Does ice drive early Maastrichtian eustasy?

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ABSTRACT

A large (30–40 m), rapid (<<1 m.y.), earliest Maastrichtian sea-level drop inferred from New Jersey sequence stratigraphic records correlates with synchronous $\delta^{18}O$ increases in deep-water benthic and low-latitude surface-dwelling planktonic foraminifera. The coincidence of these events argues for the development of a moderate-sized ice sheet during the early Maastrichtian.

INTRODUCTION

Early studies interpreted low $\delta^{18}O$ values of Late Cretaceous to Eocene deep-water benthic foraminifera as reflecting warm high latitudes and suggested that large continental ice sheets did not develop until the middle Miocene (Savin et al., 1975; Shackleton and Kennett, 1975). This interpretation was consistent with early results showing that ice-rafted debris near Antarctica was primarily late Miocene and younger (e.g., Tucholke et al., 1976). Subsequent records of Oligocene glaciomarine sediments near Antarctica and global $\delta^{18}O$ variations established that a large ice sheet was present since at least 33 Ma (see summary in Miller et al., 1991; Barrera and Huber, 1991) and perhaps since ca. 42 Ma (Browning et al., 1996). Nevertheless, floral and faunal evidence for relatively warm conditions on coastal Antarctica (Askin, 1992; Case et al., 1998) and the lack of firm isotopic or sedimentological evidence for glaciation has led to the interpretation that the Earth was largely ice free during the Late Cretaceous to early Eocene.

The interpretation of an ice-free Earth before ca. 42 Ma is not consistent with evidence for brief (<<1 m.y.), high-amplitude (10 to ~70 m) global sea-level (eustatic) changes during the Late Cretaceous to early Eocene that have been inferred using sequence stratigraphy (Haq et al., 1987). The Haq et al. (1987) record includes a large (~60 m) early Maastrichtian sea-level fall at 71 Ma. Although the Haq et al. (1987) record has been questioned, large, rapid eustatic events occurred during the Late Cretaceous to early Eocene, an interval assumed to be ice free (e.g., Browning et al., 1996). However, the only known mechanism that can produce such large, rapid sea-level variations is glacioeustasy (Pitman and Golvchenko, 1983).

Data supporting or refuting possible Cretaceous to early Eocene ice sheets are lacking, and the assumption of an ice-free Cretaceous was questioned by Stoll and Schrag (1996). Although there is ample geological evidence for the warmth of the Late Cretaceous to early Eocene Southern Ocean (with temperatures of 10–15 °C; Stott et al., 1990) and for the relative warmth (6–9 °C) of the coastal Antarctic regions (Barrera et al., 1987), it is possible that the interior of East Antarctica could have supported a moderate-sized ice sheet.

Any reconstruction of the glacial history of the planet must be consistent with the inferences from at least three major data sets. The most important and least equivocal data are provided by direct evidence for glaciations (e.g., ice-rated debris). This evidence is restricted to Antarctic sediments younger than 42 Ma (e.g., Barrera and Huber, 1991), although sampling of older Antarctic sediments is limited. A second source of evidence is the marine carbonate $\delta^{18}O$ record, which reflects an ice-volume signal as well as the effects of temperature and local precipitation on seawater $\delta^{18}O$ (e.g., Savin et al., 1975). A third source of evidence about continental glaciation, albeit circumstantial, is provided by the sequence stratigraphic record. Coincidence of sequence boundaries, $\delta^{18}O$ variations, and limited direct glacial evidence over the past 42 m.y. has demonstrated that glacial growth and decay is an important, if not dominant, process controlling eustasy (Miller et al., 1998a).

Recently published isotopic (Barrera and Savin, 1999) and sequence stratigraphic data (Sugarman et al., 1995; Miller et al., 1998b) provide a new perspective on sea-level and isotopic changes during the Maastrichtian, an interval of notable oceanographic and climatic variability. The Maastrichtian was the coolest interval of the Cretaceous following the peak warm in the early Late Cretaceous (Barrera, 1994). We focus in this paper on a major global $\delta^{18}O$ increase and sea-level lowering that occurred in the early Maastrichtian.

ISOTOPIC METHODS AND RESULTS

Oxygen isotope data (Fig. 1; Kennett and Stott, 1990; Barrera and Huber, 1990; Barrera and Savin, 1999) were obtained from benthic foraminifera (*Gavelinella beccariiformis* and *Nuttallides truempyi*) and shallow-dwelling planktonic foraminifera *Pseudoguembelina excolata* and *P. costulata).* The *G. beccariiformis* record is from Antarctic Site 690 (65°10′S, 01°12′E; paleolatitude ~65°S; paleodepth 1800 m), the *N. truempyi* record is from Pacific Deep Sea Drilling Project (DSDP) Site 463 (21°21′N, 174°40′E; paleolatitude ~3°S; paleodepth 1500 m), and the planktonic foraminiferal record is from Pacific DSDP Sites 463 and 305 (32°13′S, 157°51′E; paleolatitude ~10°N; paleodepth ~1000 m). These upper Campanian–Maastrichtian sections were correlated using magnetostratigraphy, biostratigraphy, Sr, and $\delta^{13}C$ stratigraphy, with age uncertainties of 0.5 m.y. or less (Barrera and Savin, 1999). Numerical ages were assigned based on the magnetochronology of Cande and Kent (1995) and a 71.3 Ma Campanian-Maastrichtian boundary (chron C32n; Gradstein et al., 1994).

Benthic foraminiferal $\delta^{18}O$ values at tropical Pacific Site 463, as well as other sites in the Indian-Pacific and South Atlantic basins (Barrera et al., 1997; Barrera and Savin, 1999), were higher by 1.0‰ between 71 and ca. 70 Ma than during the preceding and subsequent 2 m.y. intervals (Fig. 1). The 71 Ma isotopic increase at Site 463 seems to have occurred in a stepwise manner, and the termination of the cool episode between 70 and 69 Ma was characterized by high isotopic variability. At Antarctic Site 690, benthic foraminiferal $\delta^{18}O$ values increased gradually from 74 to 71 Ma; the 71 Ma increase was smaller than at Site 463, and values did not decrease after 70 Ma. A second $\delta^{18}O$ increase that occurred at 68 Ma at Sites 690 and 463 appears to have been global (Barrera and Savin, 1999).

The $\delta^{18}O$ values of shallow-dwelling planktonic foraminifera from low-latitude Pacific Sites 463 and 305 follow the same trend: values increased at 71 Ma, coinciding with a similar change in benthic foraminiferal $\delta^{18}O$ values, and remained higher after this time. The 71 Ma planktonic $\delta^{18}O$ increase was ~0.3‰ at Site 463, and 0.5‰ at Site 305.

SEQUENCE STRATIGRAPHIC METHODS AND RESULTS

Upper Cretaceous sequence boundaries in the New Jersey coastal plain are distinct surfaces of erosion associated with base-level shifts (Sugarman et al., 1995). Sequences generally consist of basal transgressive system tract glauconite sands (e.g., the Navesink Formation, Fig. 2) and overlying regressive (highstand systems tract) silts and sands (i.e., the Red Bank Formation; Fig. 2). Olsson (1991) and Sugarman et al. (1995) dated outcrops (Big Brook, Poricy Brook, and New Egypt; Fig. 2) and updip boreholes (Freehold and Clayton; Fig. 2) using biostratigraphy and Sr isotopic stratigraphy. Drilling downdip at Bass River (Fig. 2) recovered

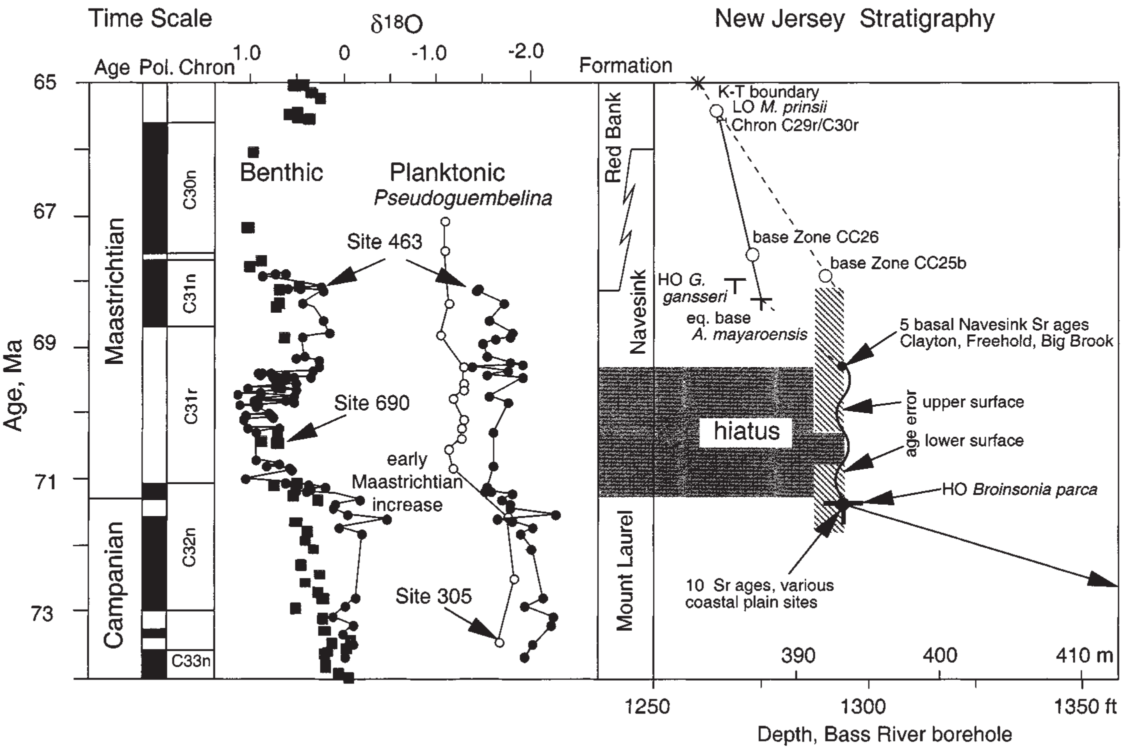

Figure 1. Comparison of late Campanian to Maastrichtian $\delta^{18}O$ records from Antarctic Site 690 and tropical Pacific Sites 463 and 305 with New Jersey coastal plain stratigraphy. HO = highest occurrence; eq. = equivalent to; LO = lowest occurrence; K-T = Cretaceous-Tertiary; Pol. = polarity.

a thick (212 m) Upper Cretaceous section dated with planktonic foraminiferal and nannofossil biostratigraphy and limited magnetostratigraphy (Miller et al., 1998b). We integrated results from outcrops and the Bass River borehole (Fig. 1) to quantify age estimates of uppermost Campanian to Maastrichtian sequences.

The sequence boundary separating the base of the Navesink Formation from the underlying Mount Laurel Formation (Fig. 2) is one of the most dramatic of the 30 Cenozoic and 6–7 Late Cretaceous sequence boundaries identified in the New Jersey coastal plain (Miller et al., 1998a, 1998b). In outcrop, the Mount Laurel Formation is a well-sorted medium sand with *Ophiomorpha* and *Asterosoma* burrows and tabular planar cross-bedding, indicating bimodal (onshore-offshore) current directions; it records lower shoreface deposition (Martino and Curran, 1988). The sequence boundary is marked by a distinct erosional surface and a phosphorite layer that is often iron cemented at its base.

The sequence above the Mount Laurel Formation consists of a heterolithic interval, overlain successively by the Navesink and Red Bank Formations. The heterolithic interval (to 1.5 m thick) consists of a blue-gray silt with pods and beds of reworked Mount Laurel sand and rip-up clasts (Fig. 2). This reworked zone is regressive and we interpret it as one of the few lowstand deposits in the coastal plain. A cemented erosional surface at the top of the lowstand section is interpreted as a transgressive surface (Fig. 2). Above the transgressive surface, the Navesink Formation is a clayey glauconite sand; benthic foraminiferal data indicate that outcrop localities were deposited in ~60 m paleodepth (Olsson, 1991). The overlying silts and fine sands of the Sandy Hook Member of the Red Bank Formation were deposited in shallower conditions (below wave base, ~30 m paleodepth), whereas the overlying sands of the Shrewsbury Member represent an offshore sand bar deposited parallel to shore.

The age of the sequence boundary and the overlying Navesink and Red Bank Formations has been firmly established. Below the sequence boundary, 10 Sr isotopic age estimates from the upper Mount Laurel Formation at various coastal plain localities (Fig. 2) yield an average age of 71.3 Ma (Sugarman et al., 1995). Whereas the errors on Sr isotope age estimates are large for this interval (±1 m.y.), the highest occurrence of *Broinsonia parca* at the top of the Mount Laurel Formation at Bass River yields the same age. The age of the overlying surface is less well constrained by 5 Sr isotopic ages as 69.1 Ma; biostratigraphic data establish it as definitely older than 68 Ma (Fig. 1). We estimate the age of the hiatus as 69.1 ± 1 Ma to 71.3 ± 0.5 Ma (Fig. 1).

DISCUSSION

The simultaneous increase in benthic and low-latitude planktonic foraminiferal $\delta^{18}O$ values ca. 71 Ma indicates a decrease in both deep and tropical sea-surface temperatures and/or an increase in the $\delta^{18}O$ value of the oceans (δ_w) due to increased continental ice volume. The $\delta^{18}O$ increase is greater in benthic foraminifera, and thus part of the change probably involved a decrease in deep-water temperature. The growth of an ice sheet requires that $\delta^{18}O$ values increase globally, both in deep and tropical surface waters, as observed in the early Maastrichtian ca. 71 (Fig. 1); such covariance has been used as an ice-volume indicator for the Pleistocene and Tertiary (Shackleton and Opdyke, 1973; Miller et al., 1987). Nevertheless, because tropical sea-surface temperatures may vary (e.g., Guilderson et al., 1994), coeval benthic and tropical planktonic $\delta^{18}O$ increases do not provide unequivocal evidence for ice growth.

Benthic foraminiferal $\delta^{18}O$ values provide evidence for the development of ice sheets ca. 71 Ma. Cool, if not frigid, deep-water temperatures are indicated by early Maastrichtian $\delta^{18}O$ values of ~1.0‰ in *Nuttallides* (Fig. 1). These values would indicate bottom-water temperatures of ~5 °C assuming an ice-free world (δ_w = –1.2‰ PDB [Peedee belemnite] in the paleotemperature equation [Shackleton and Kennett, 1975; Miller et al., 1987] with *Nuttallides* values of –0.76‰ relative to equilibrium [Pak and Miller, 1992] and the paleotemperature equation in Barrera and Savin [1999]). Such high $\delta^{18}O$ values are incompatible with an ice-free world because they indicate cool conditions in the high-latitude source regions (e.g., Miller et al., 1987). Temperatures would have been only slightly higher (6–7 °C vs. 5 °C) if a moderate-sized ice cap was present.

The ca. 71 Ma $\delta^{18}O$ increase in benthic and tropical planktonic foraminifera correlates with the sequence boundary found in New Jersey (Fig. 1) and elsewhere (Haq et al., 1987). The hiatus associated with the sequence

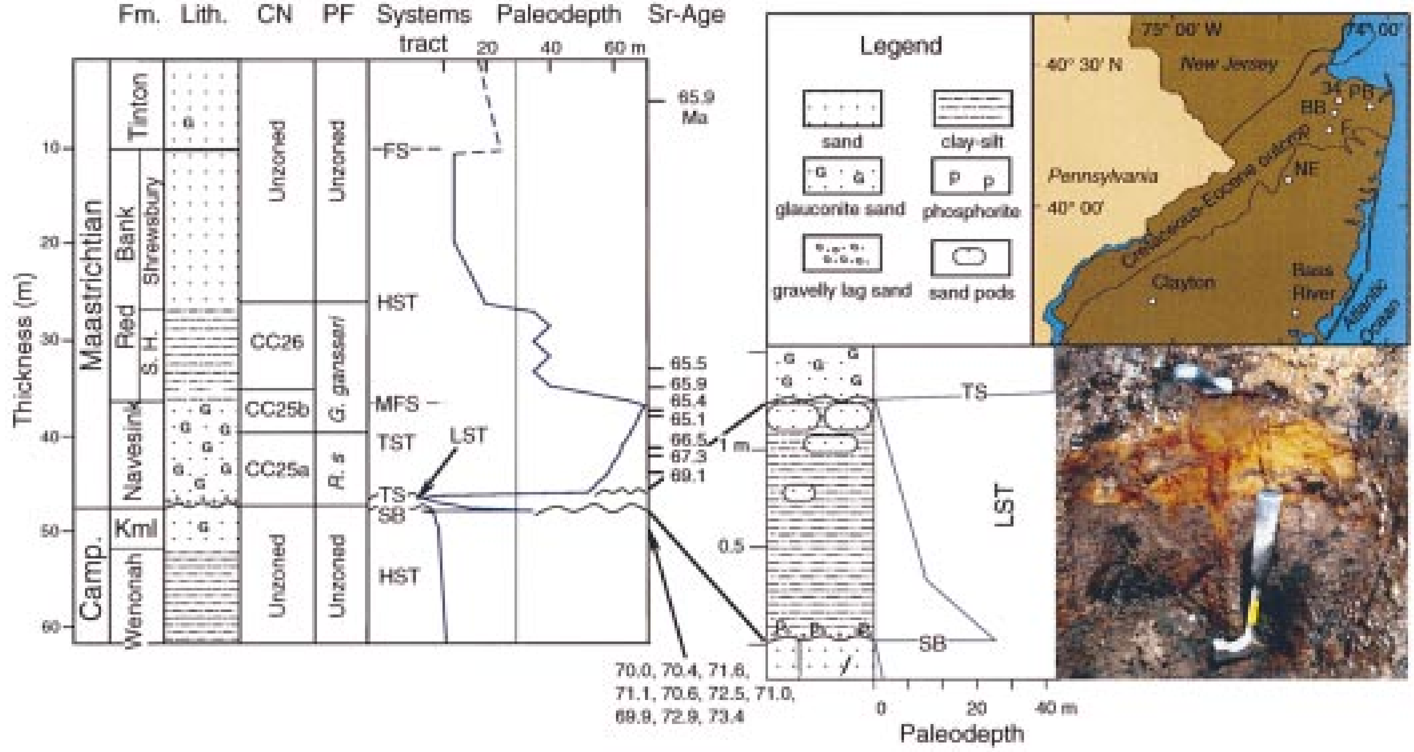

Figure 2. Upper Campanian (Camp.)–Maastrichtian stratigraphy for outcrops at Big Brook (BB), Matawan Route 34 (34), and Poricy Brook (PB). Stratigraphic section (thickness, lithology [Lith.], formation [Fm.], biostratigraphy, and Sr isotopic ages) is modified after Sugarman et al. (1995). Sr isotopic ages (10) from upper Mount Laurel Formation are derived from various coastal plain localities. S.H. = Sandy Hook Member, Kml = Mount Laurel. Lower inset shows enlargement of interval between sequence boundary (SB) and transgressive surface (TS) at Route 34 shown in photo; top right inset is location map (F = Freehold borehole; NE = New Egypt). CN = calcareous nannofossil zones; PF = planktonic foraminiferal zones; *R.s.* = *R. subcircumnodifer* assemblage zone. Systems tracts interpretation: LST = lowstand, TST = transgressive, HST = highstand, MFS = maximum flooding surface, FS = flooding surface.

boundary began shortly after the δ^{18}O increase, implying rapid response to a glacioeustatic lowering. Subsequent transgression during Navesink deposition can be attributed to continued tectonic subsidence.

We estimate that a 40 m eustatic lowering occurred ca. 71 Ma on the basis of the New Jersey record. Benthic foraminifera indicate 60 m of shallowing within the Navesink and the underlying sequence (Olsson, 1991). Although not rigorously backstripped, these water-depth variations correspond to ~40 m of eustatic lowering correcting for water loading, the primary effect in this setting (e.g., Kominz et al., 1998).

A 40 m eustatic lowering ca. 71 Ma is consistent with estimates derived from δ^{18}O studies. The minimum increase observed in planktonic foraminiferal δ^{18}O records is ~0.3‰ (Fig. 1). Taking this as an upper limit for global δ_w changes and using the Pleistocene sea-level/δ^{18}O calibration of 0.11‰/10 m (Fairbanks and Matthews, 1978), a glacioeustatic change would be ~27 m. It is probably not correct to apply the Pleistocene calibration to the Cretaceous; however, the freezing point of permanent ice (–17‰) provides an upper limit for the sea-level/δ^{18}O calibration of 0.055‰/10 m (Miller et al., 1987). Using this as an upper limit, a maximum eustatic change would be ~55 m, although this upper limit is probably not realistic (Miller et al., 1987). An intermediate value of 0.08‰/10 m yields a sea-level change of ~40 m. It is probably not correct to assume that the entire tropical planktonic foraminiferal δ^{18}O signal reflects ice-volume changes, because as much as half of the late Pleistocene tropical δ^{18}O change is due to cooling (Guilderson et al., 1994). Assuming that half of the planktonic δ^{18}O signal is due to cooling yields a lower limit of 20 m.

Thus, the best estimate derived from margin and δ^{18}O records is that sea level dropped by ~20–40 m, equivalent to 25%–40% of the volume of the modern Antarctic ice cap (Savin and Douglas, 1985). This is significantly less than the ~60 m drop estimated by Haq et al. (1987), equivalent to ~100% of the modern Antarctic ice sheet.

The early Maastrichtian sea-level lowering was associated with a major reorganization in deep-water circulation. Barrera et al. (1997) concluded that a coeval early Maastrichtian paleoceanographic event reflected a short-term influx of cool water from a high-latitude source into intermediate depths in the tropical Pacific. They also noted a negative global δ^{13}C excursion that is clearly correlated with a sea-level lowering indicated by New Jersey records. One possible mechanism for this excursion is an increase in the ratio of organic to inorganic carbon in oceanic input due to increased weathering of organic-rich sediments exposed on continental shelves during a sea-level lowering (Barrera, 1994; Barrera et al., 1997).

The development of an ice sheet in the early Maastrichtian may not have been permanent. By 69 Ma and again by 65.6 Ma, benthic foraminifera δ^{18}O values were ~0.25‰, corresponding to temperatures of 8 °C, assuming an ice-free world. These relatively warm deep-water and high-latitude temperatures are likely consistent with the absence of large ice sheets. Similarly, the sequence stratigraphic record shows continuous deposition and little sea-level change over a several million year interval crossing the Cretaceous-Tertiary boundary (Olsson et al., 1997).

CONCLUSIONS

We present here isotopic and sequence stratigraphic evidence for an early Maastrichtian growth of an ice sheet and attendant glacioeustatic lowering. We lack definitive glaciological evidence for this ice sheet, but conclude that isotopic evidence and sequence stratigraphic data are suggestive: an ice sheet equivalent to 25%–40% of the volume of the present-day Antarctic ice cap caused a 20–40 m glacioeustatic lowering at 71 Ma in the earliest Maastrichtian. While we interpret the early Maastrichtian sea-level lowering as glacioeustatic, we are not ready to extend glacioeustasy to intervals that were most probably ice free (e.g., the middle Cretaceous); thus, it is possible that some as-yet unidentified mechanism of eustatic change caused these lowerings.

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