Oligocene glacio-eustasy and erosion on the margins of the North Atlantic

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ABSTRACT

Oligocene foraminiferal $\delta^{18}O$ records suggest the development of ice caps (= inferred glacio-eustatic falls) at ca. 36, 31, and 25 Ma. Biostratigraphic analyses of wells from the United States east coast and Irish continental margins consistently show that upper Oligocene sediments overlie a diachronous erosional surface, underlain by lower Oligocene to Eocene strata. At the minimum, the hiatus extends between ca. 34 and 30 Ma. We speculate that erosion during a glacio-eustatic fall near the early/late Oligocene boundary (ca. 32–31 Ma) developed (1) an unconformity on the margins, (2) numerous canyons noted in seismic profiles from the margins of the North Atlantic, and (3) a coastal offlap event. Using $\delta^{18}O$ data, we apply a model for eustatic changes and margin response that explains the relationships of sea level, unconformities, and coastal onlap/offlap events.

INTRODUCTION

Publication of a "global cycles of sea level" curve (Vail et al., 1977; Vail and Hardenbol, 1979; among others) has intensified a long-standing geological controversy over the history of and causal mechanisms for global sea-level (eustatic) changes and their relation to unconformities. Mesozoic-Cenozoic cycles represent stratal patterns of coastal onlap (inferred sea-level rise) and "offlap" (inferred sea-level fall determined from downward shifts in onlap) observed primarily on passive continental margins. Pitman (1978) and Watts (1982) questioned the direct relationship between onlap/offlap patterns and eustatic changes and noted that glaciation is the only known mechanism for producing sea-level falls rapid enough to cause the widespread development of margin unconformities. We present evidence for three Oligocene glacio-eustatic lowerings; compare these with seismic, biostratigraphic, and chronostratigraphic records (Figs. 1, 2) from the United States and European margins; and attempt to explain margin response to these glacio-eustatic events (Fig. 3).

ISOTOPIC EVIDENCE FOR OLIGOCENE GLACIO-EUSTASY

Except for that of Matthews and Poore (1980), most isotopic studies have assumed that no significant continental ice existed prior to a $\delta^{18}O$ increase in the middle Miocene (Savin et al., 1975; Shackleton and Kennett, 1975; among others). Recent studies indicate that significant continental ice was present in the Oligocene (Miller and Fairbanks, 1983, 1984; Keigwin and Keller, 1984; Miller and Thomas, 1984; Poore and Matthews, 1984; Shackleton et al., 1984). These authors measured high $\delta^{18}O$ values (>1.9‰) in the benthic foraminifera *Cibicidoides* above the Eocene/Oligocene boundary (ca. 36–35 Ma), near the lower/upper Oligocene boundary (ca. 31–28 Ma), and below the Oligocene/Miocene boundary (ca. 25–24 Ma) at Atlantic and Pacific Deep Sea Drilling Project (DSDP) sites (time scale of Berggren et al., 1984, used throughout). Modern $\delta^{18}O$ values for *Cibicidoides* spp. at these locations would be only 1.6‰ if no continental ice sheets existed. Thus, if the Oligocene were ice free, bottom waters at these abyssal sites must have been 1–2 °C colder than today at ca. 36–35, 31–28, and 25–24 Ma. Such cold bottom-water temperatures are incompatible with an ice-free world and suggest that the $\delta^{18}O$ increases resulted from the combination of increased ice-volume and decreased bottom-water temperature.

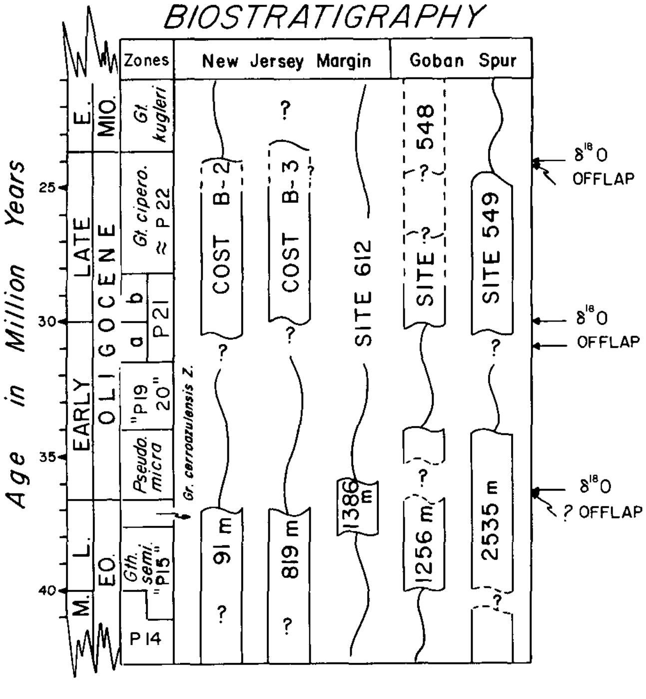

Figure 1. Stratigraphic record of wells (New Jersey margin), and DSDP boreholes (U.S. east coast and Irish margins). Present water depths for wells and boreholes indicated on columns. Planktonic foraminiferal zonal assignments: COST wells modified (see text) after Olsson et al. (1980) and Poag (1980); Site 612 after Graciansky, Poag et al. (1984); Irish continental margin (Goban Spur) sites after Miller et al. (1984) and Snyder and Waters (1984). Offlap and $\delta^{18}O$ events from Figure 3.

Covariance of planktonic and benthic foraminiferal $\delta^{18}O$ has been shown to be an indicator of Quaternary ice-volume changes (Shackleton and Opdyke, 1973). About 0.3‰–0.5‰ $\delta^{18}O$ covariance occurred at ca. 36, 31, and 25 Ma, and if we apply the Quaternary sea-level $\delta^{18}O$ calibration, sea level fell by 30 to 50 m at these times. A conservative value of 35 m is used in our eustatic curve (Fig. 3; summary in Miller and Fairbanks, 1984).

BIOSTRATIGRAPHY AND CHRONOSTRATIGRAPHIC RECORDS

We document that an upper lower Oligocene unconformity is present on two continental margins of the North Atlantic (minimum hiatus extends between 34 and 30 Ma). Unconformities apparently occur near both the top and the base of the Oligocene as well, but these are poorly constrained by existing data.

U.S. Middle Atlantic Continental Margin. Olsson et al. (1980) noted that a major unconformity separated a truncated Eocene surface from upper Oligocene strata in the middle Atlantic coastal plain. They dated the youngest sediments below this unconformity as upper Eocene (*G. cerroazulensis* Zone), and suggested that the oldest sediments above correlated with the *G. ampliapertura* Zone (≅ Zone P19/20, 34.0–31.4 Ma). The exact age of the sediments above the unconformity is uncertain. Our biostratigraphic studies indicate that Zone P21a (31.4–30.0 Ma) lies above the unconformity at the COST B-2 well (Fig. 1); Poag (1980) would suggest an even younger age (Zone P21b). At the COST B-3 well (Fig. 1) sediments above the unconformity apparently are slightly younger (Zone P21b, 30.0–28.2 Ma; this study, Poag, 1980). Although biostratigraphy at these wells remains equivocal, we affirm that lower Oligocene sediments are largely missing from the U.S. Atlantic coastal plain and continental shelf because of one or more erosional events.

Uppermost Oligocene biostratigraphy is even more uncertain on the U.S. margin. At the COST B-3 well, the Oligocene/Miocene contact may be either conformable (Poag, 1980) or disconformable (R. K. Olsson and A. J. Melillo, 1984, personal commun.). At the COST B-2 well (Fig. 3), upper Oligocene strata are separated from middle Miocene strata by an unconformity (Poag, 1980).

Irish Continental Margin (Goban Spur). Drilling at DSDP Sites 548 and 549 revealed an unconformity near the lower/upper Oligocene boundary at both sites. At Site 549, sediments assigned to Zone P21a overlie the *P. micra* Zone (Fig. 1); at Site 548, Zone P21a (possibly Zone P20) overlies the *P. micra* Zone (Fig. 1). These data suggest that the unconformity was formed in the late early Oligocene between ca. 34 and 30 Ma. Other Oligocene unconformities were noted on the Irish margin (Fig. 1). An unconformity occurs near the Eocene/Oligocene boundary at Site 548; however, the Eocene/Oligocene contact at Site 549 is apparently conformable. A major unconformity occurs between the upper Oligocene and middle Miocene at Site 549. Stratigraphic resolution is limited in the upper Oligocene at Site 548 (Graciansky, Poag et al., 1984; Miller et al., 1984; Snyder and Waters, 1984).

SEISMIC STRATIGRAPHIC RECORD OF THE OUTER MARGINS

Deep-sea erosion by bottom currents developed a widespread angular unconformity [Horizon A^u; Fig. 2 southeast of shotpoint (SP) 5000] in the western North Atlantic basin between the latest Eocene and early Miocene (e.g., Tucholke and Mountain, 1979). In the northern North Atlantic, Miller and Tucholke (1983) noted an unconformity with similar stratigraphic relations (reflector R4) and correlated it with uppermost Eocene to lowermost Oligocene sections. Consequently, they suggested that reflectors R4 and A^u mark equivalent horizons eroded by bottom currents near the Eocene/Oligocene boundary.

Although Horizon A^u is easily recognized in profiles from the deep-water North Atlantic margin, it is difficult to trace this reflector landward without ambiguity. For example, in Figure 2 seaward of SP 5000, Horizon A^u is a single, discrete unconformity. However, landward of SP 3700 there are two unconformities. The shallower, undulating interface (labeled mO) has been equated with Horizon A^u (Markl and Bryan, 1982). Mapping of this surface demonstrates unequivocally that it constitutes the surface of a series of canyons cut into the continental margin (Dillon et al., 1984). By analogy with the Pleistocene, these canyons developed during lowered sea level. The mO canyons are the largest canyons yet identified beneath the U.S. Atlantic margin, and Dillon et al. (1984) correlated their formation with the large "middle" Oligocene offlap event of Vail et al. (1977).

Correlation of reflector mO with Horizon A^u implies that A^u, the bottom currents that eroded it, and the canyon cutting are all simultaneous expressions of a sea-level event. However, Mountain and Tucholke (1984) argued against this correlation. They interpreted the relatively flat reflector that can be traced beneath reflector mO landward of SP 3700 (Fig. 2) as Horizon A^u. The canyon-cutting event represented by reflector mO merges with A^u seaward of SP 3700. The lower surface (= Horizon A^u) probably was eroded by bottom currents near the Eocene/Oligocene boundary. Subsequently, a seaward-prograding wedge of sediments was deposited along the southeastern U.S. margin (Fig. 2). A major downslope erosional event followed and developed numerous canyons and reflector mO. We agree with Markl and Bryan (1982) and Dillon et al. (1984) that this canyon-cutting event is the

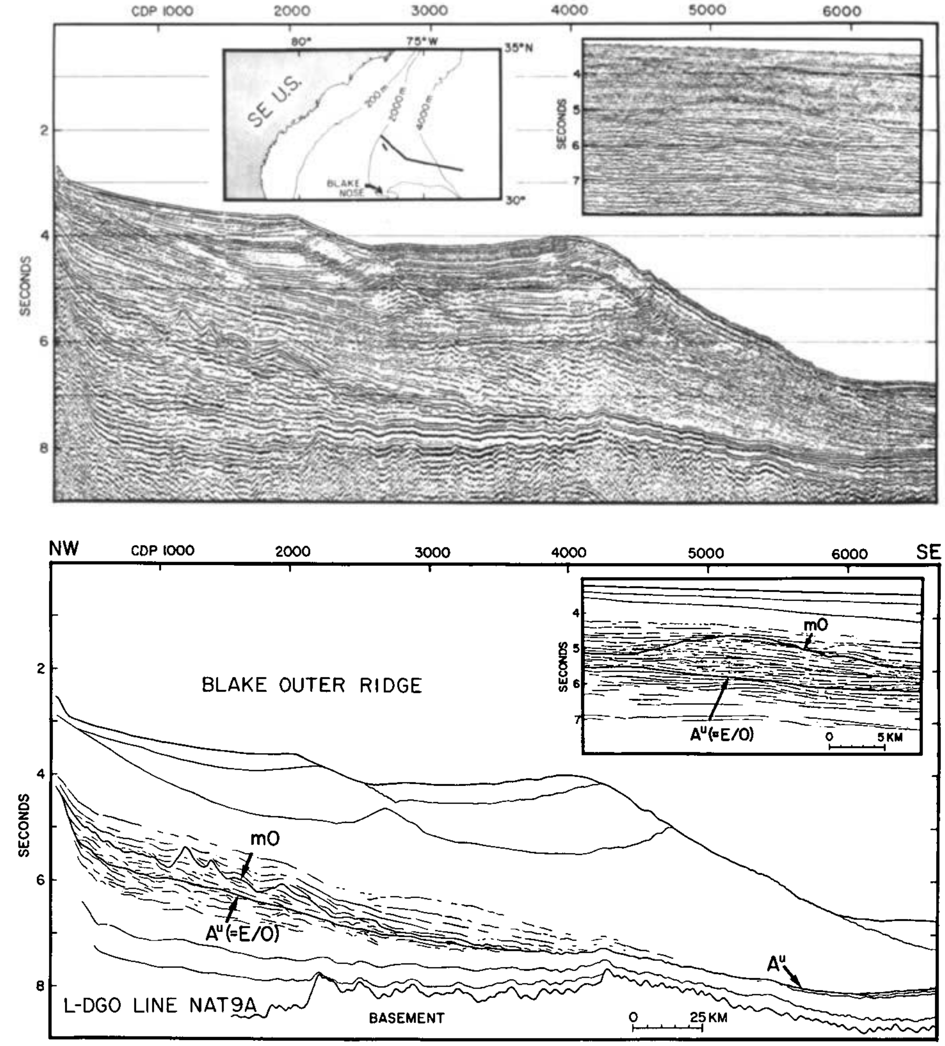

Figure 2. Multichannel seismic reflection profile NAT-9A and interpretation across Blake Outer Ridge, southeastern U.S. continental margin. Reflector identifications described in text. Inset shows details of same reflector relationships in LDGO Line 83, 25 km to southwest. Inset map shows location of profiles.

"middle" Oligocene event of Vail et al. (1977), although no borehole data exist yet to test this conclusion.

An erosional surface similar to reflector mO appears on the Norwegian margin; correlation of this surface to DSDP Site 338 suggests a "middle" Oligocene age (Mutter, 1984). On the Jan Mayen Ridge in the Norwegian-Greenland Sea, we also observe a similar erosional unconformity that separates upper Oligocene from Eocene strata at Site 346. Unfortunately, biostratigraphic resolution is limited at both these high-latitude drill sites, and the precise timing of the erosion is unclear. The same kind of dissected erosional surface has been observed beneath the New Jersey continental slope (Schlee and Grow, 1980). However, recent drilling and seismic-correlation studies show that this canyon-cutting event predates Horizon A^u and occurred in the middle Eocene (Mountain and Tucholke, 1984).

MODEL FOR OLIGOCENE GLACIO-EUSTASY AND MARGIN RESPONSE

We have constructed an Oligocene eustatic curve (Fig. 3) by subtracting our glacio-eustatic estimates from long-term tectono-eustatic estimates based upon ridge volume changes (Pitman, 1978; Kominz, 1984). We compare our eustatic curve with a generalized margin section and changes in coastal onlap/offlap in Figure 3.

The latest Oligocene (ca. 25–24 Ma) glacio-eustatic lowstand either correlates with or slightly postdates an offlap event (Fig. 3). However, the chronostratigraphic record is too uncertain to establish whether this lowstand caused widespread erosion on the margins examined here (Figs. 1, 3).

A major glacio-eustatic event also is suggested in the earliest Oligocene (Fig. 3). However, the record of coastal onlap in the latest Eocene to earliest Oligocene is uncertain. Although Vail et al. (1977) noted an offlap event at the Eocene/Oligocene boundary, Loutit et al. (1983) deemphasized its importance and changed its age assignment (Fig. 3). Biostratigraphic and chronostratigraphic evidence for erosion is also uncertain at this time (Fig. 1).

Comparison of the eustatic model with the onlap curve and chronostratigraphic record near the early/late Oligocene boundary (Fig. 3) shows several intriguing relationships. Keigwin and Keller (1984) implied that peak $\delta^{18}O$ values and the offlap event (Vail et al., 1977) occurred together at 29 Ma. This may not be true. Although the highest $\delta^{18}O$ values (= lowest glacio-eustatic lowstand) occurred in Zone P21a/P21b, which has an estimated age of 31 to 28 Ma, the offlap event occurred near the base of Zone P21a (Vail et al., 1977; Vail and Hardenbol, 1979). The offlap event therefore is constrained biostratigraphically to a level with an age of 32–31 Ma.[1] The hiatus on the U.S. and Irish margins is constrained to 34–30 Ma, preceding and slightly overlapping the time of peak $\delta^{18}O$ values. Therefore, high $\delta^{18}O$ values continued for at least 2 m.y. after the offlap and erosional events.

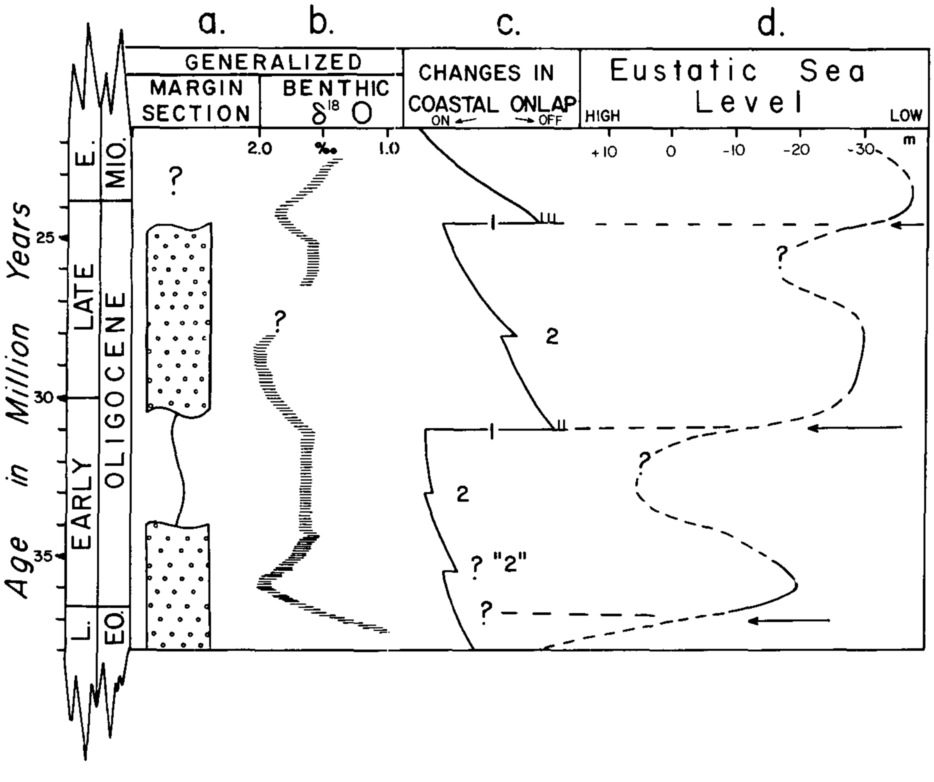

Figure 3. Model for eustatic sea-level changes. a: Generalized margin section derived from Figure 1. b: Generalized benthic foraminiferal $\delta^{18}O$ record from Miller and Fairbanks (1984), Miller and Thomas (1984), and Keigwin and Keller (1984). c: Changes in coastal onlap after Vail et al. (1977), Vail and Mitchum (1980), and Loutit et al. (1983); 1 indicates major offlap events, 2 indicates smaller offlap events, hachured area indicates erosional truncation. d: Eustatic curve constructed by subtracting 35 m at ca. 35–36 Ma, 28–31 Ma, and 24–25 Ma from Kominz's (1984) long-term record. Shape of curve (solid lines) constrained by benthic foraminiferal data; intervening sections (dashed lines) assume smoothly varying rise and fall. Sea level given in metres above present. Arrows indicate inflection points where rates of sea-level lowering were greatest.

There are uncertainties in Oligocene biostratigraphy, and the unpublished biostratigraphic data used to calibrate the onlap curve cannot be evaluated. Therefore, we cannot rigorously assess the apparent offset in the timing of the glacio-eustatic lowstand and coastal offlap. For example, we cannot ascertain whether the chronostratigraphic break (34–30 Ma) noted on the U.S. and Irish margins (Fig. 1) truly correlates with the offlap event (Fig. 3). However, we suggest that the offset is real, and we argue that this is a logical consequence of margin response to eustasy.

The increase is $\delta^{18}O$ apparently *began* near the base of Zone P21a (ca. 32–31 Ma; Miller and Fairbanks, 1984). Pitman (1978) suggested that shorelines respond not the absolute vertical position of global sea level, but to the *rate* of eustatic change. In his model, sea-level change has the greatest effect on margins when the rate of fall is fastest, not necessarily when sea level is lowest. Hence, erosion and offlap should occur at times of maximum rate of fall of eustatic sea level. We apply Pitman's model for sea-level changes using $\delta^{18}O$ data to constrain the sea-level curve. In the late early Oligocene (Fig. 3, column d) the greatest rate of eustatic sea-level fall occurred at ca. 32–31 Ma. Sea level remained low from 31–28 Ma, but the rate of fall decreased and margin subsidence exceeded sea-level fall beginning about 30 Ma. Thus, we suggest that erosion, offlap, and canyon cutting occurred during the time of ice buildup (ca. 32–31 Ma). Deposition and onlap *resumed* on the margin at ca. 30 Ma while sea level was still low (Fig. 3, columns a, d).

Despite uncertainties in stratigraphic correlations, we are confident that high $\delta^{18}O$ values (= lowest glacio-eustatic lowstands) were asso-

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[1] Vail et al. (1977) and Vail and Hardenbol (1979) used Hardenbol and Berggren's (1978) time scale to assign dates of 30–29 Ma to the lower part of Zone P21a; the time scale of Berggren et al. (1984) reassigns this interval to ca. 32–31 Ma. Keigwin and Keller (1984), using *different* time scales for each event, concluded that the offlap and the peak $\delta^{18}O$ values both occurred at 29 Ma.

ciated with the late Oligocene resumption of deposition on both the Irish and New Jersey margins. This model resolves the apparent contradiction that deposition began on the margins while sea level was still low. As more data become available, we should be able to test the model for postulated glacio-eustatic lowerings at 36 Ma and 25 Ma.

CONCLUSIONS

Oxygen-isotope studies indicate that glacio-eustatic falls of 30–50 m occurred at ca. 36, 31, and 25 Ma. Sea level remained low at 36–35, 31–28, and 25–24 Ma. Stratal patterns on margins can be related not to the absolute position of sea level, but to the rate of eustatic change, while offlap events (Vail et al., 1977) correlate with the maximum rate of sea-level fall. The stratigraphic records of two passive margins discussed here suggest that a rapid eustatic fall at ca. 31 Ma eroded unconformities on the continental shelves and cut canyons into the slopes. Sea level remained low from 31 to 28 Ma, but the *rate* of fall decreased during this period. Thus, margin subsidence exceeded the rate of sea-level fall, and deposition on the margins resumed by ca. 30 Ma. Seismic studies of the U.S. Atlantic slope and rise suggest that this event was *preceded* by bottom-current erosion that formed Horizon A^u. There is no apparent causal relation between these Oligocene sea-level and deep-water erosional events. The eustatic model used here explains much of the chronostratigraphic, isotopic, and seismic stratigraphic evidence now available, but it needs rigorous testing. This model provides an illustration of the potential response of the geologic record to eustatic fluctuations.

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