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INTEGRATED LATE EOCENE-OLIGOCENE STRATIGRAPHY OF THE ALABAMA COASTAL PLAIN: CORRELATION OF HIATUSES AND STRATAL SURFACES TO GLACIOEUSTATIC LOWERINGS

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Abstract. We integrated strontium and oxygen isotopic, biostratigraphic, and magnetostratigraphic studies of two upper Eocene-Oligocene boreholes drilled near Bay Minette and St. Stephens Quarry (SSQ), Alabama. Continuous coring provided fresh, unweathered material for magnetostratigraphic studies, minimizing problems reported from nearby outcrops. Difficulties with each technique were encountered because of diagenesis, absence of marker fossils, and the presence of unconformities; however, by integrating results from isotopic stratigraphy, biostratigraphy, and magnetostratigraphy, we correlated these relatively shallow-water deposits to the geomagnetic polarity time scale (GPTS). At the SSQ borehole, the upper Eocene to lower Oligocene section is apparently complete within our stratigraphic resolution (0.2-0.5 m.y.), allowing us to estimate the ages of several stratal surfaces. Late Eocene Sr isotope age estimates are as expected at the SSQ borehole, but Oligocene ages are ~1 m.y. older than expected due to diagenesis. At the Bay Minette borehole, a latest Eocene-earliest Oligocene and a late early Oligocene hiatus were detected. We correlate these two hiatuses and stratal surfaces at SSQ with global $\delta^{18}O$ increases inferred to represent glacioeustatic lowerings and with evidence for hiatuses on other continental margins: (1) a distinct disconformity at the base of the Chickasawhay Limestone at both boreholes and a hiatus at Bay Minette correlates with a global $\delta^{18}O$ increase; we revise the age of this surface (equivalent to the TB1.1 sequence boundary) making it ~2 m.y. older than previously reported; and (2) a surface at the top of the Shubuta Member (lowermost Oligocene) has been interpreted both as a condensed section and a disconformity; this surface at SSQ and a hiatus at Bay Minette correlate with a sharp global $\delta^{18}O$ increase and with hiatuses on the New Jersey and Irish margins. The timing of the hiatuses and stratal surfaces correlates with the inflection of the $\delta^{18}O$ increases and not with the maximum values, supporting models that indicate that unconformities form during the maximum rates of sea level fall.

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INTRODUCTION

The Gulf Coastal Plain has served as the marine reference section for the Paleogene of North America since the middle of the last century, and stage names based on key outcrop sections are still in use. However, global correlations of these strata have been debatable. Materials suitable for radiometric dating (e.g., volcanic ash) are rare in the Gulf Coast Paleogene. Microfossils, which provide most interregional correlations, may be locally diachronous because of abrupt facies changes associated with relative sea level variations in these largely shallow-water (paleodepths < 200 m) sections [Miller and Kent, 1987; Loutit et al., 1988]. Magnetostratigraphy is often limited by weak, unstable magnetizations and complicated by diagenetic overprinting in coastal plain outcrops [Ellwood et al., 1986]. Previous study of the New Jersey coastal plain [Miller et al., 1990] circumvented these limitations by (1) studying cores to minimize weathering that inhibits magnetostratigraphic and isotope studies of outcrops; and (2) integrating magnetostratigraphy, biostratigraphy, and isotopic stratigraphy.

Shallow-water sections provide the most direct monitor of sea level fluctuations. However, these sections are complicated not only by the correlation problems noted above but also by erosional/nondepositional hiatuses [Miller and Kent, 1987]. Eocene to Oligocene deposition on the Gulf Coastal Plain was punctuated by hiatuses. However, the hiatuses were short and the preserved sections are complete enough to estimate the timing of the erosional events.

Recent advances in sequence stratigraphy [e.g., Haq et al., 1987; Posamentier et al., 1988] have intensified interest in the Gulf Coast Paleogene record and its relationships with unconformities (sequence boundaries) and sea level changes

[Baum and Vail, 1988; Pasley and Hazel, 1990; Mancini and Tew, 1991; Tew, 1992]. Unconformities develop on passive continental margins during global sea level (eustatic) lowerings, although they are also caused by processes other than global sea level change (e.g., local tectonics). The development of unconformities is relatively insensitive to changes in sediment supply [Christie-Blick et al., 1990]. To differentiate regional tectonic from global causes of sea level change, the age equivalency among unconformities on different continental margins must be established.

Unconformities (sequence boundaries) may be recognized by two different methods. The first is based on recognition of their physical expressions in seismic profiles or in stratigraphic sections. The disadvantage of the sequence stratigraphic approach in boreholes is that it can be difficult to differentiate among various stratal surfaces (e.g., maximum flooding surfaces (MFS) which are downlap surfaces), disconformities, transgressive surfaces [Van Wagoner et al., 1987]) in such a limited view of the section. However, J.S. Baum et al. (Stable isotopic evidence for relative and eustatic sea level changes in Eocene to Oligocene carbonates, Baldwin County, Alabama, submitted to *Geol. Soc. Am. Bull.*, 1992) (hereinafter referred to as Baum et al., submitted manuscript, 1992) have successfully applied sequence stratigraphy to the boreholes considered here.

Unconformities also may be recognized by their associated hiatuses using various chronostratigraphic tools. Hiatuses inferred from integrated chronostratigraphic studies may be associated either with paraconformities (those without obvious physical expression) or disconformities (those with a distinct physical surface). The disadvantage of this approach is that some surfaces may have global importance but no discernible hiatus in a given section (i.e., the correlative conformity of Vail et al. [1977]). Nevertheless, previous studies have shown that many hiatuses are associated with disconformities formed during inferred sea level lowerings [e.g., Aubry, 1985; Poag et al., 1985; Miller et al., 1990]. The advantage of this approach is that paraconformities can be inferred from hiatuses determined from biostratigraphy, magnetostratigraphy, and Sr isotope stratigraphy that are not obvious in the lithostratigraphy [e.g., Miller et al., 1990].

Hiatuses may also be found in "condensed sections" [Loutit et al., 1988; Baum and Vail, 1988] associated with flooding surfaces, particularly MFS [Van Wagoner et al., 1990]. (We use the term MFS [Van Wagoner et al., 1990] as equivalent to the downlap surface of Loutit et al. [1988] and the surface of maximum starvation of Baum and Vail [1988]; this surface occurs at the top of the transgressive systems tract and is associated with seismic downlap and often with sediment starvation.) Distinguishing between unconformities and MFS can be controversial in coastal plain sections and many classic unconformities have been reinterpreted as condensed sections [Baum and Vail, 1988]. For example, the top of the Shubuta Member of the Yazoo Clay in Mississippi and Alabama has been interpreted both as a subaerial unconformity [Dockery, 1982] and as a condensed section [Mancini et al., 1987]. In many coastal plain sections (e.g., New Jersey), unconformities and flooding surfaces are often concatenated and only highstand system tracts are well developed [Olsson, 1991; Sugarman et al., 1993]. This is not the case with the Alabama Paleogene, because unconformities, transgressive systems tracts, MFS, and highstand systems tracts are often all preserved [e.g., Tew, 1992]. (Lowstand deposits are generally not found on the coastal plain.) Unconformities and flooding surfaces can only be distinguished by the nature of the surface, facies associations above and below the surface, and other stratal relations (e.g., seismic evidence). The chronostratigraphic method used here cannot differentiate between hiatuses associated with sequence-bounding unconformities and hiatuses associated with flooding surfaces, although we argue that most hiatuses are associated with unconformities, not with flooding surfaces (see discussion).

In this study, we use the chronostratigraphic approach to identify hiatuses and to date stratal surfaces identified by other studies. We apply Sr and oxygen isotopic stratigraphy, magnetostratigraphy, and planktonic foraminiferal biostratigraphy to the upper Eocene to Oligocene sections at two Alabama boreholes. Our objectives are (1) to correlate these sections to the GPTS and (2) to evaluate the ages and significance of hiatuses and stratal surfaces and their relationships with other proxies of eustatic change.

METHODS

Core Locations

Two continuously cored 2.5 inch (6.4 cm) diameter boreholes were drilled in the Alabama coastal plain by the ARCO Oil and Gas Company in conjunction with the U.S. Geological Survey, Amoco, the Alabama Geological Survey, and the University of Alabama. The Bay Minette borehole was drilled in 1986 on the Faulkner State Junior College Campus (Bay Minette, Baldwin County; 30°52'N latitude, 87°47'W longitude; S21 T2S R3E; Fig. 1) with core recovery beginning in the Paynes Hammock Formation (upper Oligocene). The St. Stephens Quarry borehole was drilled in 1987 near the highest point on the southwest rim of the quarry (St. Stephens, Washington County; 31°33' N latitude, 88° 02'W longitude; S48 T7N R1W; Fig. 1) beginning in the outcropping Chickasawhay Limestone (Oligocene). This study focuses on the upper Eocene to Oligocene sections of these boreholes; other studies will report on the stratigraphic framework of the older sections cored (Baum et al., submitted manuscript, 1992).

Tectonic Setting

Although the Gulf Coast is a classic passive continental margin, lateral and vertical movement of Jurassic salt has caused tectonic movements in southern Alabama [Murray, 1961]. For example, nonproprietary seismic lines near the SSQ borehole indicate that the Hatchetigbee Anticline (Figure 1) overlies an isolated structural high due to salt withdrawal at depth (G. Baum, personal communication, 1992). Despite salt tectonics, previously reported unconformities in the Alabama Coastal Plain apparently correlate with those in other basins [Baum and Vail, 1988] and with inferred eustatic lowerings of Haq et al. [1987], indicating that local tectonics may have overprinted, but not erased, the global sea level signature.

Lithologic Units

Between eastern Mississippi and central Alabama, Paleogene sediments are composed of mixed clastics and terrigenous carbonates [Toulmin, 1977]. This overall regional facies variation (and concomitant fossil variation) has complicated lithostratigraphic and biostratigraphic correlations. We applied classic formations and members of the three upper Eocene-Oligocene Gulf Coast Stages [e.g., Toulmin, 1977; Dockery, 1982; Mancini and Tew, 1991]:

1. The Jackson Stage includes (from oldest to youngest) the Moodys Branch Formation and North Twistwood Creek Clay, Cocoa Sand, Pachuta Marl, and Shubuta Members of the Yazoo Clay; it is approximately equivalent to the upper Eocene except for the Shubuta Member which is lowermost Oligocene (this study).
2. The Vicksburg Stage includes the laterally gradational Forest Hill Sand/Red Bluff Clay/Bumpnose Limestone, the Mint Spring Formation (generally not recognized formally as a formation in Alabama, although informally it is recognized as part of the Marianna Limestone [e.g., Bybell, 1982, her Figure 4]), the Marianna Limestone, and the Glendon Limestone and Bucatunna Clay Members of the Byram Formation; it is approximately equivalent to the lower Oligocene.

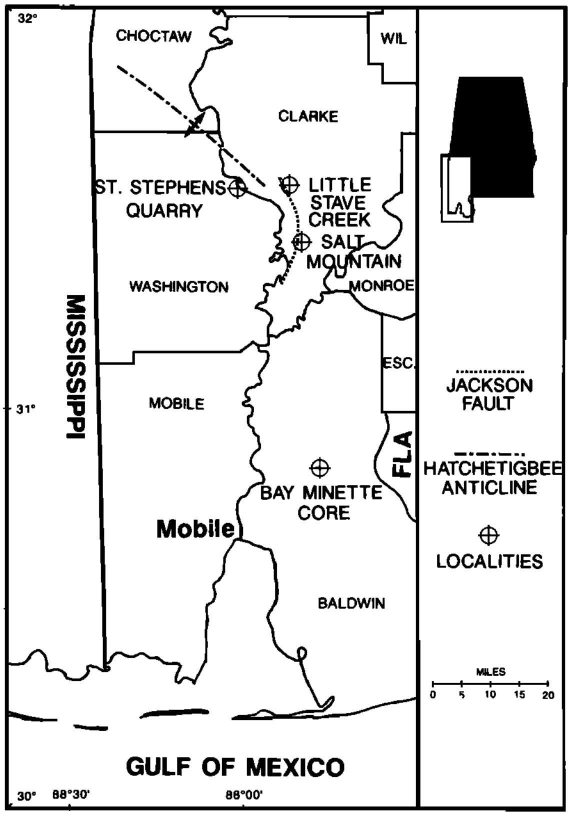

Fig. 1. Location map of Alabama showing counties, boreholes, and geological features discussed here.

3. The Chickasawhay Stage consists of Chickasawhay Limestone; it is approximately equivalent to the middle part of the Oligocene, although its precise age was poorly constrained [e.g., Eames et al., 1962; Poag, 1972; Baum and Vail, 1988] prior to this study and that of Denison et al. [1993].

These formations and members can be directly applied to the SSQ borehole because of its proximity to classic outcrops elsewhere in the quarry; minor ambiguities exist because of weathering differences at outcrop versus the borehole. Application of these units to the more basinward Bay Minette borehole is problematical. For example, lower Vicksburg carbonates in the Bay Minette core between 1480 and 1492 feet (451-455 m) could not be confidently assigned to either the Glendon or Marianna Limestone on the basis of lithology alone. Therefore, data from the Bay Minette borehole do not directly constrain the ages of lithologic units but do provide estimates of the ages of hiatuses.

Strontium Isotopes

Following the pioneering work of Burke et al. [1982], DePaolo and Ingram [1985], and Hess et al. [1986], Sr isotope stratigraphy has become a standard means for correlating upper Eocene to Recent marine carbonates. DePaolo and Ingram [1985] reported Sr isotope age estimates for Gulf Coast Paleogene sections. Denison et al. [1993] reported 105 lower Oligocene and 21 upper Eocene analyses from 42 sites in this region. Our study of the ARCO boreholes provides a detailed sampling of the upper Eocene to Oligocene sections in stratigraphic sequence. For example, we provide 24 and 23 Sr isotope analyses, respectively, for the lower Oligocene sections at SSQ and Bay Minette (Table 1). In contrast, Denison et al. [1993] reported averages for individual formations and members, obscuring intraformational Sr isotopic changes. The advantage of detailed studies in stratigraphic succession is that Sr isotope values for this interval can be expected to monotonically increase upsection; if values do not increase monotonically, then geological problems may be indicated (reworking, diagenesis, etc.). Such problems are more difficult to identify and interpret in isolated outcrop studies. Nevertheless, the high number of analyses obtained from numerous outcrop localities by Denison et al. [1993] increases the confidence in the age estimates of the formations.

Samples were processed using standard procedures [Miller et al., 1988; 1991a] with the exception of samples from the Chickasawhay Limestone, which were processed with sodium tetraphenalborate. Mixed species of foraminifera were cleaned in distilled water and dissolved in 1.5N HCl; two additional samples were analyzed for molluscan shell material and bulk carbonate (Table 1). Standard ion exchange techniques were used to separate strontium, which was analyzed on a VG Sector mass spectrometer at Rutgers University. Internal precision on the Sector is approximately ±0.000012 (mean for 49 samples in this study) and external precision is ±0.000030 or better [Miller et al., 1991a]. At the time that we analyzed the Alabama samples, NBS 987 was routinely measured as 0.710250 normalized to $^{86}Sr/^{88}Sr$ of 0.1194 [Miller et al., 1988] and values are reported as $^{87}Sr/^{86}Sr$ ratios (Table 1). We report two values for Recent marine carbonates of 0.709186±6 and 0.709196±9 measured on the modern giant clam EN-1, an informal Sr-isotope standard; this allows conversion of the data to the $\delta_{seawater}$ notation of Hess et al. [1986]. Data at site 522 [Miller et al., 1988] provide a linear relationship between the $^{87}Sr/^{86}Sr$ ratio with age determined by the magnetochronology of Berggren et al. [1985]:

$$\text{Age (Ma)} = 20080.51 - 28317.88 * {}^{87}Sr/{}^{86}Sr \quad (1)$$

Age estimates derived from (1) have an uncertainty of ±1 m.y. [Miller et al., 1988]. Data are provided in Table 1.

Foraminifera

We used planktonic foraminifera to establish biostratigraphic control. Foraminifera were obtained by disaggregating about 10 cm^3 samples over a 63 μm sieve. Additional samples were examined along with those examined for Sr isotopic studies. We used the Paleogene zonal scheme (P zones) of Berggren and Miller [1988], which is calibrated to the GPTS of Berggren et al. [1985]. The P zones applied differ in definition from those of Blow [1979] (see Berggren and Miller [1988] for discussion).

Magnetostratigraphy

We previously discussed problems in conducting magnetostratigraphic studies of shallow-water sections such as the Bay Minette and SSQ boreholes [Miller et al., 1990]. Shallow-water sediments are often characterized by complex or weak magnetizations of marginal stability. In addition, outcrop studies at SSQ demonstrate that weathering processes may severely alter the paleomagnetic record [e.g., Ellwood et al., 1986]. To address these problems, we analyzed large borehole samples (~50 cm^3) using a large access (6.8 cm) ScT cryogenic magnetometer and a large access (12 cm) solenoid for AF demagnetization. Because the boreholes were not azimuthally oriented, the directions of magnetization were referred to the vertical orientation and an arbitrary horizontal. We measured upper Eocene to Oligocene sections at the Bay Minette and SSQ boreholes (93 and 90 samples, respectively). The natural remanent magnetization (NRM) intensities are low

TABLE 1. Sr isotope Values and Age Estimates

Depth, feet	Age, Ma	$^{87}Sr/^{86}Sr$	Error	Unit
		Bay Minette Core		
1251	29.38	0.708075	0.000004	Chickasawhay
1272	32.91	0.707952	0.000018*	Chickasawhay
1272	33.63	0.707927	0.000006**	Chickasawhay
1302	32.28	0.707974	0.000013	Chickasawhay
1398	33.23	0.707941	0.000006	undifferentiated Byram
1425	35.01	0.707879	0.000010	undifferentiated Byram
1425	34.75	0.707888	0.000009	undifferentiated Byram
1451	34.95	0.707881	0.000021	undifferentiated Byram
1479	35.47	0.707863	0.000008	undifferentiated Byram
1485	35.56	0.707860	0.000017	?Marianna
1489.2	35.59	0.707859	0.000006	?Marianna
1491	35.76	0.707853	0.000006	?Marianna
1491.5	36.23	0.707835	0.000006	?Marianna
1493	36.71	0.707820	0.000006	?Bumpnose
1497	37.74	0.707784	0.000009	?Bumpnose
1502	36.88	0.707814	0.000006	Yazoo
1504	38.12	0.707771	0.000018	Yazoo
1505	37.71	0.707785	0.000006	Yazoo
1512	38.26	0.707766	0.000004	Yazoo
1516	38.09	0.707772	0.000037	Yazoo
1516	38.69	0.707751	0.000029	Yazoo
1572	37.80	0.707782	0.000011	Upper Moodys
1593	38.43	0.707760	0.000011	Upper Moodys
		St. Stephens Quarry Core		
10	32.39	0.707970	0.000011	Chickasawhay
18	32.63	0.707962	0.000084	Chickasawhay
18	32.68	0.707960	0.000011	Chickasawhay
60	34.58	0.707894	0.000007	Marianna
60	34.58	0.707894	0.000007	Marianna
67	34.84	0.707885	0.000009	Marianna
67	35.07	0.707877	0.000015	Marianna
71.8	35.15	0.707874	0.000006	Marianna
80	34.78	0.707887	0.000013	Marianna
80	34.84	0.707885	0.000010	Marianna
88	35.30	0.707869	0.000013	Marianna
103.8	33.31	0.707938	0.000008	Marianna
107	35.36	0.707867	0.000010	Marianna
114.6	36.07	0.707837	0.000008	Marianna
122.4	36.48	0.707828	0.000011	Marianna
129	37.40	0.707796	0.000005	Marianna
132	36.65	0.707822	0.000017	Mint Spring
147	38.61	0.707754	0.000005	Bumpnose
149	37.83	0.707781	0.000015	Bumpnose
155	38.12	0.707710	0.000004	Shubuta
163	37.11	0.707806	0.000006	Pachuta
163	37.31	0.707799	0.000008	Pachuta
163	36.71	0.707820	0.000008	Pachuta
183	37.03	0.707803	0.000015	North Twistwood
		St. Stephens Quarry Outcrops		
AL49	36.32	0.707828	0.000006	6 m below Glendon
AL52	33.18	0.707939	0.000013	basal Chickasawhay

*limestone, not plotted; **shell, not plotted

(median values 0.7 and 0.2 mA/M at Bay Minette and SSQ, respectively). IRM acquisition and thermal demagnetization experiments on several samples from the Bay Minette borehole (1326 feet 494 m), 1485 feet *453 m), and 1516 feet (462 m)) showed an approach to saturation remanence by 200 mT and unblocking temperatures distributed rather uniformly up to 600° C. These properties are usually associated with magnetite, although a small fraction of the IRM sometimes persisted to higher demagnetization temperatures, suggesting the presence of hematite in some samples.

Each sample was subjected to several (typically five) progressive levels of AF demagnetization to 45 mT to remove a dominant steeply downward overprint that is evidently drilling-induced. At the SSQ and Bay Minette boreholes, 27 and 28 samples, respectively, displayed erratic magnetic behavior and were not considered in the polarity interpretation (indicated with pluses in Figures 2 and 4; those plotted on 0° had unstable magnetizations and showed no linear demagnetization trajectories; those pluses plotted ≠ 0° had demagnetization trajectories that did not converge toward the origin, indicating that the characteristic component was not isolated; inclinations for the latter were plotted at highest AF demagnetization step). Data are provided in the Appendix.

We use the GPTS of Berggren et al. [1985] throughout. Cande and Kent [1992] have made substantial revisions (>2 m.y. in some cases) to much of the Paleogene GPTS. For example, the age of the Eocene/Oligocene boundary was estimated as 36.6 Ma by Berggren et al. [1985], while the revised age estimate is closer to 34 Ma [e.g., Premoli-Silva et al., 1988; Cande and Kent, 1992]. The calibrations of the boundaries of chronostratigraphic units, zones, and isotopic correlations are being revised, and it is not possible to use the revised GPTS at this time. However, we note that changes in the ages will not affect the relative timing of the events discussed here.

RESULTS

SSQ Borehole

Magnetozones observed in the upper Eocene to lower Oligocene interval at the SSQ borehole (Figure 2) are interpreted as Chronozones C16 partim to C11 partim (equivalent to Chrons C16-C11 of the GPTS). In particular, the long reversed polarity magnetozone between 54.3 and 132.9 feet (16.6-40.5 m) can be matched with the long reversed interval of Chron C12r (35.54-32.90 Ma). Assuming that there are no major gaps, the underlying magnetozones are identified as Chronozones C13, C15, and C16 (there is no Chron C14 in the GPTS) and the overlying magnetozones are correlated with C12n to C11 (Figure 2). Detailed biostratigraphic criteria support this interpretation of the chronozones and that there are no major hiatuses (Figures 2 and 3):

1. The last occurrences (LO) of *Turborotalia cerroazulensis cerroazulensis* and *Hantkenina* spp. are in Chronozone C13r (163 feet; 49.7 m), as predicted.
2. The first occurrence (FO) of *Cassigerinella chipolensis* is at 159 feet (48.5 m) at the base of the Shubuta Member in Chronozone C13r; the FO of this taxon is generally considered to be basal Oligocene (Chron C13r).
3. The LO of *Pseudohastigerina* spp. at 71 feet (21.7 m) is in Chronozone C12r, as is observed elsewhere [Berggren et al., 1985; Miller et al., 1985a]. Occurrences of small *Pseudohastigerina* above this level at the SSQ borehole (up to 40 feet, 12.2 m) may be the result of reworking. This is supported by the first occurrence of *Subbotina sellii* at 71 feet (21.6 m). Elsewhere, *S. sellii* first appears in Chronozone C12r at about the same level as the LO of *Pseudohastigerina* spp. [Miller et al., 1985a].
4. The LO of *"Turborotalia" ampliapertura* (which is the base of Zone P20) at 40 feet (12.2 m) is in lower

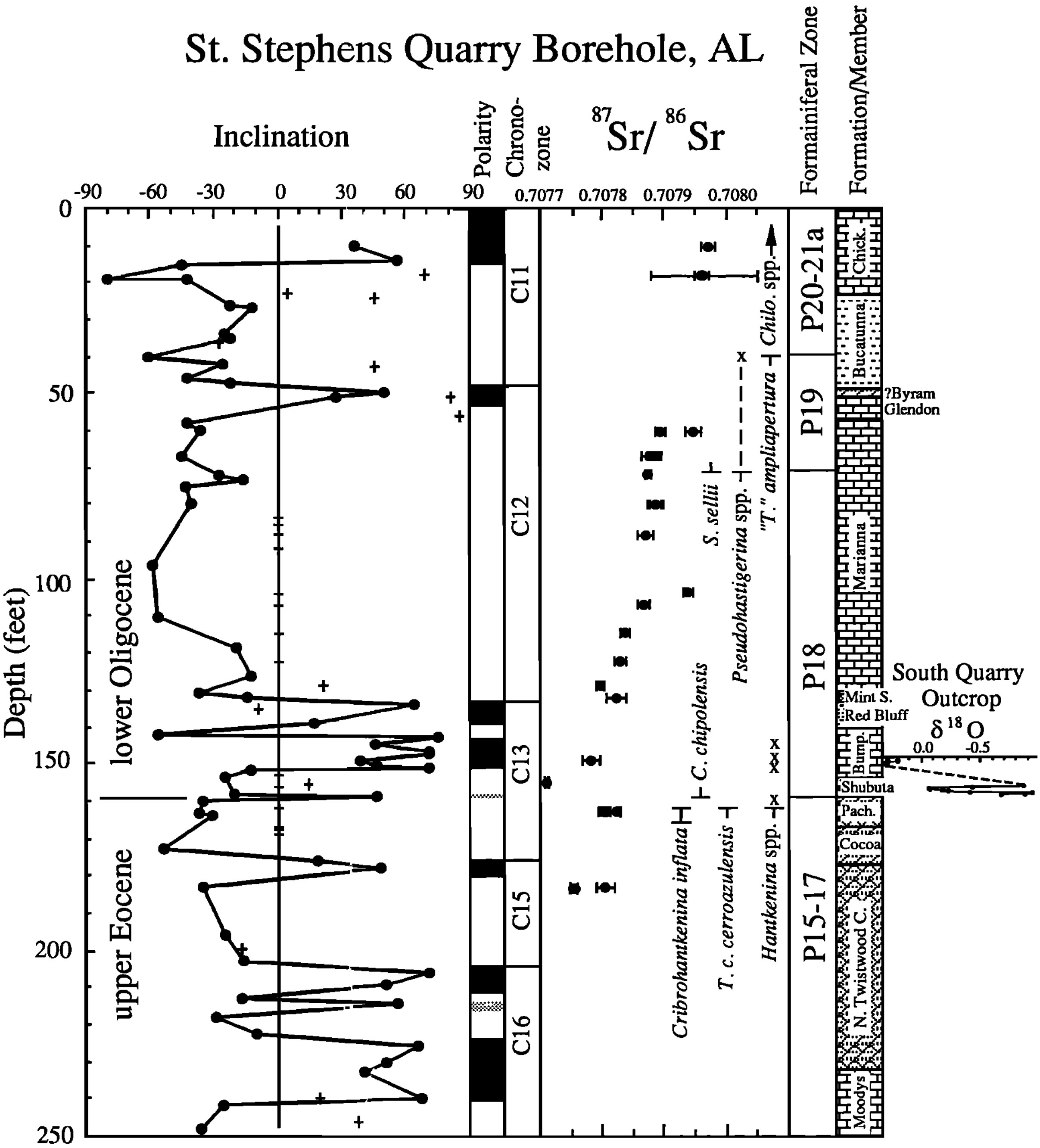

Fig. 2. Sr-isotope stratigraphy, magnetostratigraphy, and biostratigraphy, St. Stephens Quarry, AL borehole. x, isolated specimens interpreted as reworked. T, last occurrence; inverted T, first occurrence. plus indicates samples with erratic paleomagnetic behavior. Stipples indicate uncertain polarity. Sr isotope error bars reflect analytical precision (Table 1). Oxygen isotope data after Keigwin and Corliss [1986].

Chronozone C11r, as it is at sites 516 and 563 [Pujol, 1983; Miller et al., 1985a].

5. *Chiloguembelina* spp. ranges to the top of the borehole, indicating that this entire section is lower Oligocene, older than Chron C10n [Berggren et al., 1985].

The consistency between the magnetostratigraphic and biostratigraphic data is shown in an age-depth diagram (Figure 3). Three biostratigraphic levels fall on the sedimentation rate curve (solid line, Figure 3) determined by interpolating between magnetochron boundaries (open circles, Figure 3); the FO of *S. sellii* falls slightly above the curve, indicating a slightly delayed (<0.5 m.y.; Figure 3) first occurrence. The LO of *Pseudohastigerina* spp. falls slightly above the curve (Figure 3), perhaps indicating a change in sedimentation rate within Chron C12r. Sedimentation rates ranged from ~9 to 25 m/m.y. (Table 2; Figure 3), except for Chronozones C13r and

C12n, which had sedimentation rates of ~4 m/m.y. Slow sedimentation rates in Chronozones C13r and C12n indicate that these sections may contain undetected hiatuses.

Sr isotopic values yield age estimates for the upper Eocene section (163 feet (49.7 m) and below; North Twistwood Creek Clay and Pachuta Marl Members) that are consistent with magnetochronologic age estimates within the ±1 m.y. error (Figure 3). In contrast, Sr isotope age estimates for the lower Oligocene section diverge by 1-2 m.y. from the magnetochronologic ages (Figure 3), substantially larger than the margin of error (±1 m.y.). We interpret this systematic offset to reflect the diagenetic lowering of $^{87}Sr/^{86}Sr$ values (see discussion).

At the SSQ borehole (Figure 3), the Moodys Branch Formation to the Pachuta Marl Member are upper Eocene, the Shubuta Member is interpreted as lower Oligocene (see below), and the Bumpnose Limestone to Chickasawhay Limestone are lower Oligocene (the top of the latter formation is not exposed at the SSQ). Most importantly, we establish that the base of the Chickasawhay Limestone is lower Oligocene and correlates with upper Chronozone C11r (~32.2 Ma). The Bucatunna Clay/Chickasawhay Limestone contact is the major "middle" Oligocene sequence boundary of Baum and Vail [1988, Figure 13]. We establish that there is no discernible hiatus associated with this contact at the SSQ borehole, although a distinct unconformity occurs at the Quarry. Similarly, we establish that there is no discernible hiatus associated with the Eocene/Oligocene boundary at this borehole, although Chronozone C13r has low sedimentation rates (<4 m/m.y.), indicating a possible undetected hiatus (Table 2).

Placement of the Eocene/Oligocene boundary at the SSQ borehole is equivocal. The boundary is defined in the boundary stratotype at Massignano, Italy, at a level coincident with the LO of *Hantkenina* spp. and 14% of the stratigraphic distance downsection within Chronozone C13r [Premoli-Silva et al., 1988]. This level is called C13r.14 [Cande and Kent, 1992]. At the SSQ borehole, this places the boundary either within the upper Pachuta Member (163 feet (49.7 m); the last occurrence of *Hantkenina* spp.) or in the uppermost Shubuta Member (154.76 feet. (47.1 m) at the level of C13r.14). We place the boundary between these levels at the FO of *C. chipolensis* at the base of the Shubuta (see discussion). Although this uncertainty only slightly affects our chronostratigraphic interpretation, it is significant in the regional correlation of the Eocene/Oligocene boundary.

Bay Minette Borehole

We interpret a thick reversed polarity magnetozone between 1300.65 and 1478.5 feet (396.5-450.8 m; Table 3) in the Bay Minette borehole as lower Oligocene Chronozone C12r (35.39-32.9 Ma; Figures 4 and 5). This identification is based on Sr isotope age estimates of ~35 Ma for the lower part of this magnetozone (Table 1, Figure 5). *Subbotina sellii* first occurs below this reversed magnetozone (Figure 4), about 0.5 m.y. older than predicted (Figure 5). An alternative interpretation (Option 3, Figure 5) correlates this reversed magnetozone with Chronozone C11r and the underlying normal as Chronozone C12n. This option correlates the thick, undifferentiated clay of the Byram Formation at Bay Minette (Figure 4) with the Bucatunna Member of the Byram

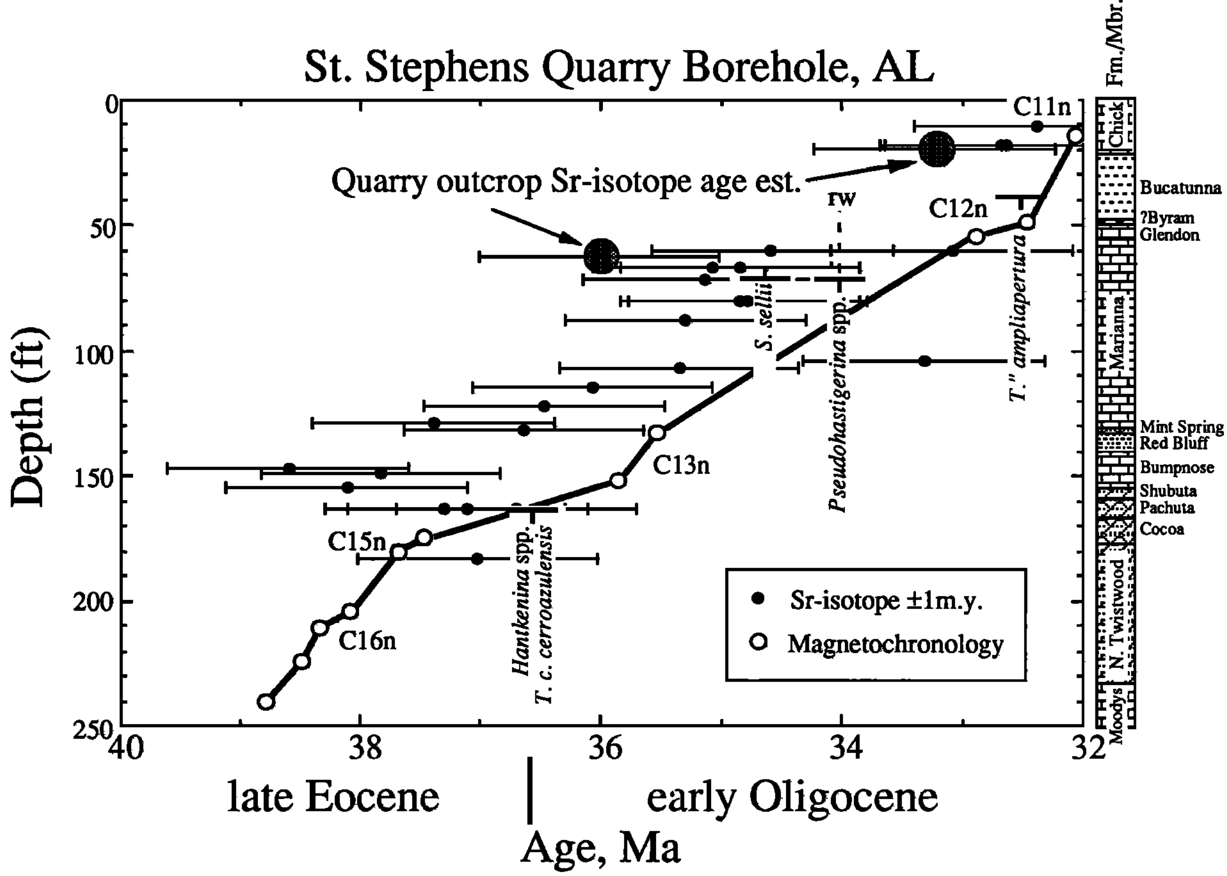

Fig. 3. Age-depth diagram showing magnetostratigraphic, Sr-isotope, and biostratigraphic age estimates, SSQ borehole. rw, specimens of *Pseudohastigerina* spp. interpreted as reworked; range is dashed above last common occurrence.

TABLE 2. Chronozonal Boundaries and Sedimentation Rates, SSQ Borehole

Chronozone	Age, Ma	Depth, m	Depth, feet	Rate, feet/m.y.	Rate, m/m.y.
base C11n	32.06	4.42	14.50		
base C11r	32.46	14.86	48.75	84.1	25.7
base C12n	32.90	16.55	54.30	14.0	4.3
base C12r	35.54	40.51	132.90	29.5	9.0
base C13n	35.87	46.49	152.50	59.4	18.1
base C13r	37.48	53.20	174.50	13.7	4.2
base C15n	37.68	55.03	180.50	30.0	9.2
base C15r	38.10	62.35	204.50	57.1	17.4
base C16n.1	38.34	64.40	211.25	28.1	8.6
base C16r.1	38.50	68.36	224.25	82.8	25.3
base C16n.2	38.79	73.46	240.95	56.0	17.1

Formation at SSQ (Figure 2). However, this interpretation requires extremely high sedimentation rates during Chron C11r (> 100 m/m.y.; dashed line Figure 5) and is contradicted by the Sr isotope data (Figure 5). In addition, it requires that *Pseudohastigerina* spp. last occurs in Chronozone C12n, 1 m.y. after its global last occurrence. Therefore, we favor correlation of this magnetozone with Chronozone C12r, not C11r.

A normal magnetozone at 1478.5-1483.0 feet (450.8-452.1 m) is interpreted as Chronozone C13n by superposition. However, a distinct unconformity lies at 1492.1 feet (454.9 m) at the top of a unit tentatively identified as the Bumpnose Limestone (Figure 4) (Baum et al., submitted manuscript, 1992). In addition, based on our Sr isotope and biostratigraphic data, we infer a paraconformity at 1503.6 feet (458.4 m) at the top of the undifferentiated Yazoo Formation. We suggest that Chronozones C15n and C13n are concatenated (indicated on Figure 4), and that Chronozone C15r lies from 1505-1514 feet (458.8-461.6 m; Figures 4 and 5). This interpretation is supported by the Sr isotope age estimates (Figure 5). The two thick magnetozones found from 1514-1560 feet (461.6-475.6 m) and 1563 feet (476.5 m) to below 1600 feet (487.8 m) are therefore interpreted as Chronozones C16n.1 and C16n.2, respectively.

There is a long hiatus (~ 2 m.y.) associated with the 1503.6 feet (458.4 m) paraconformity (i.e., spanning the Eocene/Oligocene boundary). However, the distinct 1492.1 feet (454.9 m) lower Oligocene disconformity only has a possible short hiatus (<0.5 m.y.) associated with it. These conclusions are based on the following data.

1. Sr isotope values and age estimates show a break from ~36 Ma at 1491 feet (454.6 m) to 36.7 Ma at 1493 feet (455.2 m; Table 1). Sr isotopes also indicate a break between 1502 feet (457.9 m; 36.9 Ma) and 1504 feet (458.5 m; ~38 Ma). (An intermediate age estimate of 37.7 Ma at 1497 feet is interpreted to reflect reworking of older material.) Although both hiatuses are within the resolution of Sr isotope stratigraphy (±1 m.y.), the isotope data require at least one hiatus of ~2 m.y. in the interval from 1493 feet (455.2 m) to 1504 feet (458.5 m).
2. *Hantkenina* spp., *Turborotalia cerroazulensis cunialensis*, *T. cerroazulensis cerroazulensis*, and *T. cerroazulensis cocoaensis* last occur at 1504 feet (458.5 m) in a normal magnetozone, while they disappeared globally in Chron C13r [Berggren et al., 1985]. This indicates that Chronozone C13r is missing. It is unlikely that these last occurrences are premature because the FO of *Cassigerinella chipolensis* at 1497 feet (~Zone P18; Figure 4) is immediately above this level as expected.
3. *C. chipolensis* first occurs in the middle of a normal magnetozone at Bay Minette; at the SSQ borehole, this taxon first occurs in Chronozone C13r, also indicating the absence of Chronozone C13r at Bay Minette.
4. The LO of *Discoaster saipanensis* at 1493 feet (455.2 m; M.P. Aubry, personal communication, 1988) occurs

TABLE 3. Chronozonal Boundaries and Sedimentation Rates, Bay Minette Borehole

Chronozone Base	Age, Ma	Depth Option 1		Depth Option 2		Rate Option 1		Rate Option 2	
		m	feet	m	feet	feet/m.y.	m/m.y.	feet/m.y.	m/m.y.
C10N	30.33	375.61	1232.00	na	na				
C10r	31.23	389.48	1277.50	na	na	51.7	15.8	na	na
C11N	32.06	396.54	1300.65	375.61	1232.00	27.8	8.5	na	na
C11r	32.46	missing	missing	398.48	1277.50	0	0	113.8	34.7
C12n	32.90	missing	missing	396.54	1300.65	0	0	52.6	16.0
C12r	35.29	450.76	1478.50	450.76	1478.50	71.3	21.7	71.3	21.7
C13n.1	35.47	452.13	1483.00	452.13	1483.00	25.0	7.6	25.0	7.6
C15n	37.68	458.98	1505.45	458.98	1505.45	na	na	na	na
C15r	38.10	461.59	1514.00	461.59	1514.00	20.4	6.2	20.4	6.2

na, not applicable

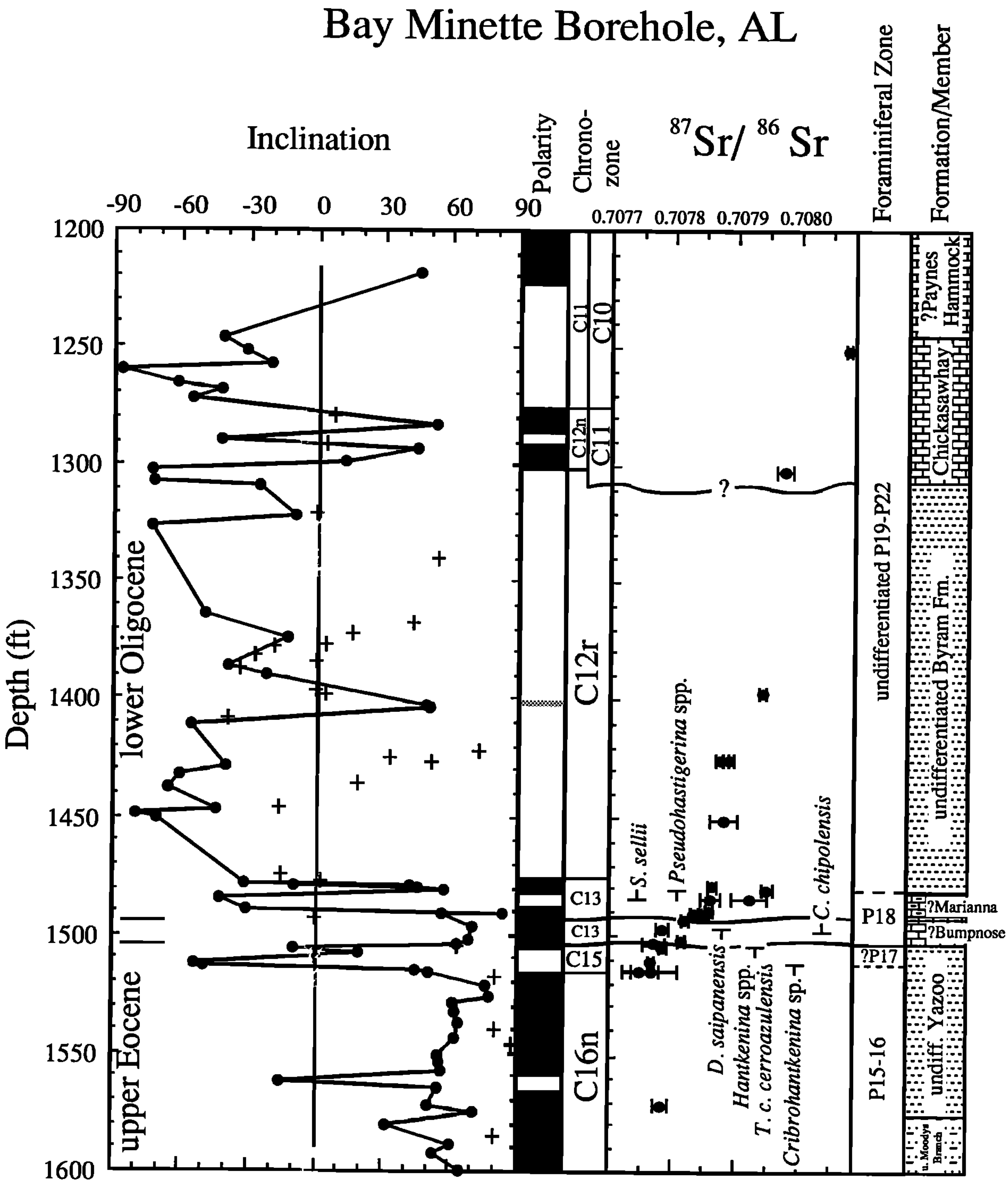

Fig. 4. Sr isotope stratigraphy, magnetostratigraphy, and biostratigraphy, Bay Minette, AL borehole. T, last occurrence; inverted T, first occurrence. Stipples indicate uncertain polarity. Chronozones show option 1 (right column) and option 2 (left column); option 3 is not shown.

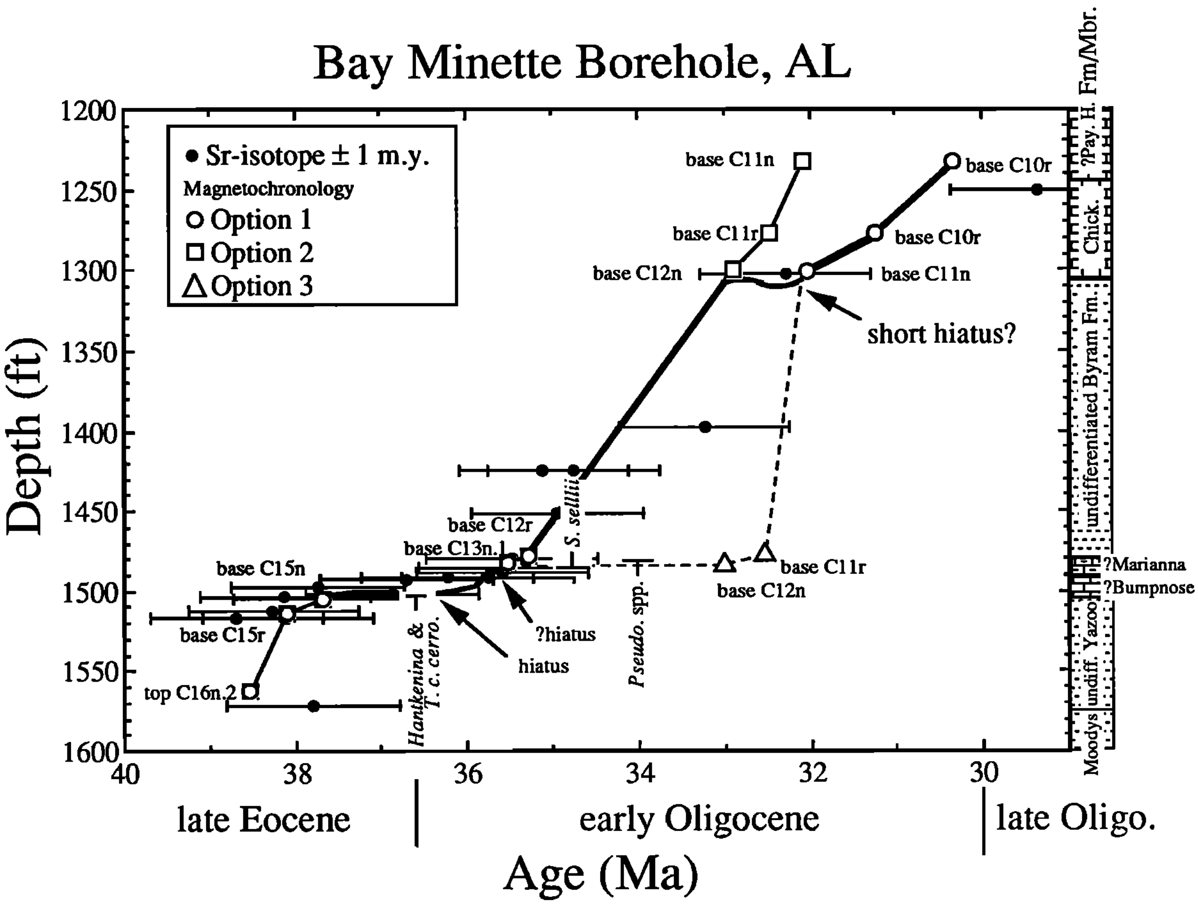

Fig. 5. Age-depth diagram showing magnetostratigraphic, Sr-isotopic, and biostratigraphic age estimates. *T. c. cerro.*, *Turborotalia cerroazulensis cerroazulensis*. *Pseudo.*, *Pseudohastigerina* spp. Sr-isotope error bars reflect analytical precision (Table 1). Fig. 6. Comparison of chronology of deposition at the SSQ and Bay Minette boreholes with the deep-sea benthic foraminiferal oxygen isotope record. Solid intervals indicate section represented by time; wavy lines indicates hiatuses. The benthic foraminiferal oxygen isotope values are those of site 522 (solid circles, [Miller et al., 1988]) and site 529 (open circles, [Miller et al., 1991]). Oi1, Oi2 are the maximum oxygen isotope values used to define the bases of oxygen isotope zones. Slashes indicate age uncertainties in the correlation of the oxygen isotope increase associated with Oi2 (see text). Horizontal arrows indicate the times of most rapid oxygen isotope increase (= inferred glacioeustatic falls). The New Jersey coastal plain record is derived from studies of the ACGS 4 borehole by Miller et al. [1990] and Christensen et al. [submitted manuscript, 1993].

within a normal magnetozone above the LO of *Hantkenina* spp.; its global LO was in Chron C13r before the LO of *Hantkenina* spp. [Berggren et al., 1985]. We attribute this anomalous distribution to reworking upward from in situ Eocene sediments at 1504 feet (458.5 m), based on the consistency of the foraminiferal first occurrences (points 2 and 3 above).

These data clearly indicate that Chronozone C13r is missing (37.2-35.9 Ma), and that Chronozone C15n is only partly represented; this long hiatus (~37.8-35.8 Ma) is associated with a paraconformity at 1504 feet (458.5 m). The data indicate that the normal interval between 1493 feet (455.2 m) and 1504 feet (458.5 m) is correlated with Chronozone C13n; thus the disconformity at the top of the ?Bumpnose Limestone (1492.1 feet, 454.9 m) occurs within Chronozone C13n (35.54-35.87 Ma) and the hiatus was quite short (<0.5 m.y.).

There are two possible interpretations of the upper lower Oligocene magnetozones (Figures 4 and 5). The first assumes stratigraphic continuity and interprets the two normal magnetozones at 1277.5 to 1300.7 feet (396.6-389.5 m) and 1232 feet (375.6 m) to the top of the borehole as Chronozones C12n and C11n partim, respectively. There are several problems with this interpretation:

1. It requires that the basal Chickasawhay Limestone be correlated with uppermost Chron C12r (~33 Ma) at the Bay Minette borehole; in contrast, we correlate the basal Chickasawhay Limestone at the SSQ borehole with uppermost Chronozone C11r (Figures 2 and 3) (see discussion).
2. It places a reversed interval within Chronozone C12n. Although this reversed interval is based on data from a single reversed sample, we place greater credence in reversed polarity determinations than in normal ones, which may reflect complete present-day overprinting. Therefore, we suggest that the magnetozones between 1277.5 and 1300.7 feet (396.6-389.5 m) better match Chron C11n, which has a distinct reversed interval within it.
3. Our Sr isotope age estimates for the basal Chickasawhay Limestone of 32.3 Ma (Table 1) are the same as

uppermost Chron C11r, although they are consistent with uppermost Chron C12r within the ±1 m.y. error.

We favor an alternative interpretation that Chronozones C11n and C10 partim are represented above 1300 feet (396.3 m) at the Bay Minette borehole (Figure 4; indicated with heavy lines, Figure 5). This requires that Chron C12n and most of Chron C11r are not represented due to a hiatus between ~33.0 and 32.0 Ma (Figure 5), consistent with distinct disconformity at 1307 feet (398 m which is the base of Chickasawhay Limestone; Baum et al. (submitted manuscript, 1992) and the interpretation of the basal Chickasawhay Limestone as a sequence boundary [Baum and Vail, 1988](see discussion).

In general, Oligocene planktonic foraminiferal biostratigraphic control is weaker at the downdip Bay Minette borehole than it is in the SSQ sections; the opposite is true for the upper Eocene. We attribute the poor Oligocene faunas at Bay Minette to lithologic control because this borehole has a thick clay section (assigned to the undifferentiated Byram Formation), while the limestones and marls of the undifferentiated Vicksburg Group are not well developed at the Bay Minette borehole. It is these latter units which yielded the better planktonic foraminiferal control at the SSQ borehole (Figure 2).

We found good agreement between Sr isotope age estimates and biostratigraphic ages at the Bay Minette borehole (Figure 5). In general, Sr isotope age estimates monotonically decrease upsection as expected with the following exceptions: (1) values below 1514 feet (461.6 m) are older than 38 Ma, an interval when the rate of change of Sr isotope ratios was low [Burke et al., 1982; Hess et al., 1986]; (2) values at 1272 feet (387.8 m; not plotted) fall well outside of the 2 σ variation (~0.000030, 1 m.y.) of values immediately above and below; and (3) the value at 1497 feet (456 m) is attributed to reworking (see above). The values obtained at 1272 feet (387.8 m) were measured on molluscan shell fragments and bulk limestone and may reflect minor diagenetic alteration (Table 1).

Our studies indicate that a hiatus is associated with the Eocene/Oligocene boundary at the Bay Minette borehole, placing an inferred paraconformity at 1504 feet (458.5 m). On the basis of lithologic and geochemical criteria, Baum et al. (submitted manuscript, 1992) interpreted sequence boundaries at 1512 feet (461.3 m) and 1492.1 feet (454.9 m) at this borehole. As noted above, there may be a short hiatus (<0.5 m.y.) associated with the unconformity at 1492.1 feet (461.3 m). We cannot detect a hiatus associated with the unconformity at 1512 feet (460.0 m, which is the top of the North Twistwood Creek as recognized at this borehole), although Chronozone C15r (1505-1514 feet; 458.8-461.6 m) has fairly low sedimentation rates (~6 m/m.y.; Table 3) and may contain an undetected hiatus.

DISCUSSION

We discuss diagenetic problems encountered at the SSQ borehole, correlations of the local units to the time scale, correlations of hiatuses and stratal surfaces at the Alabama boreholes to an oxygen isotope proxy of glacioeustasy, and implications of our integrated studies to sequence stratigraphy.

Diagenesis

The offset of Sr isotope age estimates from magnetobiochronologic estimates at the SSQ borehole perplexed us at first. Preservation at the SSQ borehole is better than at the more deeply buried Bay Minette borehole, yet the Bay Minette borehole yielded stratigraphically consistent Sr isotope age estimates (Figure 5). The discrepant age estimates at SSQ require either substantial stratigraphic reworking or a diagenetic lowering of Sr isotope values by ~0.000030-0.000050. There is no biostratigraphic evidence for the massive and systematic reworking required to lower $^{87}Sr/^{86}Sr$ values at the SSQ borehole.

There is evidence to support diagenetic lowering of $^{87}Sr/^{86}Sr$ ratios at the SSQ borehole. First, Sr isotope values do not change monotonically upsection (Figure 2) as expected, indicating that there may be a geochemical problem. Second, there seems to be considerable scatter in the Oligocene age estimates at this borehole compared to others (cf. Figures 3 and 5). Finally, outcrop samples show greater offset from magnetochronological ages compared to the borehole (Figure 3). For example, the basal Chickasawhay Limestone yields Sr isotope ages estimates of 33.2 and 32.7 Ma for the outcrop and borehole, respectively (Table 1), while magnetochronology provides an age estimate of ~32.2 Ma (Figure 3). Values for the upper Marianna Limestone show similar offsets (Figure 3, Table 1). The lower values observed at the outcrop can be explained by their poorer preservation. SEM studies of outcrop samples show both well-preserved foraminiferal specimens (preserving pores and wall structures) and sugary coated specimens with obvious crystal overgrowths. We were able to minimize, but not eliminate, diagenetic effects by picking the best preserved foraminiferal specimens from the SSQ borehole.

The age offsets we note cannot be attributed to sample contamination or laboratory problems. Denison et al. [1993] report $^{87}Sr/^{86}Sr$ values for the Chickasawhay Limestone (~0.707900) that are even lower than ours (0.707964 using SSQ borehole data) when corrected for differences in NBS-987 of 0.00110; this yields an age estimate (34.3 Ma; using the regression of Miller et al. [1988]) that is even older than ours (32.4 Ma using borehole data alone, 32.7 Ma using SSQ borehole and outcrop data; Table 1) further complicating this geochemical problem.

Denison et al. [1993] provide the best explanation for these enigmatic Sr isotope values. They suggest that brines migrated up faults associated with the Salt Mountain anticline (Figure 1), transporting low $^{87}Sr/^{86}Sr$ Mesozoic fluids. The low $^{87}Sr/^{86}Sr$ fluids overprinted the Sr isotope signal in Oligocene sediments. The effects of this diagenetic overprint are not uniform; samples from the Shubuta Member and Bumpnose Limestone (147-155 feet; 44.8-47.3 m; Table 1) show the greatest offset, while those from the Marianna and Chickasawhay Limestones show less of an offset (Figure 3). However, diagenesis does not obliterate the Oligocene ages, indicating that recrystallization was not complete.

We estimate the degree of recrystallization using a two end-member mixing model. The potential range of lower Cretaceous to Jurassic values is 0.70680-0.70760 [Burke et al., 1982]. The predicted $^{87}Sr/^{86}Sr$ for a lower Oligocene carbonate is ~0.70792 (for 34 Ma carbonates; Miller et al. [1988]), while the measured values for 34 Ma carbonates at SSQ is 0.70787 (i.e., a 0.00005 lowering). In order to estimate the percent of recrystallization, the Sr concentrations must be known. Using a range of 300-1000 ppm Sr for lower Cretaceous limestones, and 500-1000 ppm for Oligocene foraminifera [Koepnick et al., 1985], recrystallization ranges from 2-38%. Assuming a mean Jurassic value of 0.70720 [Burke et al., 1982] and similar Sr concentrations for the Jurassic and Oligocene carbonates, Oligocene carbonates at SSQ were recrystallized by 12%.

We have previously suggested that oxygen isotopes potentially provide one means of screening deep-sea sections for Sr isotope diagenesis [Miller et al., 1991a]. This method may not work on continental sections such as SSQ. The Sr isotope values for the Shubuta Member and Bumpnose-Red Bluff Formations are clearly altered by about 0.000030-0.000050, yet oxygen isotope values apparently yield an unaltered signal [Keigwin and Corliss, 1986] (Figure 2). This may be partly due to the fact that Keigwin and Corliss [1986] analyzed well-preserved specimens of *Uvigerina* spp. from the SSQ outcrop and that we analyzed large numbers of foraminifera for Sr isotopes including moderately preserved specimens by necessity. Alternatively, the Mesozoic fluids that overprinted the carbonates may not have differed as much from Paleogene oxygen isotope values as they do from Paleogene Sr isotope values.

Gulf Coast Correlations to the GPTS

Placement of the Eocene/Oligocene boundary in Alabama outcrops has been controversial: many workers place it at the Red Bluff/Shubuta contact [e.g., Mancini, 1979; Bybell, 1982; Waters and Mancini, 1982; Baum and Vail, 1988; Mancini and Tew, 1991], within the Shubuta [e.g., Keller, 1985], or at the Shubuta/Pachuta contact [e.g., Cheetham, 1957]. The LO of *Hantkenina* spp. is generally used to correlate the Eocene/Oligocene boundary as stratotypified at Massignano, Italy [Premoli-Silva et al., 1988], although some zonations use the LO of *T. cerroazulensis cerroazulensis* as a means of approximating the boundary [Berggren and Miller, 1988]. Using these criteria (Figure 2) would place the boundary near the middle of the Pachuta Member at the SSQ borehole (162.5 feet, 49.5 m). However, there is some uncertainty regarding the last occurrences of *Hantkenina* spp. and *T. cerroazulensis cerroazulensis* at the SSQ borehole. Reworked specimens of *Hantkenina* spp. are found upsection into the Bumpnose Limestone at the SSQ borehole (Figure 2; see also Keller [1985] and Bybell and Poore [1983]). Reworking is obvious in our material from the SSQ borehole because the sample from 163 feet (49.7 m) contained numerous specimens of *Hantkenina* spp., while samples above this contained only scarce fragments. However, both the LO of *Hantkenina* spp. and *T. cerroazulensis cerroazulensis* are slightly premature in the borehole. This is indicated by (1) the simultaneous LO of *Cribrohantkenina* (which is the marker for the taxon range Zone P16), *T. cerroazulensis* lineage and *Hantkenina* spp. at the SSQ borehole; and (2) the range of in situ *Hantkenina* spp. to top of the Pachuta Marl at the SSQ outcrop [Mancini, 1979; Keller, 1985].

Mancini [1982] and Keller [1985] report *T. cerroazulensis* ssp. within the Shubuta Member at SSQ outcrop, indicating that the boundary may be above the LO of in situ *Hantkenina* spp. at SSQ (i.e., above the Pachuta Marl). However, it is not clear where to place the boundary in the outcrop (i.e., near the base, middle or top of the Shubuta) due to reworking within this thin (~4 feet thick) unit. For example, Keller [1985] suggested that the occurrences of *T. cerroazulensis* ssp. in the upper Shubuta Member are due to reworking and placed the boundary in the lowermost Shubuta. We suggest that the positive identification of *C. chipolensis* at the base of the Shubuta provides the best means of identifying the Eocene/Oligocene boundary.

We place the Eocene/Oligocene boundary at the Pachuta/Shubuta contact at the SSQ borehole based on the FO of *Cassigerinella chipolensis* at 159 feet (48.5 m). Although it is not clear where *C. chipolensis* first occurs in the Massignano boundary stratotype, this taxon has not been previously reported from Eocene strata. This level corresponds to an estimated position of Chronozone C13r.30 at the SSQ borehole. Using magnetochronologic estimates (assuming constant sedimentation rates within Chronozone C13r) alone would place the boundary near the top of the Shubuta Member (154.76 feet; 47.1 m; Chronozone C13r.14) based on the boundary stratotype at Massignano, Italy [Premoli-Silva et al. [1988]. However, the assumption of constant sedimentation rates is not warranted in these shallow water sequences [Loutit et al., 1988].

Our magnetochronology provides precise age estimates for the classic Gulf Coast lithostratigraphic units at SSQ and for stratal surfaces inferred by previous studies (Table 4)[Baum and Vail, 1988; Mancini and Tew, 1991; Tew, 1992; Baum et al., submitted manuscript, 1992]. The age of the basal Chickasawhay Limestone is established as latest Chron C11r (~32.2 Ma) at the SSQ borehole, consistent with the age of 33-32 Ma at Bay Minette. This revises the timing of the major "middle" Oligocene sequence boundary (which is the basal Chickasawhay Limestone of Baum and Vail [1988]; see below). This revision agrees with their speculation that the actual age of the sequence boundary might be older (Biochron P20). We also establish that the base of the Bucatunna Formation at SSQ is associated with the top of Chronozone C12n (32.6 Ma, Table 4), establishing the age of the condensed section (MFS) at this level [Baum and Vail, 1988; Tew, 1992].

The Sr isotope data of Denison et al. [1993] would make the Chickasawhay Formation older (34.4 Ma = middle Chron C12r = late Biochron P18 = earliest Biochron N23). This older age is contradicted by magnetobiostratigraphic data: (1) the Chickasawhay Formation was assigned to Zone P21 by Poag [1972]. Although most diagnostic foraminiferal species are missing from this unit, we show that this formation must be correlated with Zones P20-P21a (Figure 2); (2) the formation is assigned to Zone NP24 by Siesser [1983]. Although this zone is shown as younger than 30.3 Ma by Berggren et al. [1985], biostratigraphic correlations at site 522 show that the base of this zone is substantially older (Chron C11; M.-P. Aubry, personal communication); 3) we correlate it with uppermost Chronozone C11r at SSQ (Figure 2). Correlation to Chronozone C12r would require that the normal magnetozone at ~ 50 feet at SSQ is spurious and the LO of *"T." ampliapertura* and *Pseudohastigerina* spp. are premature (they should be present in the Chickasawhay Formation if it is, in fact, correlated with Biochron P18).

An interval of low sedimentation rates (< 5 m/m.y.) and a possible unresolved hiatus at the SSQ borehole is associated not with the basal Chickasawhay Formation, but with the Glendon Limestone Member and undifferentiated Byram Formation (Chronozone C12n; Figure 3; Table 2). This could indicate that the lithologic units recognized are diachronous relative to the sequence boundary (i.e., with correlative hiatuses at the base of the Chickasawhay Limestone at Bay Minette and within the Byram Formation at SSQ). However, we relate this interval of low sedimentation rates related to an older unconformity (Chron C12n versus C11r) at the top of the Glendon Limestone Member (Table 4; see Mancini and Tew [1991] and Tew [1992], and discussion below).

Our integrated stratigraphy does not discern a hiatus associated with the base of the Mint Spring Formation at the SSQ borehole (Figure 3), although there is an unconformity reported from this level in Alabama [e.g., Baum and Vail, 1988; Mancini and Tew, 1991], which we correlate as the top of Chronozone C13n (35.3 Ma; Table 4). This level correlates with the 35 Ma downlap surface (MFS) of Haq et al. [1987](Table 4). We suggest that the significance of this surface remains open.

We correlate the surface at the top of the Shubuta Member (top Chronozone C13r, 36 Ma) with the TA4.4 sequence boundary of Haq et al. [1987](Table 4). Although previous studies have suggested that this is a condensed section [Baum and Vail, 1988; Loutit et al., 1988; Pasley and Hazel, 1990; Mancini and Tew, 1988; Tew, 1992], we suggest that it is a sequence boundary based on the correlation to TA4.4 and to the oxygen isotope increase (see below). The top of the Shubuta Member is overlain by an unnamed blue clay layer at the SSQ outcrop that represents deepest water deposition [Loutit et al., 1988]. We suggest that the disconformity at the top of the Shubuta as recognized here is a sequence boundary and that the overlying unnamed blue clay of Loutit et al. [1988, p. 199] represents the condensed section of the overlying sequence. This requires that the MFS and sequence boundary are close together or concatenated and interprets the blue unnamed clay as late transgressive systems tract or early highstand systems tract (i.e., the transgressive systems tract is very thin or missing).

We establish that the top of the North Twistwood Creek Clay is upper Chronozone C15n (37.3 Ma, Table 4). This is the level of a sequence boundary of Baum and Vail [1988], Mancini and Tew [1991], and Tew [1991]. Our data show low

TABLE 4. Correlation of Stratal Surfaces Reported by Previous Studies to the GPTS and the Sequences of Haq et al. [1987]

Surface (Reference)	Lithologic Unit	Chronozone	Age, Ma	Correlation to Haq et al. [1987] Sequence	Correlation to Haq et al. [1987] Age, Ma
SB-1 (1, 2)	base Chickasawhay	upper C11r	32.2	TB1.1	30
MFS/CS (1, 2)	base Bucatunna	top C12n	32.6	downlap	32
SB-2 (1, 2)	top Glendon	basal C12n	32.8	TA4.5	33
MFS/CS (1)	base Glendon	uppermost C12r	33.0		?
SB-2 (1, 2)	base Mint Spring	top C13n	35.3	downlap	35
MFS/CS (1, 2) interpreted as a SB by this study	top Shubuta	top C13r	36.0	TA4.4	36
SB-2 (1, 2)	top N. Twistwood Cr.	upper C15n	37.3	TA4.3	37

SB-1, Type 1 sequence boundary; SB-2, Type 2 sequence boundary; 1, Mancini and Tew [1991] and Tew [1992]; 2, Baum and Vail [1988]; MSF, Maximum Flooding Surface; CS, condensed section. Correlations of lithologic units and surfaces to chronozones are largely derived from the SSQ borehole (Figure 2). Correlations to the Haq et al. [1987] sequences are from this study and differ from Baum and Vail [1988], Mancini and Tew [1992], and Tew [1992] in the interpretation of the top of the Shubuta Member as a sequence boundary and the base of the Mint Spring as a possible downlap surface (= MFS).

sedimentation rates in Chronozone C15n at SSQ, consistent with there being an undetected hiatus associated with this surface (Table 2).

Our planktonic foraminiferal zonal assignments are generally consistent with the foraminiferal studies of Mancini and Tew [1991]. However, we place the top of *Pseudohastigerina* spp. (which is the top of P18 of Berggren and Miller [1988]) within the Marianna Limestone rather than the Glendon Limestone Member as suggested by Hazel et al. [1980](Figure 2) and support this by its association with the FO of *S. sellii*. Our interpretations are consistent with the nannoplankton biostratigraphy of Bybell [1982].

Comparisons to the $\delta^{18}O$ record

Summary of late Eocene-early Oligocene $\delta^{18}O$ changes. We previously summarized Cenozoic oxygen isotope data and discussed relationships among $\delta^{18}O$, sea level, and erosion on passive continental margins [Miller et al., 1985b, 1987, 1991b]. We argue for the presence of large continental ice sheets on Antarctica at least intermittently beginning in the Oligocene (ca. 35 Ma). Two major $\delta^{18}O$ increases occurred in the 38-30 Ma interval covered by the SSQ and Bay Minette boreholes, one in the earliest Oligocene and one in the late early Oligocene (Figure 6). Although there are uncertainties in correlation, the SSQ and Bay Minette boreholes provide the best chronologic control available for shallow-water upper Eocene to Oligocene sections, allowing us to evaluate the link between unconformities and the $\delta^{18}O$ record.

The earliest Oligocene (latest Chron C13r to Chron C13n) $\delta^{18}O$ increase occurred in benthic and planktonic foraminifera in every ocean [e.g., Savin et al., 1975; Shackleton and Kennett, 1975; Keigwin, 1980; Oberhansli et al., 1984; Miller et al., 1988, 1991a; Zachos et al., 1992] and was linked with a pulse of glaciomarine sediments deposited near Antarctica (see discussions in the work of Miller et al. [1991b]). We estimated that this $\delta^{18}O$ increase may represent as much as 90 m of glacioeustatic lowering or as little as 30 m (assuming a global compositional change of 0.3-0.9‰ and the Pleistocene $\delta^{18}O$ sea level calibration of Fairbanks and Matthews [1978]; see Miller et al. [1991b] and Zachos et al. [1992] for discussion). Miller et al. [1991b] used the maximum $\delta^{18}O$ value in Chron C13n (35.8 Ma) following this increase to define Oligocene oxygen isotope Zone Oi1 (Figure 6). This increase correlates with the TA4.4 sequence boundary of Haq et al. [1987]. Although sequence stratigraphers have long debated the significance of this event (i.e., whether it is a "Type 1" or "Type 2" event), it is clearly a glacioeustatic lowering of at least 30 m.

A "middle" Oligocene $\delta^{18}O$ increase has been reported in benthic foraminiferal records from Atlantic and Pacific locations [Keigwin and Keller, 1984; Miller and Thomas, 1985]. The age of this increase is still poorly constrained; our best estimate is that it occurred in Chronozone C11r (32.5-31.5 Ma) [Miller et al., 1991b]. While we have not been able to confirm the significance of this event because low-latitude planktonic foraminiferal $\delta^{18}O$ records are poor for this interval, we previously suggested that it represents a major glacioeustatic lowering [Miller et al., 1985b, 1991b]. This is supported by the association of the $\delta^{18}O$ increase with the deposition of widespread glaciomarine sediments near Antarctica (summary in the work by Miller et al. [1991b]). We used the maximum $\delta^{18}O$ values following this increase to define Oligocene oxygen isotope Zone Oi2 (Figure 6) [Miller et al., 1991b] and suggested that it correlates with the TB1.1 sequence boundary of Haq et al. [1987].

The "middle" Oligocene eustatic lowering of Vail et al. [1977] and Haq et al. [1987] (i.e., the TB1.1 sequence boundary) has caused considerable discussion because of its high amplitude and timing. Sequence stratigraphers have long debated the amplitude of this TB1.1 "middle Oligocene" event, with early estimates as high as 400 m [Vail et al., 1977], although amplitude was subsequently estimated as ~130 m by Haq et al. [1987]. Its estimated age has ranged from 29 Ma [Vail et al., 1977] to older than 31 Ma [Olsson et al., 1980; Miller et al., 1985b, 1991b]. Although Baum and Vail [1988] adopted the Exxon age estimate of 29 Ma for this event, they noted that there is evidence for its being older. We have previously suggested that this event is older in New Jersey (older than Zone P21, older than ~31.6 Ma [Olsson et al., 1980, Miller et al., 1985b]). The age of the base of the Chickasawhay Formation (latest Chron C11r; 32.2 Ma; see above) is critical to evaluating the timing of this event, because this disconformity is the expression of the TB1.1 event in the Alabama coastal plain [Baum and Vail, 1988].

Comparison with Bay Minette. Comparison of the two best benthic foraminiferal $\delta^{18}O$ records available for this interval (earliest Oligocene, site 522, Miller et al. [1988] and "middle" Oligocene site 529, Miller et al. [1991b], with the depositional records at Bay Minette shows that:

1. The 37.8-35.8 Ma hiatus at Bay Minette correlates with the earliest Oligocene $\delta^{18}O$ increase associated with Oi1 (Figure 6).
2. The 33.0-32.0 hiatus at Bay Minette is associated with the unconformity at the base of the Chickasawhay Formation (Figure 4); it correlates with the $\delta^{18}O$ increase associated with Oi2, considering the ±0.5 m.y. uncertainty in the age of the increase (Figure 6).

Thus the two major (i.e., durations >1 m.y.) hiatuses at the Bay Minette borehole agree well with the inferred glacioeustatic record. What is particularly intriguing is the observation that maximum $\delta^{18}O$ values associated with the bases of Zones Oi1 and Oi2 correlate not with stratigraphic breaks, but with the resumption of deposition at Bay Minette (Figure 4). Miller et al. [1985b] similarly noted that deposition resumed in New Jersey at the time of maximum $\delta^{18}O$ values (which is the lowest global sea level). We discuss the implications of this below.

Comparison with SSQ. Deposition at SSQ was more continuous than at Bay Minette (Figure 6), and no definite hiatuses were documented. By correlating the disconformity at the base of the Chickasawhay Formation at SSQ with uppermost Chronozone C11r (Figure 2), we establish the age of the major "middle" Oligocene lowering of Vail et al. [1977]) as 32.2 Ma and correlate this event with the oxygen isotope increase associated with Oi2 (Figure 6). We also firmly establish the age of a stratal surface at the top of the Shubuta Member.

Considerable debate has ensued over this stratal surface and placement of unconformities and hiatuses near the Eocene/Oligocene boundary in the Alabama coastal plain particularly at SSQ. One group has placed an unconformity at the base of the Mint Spring Formation and associated a condensed section (MFS) with the top of the Shubuta Member

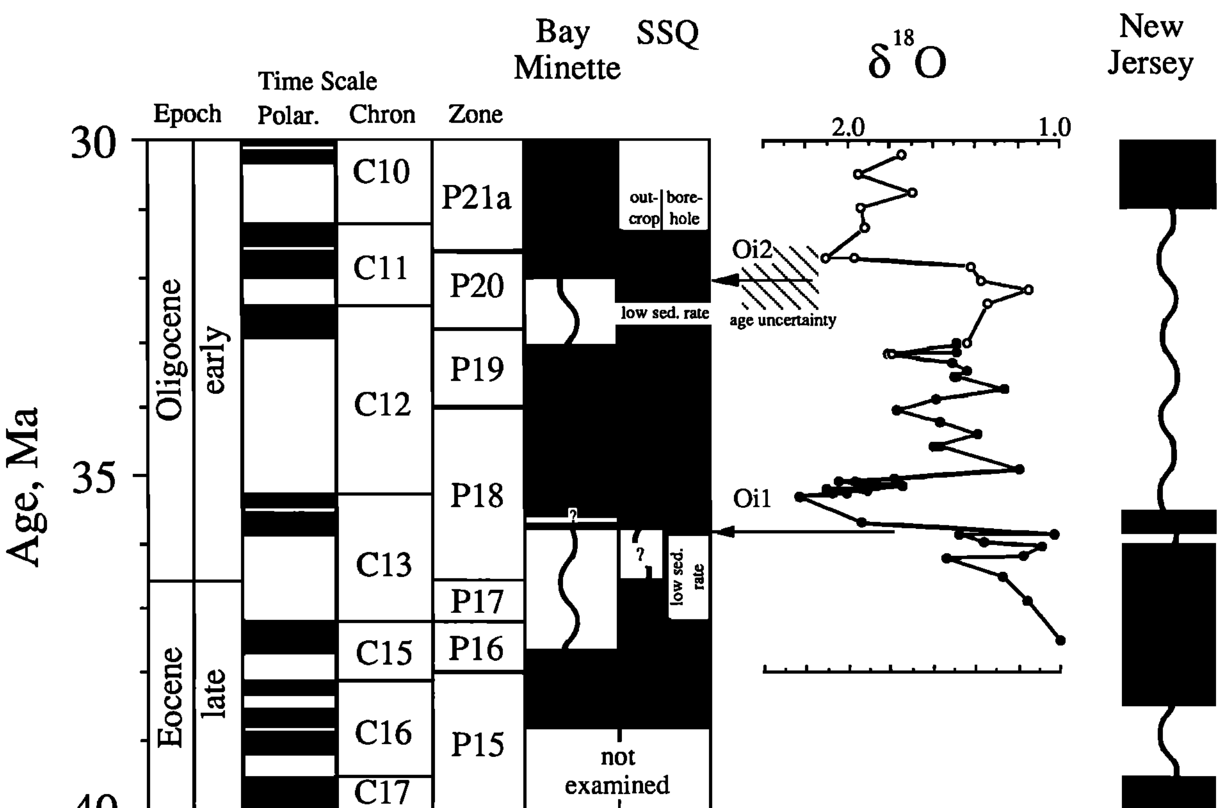

Fig. 6. Comparison of chronology of deposition at the SSQ and Bay Minette boreholes with the deep-sea benthic foraminiferal oxygen isotope record. Solid intervals indicate section represented by time; wavy lines indicates hiatuses. The benthic foraminiferal oxygen isotope values are those of site 522 (solid circles, [Miller et al., 1988]) and site 529 (open circles, [Miller et al., 1991]). Oi1, Oi2 are the maximum oxygen isotope values used to define the bases of oxygen isotope zones. Slashes indicate age uncertainties in the correlation of the oxygen isotope increase associated with Oi2 (see text). Horizontal arrows indicate the times of most rapid oxygen isotope increase (= inferred glacioeustatic falls). The New Jersey coastal plain record is derived from studies of the ACGS 4 borehole by Miller et al. [1990] and Christensen et al. [submitted manuscript, 1993].

[Baum and Vail, 1988; Loutit et al., 1988; Pasley and Hazel, 1990; Mancini and Tew, 1992; Tew, 1992]; another has mapped an unconformity at the top of the Shubuta Member and interpreted this as a sequence boundary, not a MFS [Dockery, 1982]. We establish that Shubuta Member is lowermost Oligocene and that the top of the Shubuta correlates precisely with the $\delta^{18}O$ increase: (1) the $\delta^{18}O$ increase began in latest Chron C13r at site 522 (i.e., in reversely magnetized sediments) and culminated in Chron C13n (base Zone Oi1; Figure 6); and (2) the top of the Shubuta Member (154 feet) is in uppermost Chronozone C13r, while the base of overlying Bumpnose Formation records the polarity transition from Chronozone C13r (152.1 feet) to C13n (151 feet)(Figure 2). Thus we conclude that the top of the Shubuta Member at SSQ correlates with the earliest Oligocene (Zone Oi1) glacioeustatic lowering. This conclusion is supported by Dockery's [1982] observation of channeling on the top of the Shubuta Member in Mississippi, and supports the interpretation of this surface as a sequence-bounding unconformity

Pasley and Hazel [1990] and Mancini and Tew [1991] infer a hiatus at the top of the Shubuta, although they interpret it as a submarine hiatus (condensed section = MFS). Integration of Keigwin and Corliss' [1986] $\delta^{18}O$ studies of the SSQ outcrop with our magnetostratigraphy (Figure 2) indicates that a hiatus may be associated with the top of the Shubuta Member (lowermost Oligocene; Chronozone C13r.14-C13r.32) at the outcrop based on the following:

1. Keigwin and Corliss [1986] report a sharp ~1‰ $\delta^{18}O$ increase in *Uvigerina* spp. across the Shubuta Member/Bumpnose Limestone contact at the SSQ South quarry.
2. This increase at SSQ correlates with the earliest Oligocene $\delta^{18}O$ increase [Keigwin and Corliss, 1986] that occurred across the Chron C13r/C13n boundary (~36.0-35.8 Ma; [Oberhansli et al., 1984; Miller et al., 1988]).
3. The basal Bumpnose Limestone at the SSQ borehole is uppermost Chronozone C13r; Chronozone C13n lies 2-4 feet (0.6-1.2 m) above the contact (Figure 2).

From this, we suggest that the base of the Bumpnose Limestone may be slightly younger at the South Quarry outcrop than it is in the borehole (~1.6 km apart), and that ~2-4 feet (0.6-1.2 m) is missing from the basal Bumpnose Limestone at outcrop (~0.3 m.y.). (We adjusted the depths of the $\delta^{18}O$ data on Figure 2 accordingly and show this as a hiatus on Figure 6).

Comparison to other margins. Miller et al. [1990] and B.A. Christensen et al. (Eocene-Oligocene benthic foraminiferal biofacies and depositional sequence at the ACGS#4 borehole, New Jersey coastal plain, submitted to *Palaios*, 1993) (hereinafter referred to as Christensen et al., submitted manuscript, 1993) have interpreted three major hiatuses in the upper Eocene to lower Oligocene strata of New Jersey:

1. The oldest is a late early Eocene hiatus (39.5-38.5 Ma) that apparently correlates with the TA4.2 sequence boundary of Haq et al. [1987]. The age of this hiatus is poorly constrained, and the coeval section was not examined in this study (Figure 6).
2. A short hiatus is associated with the lowermost Oligocene (~36.5-36.0 Ma) in New Jersey (Christensen et al., submitted manuscript, 1993). It correlates with the TA4.4 sequence boundary of Haq et al. [1987], with the hiatus that spans the Eocene/Oligocene boundary at Bay Minette (Figure 6), and with the top of the Shubuta Member. It also correlates with the $\delta^{18}O$ increase associated with Oi1 (Figure 1). We suggest that the development of this hiatus on two different margins correlates remarkably well with the inferred glacioeustatic record.
3. A long hiatus occurred associated with the interval between ~35 and 31 Ma. This break is consistent with correlation to the $\delta^{18}O$ increase associated with Oi2 and the TB1.1 sequence boundary, the 33-32 Ma hiatus at Bay Minette, and the base of the Chickasawhay Formation at both boreholes. However, the long duration of this hiatus prevents us from demonstrating that this gap in New Jersey is related to the others.

Lower Oligocene strata are missing or thin in many areas of the world [e.g., Eames et al., 1962; Olsson et al., 1980]. A major "middle" Oligocene hiatus occurred on passive margin and epicontinental locations from many regions including Israel [Martinotti, 1981], Australia [Loutit and Kennett, 1981], eastern Canadian margin [Gradstein and Agterberg, 1982], and northwest Europe [Aubry, 1985]. Long hiatuses on other margins prevent us from demonstrating that these hiatuses were all caused by the same event. The Gulf Coast sections, particularly at SSQ, contain one of the most complete lower Oligocene records. By dating the distinct disconformity at the base of the Chickasawhay Limestone at SSQ and Bay Minette [Baum and Vail, 1988; Baum et al., submitted manuscript, 1992], we magnetochronologically estimate an age for this event of latest Chron C11r (~32.46-32.06 Ma with a best estimate of 32.2 Ma on time scale of Berggren et al. [1985]; ~30.2 on the time scale of Cande and Kent [1992]; = the base of TB1.1 of Haq et al. [1987]). We suggest that this is the age of the major event which caused such widespread erosion on passive margins and epicontinental seas.

A stratigraphic break occurred on many margins in the earliest Oligocene, although the number and timing of events have been difficult to determine because of the lower Oligocene is poorly represented in most neritic sections. We attribute poor lower Oligocene sections to concatenation of a late early Oligocene event with an earliest Oligocene event [e.g., Miller et al., 1985b, 1991b]. The Bay Minette borehole provides on example of two early Oligocene hiatuses (Figure 6). Another is provided by site 548 on the Irish continental slope (Goban Spur) where two distinct early Oligocene hiatuses were recorded. One hiatus spanning the Eocene/Oligocene boundary (~36.6-35.5 Ma)[Snyder and Waters, 1985; Miller et al.,1985b] correlates with the earliest Oligocene oxygen isotope increase (=Oi1; Figure 6), the T4.4 sequence boundary of Haq et al. [1987], the hiatus at Bay Minette, and the top of the Shubuta Member at SSQ. The second, longer hiatus (~34-31 Ma) at site 548 [Snyder and Waters, 1985; Miller et al.,1985b] is consistent with correlation to the late early Oligocene oxygen isotope increase (Oi2), the TB1.1 sequence boundary of Haq et al. [1987], the hiatus at Bay Minette, and the base of the Chickasawhay Limestone at both boreholes. While site 548 was deposited in deep water (~1-1.5 km paleodepth) and did not record the direct effects of sea level change, hiatuses and stratigraphic gaps at such slope sites appear to correlate with sea level lowerings [e.g., Miller et al., 1990].

Implications to Sequence Stratigraphy

As noted in the introduction, the chronostratigraphic method used here cannot be used to differentiate between hiatuses associated with sequence-bounding unconformities and hiatuses associated with flooding surfaces. The significance and pervasiveness of hiatuses associated with flooding surfaces remains debatable. J.S. Baum (personal communication, 1992) suggests that hiatuses associated with flooding surfaces at the boreholes studied here should be longer than those associated with unconformities. We suggest that if the section remained under pelagic influence (neglecting local riverine influences) during maximum flooding (i.e., during the maximum rate of relative sea level rise [Posamentier et al., 1988]), the carbonate pump of pelagic productivity would maintain slow (< 10 m/m.y.), continuous sedimentation. In fact, in formulating the concept of the condensed interval, Loutit et al. [1988] considered the entire deep sea a "condensed section." Local parameters (e.g., currents, sediment supply, surface ocean productivity) will determine if there is actually a

TABLE A1. Magnetostratigraphic data

Depth, feet	Inclination, degrees	Chronozone	Formation
	SSQ Borehole		
10.2	36.2	C11n	Chickasawhay
14.0	55.7	C11n	Chickasawhay
15.0	-44.8	C11n	Chickasawhay
18.0	(68.9)		Chickasawhay
19.3	-79.8	C11n	Chickasawhay
19.4	-41.8	C11n	Chickasawhay
23.1	(5.1)		Bucatunna
24.0	(45.8)		Bucatunna
26.0	-22.4	C11r	Bucatunna
27.0	-11.8	C11r	Bucatunna
33.5	-24.4	C11r	Bucatunna
35.0	-22.2	C11r	Bucatunna
36.5	(-27.8)		Bucatunna
40.1	-60.3	C11r	Bucatunna
42.0	-25.6	C11r	Bucatunna
42.5	(45.9)		Bucatunna
45.8	-42.4	C11r	Bucatunna
47.5	-22.3	C11r	Bucatunna
50.0	49.6	C12n	Bucatunna
50.8	27.0	C12n	?Byram
51.0	(81.3)		Glendon
56.0	(85.4)		Glendon
57.8	-41.9	C12r	Glendon
60.0	-36.0	C12r	Marianna
67.0	-44.9	C12r	Marianna
72.0	-27.7	C12r	Marianna
73.5	-15.7	C12r	Marianna
75.0	-43.0	C12r	Marianna
80.0	-40.4	C12r	Marianna
83.8	unstable		Marianna
85.6	unstable		Marianna
88.0	unstable		Marianna
92.0	unstable		Marianna
96.0	-58.6	C12r	Marianna
103.8	unstable		Marianna
107.0	"		Marianna
110.6	-56.1	C12r	Marianna
114.6	unstable		Marianna
118.6	-19.6	C12r	Marianna
122.4	unstable		Marianna
126.4	-12.6	C12r	Marianna
129.0	(21.2)		Marianna
131.0	-36.8	C12r	Marianna
132.0	-14.6	C12r	Mint Spring
133.8	63.7	C13n.1	Red Bluff
135.0	(-9.3)		Red Bluff
139.0	16.8	C13n.1	Red Bluff
142.0	-56.4	C13n.1	Bumpnose
143.0	75.2	C13n.1	Bumpnose
145.0	45.6	C13n.1	Bumpnose
147.0	71.1	C13n.1	Bumpnose
147.5	71.2	C13n.1	Bumpnose
149.0	39.1	C13n.1	Bumpnose
150.2	46.9	C13n.1	Bumpnose
151.0	70.8	C13n.1	Bumpnose
152.1	-12.5	C13r.1	Bumpnose

TABLE A1. (continued)

Depth, feet	Inclination, degrees	Chronozone	Formation
153.0	unstable		Bumpnose
154.0	-25.2	C13r.1	Shubuta
155.5	(14.5)		Shubuta
156.5	unstable		Shubuta
158.0	-20.6	C13r.1	Shubuta
159.0	46.4	?	Shubuta
160.3	-35.6	C13r.1	Pachuta
162.0	unstable		Pachuta
163.0	-37.2	C13r.1	Pachuta
164.1	-30.7	C13r.1	Pachuta
166.8	unstable		Pachuta
167.5	unstable		Pachuta
169.0	unstable		Pachuta
173.0	-53.9	C13r.1	Cocoa
176.0	19.0	C15n	Cocoa
178.0	48.4	C15n	N. Twistwood Creek
183.0	-35.2	C15r	N. Twistwood Creek
195.5	-24.8	C15r	N. Twistwood Creek
199.4	(-17.0)	C15r	N. Twistwood Creek
203.0	-15.7	C15r	N. Twistwood Creek
206.0	70.5	C16n.1	N. Twistwood Creek
209.5	51.1	C16n.1	N. Twistwood Creek
213.0	-17.2	C16r.1	N. Twistwood Creek
214.0	55.9	C16r.1	N. Twistwood Creek
218.0	-29.0	C16r.1	N. Twistwood Creek
222.5	-10.2	C16r.1	N. Twistwood Creek
226.0	65.7	C16n.2	N. Twistwood Creek
230.0	50.7	C16n.2	N. Twistwood Creek
232.5	40.8	C16n.2	Moodys Branch
239.5	67.2	C16n.2	Moodys Branch
239.9	(19.7)		Moodys Branch
242.0	-25.4	C16r.2	Moodys Branch
246.0	(37.8)		Moodys Branch
248.0	-36.0	C16n.2	Gosport
	Bay Minette Core		
1218	45.7	C10n	?Paynes Hammock
1246	-41.7	C10r	?Paynes Hammock
1251.8	-31.3	C10r	Chickasawhay
1257	-20.2	C10r	Chickasawhay
1260	-86.7	C10r	Chickasawhay
1265	-61.7	C10r	Chickasawhay
1268	-42.5	C10r	Chickasawhay
1272	-54.8	C10r	Chickasawhay
1279	(8.0)		Chickasawhay
1283	53.4	C11n	Chickasawhay
1289	-42.4	C11n	Chickasawhay
1291	(4.4)		Chickasawhay
1293	44.7	C11n	Chickasawhay
1299	12.4	C11n	Chickasawhay
1302.3	-72.8	C11r	Chickasawhay
1307	-72.1	C11r	Chickasawhay
unconformity			
1309	-25.5	C12r	undifferentiated Byram
1321	unstable		undifferentiated Byram
1322	-9.4	C12r	undifferentiated Byram
1326	-72.6	C12r	undifferentiated Byram
1340	(54.7)		undifferentiated Byram
1364	-48.9	C12r	undifferentiated Byram

TABLE A1. (continued)

Depth, feet	Inclination, degrees	Chronozone	Formation
1368	(43.1)		undifferentiated Byram
1372	(15.8)		undifferentiated Byram
1374	-13.1	C12r	undifferentiated Byram
1376.5	(4.5)		undifferentiated Byram
1378	(-18.9)		undifferentiated Byram
1382	(-27.2)		undifferentiated Byram
1384	unstable		undifferentiated Byram
1386	-39.1	C12r	undifferentiated Byram
1387	(-33.6)		undifferentiated Byram
1390	-22.4	C12r	undifferentiated Byram
1396	unstable		undifferentiated Byram
1398	(4.5)	C12r	undifferentiated Byram
1403	49.4	?	undifferentiated Byram
1404	50.8	?	undifferentiated Byram
1408	(-38.9)		undifferentiated Byram
1411	-55.0	C12r	undifferentiated Byram
1422	(73.2)		undifferentiated Byram
1425	(33.3)		undifferentiated Byram
1427	(52.1)		undifferentiated Byram
1429	-40.3	C12r	undifferentiated Byram
1432	-60.6	C12r	undifferentiated Byram
1436	(18.9)		undifferentiated Byram
1438	-65.1	C12r	undifferentiated Byram
1446	(-16.5)		undifferentiated Byram
1447	-43.8	C12r	undifferentiated Byram
1449	-79.9	C12r	undifferentiated Byram
1451	-70.5	C12r	undifferentiated Byram
1475	(-15.6)		undifferentiated Byram
1477	(2.9)		undifferentiated Byram
1478	-31.5	C12r	undifferentiated Byram
1479	42.7	C13n.1	undifferentiated Byram
1479.5	-9.3	?	undifferentiated Byram
1480	45.7	C13n.1	?Marianna
1481	57.9	C13n.1	?Marianna
1485	-42.4	C13r.1	?Marianna
1489.2	-30.4	C13r.1	?Marianna
1491	84.1	C13n.2	?Marianna
1491.5	56.9	C13n.2	?Marianna
unconformity			
1493	unstable		?Bumpnose
1497	70.2	C13n.2	?Bumpnose
			?Bumpnose
unconformity			
1502.5	68.6	C15n	Yazoo Formation
1504	64.1	C15n	Yazoo Formation
1505	(63.4)		Yazoo Formation
1505.9	-9.5	C15r	Yazoo Formation
1508	19.7	?	Yazoo Formation
1510	unstable		Yazoo Formation
1512	-53.3	C15r	Yazoo Formation
1513	-49.0	C15r	Yazoo Formation
1515	45.3	C16n.1	Yazoo Formation
1516	50.7	C16n.1	Yazoo Formation
1518	(80.5)		Yazoo Formation
1522	76.6	C16n.1	Yazoo Formation
1526	78.5	C16n.1	Yazoo Formation
1529	62.0	C16n.1	Yazoo Formation
1533	62.6	C16n.1	Yazoo Formation
1537	64.7	C16n.1	Yazoo Formation
1540	(80.8)		Yazoo Formation

TABLE A1. (continued)

Depth, feet	Inclination, degrees	Chronozone	Formation
1544	62.6	C16n.1	Yazoo Formation
1547	(88.0)		Yazoo Formation
1551	54.8	C16n.1	U. Moodys Branch
1554	56.3	C16n.1	U. Moodys Branch
1558	56.9	C16n.1	U. Moodys Branch
1562	-14.9	C16n.1	U. Moodys Branch
1565	55.2	C16n.2	U. Moodys Branch
1572	50.6	C16n.2	U. Moodys Branch
1575	71.6	C16n.2	U. Moodys Branch
1581	32.3	C16n.2	U. Moodys Branch
1585	(80.9)		U. Moodys Branch
1589	61.1	C16n.2	U. Moodys Branch
1593	53.8	C16n.2	U. Moodys Branch
1600	65.2	C16n.2	U. Moodys Branch

hiatus developed in association with flooding surfaces or if deposition was continuous.

A good example of problems with differentiating flooding surfaces from unconformities is provided by the Glendon-?Byram-lowermost Bucatunna Formations at SSQ (Figure 2), an interval that we noted above has very low sedimentation rates. Tew [1992] interprets the Glendon Limestone Member as a highstand systems tract separated from the "Byram Marl" transgressive systems tract by a type 2 sequence boundary; the overlying Bucatunna is interpreted as the highstand systems tracts. Our one ?Byram sample and overlying Bucatunna sample are normally magnetized and are interpreted as Chronozone C12n. A short hiatus is implied by the low sedimentation rates in Chronozone C12n (< 5 m/m.y.). This inferred hiatus could be associated either with the unconformity at the top of the Glendon or with the MFS that separates the transgressive from highstand systems tracts (i.e., between the ?Byram and Bucatunna).

Our correlations at Bay Minette show that hiatuses are associated with the inflections of the $\delta^{18}O$ increases (= inferred glacioeustatic lowerings); resumption of deposition is associated with the maximum $\delta^{18}O$ values (which are inferred glacioeustatic minima). At the SSQ borehole, we show that two previously reported surfaces correlate with the inflections of the $\delta^{18}O$ increase: the distinct disconformity at the base of the Chickasawhay Formation [e.g., Baum and Vail, 1988] and the surface at the top of the Shubuta Member (a disconformity of Dockery [1982]; a condensed section (MFS) of Mancini and Tew [1991] and Tew [1992]). These observations are consistent with models [Pitman, 1978; Christie-Blick et al., 1990; Christie-Blick, 1991] in which unconformities form during the maximum rates of eustatic fall (i.e., during the $\delta^{18}O$ increases).

Our observation that deposition resumes on the coastal plain at times when global (glacioeustatic) sea level is actually at its lowest point (i.e., during the intervals of maximum $\delta^{18}O$ values, Figure 6) appears counterintuitive. However, it is at this point in an oscillating sea level record that the rate of eustatic fall is at its minimum; subsidence can then create enough accommodation for sedimentation to resume updip. It is at this point in the sea level curve that the transgressive systems tracts develop [Posamentier et al., 1988]. Our interpretation of deposition resuming at the time of lowest global sea level predicts that transgressive system tracts should be observed above the unconformity in this setting (lowstand deposits are not generally represented in the coastal plain). This is consistent with the interpretations of Mancini and Tew [1991], who note development of transgressive systems tracts in the Alabama coastal plain above major sequence boundaries.

CONCLUSIONS

Although the upper Paleogene sediments of the Gulf Coast have figured prominently in the understanding of global sea level changes, their ages have been debated because of the problems in correlating these shallow-water sections. We applied standard tools of magnetostratigraphy, isotopic stratigraphy, and biostratigraphy to two Alabama boreholes, circumventing problems with weathering in outcrops. Each tool has difficulties in a different part of the section: (1) magnetostratigraphic correlation is uncertain across the Eocene/Oligocene boundary at Bay Minette due to unconformities; (2) diagenesis lowered $^{87}Sr/^{86}Sr$ values in the Oligocene section at SSQ by about 0.000030-0.000050; and (3) planktonic foraminiferal biostratigraphic markers are rare in the upper Eocene at SSQ and in the Oligocene at Bay Minette. Despite these failings, each tool proved invaluable: (1) an excellent polarity record was constructed for the SSQ borehole; (2) biostratigraphy was remarkably good for the Oligocene at SSQ; and (3) Sr isotopes provided the control needed to interpret the polarity history across the Eocene/Oligocene boundary at Bay Minette. Only by integrating all three stratigraphic tools could a reliable chronology be developed. This chronology provides the necessary first step in evaluating the global synchrony of erosional events and the role of eustatic changes on passive margin sedimentation. Using this chronology, we compared late Eocene to Oligocene hiatuses in Alabama with an oxygen isotopic proxy for glacioeustatic changes. We conclude that erosion on the Alabama coastal plain occurred during intervals of increasing $\delta^{18}O$ values (i.e., during the time of most rapid inferred glacioeustatic fall) and that deposition resumed at the time of peak $\delta^{18}O$ values (during times of lowest sea level).

APPENDIX: MAGNETOSTRATIGRAPHIC DATA

Values in parentheses are for the highest AF demagnetization step; they were not considered in polarity interpretation because the characteristic component was not isolated. They are plotted on Figures 2 and 4 as pluses with values $\neq 0°$. Values are not given for samples with unstable magnetization; these are plotted on Figures 2 and 4 as pluses with values = 0°.

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