedimentation, and less light in its deeper parts, being both factors unfavorable for the large foraminifera. In contrast, most of the shelf border of islands, such as Cuba and Puerto Rico, are between 30 and 100m of water depth and the shelf has a low sedimentation rate. When a major sea-level highstand occurs, the colonization area actually increases, because the length of rivers and the sedimentation rate is reduced, and the light is sufficient in all the parts of the shelf to maintain a favorable environment for large foraminifera. Minor sea-level changes would produce minor effects in the assemblages of large foraminifera.

Regarding the sea-level highstands separately, at least thirty two genera of shallow-water large foraminifera evolved during the Middle Eocene, which is probably the largest number of newly evolved genera of shallow-water large foraminifera in any sea-level high stand (Fig. 2). The planktonic foraminiferal species also reached their peak of evolutionary appearances during the Middle Eocene (Olsson, 1982).

Haq et al. (1987) show in their eustatic sea-level changes that a major sea-level lowstand occurred between the Middle and Late Eocene. This major lowstand is not shown in the sea-level fluctuations used in this paper, however, its presence was inferred from the foraminiferal faunal break between the Middle and Late Eocene, (Figs. 1a, 1b, and 2).

Large foraminifera flourished during the Middle and Late Eocene forming extensive

Table 1: Ratio between the number of newly evolved foraminifers to the duration of the highstand of the sea level. The highstand numbers are referred to in Figure 2.

Highstands	My	NF(*)	NF/My
I	4	9	2.25
II	2.5	7	2.8
III	5	8	1.6
IV	11	32	2.9
V	2	7	3.5
VI	6	3	0.5
VII	2	2	1
VIII	6	6	1
IX	2	3	1.5

(*) Number of newly evolved foraminifera.

biostromes covering great areas of the shelves. Most of them represented the families Nummulitidae, Discocyclinidae, Orbitoclypeidae, and Lepidocyclinidae. The high proportion of clastic sediments on the shelves during the Paleocene and Eocene may have produced turbid waters which covered larger parts of the shelves than in the remaining epochs of the Cenozoic. They probably affected the corals more adversely than the large foraminifera. This gave the foraminifera an opportunity to occupy considerable areas of the shelves that otherwise would have been occupied by corals.

In contrast, only three genera appeared during the sea-level highstand of the Early Oligocene after the Late Eocene extinctions. This was the consequence of a modified Oligocene environment. Clastic sedimentation decreased during the Oligocene and in succeeding epochs, and turbid waters became less common. This could have given the corals an advantage to spread faster displacing the large foraminifera from a great part of the tropical shelves. Also, the genus Lepidocyclina diversified and its different species occupied the ecological niches that otherwise possibly would have been occupied by newly evolved foraminiferal genera. Another important ecological change occurred at the beginning of the Oligocene with the onset of cooler climates. At this time the Antarctic ice-cap formed and the oceanic waters became cooler (Shackleton and Kennett, 1975; Kennett and Shackleton, 1976). Larger foraminifera do not occur below the shelf, while solitary type of corals may live on the slope down to 800m of water depth Wells (1956). This strongly suggests that the cooler waters favored the corals more than the calcareous large foraminifera.

The relationships between sea-level highstands and shallow-water large foraminifera are evident from Figure 2 and Table 1. The number of newly evolved large foraminifera was greater during sea-level highstands, but the ratio is higher in the Paleocene than during the succeeding Cenozoic epochs. The cause, as explained above for the Oligocene, was the greater clastic sedimentation during the Paleocene and Eocene and the spreading of the corals. Furthermore, during the highstands the the Paleogene waters were deeper than the Neogene waters

This clear association of highstands with the evolutionary appearance of shallow-water large foraminiferal genera throws into question the ranges of Ripacubana, Riveroina,

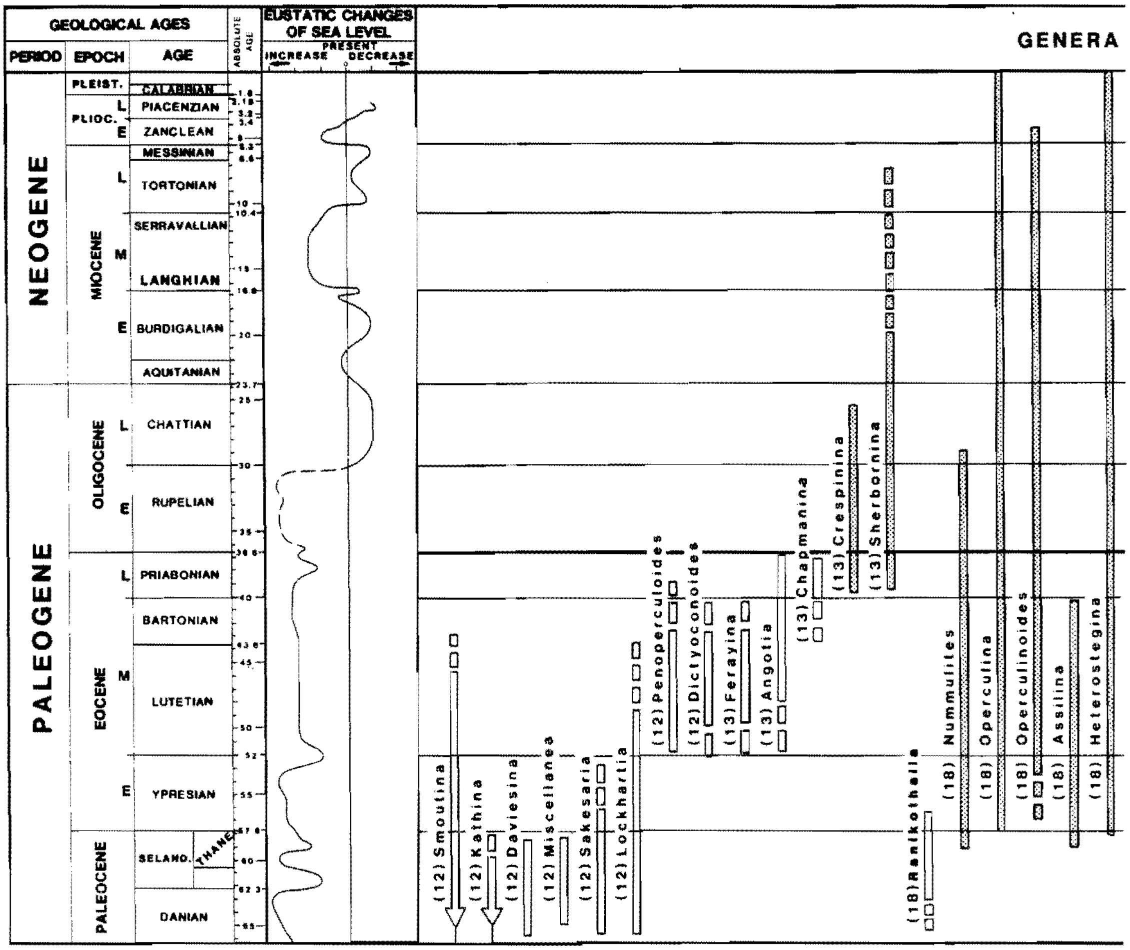


Figure 3a. Genera crossing the

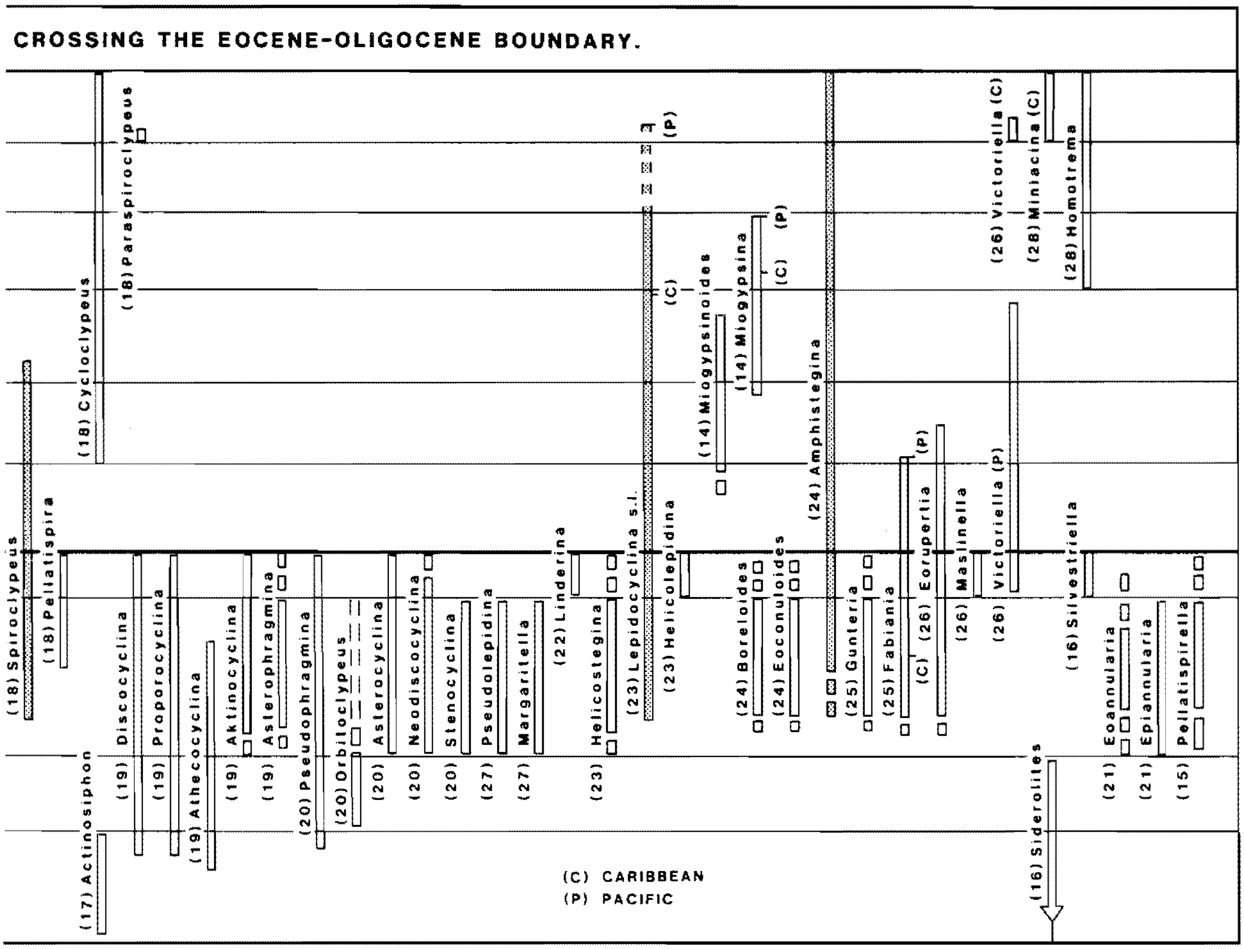
Craterites, and Keramosphaera, all of which are reported only from the Holocene. These genera probably evolved during pre-Pleistocene highstands. Robinson (1974a), for example, has reported Keramosphaera? in the Middle Eocene of Jamaica.

SURVIVAL AND EXTINCTIONS THROUGH THE CRETACEOUS-PALEOCENE AND EOCENE-OLIGOCENE BOUNDARIES

The cooling of the oceans and the sea-level changes probably had a greater direct influence on the extinctions of the large-foraminiferal assemblages in the Cenozoic than any other geological event. Two major crises, the Cretaceous - Paleocene and the Eocene - Oligocene transitions, determined most of the extinctions and further evolution of large foraminifera during the Cenozoic. The shelf border of the Antilles and Bahamas is shallower than the border of the shelf in Indonesia. The sea-level fluctuations, therefore, had a more dramatic effect than in Indonesia. Regarding again the Antilles model for the extinctions, a sea-level drop of 100m would leave the Antilles shelves subaerially exposed, and most of Bahamas, Nicaragua and Yucatàn shelves. The northern South America shelf, reduced to a narrow belt, subjected to upwelling, which would adversely affected shallow water large foraminifera.

The first and most severe extinction crisis occurred during the Cretaceous-Paleocene transition. The changes in the oceanic waters affected all marine faunas but with different intensities. For example, according to Smit (1981) the Cretaceous planktonic foraminifera became extinct with the exception of only one species, Guembelitria cretacea, which is the ancestor of all the Cenozoic planktonic foraminifera. This contrasts with 9 genera (about 25% of the 34 Maestrichtian genera of shallow-water large foraminifera) that crossed this boundary. Four of them are Textulariina, three Miliolina, and two Rotaliina (Figs. 3a, 3b).

The second largest crisis occurred at the Eocene-Oligocene boundary. Cavelier et al. (1981) considered this crisis of similar magnitude to the Cretaceous-Paleocene crisis, however, this is not consistent with the number of extinctions and survivals of planktonic and large foraminifera during both crises. Blow's (1979) stratigraphic tables show that about 9 genera and 26 species of planktonic foraminifera crossed this boundary compared to one species crossing the


CROSSING THE EOCENE-OLIGOCENE BOUNDARY.

Eocene-Oligocene boundary; Rotaliina.

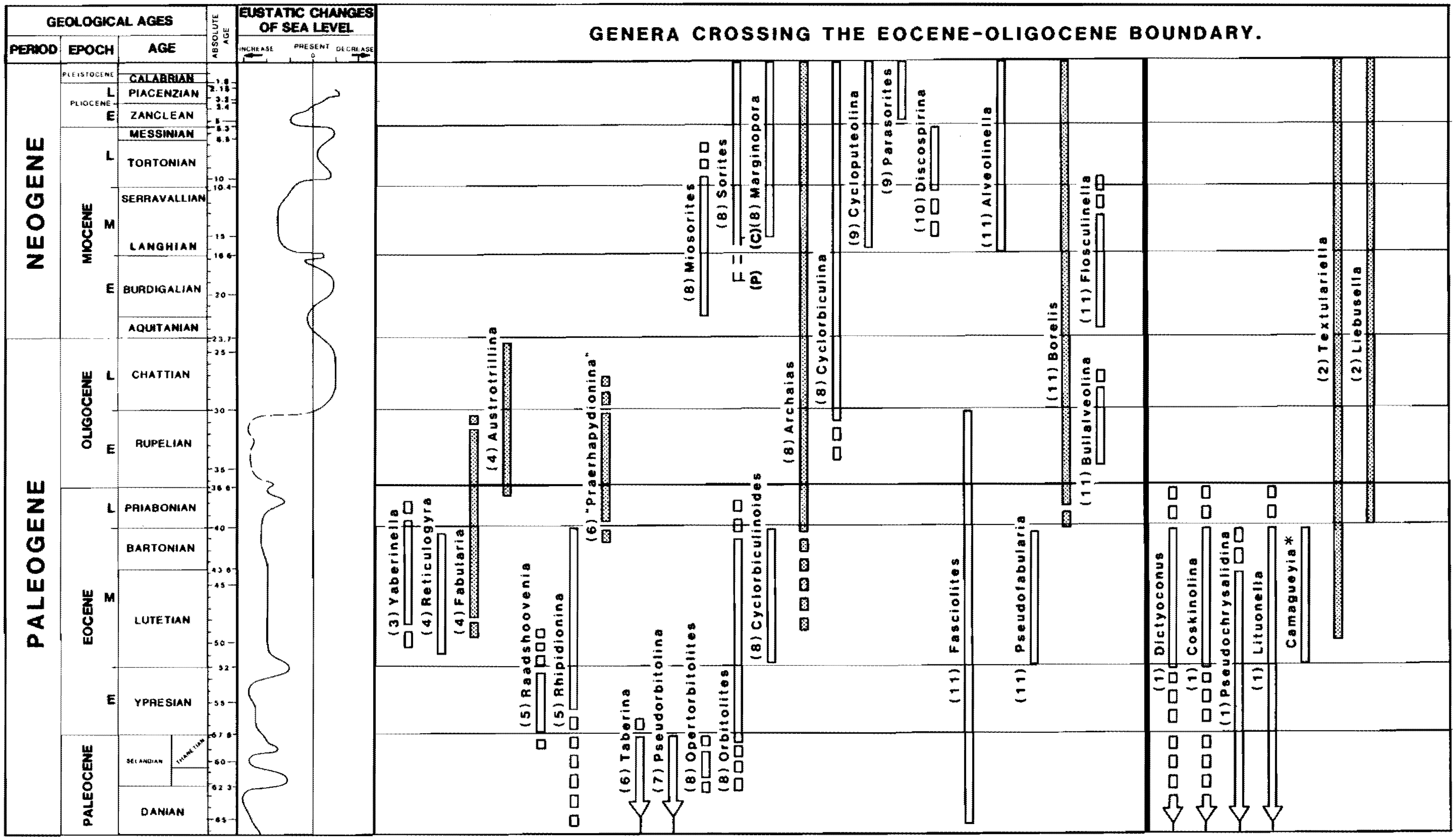


Figure 3b. Genera crossing the Eocene-Oligocene boundary; Miliolina and Textulariina. *Possibly not agglutinated.

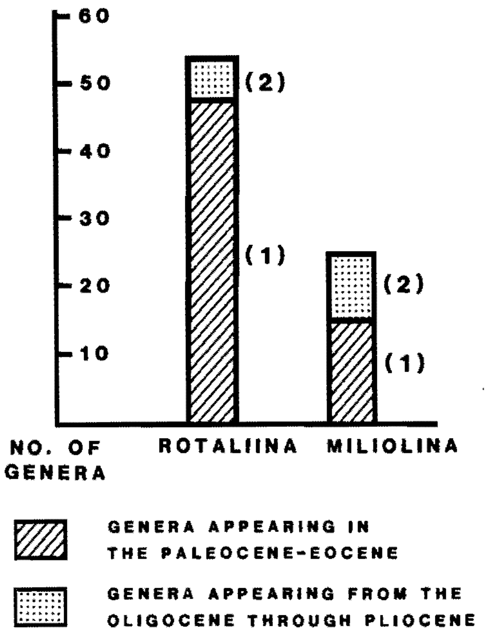


Figure 4. Number of foraminiferal genera that appeared during the (1) Paleocene and Eocene compared to the genera that appeared from the (2) Oligocene to Holocene.

Cretaceous - Paleocene boundary (Smit, 1981). Corliss and others (1984) and Pomerol (1985) demonstrated that several species of planktonic foraminifera became extinct one after the other during a relatively long span of time through the Eocene-Oligocene transition, and not simultaneously.

The genera of shallow-water large foraminifera which crossed the Eocene-Oligocene boundary are shown in Figures 3a and 3b. Four of the shallow-water genera of Textulariina became extinct before the end of the Eocene and two crossed the Eocene-Oligocene boundary.

Approximately 10 genera of large Miliolina lived during the Late Eocene, and 8 of them (80%), crossed the Eocene-Oligocene boundary (Fig. 4). The Rotaliina were relatively less successful: 31 genera were living during the Late Eocene but only 12 (39%) crossed the boundary (Fig. 5). Hence, a total of 20 genera of calcareous large foraminifera crossed this boundary, which contrasts with only 8 calcareous large foraminiferal genera crossing the Cretaceous-Paleocene boundary (Fig. 5).

The assemblages of Cenozoic calcareous large foraminifera were divided in two different groups by the transitional crisis from the Eocene to Oligocene and the subsequent ecological changes. During the Paleocene and Eocene newly 48 genera of large Rotaliina and 15 of Miliolina evolved, while from the Oligocene through the Pliocene only genera of Rotaliina and 10 of Miliolina evolved (Figs. 6a, 6b). In terms of the number of genera living before and after the crisis, 50 large Rotaliina and 15 Miliolina genera inhabited the shelves during the Paleocene and Eocene, while 16 large Rotaliina and 14 Miliolina genera lived from the Oligocene through Pliocene. The large Rotaliina were displaced, as explained above, by the corals which took over their space in the shelves. This obviously did not occur with the large Miliolina which are commonly back-reef dwellers where they may have received small amounts of terrigenous sediments or lived on the fine calcareous muds of the back-reef lagoon or on the algae living in the lagoons.

DEEP WATER LARGE AGGLUTINATED FORAMINIFERA WITH INNER STRUCTURES AND THE EOCENE-OLIGOCENE TRANSITION

Most of the genera of the Cenozoic agglutinated foraminifera with inner structures are, and were, deep-water dwellers. Liebusella is one of the exceptions as it has been reported in present seas from 40m (Seiglie, 1971) to abyssal (Brady, 1884) depth, mostly in the east side shelves of the continents. The cold anoxic upwelling waters dominate the west side where Liebusella has been rarely reported (Seiglie et al., 1985). This suggests that the species of this genus have lived on the shelf, in tropical and subtropical environments and that the deeper, rare occurrences may be transported tests. Textulariella includes species whose depth habitats range from middle sublittoral to abyssal. Small species of Reticulophragmium may be as shallow as inner sublittoral. Other shallow-water agglutinated foraminifera are relict Cretaceous genera. The inner structure of the genus Matanzia which ranges from Late Cretaceous to Pliocene has not been studied in detail and the different species attributed to it

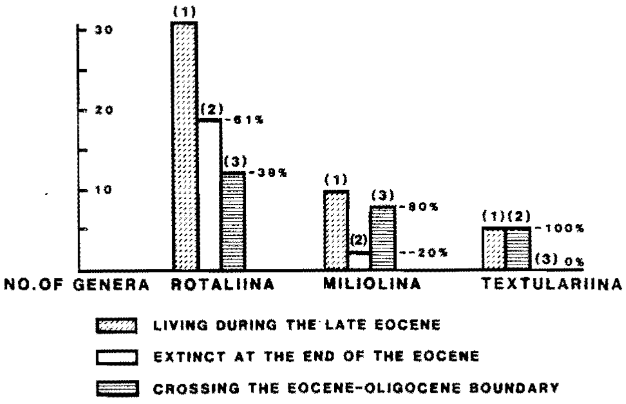


Figure 5. Number of larger foraminiferal genera (1) living during the Late Eocene, (2) extinct at the end of the Eocene, and (3) passing the Eocene- Oligocene boundary. Percentages given in relation to the number of foraminifera living during the Late Eocene.

may actually represent more than one genus.

The stratigraphic distribution of the genera of Cenozoic agglutinated foraminifera with inner structures is shown in Figure 7. Their distribution is mainly based on Gradstein (1983), Loeblich and Tappan (1985), Seiglie and Baker (1983) and Seiglie et al. (1986). The stratigraphic range chart of Figure 7 shows that none of the genera of internally complex agglutinated foraminifera appear to be affected by sea-level changes.

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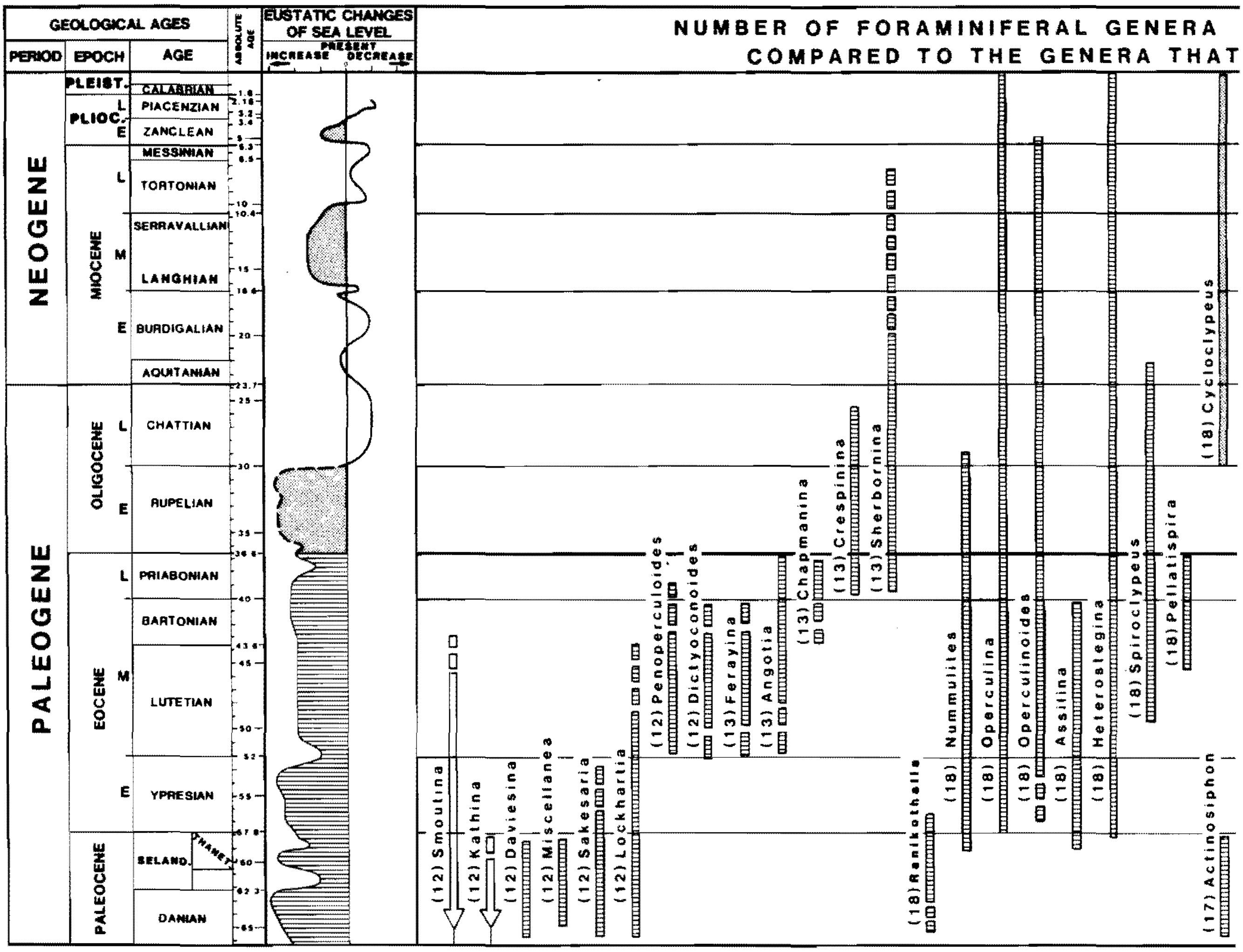

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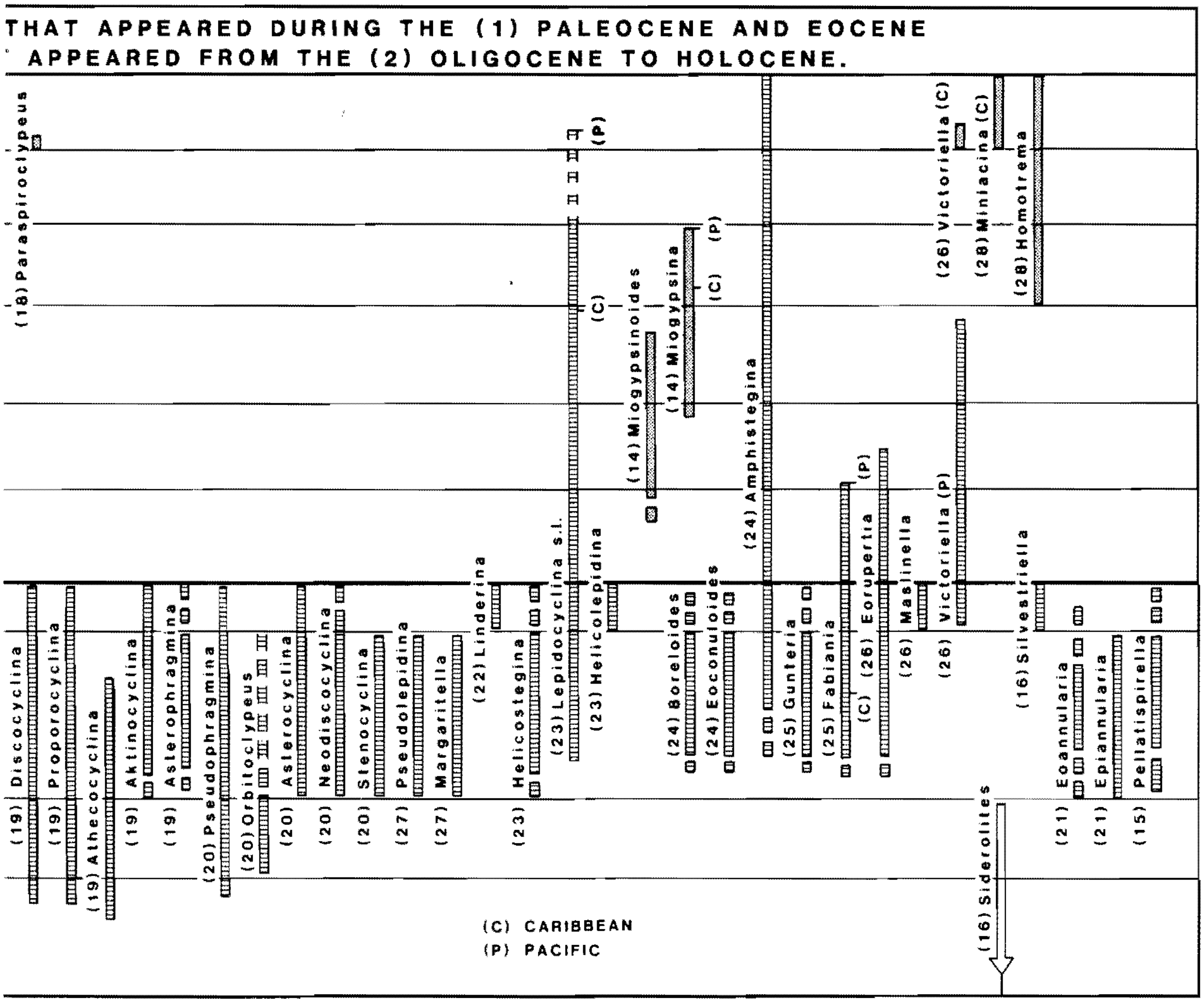

the genera that appeared from the Oligocene to Holocene; Rotaliina.

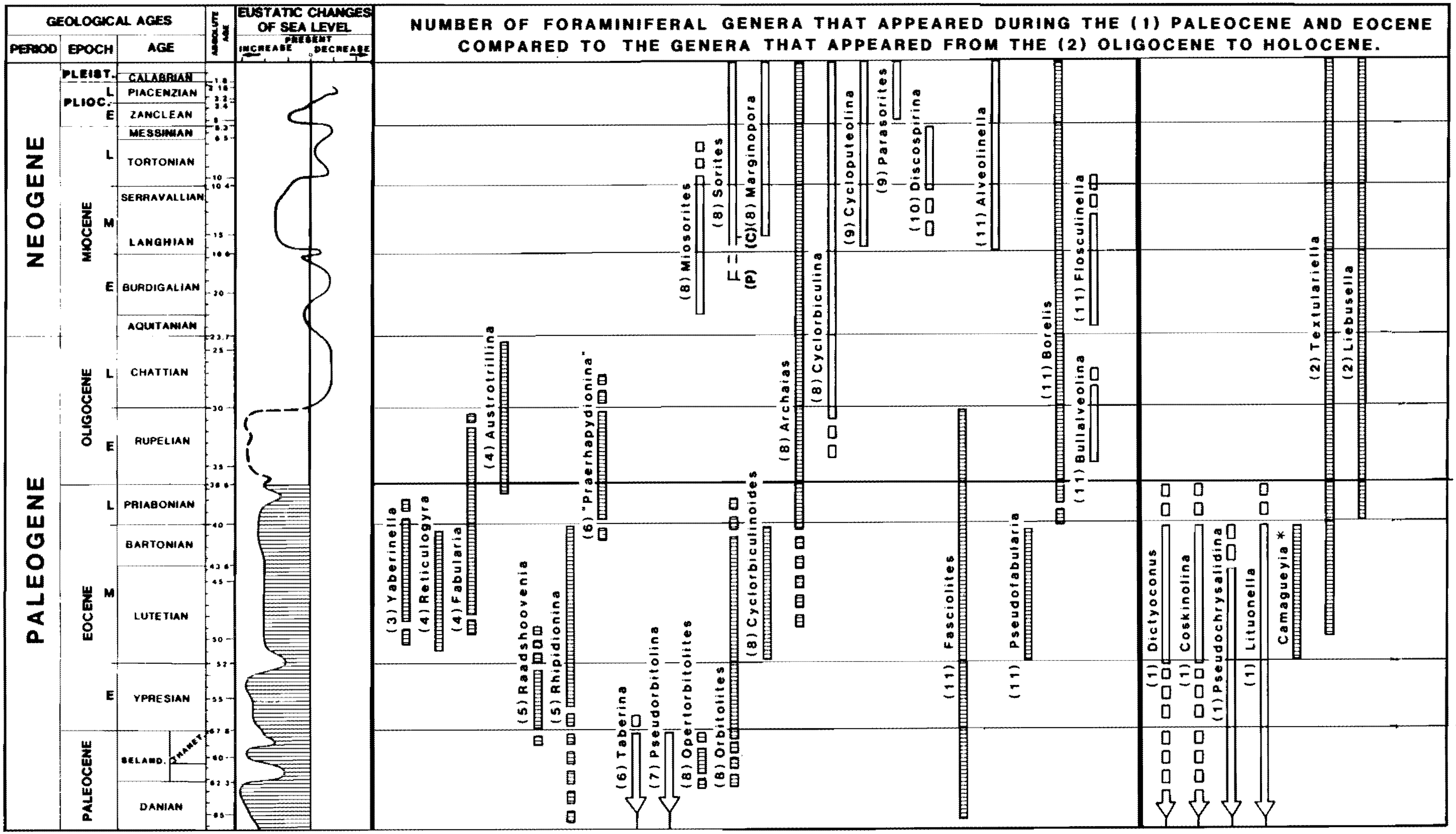

Figure 6b. Number of foraminiferal genera that appeared during the Paleocene and Eocene, compared to the genera that appeared from the Oligocene to Holocene; Textulariina and Miliolina. *Possibly not agglutinated.

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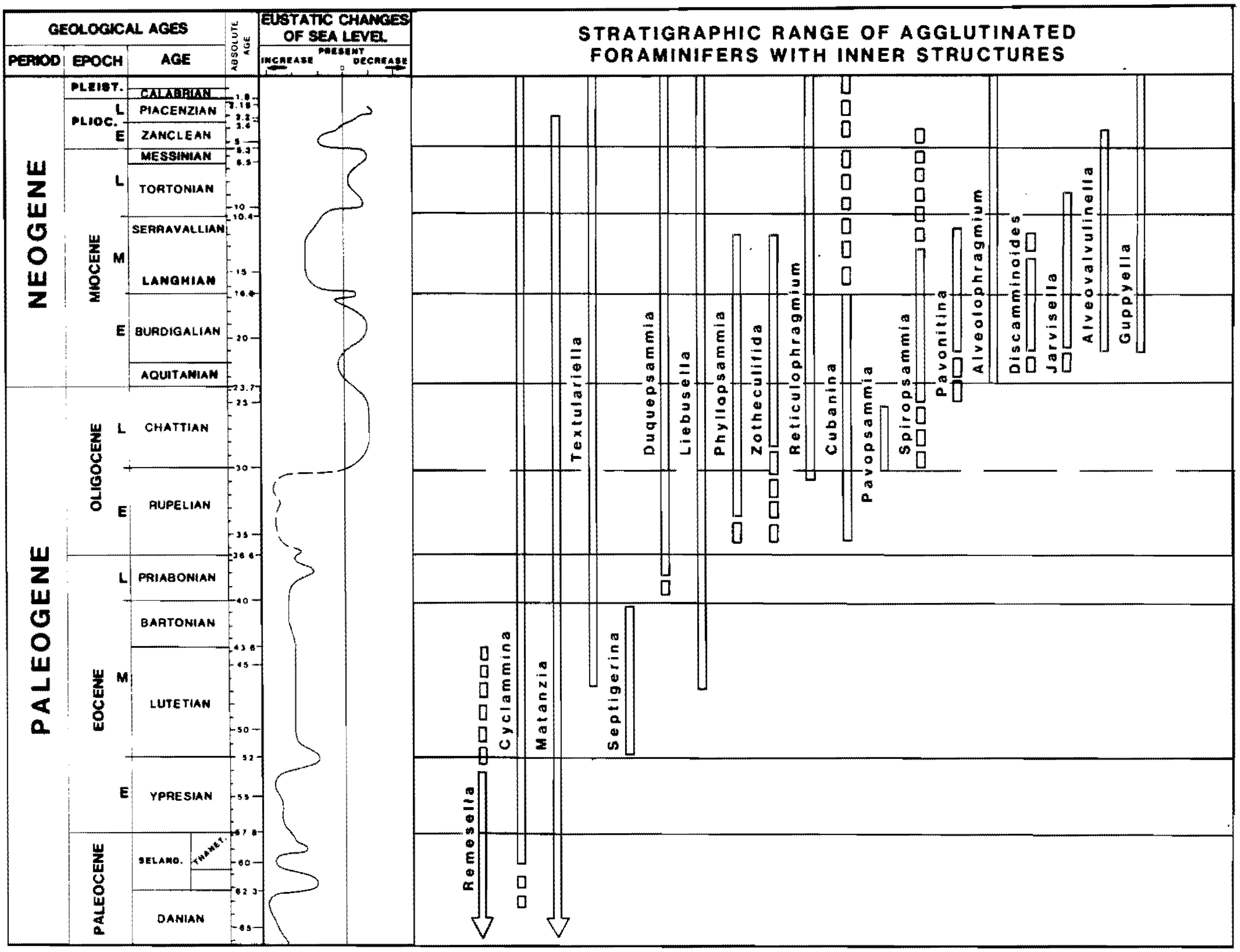

Figure 7. Statigraphic range of agglutinated foraminifera with inner structures.

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APPENDIX

LIST OF SHALLOW WATER LARGER FORAMINIFERAL GENERA AND FAMILIES

The number assigned to the foraminifers in the chart (Figs. 3a, b) are matched with the families on this list. The genera with an asterisc(*) were not included in the chart because their stratigraphic range may be doubtful.

SUBORDER TEXTULARIINA

(1) Shallow-water agglutinated foraminifera:
- Coskinolina Stache, 1875
- Lituonella Schlumberger, 1905
- Pseudochrysalidina Cole, 1941

(2) Deep-water and shallow-water agglutinated foraminifera
- Textulariella Cushman, 1927
- Liebusella Cushman, 1933

SUBORDER MILIOLINA

Cornuspiracea:
- (10) Nubeculariidae:
  - Discospirina Munier-Chalmas, 1902

Miliolacea:
- (4) Fabulariidae:
  - Austrotrillina Parr, 1942
  - Fabularia De France, 1820
  - Lacazinella Crespin, 1962*
  - Reticulogyra Adams & Belford, 1979

(5) Rhapydioninidae:
Raadshoovenia van den Bold, 1946
Rhapydionina Stache, 1913

(6) Praerhapydioninidae
Praerhapydionina Van Wessen, 1947

(7) Meandropsinidae
Pseudorbitolina Douvillé, 1910

Keramospheridae
Kanakia Hanzawa, 1957*

(8) Soritidae
Sorites Enrenberg, 1839
Marginopora Quoy and Gaymard, 1830
Miosorites Seiglie and Grove, 1976
Cyclorbiculina A. Silvestri, 1937
Archaias de Montfort, 1808
Fusarchaias Reichel, 1952*
Opertorbitolites Nuttall, 1925
Orbitolites Lamarck, 1801
Cyclorbiculinoides Robinson, 1974

(9) Peneroplidae
Cycloputeolina Seiglie and Grove, 1976
Parasorites Seiglie and Rivera, 1976

(11) Alveolinidae
Fasciolites Parkinson, 1811
Bullalveolina Reichel, 1936
Borelis de Montfort, 1808
Flosculinella Schubert, 1910
Alveolinella Douvillé, 1906
Rhipidionina Stache, 1913
(3) Incerta sedis
Yaberinella Vaughan, 1928
Somalina A. Silvestri, 1939*

SUBORDER ROTALIINA

"Rotaliacea"
(12) "Rotaliidae"
Daviesina Smout, 1954
Dictyoconoides Nuttall, 1925
Dictyokathina Smout, 1954*
Kathina Smout, 1954
Lockhartia Davies, 1932
Peneroperculoides Cole & Gravell, 1932
Sakesaria Davies, 1937
Smoutina Drooger, 1960
Miscellanea Pfender, 1935
Pseudowoodella Haque, 1956*
Storrsella Drooger, 1960*
Cuvillierina Debourle, 1955*

(13) Chapmaninidae
Sherbornina Chapman, 1922
Ferayina Frizzell, 1949
Crespinina Wade, 1955
Chapmanina A. Silvestri, 1931
Angotia Cuvillier, 1963

(14) Miogypsinidae
Miogypsina Sacco, 1893
Miogypsinoides Yabe and Hanzawa, 1928

(15) Elphidiidae
Pellatispirella Hanzawa, 1937

(16) Calcarinidae
Baculogypsina Sacco, 1893*
Baculogypsinoides Yabe and Hanzawa, 1930*
Siderolites Lamarck, 1801
Silvestrella Hanzawa, 1952

(17) Lepidorbitoididae
Actinosiphon Vaughan, 1929

Nummulitacea:
(18) Nummulitidae
Assilina d'Orbigny, 1839
Biplanispira Umbgrove, 1937*
Cycloclypeus W. B. Carpenter, 1856
Heterostegina d'Orbigny, 1826
Nummulites Lamarch, 1801
Operculina d'Orbigny, 1826
Operculinoides Hanzawa, 1935
Paraspiroclypeus Hanzawa, 1935
Pellatispira Boussac, 1906
Ranikothalia Caudri, 1944
Spiroclypeus Douvillé, 1905

(19) Discocyclinidae
Discocyclina Gumbel, 1870
Proporocyclina Vaughan & Cole, 1940
Asterophragmina Rao, 1942
Athecocyclina Gumbel, 1870
Aktinocyclina Gumbel, 1870

(20) Orbitoclypeidae
Pseudophragmina Douvillé, 1940
Orbitoclypeus A. Silvestri, 1907
Stenocyclina Caudri, 1972
Neodiscocyclina Caudri, 1972
Asterocyclina Vaughan & Cole, 1940
Orbitoidacea:
(22) Linderinidae
Linderina Schlumberger, 1893

(23) Lepidocyclinidae
Lepidocyclina (Lepidocyclina) Gumbel, 1870
L. (Eulepidina) Douvillé, 1911
L. (Nephrolepidina) Douvillé, 1911
L. (Polylepidina) Vaughan, 1924
Helicolepidina Tober, 1922
Helicostegina Barker & Grimsdale, 1936

Planorbulinacea:
(21) Planorbulinidae
Epiannularia Caudri, 1974
Eoannularia Cole and Bermúdez, 1944

(25) Cymbaloporidae
Fabiania A. Silvestri, 1924
Gunteria Cushman and Ponton, 1933
Halkyardia Heron Allen & Earland, 1918

(26) Victoriellidae
Eorupertia Yabe & Hanzawa, 1925
Maslinella Glaessner & Wade, 1959
Victoriella Chapman & Crespin, 1930

(28) Homotrematidae
Homotrema Hickson, 1911
Miniacina Galloway, 1933

Asterigerinacea:
(24) Amphisteginidae
Amphistegina d'Orbigny, 1826
Boreloides Cole and Bermúdez, 1944
Eoconuloides Cole and Bermudez, 1944
Tremastegina Brönnimann, 1951*

Acervulinacea:
(26) Acervulinidae
Gypsina Carter, 1877*
Planogypsina Bermúdez, 1952*
Sphaerogypsina Galloway, 1933*

Rotaliina incerta sedis:
Pseudolepidina Barker and Grimsdale, 1937
Margaritella Caudri, 1974
Sporadotrema Hickson, 1911*

Originated in the Holocene?
Ripacubana Loeblich & Tappan, 1964
Riveroina Bermúdez, 1939
Craterites Heron-Allen and Earland, 1924
Keramosphaera Brady, 1882

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CRETACEOUS TRANGRESSIVE AND REGRESSIVE EVENTS IN THE AVALON BASIN, GRAND BANKS OF NEWFOUNDLAND

L. B. THOMPSON

Mobil Exploration and Producing Services Inc.
P.O. Box 650232,
Dallas, Texas 75265-0232

ABSTRACT

Relative changes in sea level may result from local or regional tectonics, from eustatic cycles, or from a combination of the above mechanisms. The stratigraphy and structural history of late Jurassic to early Tertiary sediments deposited in the Avalon Basin on the Grand Banks of Newfoundland were examined in an attempt to establish the sequence of sea-level changes and to determine the mechanisms responsible for those changes. Many sea-level changes can be correlated from the North American Atlantic Coast to northern Europe, but it is unclear whether this synchroneity resulted from regional tectonics associated with the opening of the North Atlantic or is related to eustatic events. Local tectonism was responsible for the depositional sequences which created the major oil reservoirs in the Avalon Basin, and these conditions may not be repeated elsewhere on the Atlantic Shelf.

INTRODUCTION

The Avalon Basin is an extensional half-graben on the Grand Banks of Newfoundland, Atlantic Coast of Canada (Figs. 1 and 2). The basin was formed during Triassic rifting of Paleozoic basement. The basin is filled with middle to late Mesozoic synrift and post-rift sediments up to ten kilometers thick.

Within the last decade, significant amounts of hydrocarbons have been discovered in the Avalon Basin. As a result, a large data base has been amassed (over 45 wells drilled and more than 430,000 km of reflection seismic data collected), but most of these data have been analyzed only by the oil companies operating in the area. Only a handful of studies have been published on the regional stratigraphy (Jansa and Wade, 1975; Parsons and others, 1985; Tucholke and Fry, 1985), and one recent study addresses detailed tectonics and stratigraphy of the Avalon Basin (Tankard and Welsink, 1986).

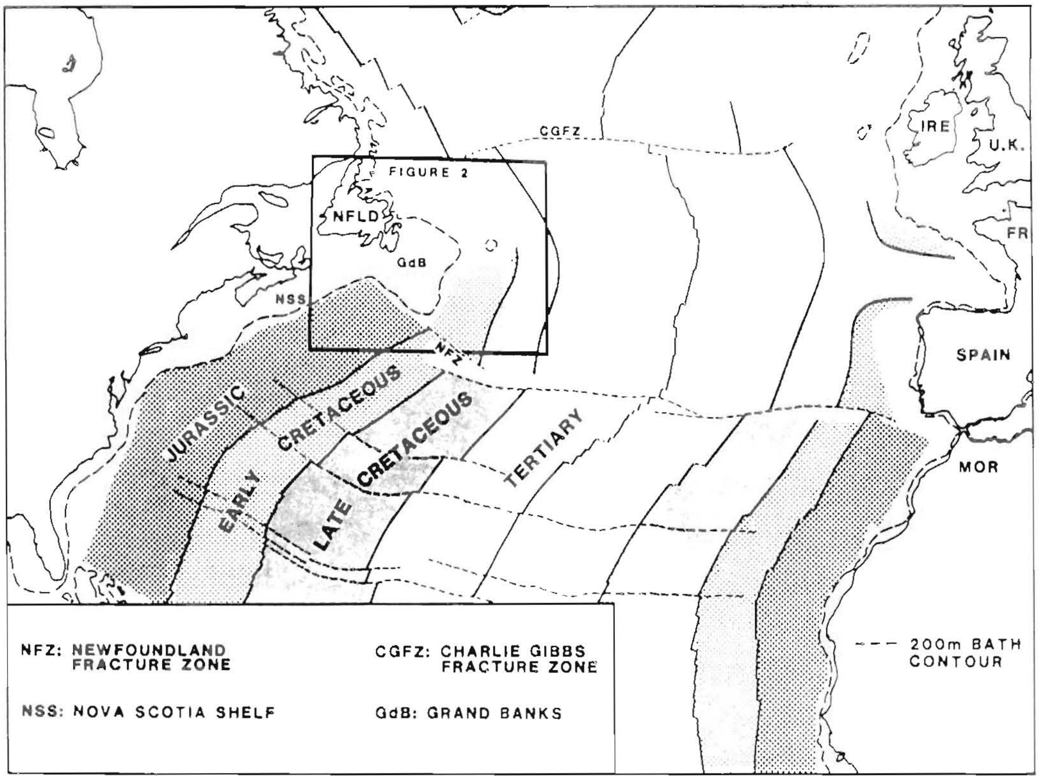


Figure 1. Map of the North Atlantic Ocean showing age of basin and location of Figure 2.

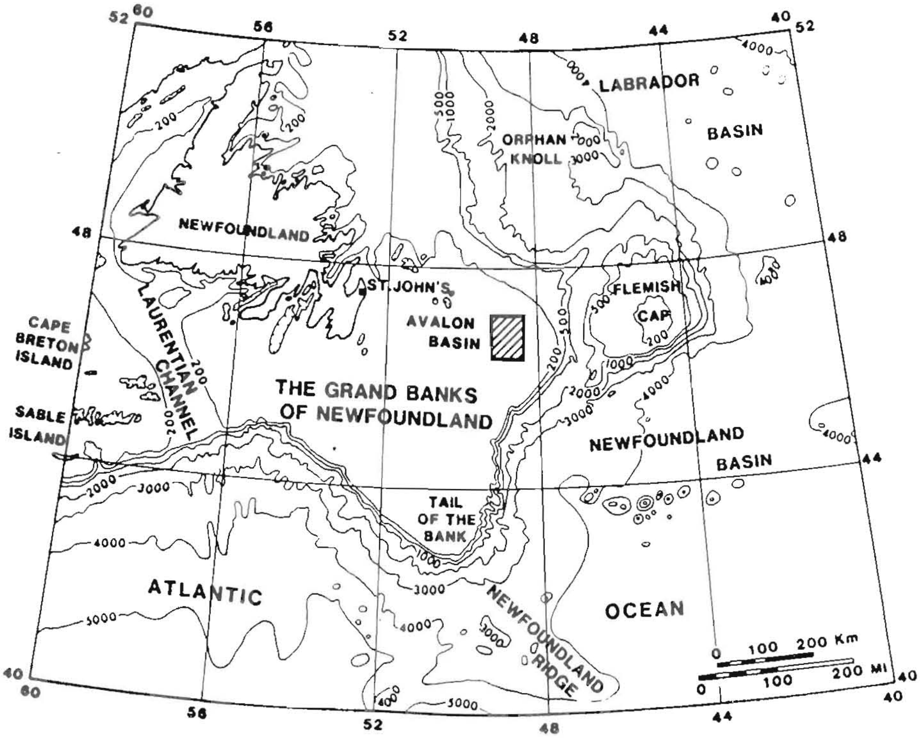


Figure 2. Location of the Avalon Basin on the Grand Banks of Newfoundland.

This study examines the transgressive and regressive events which are preserved in the rock record in the Avalon Basin. An attempt is made to examine the causes of sea-level changes and to determine whether an event is the result of local tectonism, regional tectonism or eustatic change (Montadert, 1984; Parkinson and Summerhayes, 1985). Several studies have examined the tectonic history of the continental margin of Eastern Canada (Schlee and Jansa, 1981; Keen, 1982) and of the U.S. Atlantic coast (Grow and Sheridan, 1981; Watts and Thorne, 1984) and of the stratigraphy around the North Atlantic Basin (Hardenbol and others, 1981; de Gracianski and others, 1982; Poag, 1982; Libby-French, 1984; van Harten and van Hinte, 1984). A sea-level curve is constructed for the Avalon Basin from late Jurassic to early Tertiary, and comparisons are made to data from around the North Atlantic Basin. The Avalon Basin curve is constructed using techniques similar to those of Vail and others, (1977).

TECHNIQUES

Biostratigraphic datums are derived from the examination of several microfossil groups present in ditch cuttings (composite 10 meter samples) and sidewall cores from 36 wells in the Avalon Basin (Fig. 3). Paleobathymetry is derived from integrating microfossil assemblages (primarily

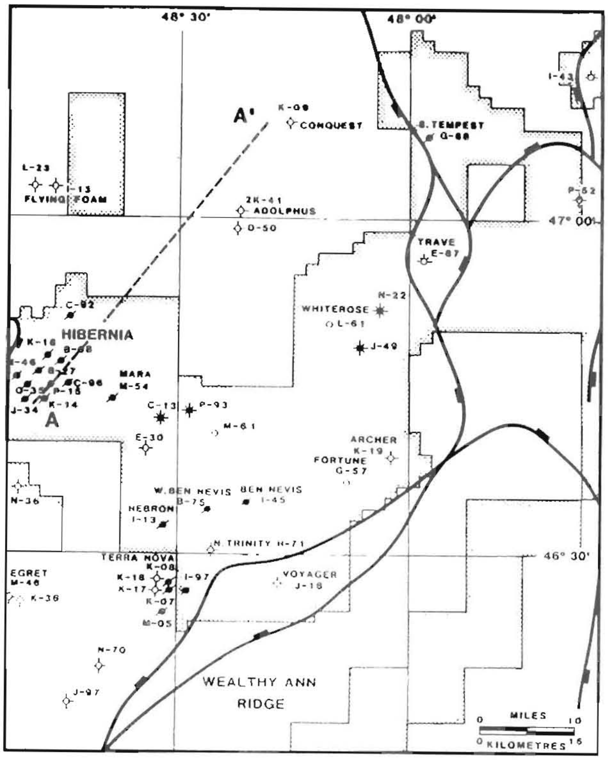


Figure 3. Principal wells in the Avalon Basin. A-A' is the location of the seismic section shown on Figure 4.

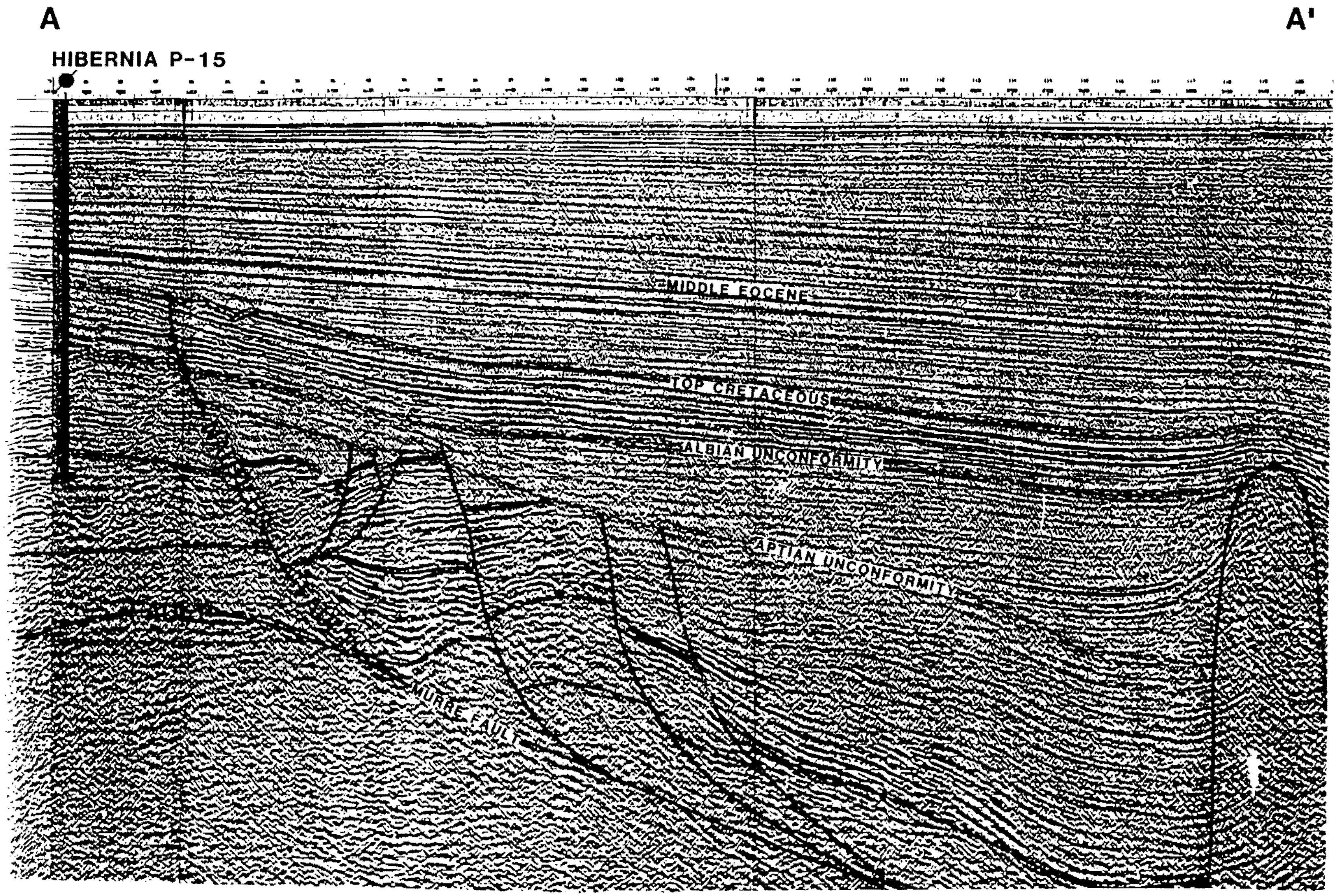
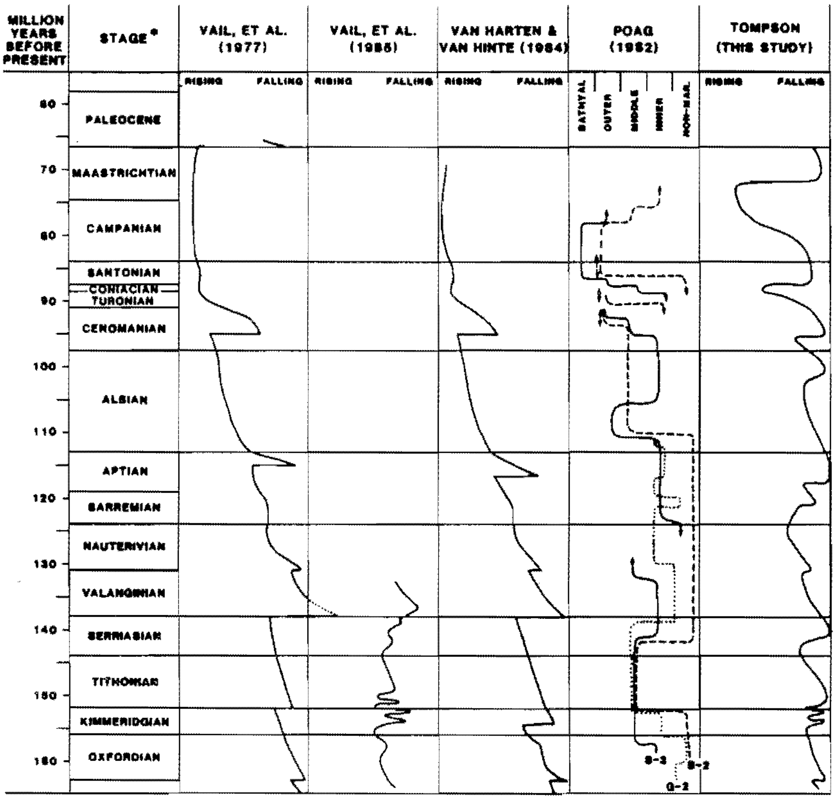
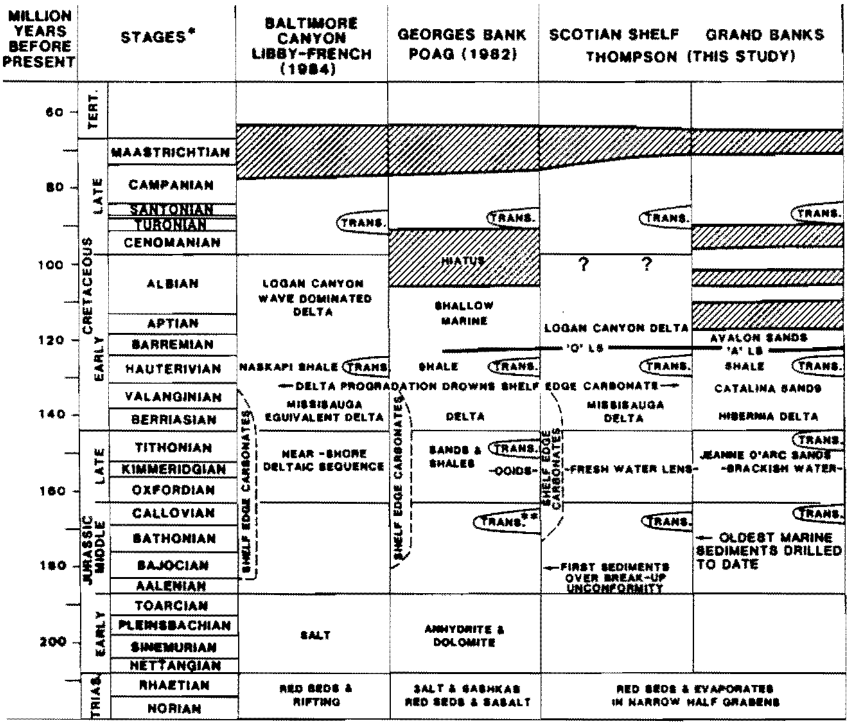


Figure 4. Seismic section from the Avalon Basin. See Figure 3 for location.



*STAGE BOUNDARIES PLACED AFTER KENT AND GRADSTEIN (1983)

Figure 5. Late Jurassic to Tertiary transgressive-regressive events in the North Atlantic Basin.



* AGES OF STAGE BOUNDARIES AFTER KENT AND GRADSTEIN (1983)
** "TRANS."=TRANSGRESSION

Figure 6. Major sediment features on the Atlantic Coast of North America.

Foraminifera and ostracodes) with ichnofossil studies, sedimentary structure analyses from conventional cores (where possible), and from lithofacies data collected from ditch cuttings. These data include lithologic sequences (such as coarsening-upward sequences) and detailed lithologic determinations (such as peloidal or ooid grainstones, siderite nodules, etc.). Note that paleobathymetries are not intended to represent absolute water depths, but approximate ranges of water depth.

Biostratigraphic data from the well bore are integrated with seismic data in order to interpolate events around the basin (Fig. 4). Structural data developed from seismic interpretations are also incorporated into this study. Several problems of detailed correlation among wells within the basin remain unresolved. The most notable problem is a precise understanding of the intricate stratigraphy of the Avalon sand reservoir of Early Cretaceous age. Nevertheless, an overall transgressive-regressive history is presented as a stimulant to further refinement in understanding Cretaceous sea level events.

RESULTS

Figure 5 summarizes the sea-level history of the Avalon Basin and compares it with histories from other areas of the North Atlantic Basin from Late Jurassic to Early Tertiary. Changes in sea level are relative and not intended to represent specific values.

LATE JURASSIC

North Atlantic Basin

Shelf-edge carbonates dominated the Atlantic margin throughout the Late Jurassic from the Blake Plateau to the Scotian Shelf (Fig. 6). Landward from the carbonate build-up a Kimmeridgian freshwater lens is indicated in the Scotian Shelf area and widespread coals were formed on Georges Bank. A Late Jurassic near-shore deltaic sequence developed in the Baltimore Canyon area. A latest Jurassic transgressive sequence is evident in the Georges Bank COST G-1 and G-2 wells (Poag, 1982).

The latest Jurassic sea-level curves of Vail and others, (1985) (Fig. 5) show a great deal of vascillation during the Kimmeridgian and early Tithonian. British ostracode assemblages (van Harten and van Hinte, 1984) also show a middle Kimmeridgian shallowing event (Fig. 5).

Avalon Basin

Very shallow marine environments were present in the Avalon Basin during the Kimmeridgian and early Tithonian. Foraminifera are restricted to small arenaceous forms and occasional miliolids and _Lenticulina_. This interval is composed of fine-grained sand and shales, including kerogen-rich source beds. For a discussion of the geographical extent of these source rocks see Powell, 1985. In the middle Kimmeridgian, normal marine conditions were interrupted by a shallowing event which resulted in brackish to freshwater conditions as seen by ostracode assemblages dominated by either _Darwinula_ or _Bisulcocypris_. Another shallowing event in the Tithonian resulted in an erosional unconformity which is present in the southern portion of the basin. A bed-load dominated river complex and delta, which is locally termed the Jeanne d'Arc Sandstone, is built out on top of the unconformity. Shallow marine sediments with a good marine dinoflagellate assemblage were deposited above the sands.

JURASSIC-CRETACEOUS BOUNDARY

The placement of the Jurassic-Cretaceous boundary on the Grand Banks is problematical. A Tithonian to Berriasian transition zone is present subjacent to the Hibernia delta. This zone includes the Berriasian nannoplankton _Polycostella senaria_ and the Tithonian palynomorphs _Amphorula metaelliptica_ and _Systematophora areolata_. Ascoli, Poag, and Remane (1984) have published an integrated calpionellid, foraminifer, and ostracode zonation of the Tithonian to Berriasian interval. This zonation is applicable on the Atlantic shelf as far north as the Grand Banks Bonnition H-32 well, but efforts to firmly fix the Jurassic-Cretaceous boundary within the Avalon Basin have been unsuccessful.

BERRIASIAN TO VALANGINIAN

North Atlantic Basin

A delta-building episode was widespread from the Baltimore Canyon to the Grand Banks (Fig. 6) during Berriasian to Valanginian. By the end of the Valanginian, these complexes had prograded across the shelf in many areas and drowned the extensive shelf edge carbonate unit.

The generalized Vail and others, (1977) sea-level curve and the sea-level curve of van Harten and van Hinte (1984) show a pronounced regression at the base of the Valanginian (see Figure 5). The Hibernia delta was deposited during a fall in sea level, and in the Avalon Basin the regression appears to be an intra-Berriasian event.

Avalon Basin

The Hibernia delta was the dominant depositional feature of this interval. A bed-load dominated river and delta complex, this feature built out rapidly over a marine shale. This event was Berriasian to early Valanginian as indicated by the palynomorphs _Muderongia simplex microperforata_ and _Imbatodinium kondratjevi_. Valanginian deposition was rapid (up to 2500m) and complex (at least two minor transgressive pulses and two late Valanginian sand and carbonate sequences). These strata contain

Foraminifera, ostracode and palynomorph assemblages that allow detailed local zonation.

HAUTERIVIAN

North Atlantic Basin

The Naskapi Shale extends from the Scotian Shelf to the Baltimore Canyon. The Naskapi was deposited during a transgression that is recognizable along the entire coast (Fig. 6). Note the presence of the "O" Limestone northward from Georges Bank. This lithologic unit may be equivalent to the "A" Limestone in the Avalon Basin.

Avalon Basin

The Hauterivian section in the Avalon Basin is comprised of a transgressive shale sequence and contains a rich assemblage of Lower Cretaceous Foraminifera. Several species of *Epistomina* and *Lenticulina* are present. The upper Hauterivian to lowest Barremian is marked by a shallower-water carbonate lithology, locally referred to as the "A" Limestone.

BARREMIAN

North Atlantic Basin

Generalized data from the Atlantic offshore show shallow marine conditions during this time (Poag, 1982). Ostracode data from England (van Harten and van Hinte, 1984) indicate a slight shallowing at this time.

Avalon Basin

Shallowing conditions continued above the "A" Limestone and, around the basin margins, resulted in an unconformity dated as middle Barremian. The Barremian palynomorphs *Muderongia simplex* and *M. imparilis* are above the unconformity, and the Barremian to Hauterivian palynomorphs *Ctenidodinium elegantulum* and *Phoberocysta* sp. and microfossils *Hutsonia* sp. (3) (Ostracod) and *Choffatella decipiens* (Foraminifera) occur below the unconformity.

APTIAN

North Atlantic Basin

A major Aptian regression was documented by Vail and others, (1977) and van Harten and van Hinte (1984) (Fig. 5). This regression is represented by an unconformity on the North American Atlantic continental shelf as seen in COST wells GE-1 and B-2 (Poag, 1982). In England this regression is dated as middle Aptian (van Harten and van Hinte, 1984). An Aptian hiatus is also noted (de Graciansky and others, 1982) across the entire deeper portion of the North Atlantic Basin.

Avalon Basin

Most of the Aptian is missing from the margins of the Avalon Basin (Fig. 4) as a result of a large angular erosional unconformity. Early Aptian to Barremian palynomorphs occur below the unconformity, and some Aptian as well as Albian and younger forms such as *Xiphophoridinium alatum* and *Rugubivesiculites rugosus* occur above this datum. Seismic data indicate that the oldest sediments on top of this unconformity are in the central portion of the basin. Although these sediments have not been drilled, they are projected as early Aptian in age. The youngest sediments below the unconformity are believed to be late Barremian to early Aptian.

ALBIAN

North Atlantic Basin

The Logan Canyon delta built out in the Baltimore Canyon area during the Albian. On Georges Bank shallow marine sediments and a local unconformity near the top of the Albian are present. (Libby-French, 1984). In England there is no apparent unconformity at this time (van Harten and van Hinte, 1984).

Avalon Basin

A second angular unconformity, locally named the "mid-Cretaceous unconformity" in the Avalon Basin, occurred during Albian time (Fig. 4). The duration of this unconformity is unclear since it is bracketed by Albian microfossils.

CENOMANIAN

North Atlantic Basin

A Cenomanian regression and erosional event of major importance in the history of the North Atlantic Basin is found all along the North American Atlantic shelf (Libby-French, 1984), in England (van Harten and van Hinte, 1984) and also in the central portion of the North Atlantic basin (de Gracianski and others, 1982).

Avalon Basin

Most of the Cenomanian appears to be absent in the Avalon Basin. The uppermost Albian and lower Cenomanian are present, as indicated by a succession of Albian microfossil assemblages containing *Epistomina spinulifera* and *Neocythere mertensi*, which are overlain by an assemblage that includes *Rotalipora cushmani*. An unconformity at the base of the Petrel Limestone separates the Cenomanian assemblages from prolific Turonian microfaunas. Some reworking of Cenomanian fossils is seen in the Petrel.

TURONIAN

North Atlantic Basin

A major Turonian transgression is noted all along the coasts of the North Atlantic basin. It has been well studied in both the North American coast and in England (Fig. 5).

Avalon Basin

A major transgression is represented by the widespread deposition of the Petrel Limestone. The Petrel contains a prolific planktonic foraminiferal fauna with abundant Marginotruncana pseudolinneiana, indicating middle to outer shelf paleoenvironments.

CONIACIAN-SANTONIAN

North Atlantic Basin

A regressive event is indicated in the Baltimore Canyon area at this time (Poag, 1982), and a minor decrease in sea level is noted in England (van Harten and van Hinte, 1984).

Avalon Basin

Coniacian to Santonian sediments appear to indicate a regression in the Avalon Basin also at this time. Although foraminiferal faunas are less prolific than in the Turonian, this section appears to be missing in a number of wells drilled near the margin of the basin. This shallowing event lasted into the early Campanian where massive sand bodies are seen prograding over top of older sediments (Fig. 4).

CAMPANIAN-MAASTRICHTIAN

North Atlantic Basin

Abundant late Campanian to early Maastrichtian assemblages are found on the North American shelf (Poag, 1982), and deep water conditions are indicated in England (van Harten and van Hinte, 1984). No late Maastrichtian microfossils have been found on the North American Atlantic shelf or in England.

Avalon Basin

An abundant outer neritic foraminiferal fauna is present in late Campanian and early Maastrichtian. A diverse Globotruncana assemblage is the most prolific component of this fauna. No Cretaceous sediments younger than early Maastrichtian are found in the Avalon Basin, thus the terminal Cretaceous unconformity resulted in a hiatus lasting from early Maastrichtian into the Paleocene.

DISCUSSION

Figure 5 summarizes the transgressive-regressive cycles for the Avalon Basin (this paper), the North American Atlantic offshore (Poag, 1982), England (van Harten and van Hinte, 1984) and the Vail and others, (1977) sea level curve. Local tectonism can be determined from these curves either as an event which is not found outside the basin or as an event of large magnitude which cannot be caused by sea-level changes on broad regional tectonic events. It is more difficult to differentiate between eustatic cycles and regional tectonics as both events would yield the same patterns on the curves. With this in mind, the following observations are made:

1) Crustal thinning and movement along a major listric fault (Murre fault in Fig. 4) in the Avalon Basin occurred from the Callovian through the end of the Jurassic (Tankard and Welsink, 1986). Synrift sediment deposition during this time kept pace with fault movement as indicated by paleobathymetry data. Extremely shallow marine conditions, punctuated by brackish water episodes, dominated this interval. This pattern is seen around the entire margin of the proto-North Atlantic (Fig. 5), so the minor fluctuations in sea level during this period were not related to local fault activity but to either eustatic sea-level changes or to regional tectonism that was felt throughout the North Atlantic basin.

2) The Berriasian to early Valanginian section in the Avalon Basin shows a marked regression followed by rapid sediment deposition. The Hibernia delta and its associated facies deposited up to 2500 meters of material in as little as five to seven million years. Major regressive events are documented around the margin of the proto-North Atlantic Basin (Fig. 5), but their timing is problematical. They may represent a single event which has been poorly calibrated as a result of vagaries in chronotratigraphic data existing near the Jurassic-Cretaceous boundary, or by a tectonic event which affected disparate portions of the basin at slightly different times. In any case, the magnitude of the event preserved in the Avalon Basin as seen by the thick sediment wedge is indicative of local tectonism and crustal loading, not global eustatic conditions.

3) A widespread Valanginian to Hauterivian transgression (Fig. 5) is reflected in the sediments in the Avalon Basin. The local tectonic regime remains unchanged from early Valanginian as rifting continued, so this event appears to represent a North Atlantic Basin eustatic event.

4) A middle Barremian unconformity is present in the Avalon Basin. Data from around the North Atlantic indicate a slight regression at this point, but not a major drop in sea level. Movement of salt and faulting within the basin indicate that local tectonism may have reinforced this event.

5) A major Aptian unconformity is present in the Avalon Basin within the Avalon sand sequence (Fig. 4). This event is recognized around the entire North Atlantic Basin (Fig. 5) and appears to be related to a change in tectonics at the mid-ocean ridge. Prior to this time, drift was occurring south of the Newfoundland Fracture Zone and extensional rifting north of the fracture (Tankard and Welsink, 1986) (note that the Newfoundland Fracture Zone is seen on Figs. 1 and 2 just south of Tail of the Bank and through the Newfoundland Ridge). In the Aptian, a "propagating rift" (Vink, 1982) began moving northward towards Flemish Cap; and the Avalon Basin tectonic regime changed from extensional rift to drift, but probably went through a transitional phase which lasted well into the Albian.

The Aptian sea level drop in the North Atlantic Basin appears to be eustatic in the classic Vail and others (1977) sense. However, as the above discussion demonstrates, it was apparently related to a major shift in regional tectonic aspect as the Grand Banks area went from rift to drift. Therefore, this Aptian event should not necessarily be viewed entirely as a "global" event. There is ample evidence of local tectonism within the Avalon Basin at this time which further accentuated this feature. This evidence can be seen on Figure 4 where a number of faults are terminated by the Aptian unconformity. Note also on the curves that the timing of this event appears to be early Aptian not late Aptian as noted by Vail and others (1977).

6) A second major unconformity is present in the Avalon Basin during the middle to late Albian (Fig. 4). This event does not appear to be regional (Fig. 5) and was apparently controlled by local tectonic activity (Tankard and Welsink, 1986).

7) An unconformity in the Cenomanian is documented from the margins to the center of the North Atlantic Basin (Fig. 5 and de Graciansky, and others, 1982). Like the Aptian "breakup" event, this regression may be the result of regional tectonic change (Masson and Miles, 1984) affecting the entire basin rather than a passive global eustatic event. There is no unusual tectonic activity occurring in the Avalon Basin at this time (Tankard and Welsink, 1986).

8) Late Cretaceous sea-level changes in the Avalon Basin appear to coincide with global events as almost all local tectonic activity ceased. From that time to the present, the basin is seen as part of a passive continental margin undergoing thermal cooling and compaction.

ACKNOWLEDGEMENTS

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SEQUENCE STRATIGRAPHY, BIOSTRATIGRAPHY, AND CORRELATION OF EOCENE THROUGH LOWER MIOCENE STRATA IN NORTH CAROLINA

V. A. ZULLO AND W. B. HARRIS

Department of Earth Sciences, University of North Carolina at Wilmington, North Carolina 28403

ABSTRACT

Exposed Eocene through lower Miocene rocks of the North Carolina Coastal Plain are assigned to ten depositional sequences representing Coastal Onlap Supercycles TA3, TA4 and TB1 (Lutetian - Aquitanian Stages). Five depositional sequences and four megafossil zones are recognized in Eocene rocks assigned to Castle Hayne Limestone and New Bern Formation. The oldest depositional sequence, Sequence 0 has not yielded age-diagnostic fossils. Sequence 1, characterized by the Protoscutella mississippiensis Assemblage Zone, is correlated with the lower Lisbon Formation and equivalents in the Gulf Coastal Plain (middle Lutetian). Sequence 2 contains the P. conradi Assemblage Zone, and is correlated with the upper Lisbon Formation and equivalents in the Gulf Coast (upper Lutetian to lower Bartonian). The lowermost part of Sequence 3 contains the P. plana Assemblage Zone, and is correlated with the Gosport Sand in the Gulf Coast (Bartonian). The remainder of Sequence 3 and Sequence 4 is characterized by the Periarchus lyelli Assemblage Zone. Upper Sequence 3 is correlated with the Moodys Branch Formation, the North Twistwood Creek Clay and the Cocoa Sand in the Gulf Coast (upper Bartonian(?), lower Priabonian). Sequence 4 includes the New Bern Formation and the uppermost part of the Castle Hayne Limestone in North Carolina, and is correlated with the Pachuta Marl and, possibly, the Shubuta Marl in the Gulf Coast (upper Priabonian). Based on megafossil and calcareous nannofossil evidence, the Protoscutella plana-Periarchus lyelli zonal boundary is equivalent to the Claibornian-Jacksonian Stage boundary in the Gulf Coastal Plain.

Five depositional sequences and two megafossil zones are recognized for Oligocene and lower Miocene rocks. Sequence 5, characterized by the Lophobalanus kellumi Assemblage Zone, is correlated with lower Vicksburgian strata of the eastern Gulf Coastal Plain (Rupelian). Sequences 6 through 9, characterized by the L. baumi Assemblage Zone, are correlated with Chickasawhayan and Tampan strata of the Gulf Coast (Chattian - lower Aquitanian).

INTRODUCTION

It is the object of this paper to document temporal and spatial distribution of Eocene through lower Miocene strata in the North Carolina Coastal Plain, and to provide new data from invertebrates (oysters, pectinids, echinoids and barnacles), calcareous nannofossils and planktonic foraminifers that serve to correlate these strata with deposits of the southeastern Atlantic and eastern Gulf Coastal Plains. The goal of this paper is to emphasize that global sea level changes are reflected by specific stratigraphic signatures in coastal depositional basins. This is to say that deposition or unconformity development in one basin is recognizable in all coastal depositional basins, unless significant local tectonism can be demonstrated to have negated the effects of eustacy. Because syn- and post-depositional environmental, climatic and tectonic factors can alter stratigraphic signatures preserved on basin margins, the sequence stratigraphic approach must be integrated with biostratigraphic data to correlate partially preserved depositional sequences within and between basins.

The stratigraphy and age of exposed Eocene, Oligocene and Miocene rocks in the North Carolina Coastal Plain have been the subject of periodic debate for 75 years. Most recently, stratigraphic revisions by Baum and others (1978) and Ward and others (1978) resulted in two different nomenclatures, lithostratigraphic interpretations and age determinations based on studies of the same exposures. Baum and others (1978) recognized two Eocene units, the middle Eocene Castle Hayne Limestone and the upper Eocene New Bern Formation, one Oligocene unit, the lower to middle Oligocene Trent Formation, and three lower Miocene units, the laterally equivalent Belgrade and Silverdale formations and the overlying Crassostrea channel deposits. Ward and others (1978) recognized a single middle Eocene unit, the Castle Hayne Formation divided into three members, one upper Oligocene unit, the River Bend Formation, and one lower Miocene unit, the Belgrade Formation, divided into two, laterally equivalent members. The relationships between these interpretations are illustrated in Figure 1. Since 1978, several papers have addressed the controversies posed by these revisions, with the majority focused on the age of the Castle Hayne Limestone (see Zullo and Harris, 1986 for summary).

The difficulties encountered in interpreting the stratigraphic relationships of Coastal Plain rocks are directly related to the scarcity of natural exposure, and the lack of subsurface control and modern biostratigraphic studies. Mapping is dependent on minor outcrops along streams and isolated quarry operations, and both types of exposure are ephemeral. Many of the classic natural localities described in the older literature are either covered or badly weathered, and older quarries are now flooded. Because of the unresolved controversies in stratigraphic interpretation, and because normal mapping procedures were untenable, we have interpreted these lower Cenozoic rocks based on concepts of sequence stratigraphy and new biostratigraphic data. Eight major quarries between Wilmington and New Bern were studied in detail (Figures 2 through 10), and the data obtained from these sections were used to reinterpret the stratigraphic

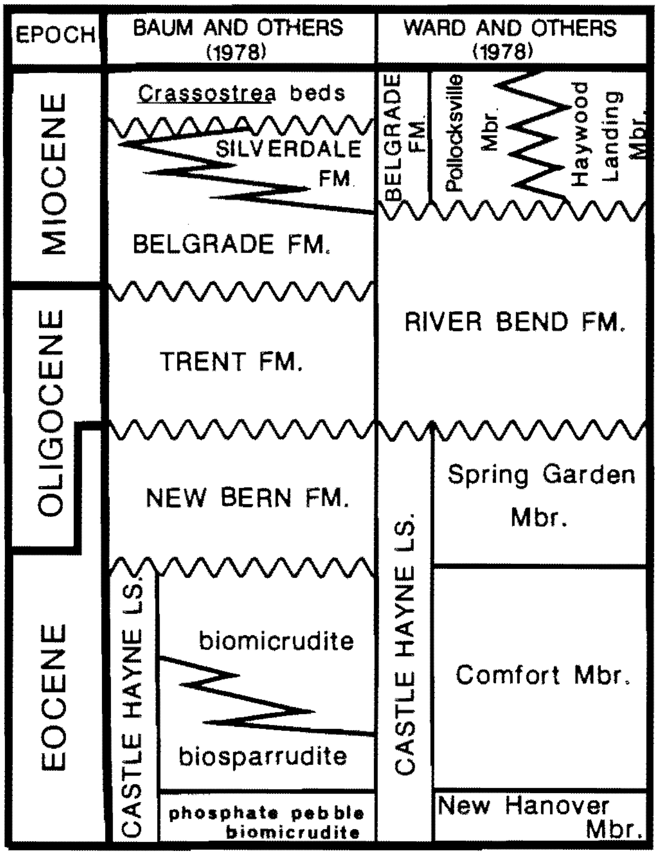

Figure 1. Comparison of the interpretations of Baum and others (1978) and Ward and others (1978).

relationships of numerous minor exposures throughout the outer Coastal Plain. Based on these studies, seven depositional sequences, representing parts of Coastal Onlap Supercycles TA3, TA4, and TB1 are recognized for exposed Eocene through lower Miocene rocks of the North Carolina Coastal Plain.

SEQUENCE STRATIGRAPHIC CONCEPTS

An outgrowth of the development of the stratigraphic interpretation of seismic data by Vail and others (1977) has been a re-emphasis of the importance of unconformities for subdividing the geologic column into depositional cycles. Recognition of unconformities, or their correlative conformities is one of the major premises for the stratigraphic interpretation of seismic data. These surfaces, or sequence boundaries, allow subdivision of seismic reflection data into genetically related packages. Mitchum and others (1977) call the relatively conformable succession of strata that occurs between sequence boundaries depositional sequences. Depositional sequences are observable units interpreted as time-rock packages or chronostratigraphic units. As depositional sequences are bounded by unconformities, each is interpreted to represent a cycle of sea level change during which a relative rise and fall occurs (Vail and Mitchum, 1979). Vail and Mitchum suggested that sequence boundaries are synchronous world wide and are related to global cycles of sea level change. In this paper depositional sequence is used to refer to genetically related sediments bounded by unconformities that are observable in the field. Cycle is used in the interpretive sense to represent a rise and subsequent fall in sea level.

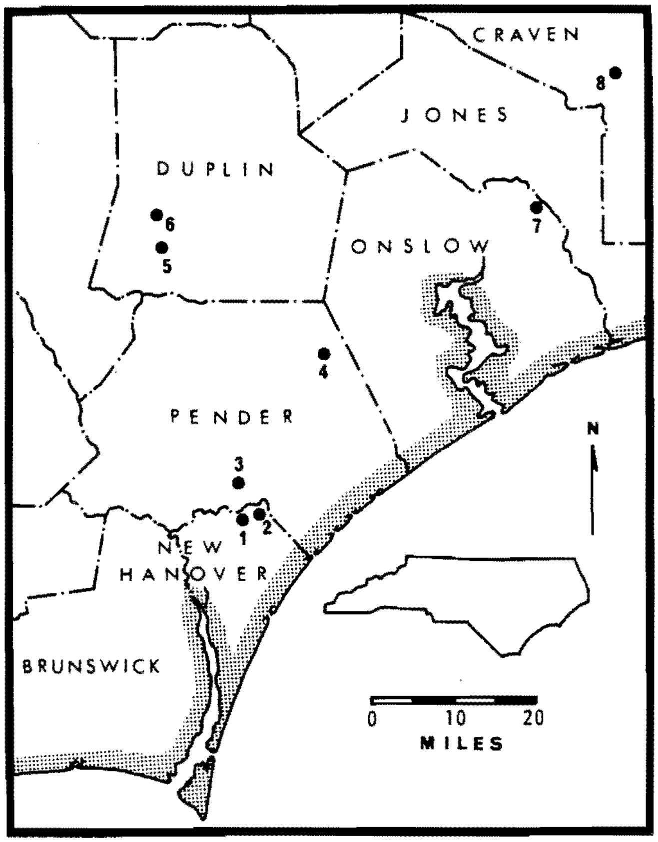

Figure 2. Location of the eight quarries used to develop the sequence-stratigraphic framework for the North Carolina Eocene to lower Miocene. (1) Martin Marietta Castle Hayne quarry, (2) Martin Marietta Ideal quarry, (3) Martin Marietta Rocky Point quarry, (4) East Coast Limestone quarry, (5) Billy B. Fussell quarry, (6) Atlantic Limestone quarry, (7) Martin Marietta Belgrade quarry, (8) Martin Marietta New Bern quarry.

LITHOSTRATIGRAPHY

Eocene

Five depositional sequences, each separated by a phosphatized and glauconitized disconformable surface, are recognized for Eocene rocks exposed in the North Carolina Coastal Plain. As these sediment packages do not conform to previously defined lithostratigraphic units, they are here designated, from oldest to youngest, Sequences 0 through 4.

Sequence 0. Recent excavations in the southwest corner of the Fussell quarry in Duplin County (Figure 3) exposed a bryozoan limestone resting

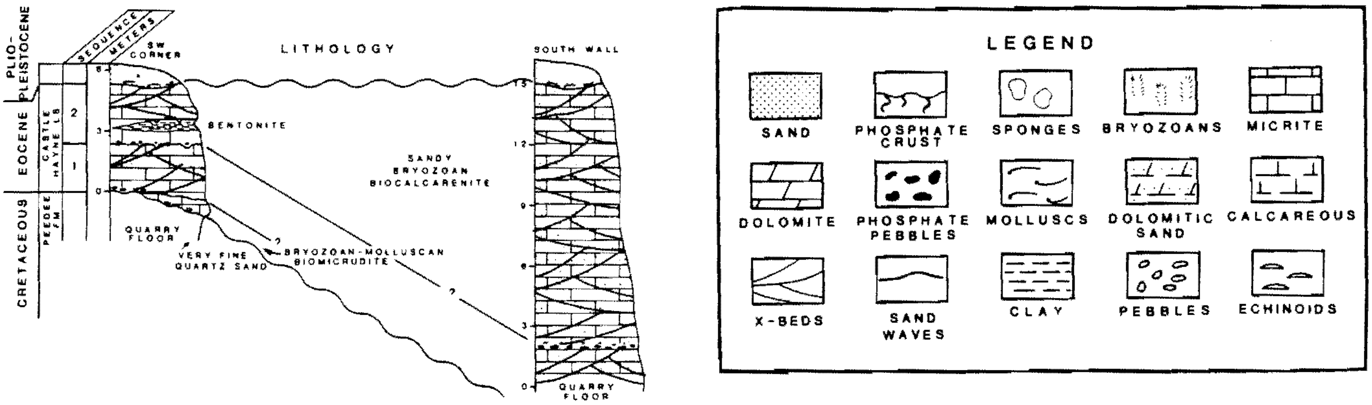

Figure 3. Billy B. Fussell Quarry. This quarry is located on the south side of State Route 1148, 1.1 km west of its intersection with U. S. Highway 117, Duplin County, in the Rose Hill 15" quadrangle. The quarry exposes almost 15 m of the Castle Hayne Limestone, and a 5 m thick sequence of surficial sand and clay. The lithologic legend applies to Figures 3 through 10.

disconformably on a Cretaceous high, and disconformably overlain by Sequence 1 sediments of the Castle Hayne Limestone. Large slump blocks of the bryozoan limestone are also present within overlying Sequence 1 deposits adjacent to and below the Cretaceous high. The bryozoan limestone consists of a lower, coarse, well-washed, cross-bedded bryozoan biosparrudite and an upper, dense, distinctively blue-gray, molluscan-bryozoan biomicrudite. Molluscs and worm tubes of the overlying Castle Hayne Limestone are attached to the highly bored and phosphatized upper surface of the molluscan-bryozoan biomicrudite. Although no age diagnostic fossils have been recovered from the limestone, we presume that it is Eocene and, possibly, early Claibornian in age.

The lower, indurated, sandy, bryozoan biomicrudite and biosparrudite exposed at the Atlantic Limestone quarry in Duplin County is tentatively referred to Sequence 0 (Figure 4). From Otte's (1979) description of the bored and encrusted disconformable upper surface, and the suggestion of post-depositional slumping, the lower unit is more similar to Sequence 0 at the Fussell quarry than it is to Sequence 1 at the Fussell quarry. Either Sequence 1 is not preserved at the Atlantic Limestone quarry or has yet to be recognized.

Sequence 1. This sequence includes the lower part of the New Hanover Member of the Castle Hayne Limestone of Ward and others (1978), the corresponding lower part of the phosphate-pebble conglomerate of Baum and others (1978), and the lower part of the Santeelampas beds of Harris and Zullo (1980). Where this sequence overlies the Rocky Point Member of the Upper Cretaceous Peedee Formation, it is represented by a sandy, phosphate-pebble biomicrudite containing reworked, often phosphatized and glauconitized, Cretaceous fossils and lithoclasts. In areas were it rests directly on Peedee sand, phosphate pebbles and reworked fossils are absent, and the dominant lithology is calcarenite containing a high percent of quartz sand and pebbles.

Sequence 1 is widespread throughout the southeastern North Carolina Coastal Plain, varying in thickness from a few centimeters to over 3 m.

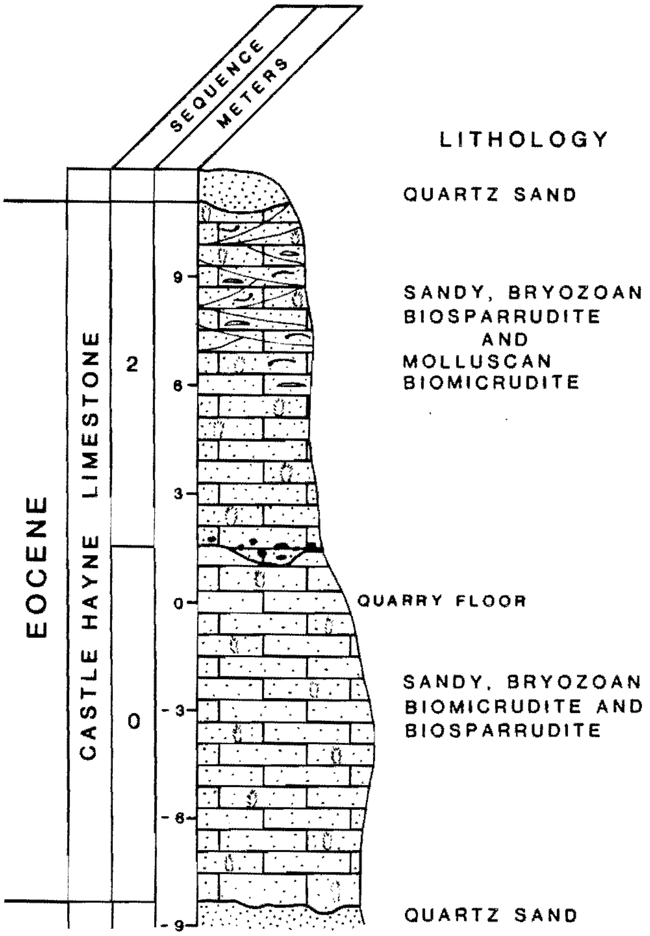

Figure 4. Atlantic Limestone Quarry. This quarry is located east of U. S. Highway 117, 3.8 km SSE of Magnolia, Duplin County, but is no longer in operation. The quarry was discussed by Otte (1979, 1981) from outcrop and core data. Here Eocene carbonates are disconformable on the Cretaceous Peedee Formation and are overlain disconformably by up to 4.5 m of surficial sand. The carbonate sequence consists of two distinctive parts separated by a disconformable surface of high relief.

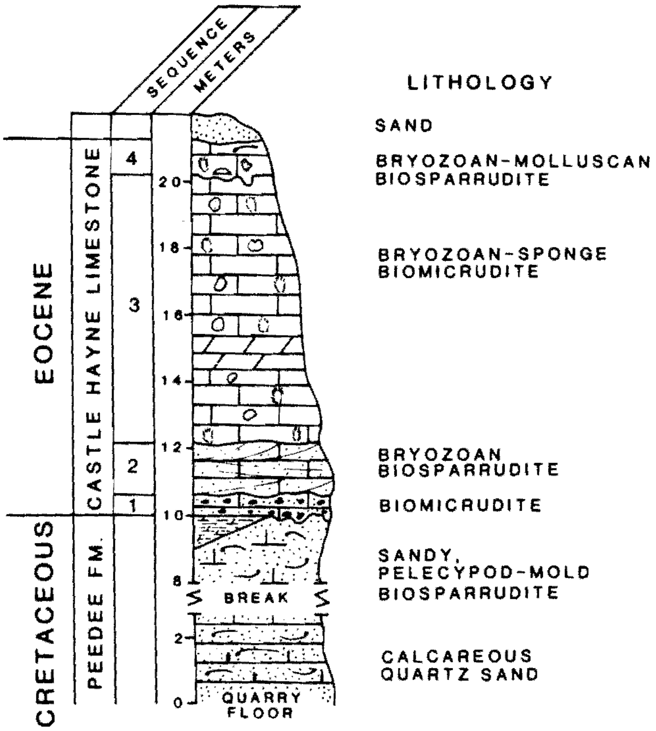

Figure 5. Composite section from the Martin Marietta Castle Hayne quarries. This abandoned quarry operation is the locality of the lectostratotype of the Castle Hayne Limestone selected by Baum and others (1978). The quarries are located 2.8 km east of the intersection of U. S. Highway 117 and Route 1002, on the north side of Route 1002, New Hanover County. The section exposed included up to 10 m of the Cretaceous Rocky Point Member of the Peedee Formation, disconformably overlain by up to 11 m of the Castle Hayne Limestone and about 1 m of surficial sand.

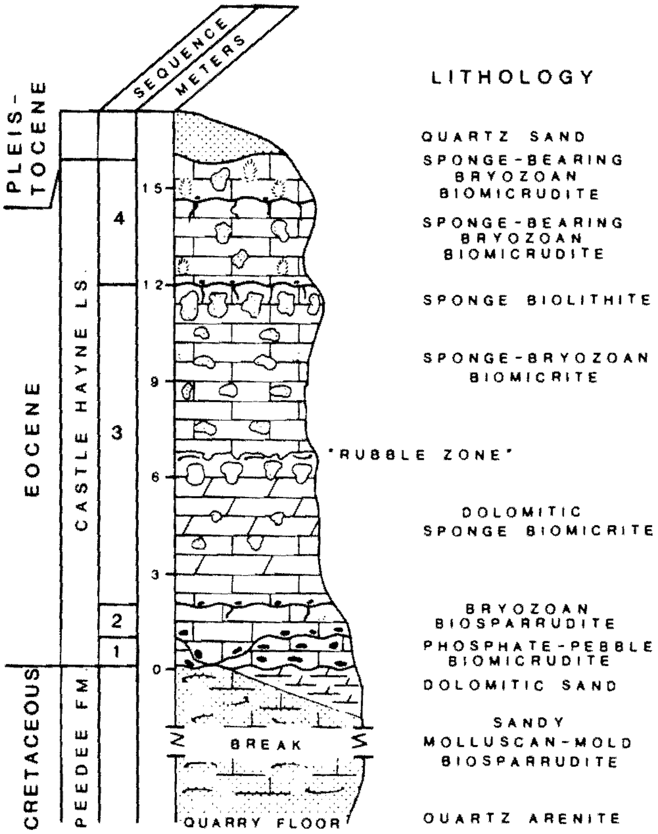

Figure 6. Martin Marietta Ideal Quarry. This quarry is located approximately 3.2 km east of the intersection of U. S. Highway 117 and State Road 1002, on the north side of Route 1002, New Hanover County. The quarry exposes about 7 m of the Cretaceous Rocky Point Member of the Peedee Formation, disconformably overlain by nearly 16 m of the Castle Hayne Limestone and 1.5 m of surficial sand and clay.

Age diagnostic fossils from Sequence 1 deposits have been found at the Fussell quarry and Natural Well in Duplin County, at the Lanier and Holly Ridge quarries in Pender County, and along Mosley Creek on the Lenoir-Craven county line. The identity of this sequence at other localities is based on lithology and superpositional relationship with overlying Sequence 2 deposits that contain a diagnostic fossil assemblage. At the Martin Marietta Castle Hayne (Figure 5) and Ideal (Figure 6) quarries in northern New Hanover County, Sequence 1 is rather thin, ranging from 0 to 0.5 m in thickness, and homogeneous throughout its observed lateral extent. It is in these quarries that Sequence 1 is represented by the phosphate pebble conglomerate typifying the New Hanover Member of the Castle Hayne Limestone. The only known North Carolina occurrence of the barnacle Euscalpellum chamberlaini is from Sequence 1 at the Castle Hayne quarry.

To the north, at the Martin Marietta Rocky Point quarry in Pender County (Figure 7), Sequence 1 attains greater thickness and is more variable lithologically. Deposits attributed to this sequence are present mainly in the central and northern part of the quarry and are absent to the east, west and south. Sequence 1 forms a discontinuous layer, ranging from 0 to 1.25 m thick, on top of the Cretaceous Rocky Point Member of the Peedee Formation and is best developed in low areas on top of the unit. It consists of varying lithologies all of which are characterized by fine- to medium-grained quartz sand. The lowermost lithology (1a) is a light gray, dense, molluscan biomicrudite with scattered phosphatized lithoclasts that range to several centimeters in diameter. Although variable in thickness, it does not exceed 35 cm. In sharp, conformable contact with this lithology is a light gray, partially indurated, cross-bedded, bryozoan biosparrudite (1b). Although no detailed studies of the cross-bed sets have been made, they generally dip to the northwest. The cross-bedded bryozoan biosparrudite grades upward into light gray, well lithified, bryozoan-molluscan biomicrudite (1c). These latter two lithologies are the dominant rock types within Sequence 1. The top of Sequence 1 is a disconformity that is marked by solution features that extend 20 to 30 cm into the top of the unit. The surface is coated with phosphate and glauconite mineralization. No age diagnostic fossils have been recovered from Sequence 1 at the Rocky Point quarry.

Sequence 1 at the Fussell quarry in Duplin County rests disconformably on the Cretaceous Peedee Formation and the contact is often exposed in dredge spoils adjacent to drainage ditches that have been cut below the quarry floor. Baum and others (1978) considered the exposures at this quarry to represent their biosparrudite facies of the Castle Hayne Limestone, and Ward and others (1978) assigned them to their Comfort Member. The Eocene section at the Fussell quarry does not contain any obvious disconformities and, in overall appearance, is lithologically homogeneous. However, about 2.3 m above the base of the section, a 0.3 m thick zone contains vein quartz pebbles, and phosphatized and glauconitized pebbles ranging in size to a maximum of 3 cm. The section below the pebble zone, here assigned to Sequence 1, consists principally of fine- to medium-grained, light gray, unconsolidated, sandy, bryozoan biocalcarenite with polydirectional cross-bed sets up to 20 cm in amplitude. Sequence 1 contains the echinoids *Protoscutella mississippiensis rosehillensis* Kier and *Santeelampas oviformis* Cooke, and a small oyster of the genus *Cubitostrea* that is conspecific with the form from the Martin Marietta Berkeley, South Carolina quarry Warley Hill Marl (=*C. lisbonensis* beds of the Santee Limestone) identified with *C. lisbonensis* by Powell and Baum (1982). Nannofossils studied by Worsley and Laws (1986) were assigned to Martini's Zone NP15.

To the east, Sequence 1 is exposed in the East Coast Limestone quarry in Pender County (Figure 8), where it is represented by a highly discontinuous, sandy, molluscan-mold biomicrudite. The unit is absent in the southern part of the quarry, but attains a thickness of 0.5 m in the northern part. Here, Sequence 1 overlies the Rocky Point Member of the Peedee Formation, and contains phosphatized and glauconitized pebbles and reworked Cretaceous fossils.

Biostratigraphic data provided by nannofossils (Worsley and Laws, 1986; Martini Zone NP15) suggest that deposits near the base of the Lanier and Holly Ridge quarries near Maple Hill, Pender County, and at Natural Well in Duplin County correlate to Sequence 1. The identification of the echinoid *Protoscutella mississippiensis rosehillensis* by Kier (1980) from Eocene rocks overlying the Paleocene Beaufort Formation at Mosley Creek on the Lenoir-Craven county line is indicative also of Sequence 1.

Sequence 2. Deposits assigned to Sequence 2 include the upper part of the New Hanover Member of Ward and others (1978), the corresponding phosphate-pebble biomicrudite of Baum and others (1978), and the upper part of the *Santeelampas* beds of Harris and Zullo (1980). Sequence 2 deposits have a distribution similar to those of Sequence 1, but are more continuous and usually thicker. Sequence 2 is disconformable on Sequence 1, Sequence 0, older Paleogene units, or the Cretaceous Peedee Formation, and the disconformity is usually a solutioned, phosphatized and glauconitized surface. As with Sequence 1, the carbonates of Sequence 2 have a high quartz sand content. Unlike Sequence 1, most of the exposures of Sequence 2 have yielded age diagnostic fossils.

Sequence 2 at the Martin Marietta Castle Hayne and Ideal quarries in northern New Hanover County is represented by about 0.5 m of white to cream, sandy biomicrite. Ward and others (1978) reported a well-preserved specimen of the oyster *Cubitostrea sellaeformis* at the Castle Hayne quarry, which Ward (personal communication, 1986) obtained from this unit.

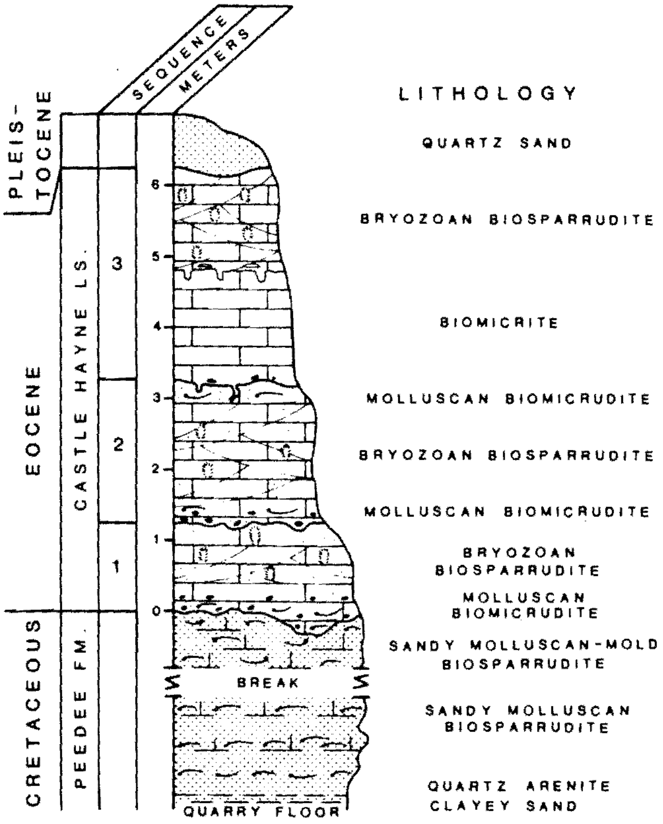

Figure 7. Martin Marietta Rocky Point Quarry. This quarry is located about 2 km southeast of Rocky Point, Pender County, in the Mooretown, North Carolina 7 1/2" quadrangle. The quarry exposes the Peedee Formation, the Rocky Point Member of the Peedee Formation, the Castle Hayne Limestone, and surficial sand and clay of unknown age.

At the Martin Marietta Rocky Point quarry to the north, Sequence 2 disconformably overlies Sequence 1 or the Rocky Point Member of the Peedee Formation and has a maximum thickness of 2 m in the central and eastern part of the quarry. Here the sequence consists of three lithologies that are similar to the those of Sequence 1. The basal lithology (2a) is a tan-gray, nonindurated to well lithified biomicrite or molluscan biomicrudite with phosphatized lithoclasts to several cm in size near the base. The middle lithology (2b) is a tan-gray, partially indurated, cross-bedded, bryozoan biosparrudite that grades upward into a light gray, dense, well lithified, molluscan biomicrudite (2c) containing the pectinid *Chlamys clarkeana*. Fine- to medium-grained quartz sand is present throughout the lithologies of Sequence 2. Sequence 2 rapidly thins to the east by onlap of the lower lithology onto Sequence 1 or the Rocky Point Member. Thickening of the middle lithology into large-scale sand waves with north-south orientation occurs at various locations in the quarry. The upper surface of Sequence 2 is marked by solution features that extend up to one meter into the unit. The solution features are coated with phosphate and glauconite mineralization which is commonly encrusted by bryozoans, oysters, and worm tubes. In addition, alternating sediment and phosphate-glauconite crusts in some of the solution features indicate episodic sedimentation. Biostratigraphically significant fossils occurring in Sequence 2 lithologies at the Rocky Point quarry include *Cubitostrea sellaeformis* (Conrad) from 2b and 2c, *Protoscutella conradi* from 2b and 2c, and *Chlamys clarkeana* from 2c.

Farther north, at the Fussell quarry, Sequence 2 is represented by over 12 m of light gray biocalcarenite that displays cross-bed sets with amplitudes to 1.5 m. The lower 7 m of the biocalcarenite is coarse-grained, whereas the upper 5 m is medium- to fine-grained. About 2.7 m above the pebble zone separating Sequence 1 from Sequence 2, a 5 cm thick lens of light gray, soft micrite forms a distinctive marker bed throughout much of the quarry. Insoluble residue studies, terrigenous percentages, and size decrease of the allochemical components indicate that the overall section fines upward above the micrite beds (Baum, 1980; Otte, 1981; this study). Sequence 2 at the Fussell quarry contains *Protoscutella conradi*, *Santeelampas oviformis*, and *Chlamys clarkeana*, indicating correlation with Sequence 2 at the Rocky Point quarry.

Sequence 2 at the nearby Atlantic Limestone quarry disconformably overlies the indurated biomicrudite and biosparrudite of Sequence 0, and is represented by an unindurated, sandy, bryozoan biomicrudite that grades upward into a loosely cemented, cross-bedded, bryozoan biosparrudite that contains thin interbeds of indurated molluscan biomicrudite in its lower part and is cross-bedded in its upper part. The lower 1.5 m of this unit contains phosphatized pebbles and cobbles, some of which appear to be clasts of the underlying indurated biosparrudite. The quartz sand and pebble content decreases upward. Sequence 2 at this quarry contains *Protoscutella conradi*, *Santeelampas oviformis* and *Eurhodia rugosa ideali* (see Kier, 1980).

At the East Coast Limestone quarry in Pender County, Sequence 2 is represented by basal molluscan biomicrudite that grades upward into porous, cross-bedded bryozoan biosparrudite, with a maximum thickness of 2 m in the northern part of the quarry. Large scale sand waves are developed in the biosparrudite, especially on the eastern side of the quarry, and the upper part is dense, non-porous, and appears to contain micrite between the allochems. The upper surface of this unit is highly irregular and phosphate- and glauconite-coated and has attached oysters, bryozoans, worm tubes and horizontal burrow structures. *Protoscutella conradi* is found locally within this unit.

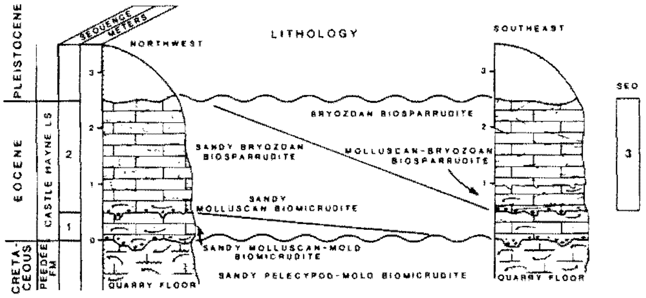

Figure 8. East Coast Limestone Quarry. The location of this quarry is approximately 4 km northwest of Maple Hill, Pender County on the north side of State Route 53. The section exposed in the quarry includes about 2 m of the Cretaceous Rocky Point Member of the Peedee Formation disconformably overlain by 2.5 m of the Castle Hayne Limestone and 1 m of surficial sand and clay.

Sequence 3. Sediments included in Sequence 3 were referred to the Comfort Member of the Castle Hayne Limestone by Ward and others (1978). This sequence is widespread in New Hanover, Pender, Onslow, Jones and southwestern Craven counties. Downdip, Sequence 3 is disconformable on sediments of Sequences 1 and 2, but updip it may overlie older Paleogene or Cretaceous units. Units attributed to Sequence 3 differ from those of Sequences 1 and 2 in their overall greater thickness and in their lower quartz sand content. Sequence 3 lithologies have a diverse and well preserved invertebrate fauna, and locally contain well preserved foraminiferal and calcareous nannofossil assemblages.

The Martin Marietta Castle Hayne quarrying operations, at which Baum and others (1978) designated their lectostratotype of the Castle Hayne Limestone, consists of three pits (Figure 4), each exposing different sets of lithologies. Pit 1a, the oldest and easternmost pit, contains the lectostratotype. Half of Pit 1 was diked and flooded in 1977, and the other half was accidentally flooded in January 1979. A second pit was developed about one-quarter mile to the west and was abandoned and flooded in 1982. Pit 3, the last pit to be worked by Martin Marietta, is to the southwest of pit 1. Pit 3 was abandoned and flooded in 1985. The description of Castle Hayne Sequence 3 deposits that follows is based on the section exposed in Pit 3. The section representing Sequence 3 is approximately 7 m thick, and disconformably overlies the upper part of the New Hanover Member of the Castle Hayne Limestone (Sequence 2). The lower 2 m is a calcarenite containing phosphatized and glauconitized pebbles at its base, and coarsening and grading upward into a 0.3 m zone of pebble- to cobble-size, glauconitized, micrite clasts (sponges?) and invertebrate fossils. This clast zone is considered to be equivalent to the rubble zone in Sequence 3 at the Ideal quarry as described below. The clast zone is conformably overlain by almost 4 m of glauconitic and phosphatic calcarenite that coarsens upward. A 1 m thick zone of dolomitic micrite occurs above the clast zone. The upper calcarenite is disconformably overlain by Sequence 4 deposits. The most diverse invertebrate assemblage occurs in the glauconitized clast zone, and is typified by the clypeasteroid echinoid *Periarchus lyelli*, the pectinid *Chlamys membranosa*, and the barnacle *Euscalpellum carolinensis*. Kier (1980) described a diverse echinoid fauna from sediments of this sequence exposed in Pit 1.

Sequence 3 deposits at the Martin Marietta Ideal quarry are similar to those of Pit 3 at Castle Hayne, and were described in detail by Harris and others (1986b). The nearly 9 m of sediments attributed to Sequence 3 are disconformable on Sequence 2 and include, from base to top, bryozoan biosparrudite, sponge biomicrite, sponge-bryozoan biomicrite, and sponge biolithite. A rubble zone consisting of glauconite-coated sponges in a matrix of dolomitic biomicrite occurs near the top of the sponge biomicrite and about 4 m above the base of the sequence. This rubble zone is a consistent marker horizon, and retains approximately the same elevation throughout the quarry. Invertebrates obtained from this section include *Periarchus lyelli*, *Chlamys deshayesii dennisoni*, and *C. cookei*. Worsley and Laws (1986) identified calcareous nannofossil assemblages indicative of zones NP 17 below, and NP 18 above the rubble zone of Sequence 3 at the Ideal quarry.

Sequence 3 is represented by the thickest and most extensive unit in the Rocky Point quarry. It is best developed in the eastern part where it attains a thickness of about 3 m, and is absent in the extreme northern and western parts of the quarry. The sequence includes two lithologies that disconformably overlie Sequence 2 sediments or the Rocky Point Member in different parts of the quarry. The lowermost lithology consists of soft, poorly consolidated, tan biomicrite (3a). The unit lacks bedding or lamination and attains a maximum thickness of 1.5 m. Conformably overlying this lithology is tan, well washed, highly cross-bedded, unconsolidated bryozoan biocalcarenite (3b). The sediment contains abundant, small, sand-size, well rounded grains of glauconite. The contact between 3a and 3b is highly burrowed with the overlying calcarenite filling the burrows. No phosphate or glauconite mineralization marks the contact. The basal part of the biomicrite (3a) has yielded calcareous nannofloras representative of Zones NP16 at its base and NP17 higher in the unit, and the barnacle *Arcoscalpellum* sp., cf. *A. jacksonense* Withers. A diverse echinoid fauna, including the clypeasteroid *Periarchus lyelli* (Conrad), occurs in 3b and is particularly abundant just above the contact with 3a. The pectinids *Chlamys cookei* and *C. membranosa* also occur in 3b.

Sequence 3 deposits are absent at the Fussell and Atlantic Limestone quarries in Duplin County. At the East Coast Limestone quarry to the east, Sequence 3 deposits consist of two distinct lithologies with a combined thickness of over 2 m. The basal part consists of about 0.5 m of coarse, bryozoan-molluscan biosparrudite. This lithology is overlain conformably by well washed, cross-bedded, bryozoan biosparrudite. The contact between the two lithologies displays shallow burrows. Sequence 3 thickens toward the south and southeast boundaries of the quarry, where it directly overlies the Cretaceous Rocky Point Member. The basal unit of Sequence 3 contains the echinoid *Protoscutella plana*, and the upper unit has yielded the barnacle *Arcoscalpellum jacksonense*.

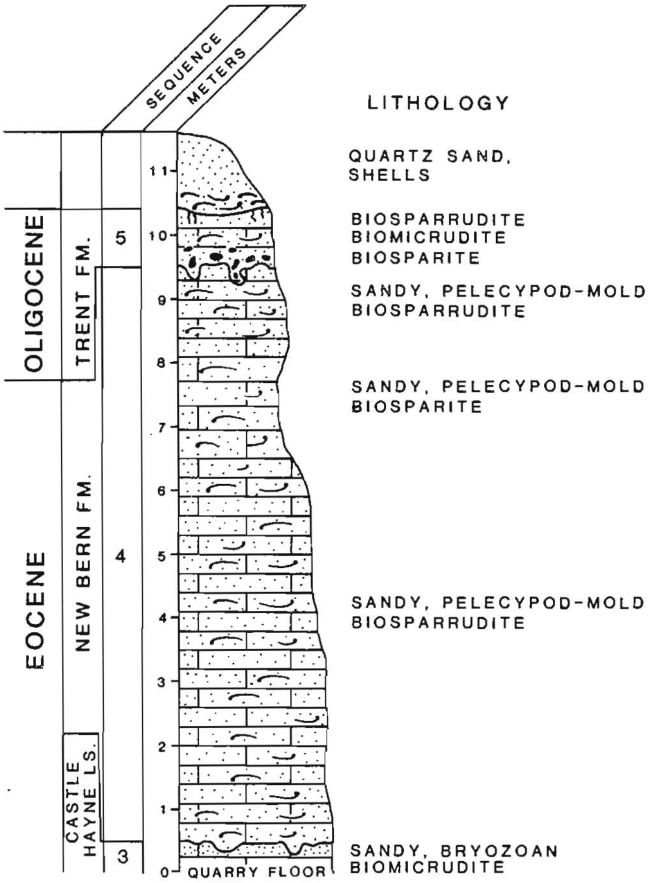

Figure 9. Martin Marietta New Bern Quarry. This quarry is located 1 km east of the intersection of State Route 55W and Route 1402 on the north side of Route 1402 in New Bern, Craven County. A composite section at this quarry includes 0.5 m of the Castle Hayne Limestone at the base of the quarry, disconformably overlain by 9 m of the Eocene New Bern Formation which, in turn, is disconformably overlain by 1 m of the Oligocene Trent Formation. The Trent Formation is overlain disconformably by 1 m of surficial sand and clay at the base of which are discontinuous lenses of the Pliocene Yorktown Formation.

One-half m of the uppermost part of Sequence 3 is exposed at the base of the Martin Marietta New Bern quarry (Figure 9). A grayish-white calcarenite on the quarry floor grades upward into a dark gray, calcareous, fine-grained sandstone containing abundant *Turritella* molds and casts, and very large, highly campanulate specimens of *Periarchus lyelli* (*pileussinensis* type). The upper surface of this sandstone is highly bored and corroded, and coated with glauconite and phosphate. The upper part of the section at the Lanier quarry at Maple Hill, Pender County is correlated to Sequence 3 based on calcareous nannofossil assemblages identified with Zone NP 17 by Worsley and Laws (1986). *Arcoscalpellum jacksonense* is represented also at the Lanier quarry.

Sequence 4. Known occurrences of Sequence 4 deposits are limited to the Martin Marietta Castle Hayne, Ideal and New Bern quarries. At the Castle Hayne quarry, Sequence 4 deposits consist of about 1 m of bryozoan-molluscan biosparrudite disconformably overlying the bored, solutioned, and phosphate- and glauconite-coated upper surface of Sequence 3. Large, campanulate specimens of *Periarchus lyelli* are common in this unit.

The section representing Sequence 4 at the Ideal quarry is approximately 4 m in thickness and includes 1 m of biomicrudite containing abundant invertebrates that grades upward into sponge-bearing biomicrite and bryozoan biomicrudite. Sequence 4 deposits disconformably overlie the phosphate- and glauconite-coated upper surface of Sequence 3. Large, campanulate *Periarchus lyelli* and the pectinid *Chlamys deshayesii dennisoni* are abundant in the basal biomicrudite.

Sequence 4 at the New Bern quarry is represented by the New Bern Formation, which consists of 6 m of sandy, pelecypod-mold biosparrudite at the base, overlain conformably by 1.5 m of sandy, pelecypod-mold biosparite, which in turn is overlain conformably by 1.5 m of sandy, pelecypod-mold biosparrudite. Large, very campanulate specimens of *Periarchus lyelli* and the pectinid *Chlamys* (*Aequipecten*) n. sp. are known from the New Bern Formation at this locality.

Oligocene - lowermost Miocene

Five sequences are recognized for Oligocene through lowermost Miocene (Rupelian through Aquitanian) rocks in the North Carolina Coastal Plain.

Sequence 5. This sequence is represented by the Trent Formation of Baum and others (1978) that crops out in a narrow, SW-NE trending belt in Onslow, Jones and Craven counties. This unit is equivalent to the lower River Bend Formation of Ward and others (1978). Baum and others (1978) described three lithologies in the Trent: a basal sandy, echinoid biosparite that grades upward into a sandy, pelecypod-mold biomicrudite that, in turn grades upward into a barnacle, pelecypod-mold biosparrudite. The archaeobalanid barnacle *Lophobalanus kellumi* is found throughout the Trent Formation. The pectinid *Chlamys trentensis* and another resembling *Pecten poulsoni* occur in the basal biosparite.

Sequences 6 through 8. These sequences are represented by the Belgrade and Silverdale formations, and the overlying *Crassostrea* channel deposits of Baum and others (1978), and include deposits laid down during several cycles of Supercycle TB1. With the present lack of subsurface control and the limited exposures in the area, all of the deposits are here referred to a single sequence.

The Belgrade Formation is equivalent to the upper River Bend Formation of Ward and others (1978). The Silverdale Formation and the *Crassostrea* channel deposits were included in the Belgrade Formation of Ward and others (1978). The only exposures of the Belgrade Formation are in the Martin Marietta Belgrade quarry, Onslow County (Figure 10) where 8 m of the formation are exposed above the quarry floor. The dominant lithology is sandy, pelecypod-mold biomicrudite with minor sand interbeds. A 1 m thick bed of quartz sand containing abundant shells of *Anomia* and barnacles occurs about 1.5 m above the base of the quarry floor. Test holes indicate that the Belgrade lithology of alternating biomicrudite and quartz sand persists for at least another 20 m below the quarry floor. Phosphatized, bored surfaces, on which large shells of *Crassostrea blanpeidi* are attached, are found both within and at the top of the Belgrade Formation.

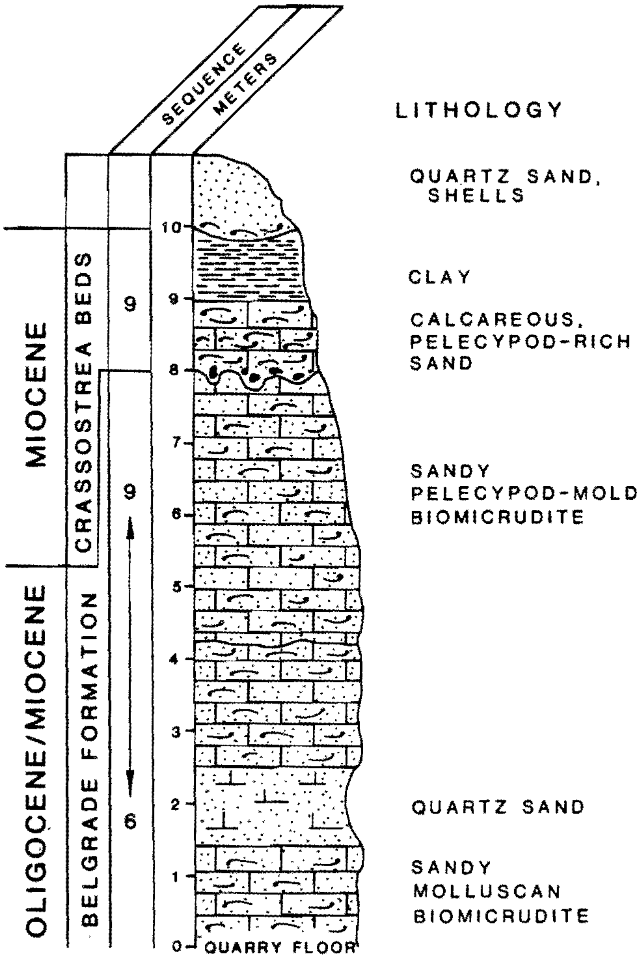

Figure 10. Martin Marietta Belgrade Quarry. This quarry is located on the southwest bank of the White Oak River just east of U. S. Highway 17 at Belgrade, Onslow County. The section exposed includes 8 m of the Oligocene Belgrade Formation of Baum and others (1978), disconformably overlain by up 2 m of *Crassostrea*-bearing channel deposits which are, in turn disconformably overlain by thin, discontinuous lenses of the Pliocene Yorktown Formation and surficial sand, clay and gravel.

Crassostrea-bearing channel deposits that grade laterally into fossiliferous clays and sand overlie the Belgrade Formation. The channel deposits are equivalent to the Pollocksville Member, and the sand and clay to the Haywood Landing Member of Ward and others (1978). Recent excavations at the Belgrade quarry have exposed over 2 m of the sand and clay lithology.

The Silverdale Formation is exposed in quarries near the town of Silverdale, Onslow County. The Jones quarry, located east of Silverdale at the intersection of County Roads 1434 and 1442, exposes 0.8 m of dense, sandy, pelecypod-mold biomicrudite which grades upward into 2.4 m of unconsolidated, sand containing a well preserved and diverse molluscan fauna.

BIOSTRATIGRAPHY

Invertebrate Biostratigraphy

Until recently, one of the major problems in correlation of facies within the Castle Hayne Limestone has been the lack of modern, systematic studies on specific invertebrate groups that are abundant and well preserved in these facies. The publication of Kier's (1980) monograph on the echinoids of Eocene deposits in North and South Carolina, the work of Cooper (1981; and unpublished data) on Carolina pectinids, and Zullo's (1984) biostratigraphic study on Gulf and Atlantic Coastal Plain cirripeds provide data for the construction of a biostratigraphic framework that is both reliable and remarkably sensitive. This framework not only serves in the correlation of regional lithofacies and sequences, but also permits rather accurate correlation of the exposed Carolina Eocene with the standard Gulf Coast section.

Cubitostrea zonation. The Eocene zonation developed by Stenzel (1949) on one lineage of *Cubitostrea* species has been used widely in the Gulf and southeastern Atlantic Coastal Plains (see Toulmin, 1977). This lineage is restricted to the Claibornian Stage, and includes *C. perplicata* (Dall) from the lower Claibornian, *C. lisbonensis* and *C. smithvillensis* (Harris) from the middle Claibornian, and *C. sellaeformis* from the upper Claibornian. A subspecies, *C. sellaeformis vermilla* (de Gregorio) is reported from the uppermost Claibornian Gosport Sand in Alabama (Palmer and Brann, 1965). *Cubitostrea perplicata* is known only from the upper Tallahata Formation in Alabama and Georgia, but occurs with *C. lisbonensis* in the Winona Formation (=lower middle Claibornian) of Mississippi (Dockery, 1980). *Cubitostrea lisbonensis* ranges from Texas to Alabama in the lower Lisbon formation and its equivalents, and has been tentatively identified from North and South Carolina. The Carolina specimens are small, with the largest one from North Carolina having a maximum length of 38 mm (Fussell quarry, Sequence 1). These North Carolina *Cubitostrea* are conspecific with specimens of similar size identified as *C. lisbonensis* by Powell and Baum (1982) from the lower part of their Santee Limestone (= Warley Hill marl) in South Carolina. *Cubitostrea smithvillensis* has been reported only from the Weches Formation of Texas and the lower Lisbon Formation of Alabama above the range zone of *C. lisbonensis*. *Cubitostrea sellaeformis* is the most widespread of the species, ranging from Mexico to Virginia in units equivalent to the upper Lisbon Formation in Alabama. This species is abundant in the upper Santee Limestone (restricted sense of Powell and Baum, 1982) in South Carolina, and occurs in Sequence 2 at the Rocky Point quarry. *Cubitostrea sellaeformis* was previously reported by Ward and others (1978) from the New Hanover Member at the Castle Hayne quarry (Sequence 2). These North Carolina occurrences provide correlation of one part of the Castle Hayne Limestone with the upper Santee Limestone of Powell and Baum (1982) in South Carolina and the upper Claibornian of the Gulf Coast.

Echinoid zonation. Kier (1980) recognized three informal echinoid assemblage zones within the Eocene carbonate sequence in the Carolinas. The oldest, or "early" zone, was characterized by the co-occurrence of *Santeelampas oviformis* and *Protoscutella mississippiensis*, and included the range zone of *P. conradi*. Kier's "middle" zone was characterized by *Linthia harmatuki* Kier, and included the range zone of *Protoscutella plana*. His "late" zone was characterized by abundant *Periarchus lyelli* and *Echinolampus appendiculata*.

Based on analysis of Kier's (1980) echinoid distribution data, recollection of sections in North and South Carolina, and Kier's (in litt., 1985) recent re-evaluation of *Protoscutella plana*, we propose an echinoid zonation based on the mutually exclusive range zones of the species of *Protoscutella* and *Periarchus*. As discussed by Kier (1980, p. 6-7), the species present in the Carolinas form an evolutionary sequence that is most readily discerned by changes in the position of the periproct. The oldest members of the sequence, *Protoscutella mississippiensis* and its subspecies, bear a marginal periproct. *Protoscutella mississippiensis* is known from the upper Tallahatta and overlying lower Lisbon formations of Alabama, the Winona Formation of Mississippi, and the Mount Selman Formation of Texas (Cooke, 1959; Dockery, 1980; Kier, 1980). With the exception of the upper Tallahatta Formation, these Gulf Coast units are included in the *Cubitostrea lisbonensis* Zone (middle Claibornian Stage). The upper Tallahata Formation represents the slightly older *C. perplicata* Zone (lower Claibornian Stage), although both *C. perplicata* and *C. lisbonensis* are found with *Protoscutella mississippiensis* at one locality in the Winona Formation of Mississippi (Dockery, 1980). In South Carolina *P. mississippiensis* occurs in the Warley Hill Formation at Wilson's Landing (Kier locality 38) and on the south bank of the Santee River, Berkeley County (Kier locality 43) with *Santeelampas oviformis*. In North Carolina this species (subspecies *rosehillensis* Kier) is found in Sequence 1 at the Fussell quarry and at Mosley's Creek (Kier locality 35) also with *Santeelampas oviformis*.

Protoscutella conradi is distinguished from *P. mississippiensis* by its inframarginal periproct, located between coronal plates 2-3 or 3-4, rather than 5-12 (Kier, 1980, p. 37). This species is known only from the Carolinas in units overlying *P. mississippiensis*-bearing beds. In South Carolina *P. conradi* occurs with *Eurhodia rugosa rugosa* (Ravanel) and, in most cases, *E. holmesi* (Twitchell) throughout the section at the Georgetown quarry (Kier locality 37), in the *Cubitostrea sellaeformis* beds of the Santee Limestone at the Berkeley quarry (Kier locality 45) and at three other minor localities in Berkeley County (Kier localities 40, 42, 47). In North Carolina *P. conradi* is found in Sequence 2 with *Cubitostrea sellaeformis* at the Rocky Point quarry, in the upper part of the section at the Atlantic Limestone quarry with *Santeelampas oviformis* and *Eurhodia rugosa ideali* Kier (Sequence 2), in Sequence 2 at the Fussell quarry with *Santeelampas oviformis* and *Eurhodia rugosa ideali*, at another Duplin County locality near Rose Hill (Kier locality 33), and at two localities near Maple Cypress, Craven County with *Eurhodia holmesi* (Kier localities 23, 24). The range zone of

Protoscutella conradi is equivalent to the *Cubitostrea sellaeformis* Zone (upper Claibornian) of the Gulf Coast.

Protoscutella plana is distinguished from *P. conradi* by its thicker test margin, lesser degree of indentation along the posterior margin and by the position of its periproct which is located about two-fifths the distance from the posterior margin to the peristome (Kier, 1980, p. 38). This species is known only from the Carolinas where its occurrences are quite limited. *Protoscutella plana* is restricted to the lowermost Cross Formation in South Carolina (Kier localities 45, 46, 48) and that part of the Castle Hayne Limestone exposed at the East Coast quarry in North Carolina (Kier locality 10). In South Carolina, *P. plana* is always found in beds overlying those bearing *P. conradi*.

Based on new material from Sequence 3b at the Rocky Point quarry, together with a re-evaluation of previous collections of *P. plana* and *Periarchus lyelli*, Kier (in litt., 1985) re-assigned specimens attributed to *Protoscutella plana* from the lower part of the Castle Hayne Limestone at the Castle Hayne quarry (Kier locality 34) to *Periarchus lyelli*, and specimens attributed to *P. lyelli* from the East Coast Limestone quarry (Kier locality 10) to *Protoscutella plana*. These re-identifications were based on the observation that the periproct in *Periarchus lyelli* migrates towards the peristome with increase in test size. Kier's (in litt., 1985) analyses of specimens from the Ideal and Comfort quarries in North Carolina are summarized in Table 1.

Table 1. Periproct position in *Periarchus lyelli* from North Carolina (Kier, in litt., 1985)

Specimen length(mm)	Number of specimens	% periprocts in plates 2a, 2b, 3a/3b	% periprocts in plates 2a-2b
Ideal Quarry (Kier locality 34)			
25-35	17	53	47
36-45	34	35	65
46-55	7	9	91
> 55	8	0	100
Comfort quarry (Kier locality 13)			
< 43	8	100	0
> 43	6	17	83

Thus, the position of the periproct as a means of identification for *P. lyelli* is only meaningful for large specimens. In these individuals the periproct is located approximately halfway between the posterior margin and the peristome on the suture between interambulacral coronal plates 2a and 2b. In smaller specimens, where the position of the periproct is similar to that of *Protoscutella plana*, the most distinguishing feature is the thinner margin.

Periarchus lyelli is widespread in North Carolina. At the Rocky Point quarry it occurs in Sequence 3b and is particularly abundant just above the contact with Sequence 3a. At the Castle Hayne and Ideal quarries, *P. lyelli* occurs throughout the Castle Hayne Limestone above the contact with the New Hanover Member. At the New Bern quarry it is found both in the upper part of the Castle Hayne Limestone and in the overlying New Bern Formation. Altogether, Kier (1980) lists twelve North Carolina localities for *P. lyelli*. In South Carolina *P. lyelli* is restricted to the upper part of the Cross Formation, and has been reported from the Giant (Kier locality 39) and Santee Portland Cement (Kier locality 46) quarries and at Pinopolis Dam, Berkeley County (Kier locality 44).

Elsewhere, *Periarchus lyelli* is widespread in units of early Jacksonian age, being reported from Georgia, Alabama, Mississippi, Louisiana and, questionably, from Texas. The majority of these records are from the Moodys Branch Formation and its equivalents. *Periarchus lyelli pileussinensis* (Ravanel), which Kier (1980) regards as synonymous with *P. lyelli*, has been reported from the Cocoa Sand in Mississippi (Dockery, 1980), and from the upper Moodys Branch Formation through the Pachuta Marl in Alabama (Toulmin, 1977). This subspecies has also been reported from Jacksonian units in Georgia (Toulmin, 1977) and the type specimen is presumably from the Cross Formation of South Carolina.

There are two reports of *P. lyelli* from the upper Claibornian upper Lisbon Formation and Gosport Sand in Alabama. Cooke (1959, p. 42, pl. 14, figs. 1-3) illustrated a typical specimen of *P. lyelli* from Gopher Hill on the Tombigbee River in Washington County. It is likely, however, that this specimen came from the Moodys Branch Formation rather than the Gosport Sand. The exposure at Gopher Hill (= Baker Hill) includes the upper Lisbon Formation with *Cubitostrea sellaeformis*, overlain by 5 to 6 m of Gosport Sand with a well-preserved molluscan fauna, which is in turn overlain by the lower part of the Moodys Branch Formation containing typical *Periarchus lyelli* (Toulmin, 1977). Clypeasteroids collected and described by Toulmin (1977, p. 306, pl. 51, figs. 3-5) from the Gosport Sand at Gopher Hill, as well as from the upper Lisbon Formation at three localities in Alabama and two on the east bank of the Chattahoochee River in Georgia differ substantially from typical *P. lyelli*. The figured specimen, from the upper Lisbon Formation in Coffee County, is over 60 mm in length, and the periproct is located less than two-fifths the distance from the posterior margin towards the peristome. Toulmin regarded these Claibornian echinoids as representative of a subspecies distinct from typical *P. lyelli*, noting differences in the position of the periproct and in the outline of the apical region of the test. The position of the periproct is suggestive of *Protoscutella* rather than *Periarchus*, and is similar to the position of the periproct in *Protoscutella plana*. Toulmin's subspecies differs from *P. plana*, however, in having a thin, rather than tumid margin.

In summary, *Periarchus lyelli* (including the subspecies *pileussinensis*) is restricted to Jacksonian strata in the Gulf Coastal Plain. Although Kier (1980) regarded *P. lyelli pileussinensis* as only a more campanulate version of typical *P. lyelli*, there is a definite trend in Gulf Coast and Georgia *Periarchus* toward increased campanulation upward through the lower Jacksonian section (Toulmin, 1977; Dockery 1980). A similar trend may be present in the Carolinas, for markedly campanulate *P. lyelli* appears to be restricted to the upper Cross Formation in South Carolina, and the New Bern Formation and presumed uppermost parts of the Castle Hayne Limestone in North Carolina. However, there is not sufficient control at present to verify this trend.

The range zones of Protoscutella-Periarchus species provide an excellent means for correlation of facies in the exposed Eocene sequences of the Carolina. In addition, the oldest and youngest species in this evolutionary lineage are also found in the eastern Gulf Coast, and allow accurate ties with the standard Gulf Coast section. Protoscutella mississippiensis is restricted to the upper lower and middle Claibornian (uppermost Cubitostrea perplicata and C. lisbonensis Zones). Protoscutella conradi, by virtue of its association with Cubitostrea sellaeformis in the Carolinas, is restricted to the upper Claibornian. The stratigraphic range of P. plana is difficult to determine because of its absence in the Gulf Coast section. Its position relative to the ranges of P. conradi and P. lyelli suggests that P. plana is latest Claibornian in age and occurs in strata equivalent to the Gosport Sand in Alabama. Associated fauna, however, suggests an early Jacksonian age and an equivalence with the lower Moodys Branch Formation of the Gulf Coast. Periarchus lyelli, in the broad sense, is restricted to Jacksonian strata in the Gulf and southeastern Atlantic Coastal Plains and is most abundant in units equivalent to the Moodys Branch Formation of Mississippi and Alabama.

Pectinid zonation. Cooper (1981; unpublished data) examined the systematics and stratigraphic distribution of species of the pectinid genus Chlamys Röding in the Carolinas. Pectinids, especially species of Chlamys, are abundant and well-preserved in the Carolina Eocene, have relatively short stratigraphic ranges, and at least a few species provide ties with the Gulf Coast section. There has been little of a comprehensive nature that has been done with the systematics and biostratigraphy of Carolina Eocene pectinids, and Cooper (in litt., 1981) has found numerous errors in species identification and stratigraphic occurrence in the earlier literature for the region. Basically, pectinid stratigraphic ranges mirror those determined for Cubitostrea and the echinoids. The oldest species, Chlamys (Chlamys) clarkeana, is found in association with Cubitostrea lisbonensis and Protoscutella mississippiensis. Chlamys clarkeana provides a tie with the eastern Gulf Coast section, being known from the upper Tallahatta Formation (Cubitostrea perplicata Zone) and the Lisbon Formation (C. lisbonensis, ?C. sellaeformis zones) in Alabama (Toulmin, 1977), and the Winona Formation (C. lisbonensis Zone) in Mississippi (Dockery, 1980).

The pectinid fauna associated with Cubitostrea sellaeformis and Protoscutella conradi in the Carolinas is depauperate and poorly understood at present. Chlamys clarkeana, however, does appear to co-occur with Protoscutella conradi in North and South Carolina.

The fauna associated with Protoscutella plana and Periarchus lyelli is more diverse, however, with three species in common between North and South Carolina, and at least one that provides ties with the Gulf Coast Jacksonian. The Chlamys (Chlamys) membranosa of Kellum (1926) is the most abundant pectinid in the upper Castle Hayne Limestone in North Carolina and is present in the Cross Formation of South Carolina. According to Muriel Hunter (in litt., 1981), this is not the same species as the type of Chlamys membranosa (Morton), which is apparently from the Santee Limestone. Cooper (in litt., 1981), however, has identified Kellum's C. membranosa from the Protoscutella conradi Zone at the Georgetown quarry in South Carolina and the P. plana Zone at the East Coast Limestone quarry (Sequence 3) in North Carolina. These records indicate that Kellum's Chlamys membranosa has a relatively long stratigraphic range. Ward and others (1978) suggested that C. membranosa was synonymous with C. wahtubbeana from the Claibornian of the Gulf Coast, but Cooper (1981, in litt.) has shown that the two species differ significantly in radial ornament, valve thickness, and valve convexity. To date, Kellum's C. membranosa has not been collected outside of the Carolinas.

Chlamys (Aequipecten) cookei and Chlamys (Aequipecten) n. sp. appear to be restricted to the Periarchus lyelli Zone both in North and South Carolina. Toulmin (1977, p. 354) indicated that Chlamys cookei occurs in the upper Jacksonian of the Gulf Coast, but did not provide locality data. Chlamys (Aequipecten) n. sp. has been confused with C. cawcawensis (Harris) from the Santee Limestone in South Carolina. The new species, restricted stratigraphically to the Cross Formation in South Carolina and the upper Castle Hayne Limestone and New Bern Formation in North Carolina has smooth rather than lamellate ribs, and relatively inconspicuous rather than well developed intercostae (Cooper in litt., 1981). According to Cooper, the specimens attributed to C. cawcawensis by Toulmin (1977, pl. 54, figs. 12-13) from the lower Jacksonian Moodys Branch Formation in Alabama probably represent Chlamys (Aequipecten) n. sp.

Chlamys (Aequipecten) deshayesii dennisoni is abundant in the upper part of the section at the Ideal quarry in North Carolina in association with Periarchus lyelli (upper Sequence 3, lower Sequence 4). This species is otherwise known from the Moodys Branch Formation in Alabama at the same localities that have yielded specimens here attributed to Chlamys (Aequipecten) n. sp. (Toulmin, 1977, p. 314).

Although the study of Carolina pectinids is still preliminary, available data support the stratigraphic conclusions derived from analyses of the distribution of Cubitostrea and echinoid species. Chlamys clarkeana occurs in middle Claibornian units containing Cubitostrea lisbonensis and Protoscutella mississippiensis, and appears to be present in upper Claibornian units characterized by Cubitostrea sellaeformis and Protoscutella conradi. Chlamys cookei, C. deshayesii dennisoni, and Chlamys (Aequipecten) n. sp. are restricted to units characterized by Periarchus lyelli.

Oligocene pectinids are not as numerous. Chlamys trentensis (Harris) and another species resembling Pecten poulsoni Morton is restricted to the Trent Formation, and a new species related to C. trentensis is known from the Belgrade and Silverdale formations.

Cirriped zonation. Zullo (1984) recognized two Eocene cirriped assemblage zones in the southeastern Atlantic and Gulf Coastal Plains. The

Arcoscalpellum subquadratum Zone is equivalent to the *Cubitostrea sellaeformis* Zone of the upper Claibornian, and has been recognized from the eastern Gulf Coastal Plain north to South Carolina. This zone is represented in South Carolina by *Arcoscalpellum subquadratum* and *Aporolepas americana* (Withers) in the Santee Limestone (*C. sellaeformis* beds). To date, no cirripeds diagnostic of this zone have been found in North Carolina.

The *Arcoscalpellum jacksonense* Zone ranges through the Jacksonian of the Gulf Coast and has been recognized both in North and South Carolina. *Arcoscalpellum jacksonense* is known from the Dry Branch and upper Cross formations in South Carolina, and from the Castle Hayne Limestone at the East Coast Limestone quarry (upper Sequence 3), the Lanier quarry (Sequence 3), the Castle Hayne quarry (Sequence 3 and 4), and the North Carolina Lime Excavating Company quarry west of Comfort, Jones County in association with *Periarchus lyelli*. *Euscalpellum carolinensis* is found in association with *Protoscutella plana* in the lowermost Cross Formation in South Carolina, and with *Periarchus lyelli* at the North Carolina Lime Excavating Company quarry in Jones County, North Carolina. *Aporolepas howei* is abundant in the Griffins Landing Member of the Dry Branch Formation in the Savannah River region. *Euscalpellum chamberlaini*, known from the Weches Formation in Texas and the Winona Formation in Mississippi, occurs in the Castle Hayne Limestone (Sequence 1) at the Martin Marietta Castle Hayne quarry.

Two cirriped zones, *Solidobalanus* B and *Solidobalanus* C, were proposed by Zullo (1979) for the Oligocene and lower Miocene of North Carolina. Since that time, these barnacles were described, as *Lophobalanus kellumi* (Zullo and Baum) and *Lophobalanus baumi* Zullo, respectively. *Lophobalanus kellumi* is restricted to the Trent Formation where it is found in association of *Chlamys trentensis*. *Lophobalanus baumi* occurs in the Belgrade and Silverdale formations and the overlying *Crassostrea* channel deposits in North Carolina and in the barnacle buhrstone found in channels cut into Eocene units of the Savannah River region in South Carolina. A species perhaps conspecific with, but certainly related to *L. baumi* is known from the Byram Formation in Mississippi. In North Carolina, *L. baumi* is found in association with the earliest occurrences of the balanid genus *Concavus*, and the ostreid *Crassostrea blanpiedi*.

Calcareous Nannofossil Biostratigraphy

Worsley and Laws (1986) described calcareous nannofossil assemblages from the Castle Hayne Limestone indicative of Martini Zones NP15, NP16, NP17 and NP18. Based on this study, Sequence 1 strata were assigned to Zone NP15, Sequence 2 to Zone NP16, and Sequence 3 to Zones NP16, NP17 and NP18. No diagnostic assemblages were obtained from Sequence 4 units.

Berggren and Aubry (1984), based on samples from the Martin Marietta Castle Hayne quarry, concluded that nannofloral evidence indicated assignment of the Castle Hayne Limestone to Zones NP16 - lower NP17. Hazel and others (1984a) reached the same conclusion using samples from the same quarry. As discussed in the section on lithostratigraphy, the three pits at the Castle Hayne quarry exposed different parts of the Castle Hayne section. The samples analyzed by Berggren and Aubry (1984) and by Hazel and others (1984a) were from Sequence 3 in pit 2 (i.e., above the New Hanover Member). Pit 1 was flooded before the samples were taken, and pit 3 was still to be dug. Only the part of Sequence 3 below the glauconitic clast zone was exposed in pit 2. As discussed above, the lower part of Sequence 3 contains a nannoflora indicative of Zones NP16 and NP17. Thus, the findings of Worsley and Laws, Berggren and Aubry, and Hazel and others are in agreement for the specific part of the Eocene section exposed in pit 2.

Jones (1983, p. 9), citing Berggren and Aubry, noted the presence of nannofossils indicative of Zone NP23 in "rocks overlying the middle Eocene section in Craven County." We presume these rocks represent the Trent Formation (Sequence 5). Harris and others (1986a) attributed nannofossils from the Silverdale Formation (lower Sequence 6) to Zone NP24, whereas those from the overlying *Crassostrea* channel and equivalent deposits (upper Sequence 10) were assigned to Zones NP25 - NN1.

Foraminiferal Biostratigraphy

The planktonic Foraminifera of the Castle Hayne Limestone have been discussed by Jones (1983), Berggren and Aubry (1984) and Hazel and others (1984a). In addition, both Paul F. Huddlestun and Garry D. Jones have provided us with data on additional samples from Sequences 2 and 3. Jones (1983) studied samples from the majority of the quarries discussed in this study, and assigned the Castle Hayne Limestone to Blow's (1969) Zones P12 and P13. Berggren and Aubry (1984) further refined Jones's data and limited the age assignment to Zone P13. Hazel and others (1984a) concluded that the Castle Hayne Limestone (i.e., the lower part of Sequence 3 in pit 2 at the Castle Hayne quarry) indicated Zones P12 and P13. Huddlestun (personal communications, 1983-1986) examined samples from Sequence 3, including a sample from the Martin Marietta Ideal quarry that yielded the calcareous nannofossil *Chiasmolithus oamaruensis* (Zone NP18), all of which he assigned to Zone P13. Jones (personal communication, 1986), who examined splits of these samples, arrived at the same conclusion.

Thus, age determinations from planktonic Foraminifera disagree significantly with those obtained from calcareous nannofossils and invertebrates. We cannot determine the reason for such widely disparate age determinations, but because calcareous nannofossil and invertebrate biostratigraphies are internally consistent and are supported by coastal onlap stratigraphy, we suspect that foraminiferal biostratigraphy for this part of the Paleogene is in need of restudy.

Jones (1983, p. 9) recognized planktonic foraminifers indicative of Blow's Zones P20 and P21 in "the oldest rocks [Trent Formation?] overlying the middle Eocene section." Zarra (1983; personal communication, 1983) assigned the Trent Formation (Sequence 5) to the *Globigerina ampliapertura* Zone (= Zone P19/20), except for one locality where he found a fauna indicative of the *G. ciperoensis* Zone (= Zone P22). Zarra identified the *G. ampliapertura* and *G. ciperoensis* Zones in the

Belgrade Formation, and the upper *Globorotalia kugleri* Zone (= upper Zone N4) in the Silverdale Formation and in the *Crassostrea* channel deposits and equivalents.

ASSEMBLAGE ZONES

Based on the faunistic data presented above, six assemblage zones can be recognized within the exposed Eocene through lowermost Miocene carbonate sequences in the Carolinas. These are, from oldest to youngest, the:

1. *Protoscutella mississippiensis Assemblage Zone*. This, the oldest recognized zone, is characterized by the range zone of *P. mississippiensis* and its subspecies, and includes the range zones of *Cubitostrea lisbonensis* and *Euscalpellum chamberlaini*. The earliest occurrences of *Santeelampas oviformis* and *Chlamys clarkeana* are in this zone. The *P. mississipiensis* Zone is recognized in the *Cubitostrea lisbonensis* beds of the Santee Limestone (=Warley Hill Marl) in South Carolina and Sequence 1 equivalents in North Carolina. Associated calcareous nannofossils are indicative of Zone NP15.

2. *Protoscutella conradi Assemblage Zone*. This zone, characterized by the range zone of *P. conradi*, includes the range zones of *Cubitostrea sellaeformis*, *Arcoscalpellum subquadratum* and *Aporolepas americana*. The earliest occurrence of *Chlamys membranosa* and the last occurrences of *Chlamys clarkeana* and *Santeelampas oviformis* are in this zone. The *P. conradi* Zone is recognized in the *Cubitostrea sellaeformis* beds of the Santee Limestone and McBean Formation of South Carolina and Sequence 2 equivalents in North Carolina. Associated calcareous nannofossils are indicative of Zone NP16.

3. *Protoscutella plana Assemblage Zone*. This zone is characterized by the range zone of *P. plana*, and includes the earliest occurrence of *Euscalpellum carolinensis*. This zone is quite localized, being recognized in the lower Cross Formation in South Carolina and at one locality in the Castle Hayne Limestone in North Carolina. Although calcareous nannofossils have not been analyzed from strata definitely referable to this zone, ancillary data suggest correlation of the *P. plana* Zone with Martini Zone NP16.

4. *Periarchus lyelli Assemblage Zone*. This zone, characterized by the range zone of *P. lyelli*, includes the range zones of *Arcoscalpellum jacksonense*, *Aporolepas howei*, *Chlamys deshayesii dennisoni*, *C. cookei*, *Chlamys (Aequipecten)* n. sp., the oyster *Pycnodonte trigonalis*, and the last occurrence of *Euscalpellum carolinensis*. The *P. lyelli* Zone is recognized in the upper Cross Formation, the Dry Branch Formation, and the Tobacco Road Sand in South Carolina, and the Castle Hayne Limestone (upper Sequence 3, Sequence 4) and the New Bern Formation (Sequence 4) in North Carolina. Calcareous nannofossils assemblages associated with this zone are indicative of Zones NP17, NP18 and, perhaps, NP19/20.

5. *Lophobalanus kellumi Assemblage Zone*. This zone is characterized by the range zone of the archaeobalanid barnacle *L. kellumi*, which is abundant throughout the Trent Formation in North Carolina, but is unknown elsewhere. Representative invertebrates include the pectinids *Chlamys trentensis* and *Pecten* aff. *P. poulsoni*. The extensive molluscan fauna of the Trent Formation, represented by molds and casts, bears considerable affinity to that of lower Vicksburgian strata in Mississippi (J. G. Carter, personal communication, 1986). This zone is tentatively considered to include calcareous nannofossil Zones NP21 and NP22 and planktonic foraminiferal Zone P19/20.

6. *Lophobalanus baumi Assemblage Zone*. This zone includes the range zones of *L. baumi*, the balanid barnacles *Concavus belgradensis* Zullo and *C. crassostricola* Zullo, the ostreid *Crassostrea blanpiedi*, and the pectinid *Chlamys* n. sp., aff. *C. trentensis*. This zone is recognized in the Belgrade and Silverdale formations, and the *Crassostrea* channel deposits of Baum and others (1978). This zone includes calcareous nannofossil Zones NP24, NP25 and NN1, and planktonic foraminiferal Zones P22 and N4.

CORRELATION TO GLOBAL CYCLES

The following discussion of the correlation of North Carolina Eocene through lower Miocene units is summarized in Figures 11 through 13. Reference to Depositional Sequence Tracts in the discussion follows Haq and others (1987).

Sequence 0 has not yielded diagnostic fossils for correlation. Sequences 0 and 1, because of their thinness and limited distribution, are difficult to place within systems tracts. As Sequences 0 and 1 do not display the parameters which allow assignment to specific parts of a depositional sequence, they are not assigned to individual tracts.

Sequence 1, characterized by the *Protoscutella mississippiensis* Zone, is correlated with the lower Santee Limestone (*C. lisbonensis* beds of Powell and Baum, 1982) of South Carolina, and the lower Lisbon Formation of Alabama. Powell and Baum (1982) and Baum (1986) assigned the lower Lisbon to Cycle TE2.2, which is equivalent to Cycle TA3.4 of Haq and others (1987). Calcareous nannofossil biostratigraphy suggests that the age of Cycle TA3.4 is older than that shown by Haq and others. Hazel and others (1984b) reported *Lophodolithus mochlophorus* and *Chiasmolithus gigas* from the lower Lisbon of Little Stave Creek, Alabama which are indicative of Okada and Buckry's (1980) Zone CP13b (≈ middle of Martini Zone NP15). The same conclusion was drawn by Worsley and Laws (1986) for the *P. mississippiensis* Zone at the Fussell quarry, based on the co-occurrence of *Chiasmolithus gigas* and *C. staurion*. Currently, Cycle TA3.4, to which both the lower Santee Limestone and the lower Lisbon Formation are assigned, is correlated with Martini Zone NP16 (= Okada and Buckry Zone CP14a).

Sequence 2, although thicker and better developed than Sequence 1, is also difficult to assign to specific systems tracts. At the Fussell quarry, Sequence 2 is interpreted to represent transgressive deposits because of the decrease in grain size and the increase in micrite upward in the section. The overall aspect of Sequence 2 at the Rocky Point quarry also appears to be transgressive. Sequence 2, characterized by the *Protoscutella conradi* Zone, is correlated with the upper Santee Limestone (*C. sellaeformis* beds of Powell and Baum, 1982) and McBean Formation of South Carolina, and the upper Lisbon Formation of

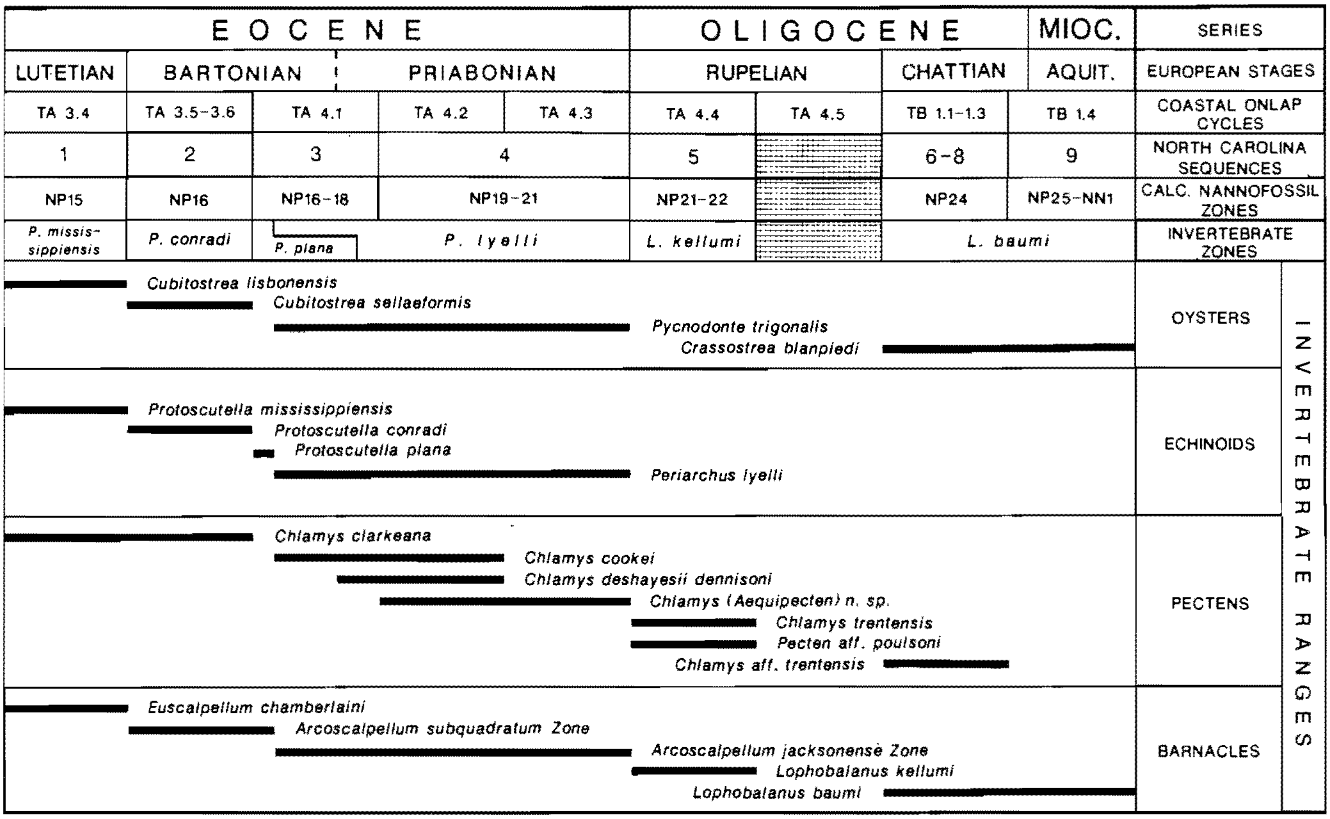

Figure 11. Ranges of age diagnostic Eocene through lower Miocene invertebrate species in North Carolina. Correlation of European stages and Coastal Onlap Cycles after Haq and others (1987). Calcareous nannofossil zonation after Worsley and Laws (1986) and Harris and others (1986a).

Alabama. Powell and Baum (1982) and Baum (1986) assigned the upper Santee Limestone and the upper Lisbon Formation to Cycle TE2.3, which is equivalent to Cycles TA3.5 and TA3.6 of Haq and others (1987). As calcareous nannofossils of Sequence 2 units in North and South Carolina are indicative of Martini Zone NP16, and as no evidence of a sequence break is seen in Sequence 2, we assign Sequence 2 to Cycle TA3.5.

Sequence 3 represents the most complete Eocene depositional sequence exposed in the Coastal Plain of North Carolina. Lowstand deposits of the sequence occur at the East Coast quarry (Protoscutella plana beds), and transgressive and highstand deposits are well developed at the Rocky Point and Ideal quarries. The "rubble zone" is interpreted as the condensed interval or surface of maximum flooding that separates the lower transgressive deposits from the overlying highstand deposits. Sequence 3, characterized by the Protoscutella plana Zone at the base, and the Periarchus lyelli Zone above, is correlated with the Cross Formation in South Carolina, the Clinchfield and Dry Branch formations in South Carolina and Georgia, and the Gosport Sand and lower Moodys Branch Formation in Alabama. Powell and Baum (1982) and Baum (1986) assigned the Alabama section from the Gosport Sand through the lower Moodys Branch Formation to Cycle TE3.1, and the upper Moodys Branch and North Twistwood Creek Clay to Cycle TE3.2, with a Type 2 unconformity between the lower and upper Moodys Branch Formation as the boundary between the two cycles. Powell and Baum (1982) placed the Cross in both cycles, noting that they were unable to differentiate these two cycles in the Cross Formation. We are restricting the Cross Formation and Sequence 3 to Cycle TE3.1 based on the presence of non-campanulate Periarchus lyelli, and calcareous nannofossils indicative of Martini Zones NP16 through NP18. Cycle TE3.1 is equivalent to Cycle TA4.1 of Haq and others (1987).

Sequence 4 represents shelf margin deposits that reflect a downward shift in coastal onlap to a position above the shelf break. The New Bern Formation may represent highstand deposits of this sequence, but otherwise neither transgressive nor highstand deposits are identified. Sequence 4, characterized by the upper part of the Periarchus lyelli Zone, is correlated with the Harleyville and Parkers Ferry members of the Cooper Formation in South Carolina, the Tobacco Road Sand in South Carolina and Georgia, and (in ascending order) the upper Moodys Branch Formation, North Twistwood Creek Clay, Cocoa Sand, Pachuta Marl, Shubuta Clay, and Red Bluff Formation/Bumpnose Limestone in Alabama. Baum (1986) assigned the upper Moodys and North Twistwood Creek to Cycle TE3.2, and the Cocoa Sand through the Red Bluff to Cycle TE3.3. These cycles are equivalent to Cycles TA4.2 and TA4.3, respectively, of Haq and others (1987). Based on biostratigraphic evidence, Sequence 4 appears to span Cycles TA4.2 and TA4.3. A disconformable surface within Sequence 4 deposits at the Ideal and New Bern quarries may represent the 37 Ma

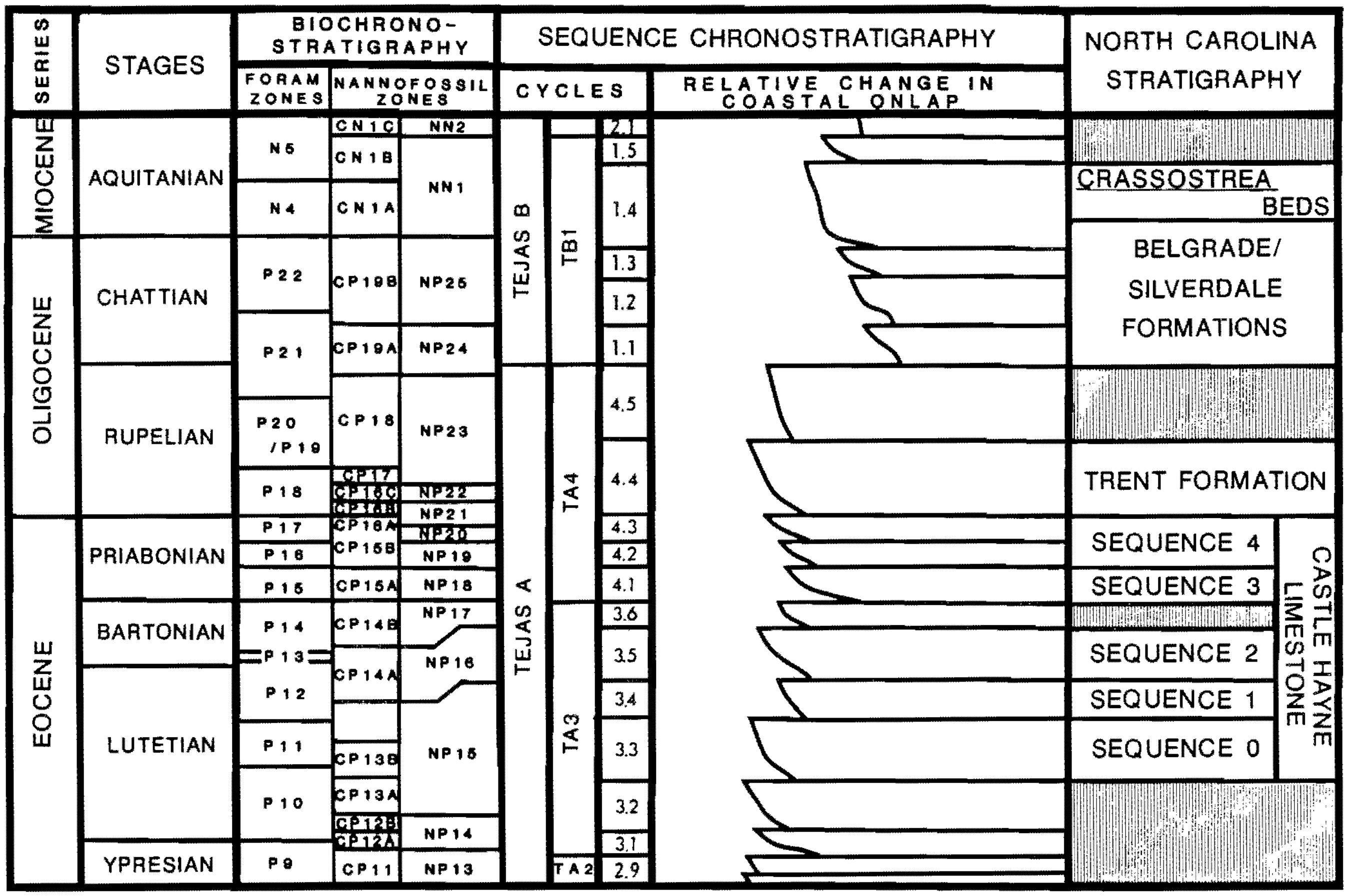

Figure 12. Suggested correlation of North Carolina sequences with the Global Coastal Onlap Chart of Haq and others (1987). The ranges of nannofossil zones NP 15, 16 and 17 have been adjusted based on the distribution of nannofossil assemblages discussed in this study.

unconformity between Cycles TA4.2 and TA4.3, but our limited knowledge of the stratigraphy and distribution of Sequence 4 precludes its subdivision at this time. Calcareous nannofossils indicative of Martini Zones NP18 and NP19/20 characterize Sequence 4 correlates in South Carolina and Alabama.

Sequence 5 illustrates a complete Oligocene depositional sequence. The basal sandy biosparite is considered to represent transgressive deposits, and the overlying pelecypod-mold biomicrudite and barnacle, pelecypod--mold biosparrudite represent highstand deposits of the sequence. The condensed interval separating the two systems tracts is not represented by the common parameters because of the updip outcrop of the unit. Sequence 5, represented by the Trent Formation and characterized by the Lophobalanus kellumi Zone, is correlated with the Mint Spring Formation and Marianna Limestone of the eastern Gulf Coastal Plain. This correlation is based on the molluscan fauna, planktonic Foraminifera representative of Blow's Zones P19/20, and calcareous nannoplankton indicative of Martin Zones NP21 and NP22. Strata of this age have not been identified in South Carolina or adjacent parts of Georgia. Baum (1986) assigned the Mint Spring Formation and Marianna Limestone to Cycle TO1.1, which is equivalent to Cycle TA4.4 of Haq and others (1987).

Sequences 6 through 8 represent transgressive deposition during three cycles, and are restricted to an incised valley that formed during the 30 Ma sea level fall. Phosphate-coated surfaces at the Belgrade quarry may represent condensed intervals or sequence boundaries. However, no highstand deposits have been identified in outcrop. Highstand deposits of these sequences probably occur offshore in Onslow Bay. Sequences 6 through 8, represented by the Silverdale Formation and lower Belgrade Formation, and characterized by the lower part of the Lophobalanus baumi Zone, are correlated with the Ashley Member of the Cooper Formation in South Carolina, the barnacle buhrstone in the Savannah River region of South Carolina and Georgia, and the Chickasawhay Formation of Alabama and Mississippi. Correlation is based on planktonic Foraminifera of Zone P22 and calcareous nannofossils of Zone NP24. Baum (1986) assigned the Waynesboro Sand and overlying Chickasawhay Formation to Cycle TO2.1, which is equivalent to Cycles TB1.1 through TB1.3 of Haq and others (1987).

Sequence 9 preserves transgressive deposits of the upper part of the Belgrade Formation and highstand deposits of the Crassostrea beds. Transgressive deposits of Sequence 9 are restricted to the same incised valley that limits the distribution of Sequences 6 through 8. Crassostrea channel and related clay and sand are separated

EPOCH	STAGE	CYCLE	ALABAMA		GEORGIA/ S. CAROLINA	SOUTH CAROLINA		NORTH CAROLINA	
MIOC.	AQUIT.	TB 1.4		PAYNES HAMMOCK FM. NP25-NN1	Upland unit	Upland unit(?)			SEQUENCE 9 Crassostrea channels upper Belgrade Fm. NP25-NN1, N4
OLIGOCENE	CHATTIAN	TB 1.3 1.2 1.1		CHICKASAWHAY FM. NP24	barnacle buhrstone	COOPER FORMATION	Ashley Mbr. NP24, P21		SEQUENCES 6-8 Silverdale Fm./ lower Belgrade Fm. NP24, P22
	RUPELIAN	TA 4.5		BUCATUNNA FM. GLENDON LS./BYRAM FM.	(Oligocene limestone downdip)		▥		▥
		TA 4.4		MARIANNA LIMESTONE MINT SPRING FM. NP21-22			▥		SEQUENCE 5 Trent Fm. NP21-22, P19/20
EOCENE	PRIABONIAN	TA 4.3		FOREST HILL/RED BLUFF/BUMPNOSE NP19-21 COCOA/PACHUTA/SHUBUTA	TOBACCO ROAD SAND		Parkers Ferry Mbr. NP19/20	CASTLE HAYNE LIMESTONE	NEW BERN FM.
		TA 4.2		NORTH TWISTWOOD CREEK CLAY NP19/20, P16 upper MOODYS BRANCH FM.	DRY BRANCH FM. P16		Harleyville Mbr. NP19/20		SEQUENCE 4
	BARTONIAN	TA 4.1		lower MOODYS BRANCH FM. NP17 GOSPORT SAND	CLINCHFIELD FM.	CROSS FM. NP17-?18			SEQUENCE 3 NP16-18, P13
		TA 3.6 3.5	LISBON FM.	upper NP16-17, P13	MCBEAN FM. NP16	SANTEE LS.	upper NP16		SEQUENCE 2 NP16
	LUTETIAN	TA 3.4		lower NP15	unnamed unit		lower		SEQUENCE 1 NP15

Figure 13. Suggested correlation of global cycles with Eocene through lower Miocene stratigraphic units in the eastern Gulf and southeastern Atlantic Coastal Plains. Alabama sequence stratigraphy after Baum (1986). Georgia stratigraphy adapted from Huddlestun and Hetrick (1986). South Carolina stratigraphy modified from Powell and Baum (1982) and Ward and others (1979).

from the underlying upper Belgrade by a phosphate-coated surface that is interpreted as the condensed interval. These highstand deposits are not confined to the incised valley. Sequence 9, represented by the upper Belgrade Formation and the Crassostrea channel and related sand and clay deposits, and characterized by the upper Lophobalanus baumi Zone, is correlated with the informal Upland unit of the Savannah River region in South Carolina and Georgia, and the Paynes Hammock Formation in Alabama and Mississippi. Correlation is based on planktonic Foraminifera of Zone N4, and calcareous nannofossils of Zones NP25 and NN1. Baum (1986) assigned the Paynes Hammock Formation to Cycle TM1.1, which is equivalent to Cycle TB1.4 of Haq and others (1987).

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