ELSEVIER

Marine Geology 217 (2005) 215–231

MARINE GEOLOGY
INTERNATIONAL JOURNAL OF MARINE GEOLOGY, GEOCHEMISTRY AND GEOPHYSICS

www.elsevier.com/locate/margeo

Visions of ice sheets in a greenhouse world

Kenneth G. Miller*, James D. Wright, James V. Browning

Department of Geological Sciences, Rutgers University, Piscataway, NJ 08854, United States

Accepted 2 March 2005

Abstract

Backstripped eustatic estimates from New Jersey and the Russian platform show large (>25 m) and rapid (<1 my) sea-level changes in the Late Cretaceous to early Eocene (99–49 Ma). The largest of these sea-level events occurred at the Campanian/Maastrichtian boundary (71.5 Ma); we infer that ice growth and attendant sea-level lowering at 71.5 Ma were as great as many Oligocene and younger events (~40 m sea-level change). Glacioeustasy is the only known mechanism that can account for Late Cretaceous to early Eocene rapid changes because other hypothesized mechanisms (steric effects, water storage in lakes, deep-water changes, groundwater, or sea ice) are too slow or too small. In contrast to this evidence for glacioeustasy, ample geological evidence points to warm high-latitude temperatures at this time. The late Cenomanian–early Turonian $\delta^{18}O$ record highlights the enigma. This was the warmest interval of the past 200 my, yet it was bracketed by two inferred eustatic falls of ~25 m that were associated with two large (>0.75‰) deep-sea $\delta^{18}O$ increases (92–93 Ma, mid-Turonian and 96 Ma, mid-Cenomanian). We reconcile records of warm high latitudes with glacioeustasy by proposing that Late Cretaceous–early Eocene ice sheets generally reached maximum volumes of $8–12 \times 10^6$ km^3 (20–30 m glacioeustatic equivalent), but did not reach the Antarctic coast; hence, coastal Antarctica (hence deep water) remained relatively warm even though there were significant changes in sea level as the result of glaciation. Unlike the Oligocene and younger icehouse world, these ice sheets only existed during short intervals of peak Milankovitch insolation, leaving Antarctica ice-free during much of the greenhouse Late Cretaceous to middle Eocene. These results highlight the need to re-evaluate the paradigm that polar ice sheets did not exist during times of warm high-latitude climates.

Keywords: ice sheets; greenhouse world; sea-level changes

1. Background

Various datasets indicate that high latitudes were significantly warmer than at present for much of the Phanerozoic (e.g., Royer et al., 2004). Modelers can explain such warm (>10–15 °C) high latitudes only

* Corresponding author.
E-mail address: kgm@rci.rutgers.edu (K.G. Miller).

0025-3227/$ - see front matter © 2005 Elsevier B.V. All rights reserved.
doi:10.1016/j.margeo.2005.02.007

by invoking high CO_2 levels (>2–4 times present levels; e.g., Royer et al., 2004). Such warm intervals have been termed the greenhouse world (e.g., Fisher, 1984) and they have been presumed to be ice-free. Large ice sheets of the modern Earth (middle to late Cenozoic) have been termed the icehouse world (Fisher, 1984), with previous "icehouses" occurring in the Pennsylvanian–early Permian, Late Ordovician, and Proterozoic.

Despite intense efforts by paleoceanographers, the timing of inception of large ice sheets in both hemispheres remains a poorly understood and constrained problem. The history of Northern Hemisphere Ice Sheets (NHIS), a.k.a. the "Ice Ages", provides a useful example applicable to the understanding of the inception of Antarctic glaciation. Paleoceanographers cite the inception of NHIS as occurring at ca. 2.6 Ma (late Pliocene; Shackleton et al., 1984). However, this "inception" really represents a major growth in amplitude of NHIS. Significant (at least Greenland-sized) NHIS extend back to at least the middle Miocene (ca. 14 Ma; see summary in Wright and Miller, 1996).

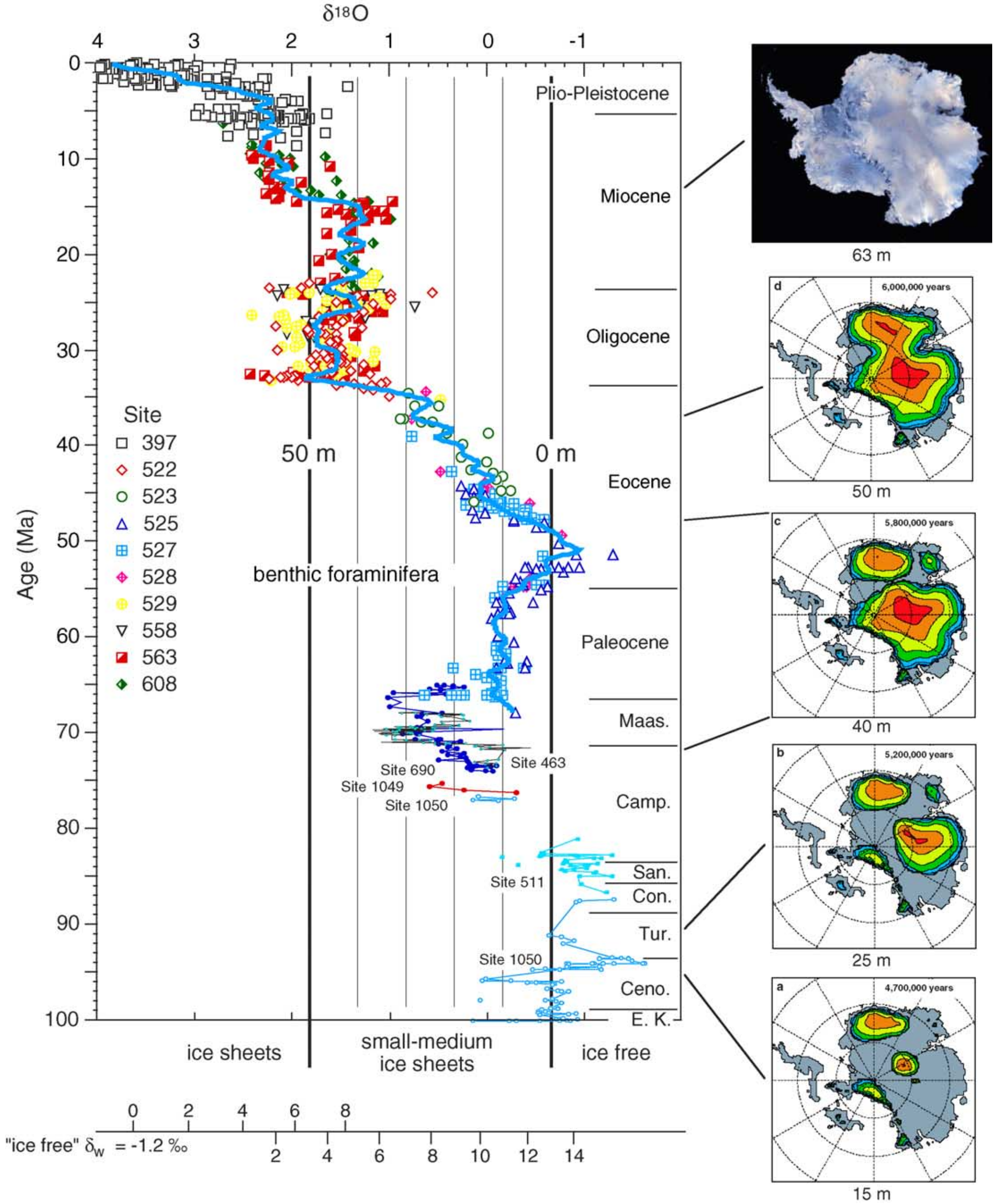
The timing of initial glaciation in Antarctica has been even more debatable. Pioneering studies of Shackleton and Kennett (1975) and Savin et al. (1975) assumed that a continent-sized ice sheet first appeared in Antarctica in the middle Miocene (ca. 15 Ma). Matthews and Poore (1980) and Miller and Fairbanks (1983) questioned this assumption and suggested large ice sheets existed in Antarctica since at least the earliest Oligocene (33 Ma), and subsequent isotopic and glaciological data firmly support this interpretation (Miller et al., 1991; Zachos et al., 1992; see below). Ice-volume changes can be firmly linked to global sea-level changes in the Oligocene and younger icehouse (Miller et al., 1998). Today, most paleoceanographers cite ca. 33 Ma as the inception of modern Antarctic glaciation, though this supposition is now being challenged (Stoll and Schrag, 1996; Miller et al., 1999, 2003).

The pre-Oligocene greenhouse presents a different world from the icehouse. Isotopic, floral, and faunal evidence indicates that high latitudes were warm during the Late Cretaceous to early Eocene (e.g., Shackleton and Kennett, 1975; Savin et al., 1975; Savin, 1977; Shackleton and Boersma, 1981; Miller et al., 1987; Zachos et al., 1996). Benthic foraminiferal $\delta^{18}O$ values indicate that deep-water (and by inference high-latitude surface-water) temperatures were warm (6 to 16 °C; Fig. 1); such warm high-latitude temperatures have been assumed to indicate warm poles, though $\delta^{18}O$ values are equivocal as to the presence of ice sheets in the interior of Antarctica. Oxygen isotopic values place few other constraints on the presence of ice sheets in the greenhouse. A 15–30 m equivalent ice-volume signal would yield a <0.3‰ $\delta^{18}O$ change (see below for discussion) that could be easily masked by normal temperature variations.

There are three major means of evaluating ice sheets: directly from sedimentological evidence, indirectly from $\delta^{18}O$ values, or inferentially from reconstruction of sea level (e.g., Barrett, 1999). We focus in this paper on estimates derived from $\delta^{18}O$ values and sea-level studies, though it is important to evaluate constraints placed by sedimentological evidence that includes tills, diamictons, and ice rafted detritus (IRD). Diamictons are often difficult to date and poorly preserved; in addition, the pre-Oligocene record of the main Antarctic continent is poorly exposed and sampled. Because IRD is found in pelagic sediments that are easy to date using

Fig. 1. Stable isotopic synthesis for the past 100 million years and maps showing maximum sizes of ice sheets during peak glaciation for several intervals. Cenozoic isotopic synthesis is modified using the data compiled by Miller et al. (1987) showing data points and Gaussian fit. The data were recalibrated to the Cande and Kent (1992) time scale using the look-up table of Wei (1994). The smoothed curve was created using methodologies identical to Miller et al. (1987); data were interpolated to constant 0.1 my interval and smoothed with a 21-point Gaussian convolution filter using Igor™ for a Macintosh. Late Cretaceous synthesis is after Miller et al. (2003, 2004) and is based on data from deep-sea oxygen benthic foraminiferal $\delta^{18}O$ records (Site 463, Barrera and Savin, 1999; Site 690, Barrera and Savin, 1999; Site 511, Huber et al., 1995; Sites 1049 and 1050, Huber et al., 2002). Maps are after DeConto and Pollard (2003) and show the amount of equivalent sea-level change proscribed for a given state (e.g., 25 m) and are calibrated to the $\delta^{18}O$ synthesis (correlation lines) using sea-level data shown in Fig. 4. Thick vertical lines are the 50 m and 0 m sea-level equivalent based on early Oligocene and early Turonian ice volumes, respectively; thin vertical lines are 10 m increments assuming linear response between these states. Though probably an oversimplification, this assumption is consistent with maximum ice volume for each state shown on the right (e.g., 40 m lowering at 71.5 Ma is consistent with the scaling indicated by $\delta^{18}O$ values).

biostratigraphy, it often provides the strongest evidence for the presence of large ice sheets calving into the ocean. For example, the finding of 33 Ma IRD at 58°S by Ocean Drilling Program Leg 120 (Zachos et al., 1992) finally convinced much of the paleoceanographic community of the presence of large Oligocene ice sheets in Antarctica (summaries in Miller et al., 1991; Zachos et al., 1994). Nevertheless, IRD has proven to be an elusive marker and its absence in the stratigraphic record cannot be readily taken as an indicator of ice-free conditions on the Antarctic continental margin, let alone the



continental interiors. For example, despite numerous expeditions to the Antarctic region by the Deep Sea Drilling Project (DSDP) in the 1970s, little pre-late Miocene IRD was found (see summary in Tucholke et al., 1976). Lower to middle Eocene IRD has been reported near Antarctica, but the importance of this has been largely ignored (see summary in Wise et al., 1991). We conclude that diamictons and well-dated IRD provide evidence for the presence of ice sheets (though not necessarily variations in ice-sheet size; see review of Barrett, 1999). In contrast, their absence does not indicate an ice-free continent, especially considering the poorly sampled stratigraphic record near Antarctica.

In this paper, we briefly review evidence for large, rapid sea-level changes that suggest glacioeustatic variations in the greenhouse world. We reconcile this apparent enigma by: 1) re-evaluating stable isotopic results; 2) comparing stable isotopes with modeling results that illustrate the geographic coverage of Antarctica under different thermal states and sea-level positions; and 3) providing a new vision of the scale and tempo of ice-volume changes in a greenhouse world. We conclude that in Late Cretaceous–early Eocene world, ice sheets were restricted in area in Antarctica, ephemeral, and paced by Milankovitch forcing.

2. The sea-level record

Sea-level studies of continental margins have long challenged our view of an ice-free planet. Pioneering work by Exxon Production Research Company (EPR) reported extremely large (up to 400 m) and rapid (<1 my) sea-level changes throughout the Phanerozoic (Vail et al., 1977). The only known mechanism for causing sea-level changes in excess of 10 m in less than 1 my is glacioeustasy (Pitman and Golovchenko, 1983). EPR reduced its eustatic estimates to less than 160 m, but still showed up to 110 major eustatic lowerings during the past 250 my (Haq et al., 1987), most of which could only be explained by glacioeustasy. Other researchers have viewed these estimates either as gospel or with a jaundiced eye (e.g., Miall, 1991). The most fundamental criticism is that the EPR sea-level estimates are not reproducible because they were based on proprietary data. Academic studies of outcrops and boreholes (e.g., Hancock, 1993; Sahagian et al., 1996; Miller et al., 1996, 1998) appear to corroborate greenhouse sea-level changes proposed by EPR. However, the amplitudes of eustatic change published by EPR appear to be much to high and independent studies are needed to determine the timing and scale of greenhouse sea-level changes.

The geological record contains a rich history of transgressions (landward movement of the strand line) and regressions, but a global sea-level record can only be obtained by accounting for the effects of subsidence/uplift, sediment loading, and other effects on the stratigraphic record. The effects of subsidence and sediment supply can be evaluated through a method of one-dimensional inverse models termed backstripping (e.g., Watts and Steckler, 1979). Backstripping first removes the effect of compaction and sediment loading from observed basin subsidence. By assuming thermal subsidence on a passive margin, a portion of tectonic subsidence can be removed. The difference between observed subsidence and a best-fit theoretical thermal curve (termed R2 for second reduction; Bond and Kominz, 1984) is the result of either eustatic change or any subsidence unrelated to two-dimensional passive margin subsidence (Kominz et al., 1998).

Previous outcrop studies have suggested rapid sea-level changes in the Late Cretaceous, though there have been very few backstripped records. Studies of western European Upper Cretaceous sections (Hancock, 1993) have documented fairly rapid transgressions and regressions, though the influence of tectonic and sedimentation changes have not been taken into account in these records. U.S. Western Interior Upper Cretaceous sections are generally overprinted by compressional tectonics, through it may possible to extract a sea-level estimate through backstripping in the future. The most complete Upper Cretaceous backstripping analyses come from studies of the Russian platform (Sahagian et al., 1996) and New Jersey records shown here (Miller et al., 2002, 2003; Kominz et al., 2003; Van Sickel et al., 2004). The Russian platform backstripped records demonstrate that Late Cretaceous sea-level changes were on the order of tens of m and occurred within <1 my. However, their temporal resolution is relatively coarse and additional studies are needed to compare with their records.

The Ocean Drilling Program (ODP) began to assemble icehouse records of sea-level change as part of a global campaign to understand eustatic variation. Oligocene–Miocene strata on the New Jersey margin (Fig. 2) were targeted for the initial efforts for offshore (Legs 150 and 174A; Mountain et al., 1994; Austin et al., 1998) and onshore drilling (Legs 150X and 174AX; Fig. 1) because of the margin's simple passive margin tectonics (Kominz et al., 1998) and abundant data available for planning coreholes. Backstripping of these relative sea-level changes provides a eustatic estimate by accounting for the effects of compaction, loading, and thermal subsidence (e.g., Steckler and Watts, 1978). Application of backstripping to the New Jersey records (Kominz et al., 1998, 2002) indicates that large (15–80 m) and rapid sea-level changes occurred during the icehouse world of the Oligocene–Miocene; these sea-level falls have been directly linked to $\delta^{18}O$ increases providing prima facie proof of a causal link (Miller et al., 1996, 1998).

Recent onshore drilling in the New Jersey coastal plain (ODP Legs 150X and 174AX; Fig. 1) has yielded a record of relative sea-level fluctuations for

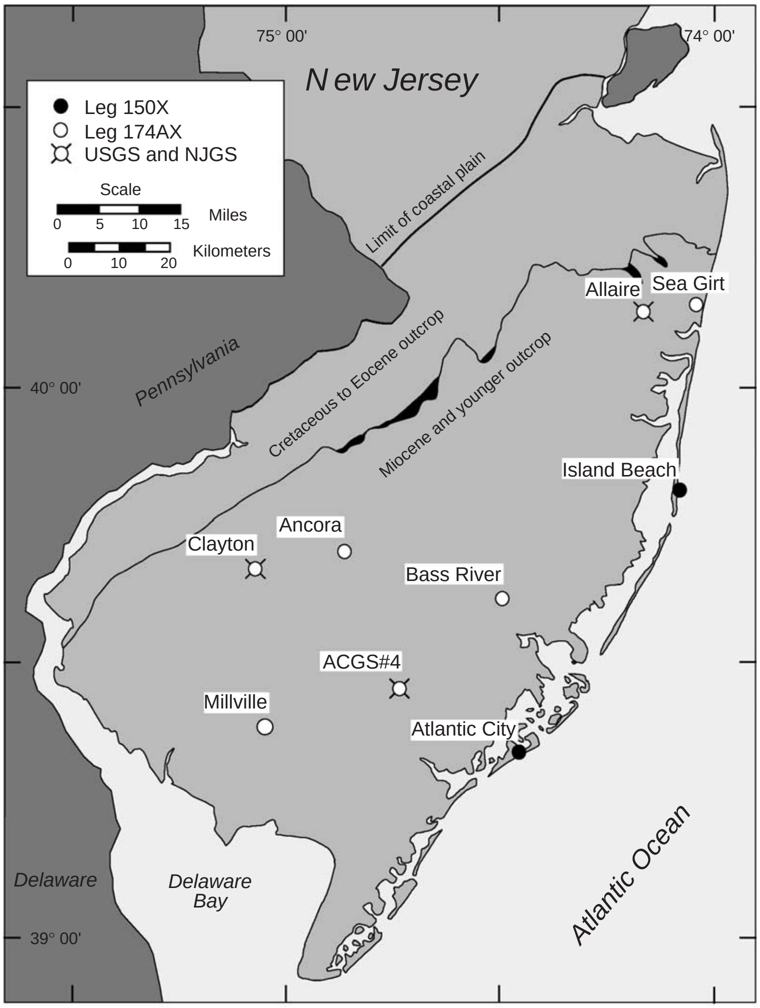


Fig. 2. New Jersey location map showing sites used in evaluating Eocene (Island Beach, Clayton, ACGS#4, Atlantic City, and Allaire) and Late Cretaceous (Ancora and Bass River) sea-level changes. Millville and Sea Girt are recently drilled sites that will be used to test the sea-level records.

the greenhouse world of 100–49 Ma (Browning et al., 1996, 1997; Miller et al., 2003, 2004). Browning et al. (1996, 1997) presented Eocene results from drilling at Island Beach, Allaire, Clayton, ACGS#4, and Atlantic City boreholes (Figs. 2 and 3). Studies of Ancora and Bass River (Fig. 2) have provided a firm and testable chronology for Late Cretaceous sequences (Fig. 4). By integrating Sr-isotopic and biostratigraphic studies, Miller et al. (2003, 2004) dated 14 Upper Cretaceous sequences with an age resolution of ± 0.5 my (Fig. 5). They showed (Miller et al., 2004) that the timing of Late Cretaceous sequence boundaries in New Jersey is similar to sea-level lowerings of EPR and the Russian Platform (Fig. 4). By dating Late Cretaceous sequence boundaries, Miller et al. (2003, 2004) provided a chronology of inferred eustatic falls (i.e., assuming minimal tectonic effects on this passive margin).

Kominz et al. (2003) and Van Sickel et al. (2004) provided backstripped eustatic estimates for the Late Cretaceous and younger strata at Ancora and Bass River. They used one-dimensional backstripping to remove the effects of compaction and sediment loading from observed basin subsidence (e.g., Watts and Steckler, 1979). Backstripping studies show that simple thermal subsidence, sediment loading, and compaction are the dominant causes of subsidence in the New Jersey coastal plain (Kominz et al., 1998; Van Sickel et al., 2004). Accommodation space in the New Jersey coastal plain is dominated by the flexural response to sediment loading of the stretched crust seaward of the basement hinge zone and subsidence is exponential in form (Steckler, 1981). Kominz et al. (1998, 2002) and Van Sickel et al. (2004) obtained a eustatic estimate (R2 or second

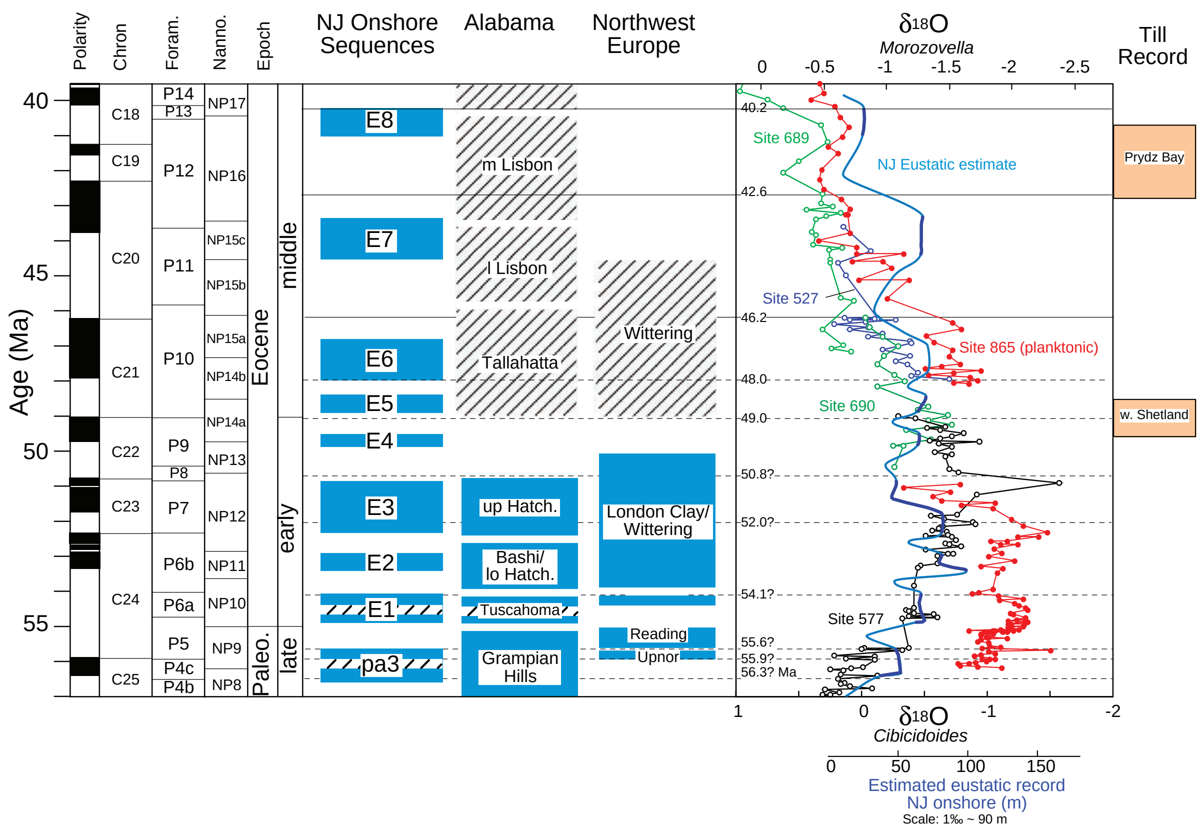


Fig. 3. Comparison of Eocene onshore sequences (New Jersey, Alabama and northwest Europe), oxygen isotopes, NJ backstripped estimates, and till records in high southern latitudes. Modified after Miller et al. (1998) using backstripped estimates of Kominz et al. (1998) (solid blue) and scaling the lowstands (hiatuses onshore) to the oxygen isotopic record. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

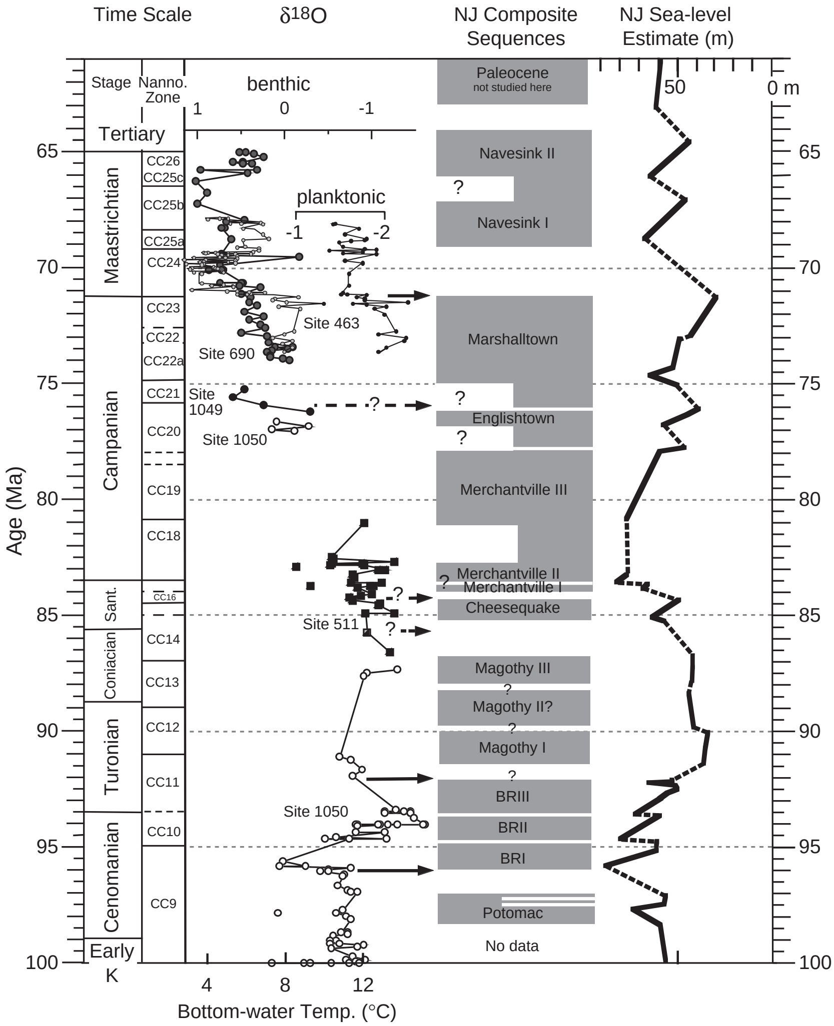


Fig. 4. Comparison of Late Cretaceous deep-sea benthic foraminiferal δ^{18}O records (Site 463, Barrera and Savin, 1999; Site 690, Barrera and Savin, 1999; Site 511, Huber et al., 1995; Sites 1049, 1050, Huber et al., 2002), planktonic foraminiferal δ^{18}O records (Site 463, Barrera and Savin, 1999), New Jersey composite sequences, the best eustatic estimate derived from backstripping Ancora and Bass River, the relative sea level curve from NW Europe (thin gray continuous line; Hancock, 1993), the backstripped record from the Russian platform (black continuous line; Sahagian et al., 1996). Arrows indicate positive δ^{18}O inflections (inferred cooling and/or ice volume increases) and inflections in the European sea-level curves. Temperatures were computed assuming an ice-free world ($\delta_w = -1.2$‰ PDB; Shackleton and Kennett (1975), with *Nuttallides* values of -0.76‰ relative to equilibrium (Pak and Miller, 1992), and the paleotemperature equation in Barrera and Savin (1999). For the composite: blue boxes indicate time represented, white areas indicate hiatus, and thin white lines indicate inferred hiatuses. E.K. = Early Cretaceous. Modified after Miller et al. (2004).

reduction; Bond and Kominz, 1984) by subtracting a best-fit thermal subsidence history from basin subsidence. Both Ancora and Bass River R2 curves are very similar, indicating that effects of thermal subsidence, loading, and water-depth variations have been successfully removed. The rates of these sea-level changes are 15–40 m/my. Miller et al. (2004) noted that backstripping, seismicity, seismic stratigraphic data, and distribution patterns of sediments all indicate minimal tectonic effects on the Late Cretaceous to Tertiary New Jersey coastal plain. Having eliminated tectonics as a source of these events, the only mechanism than can explain such rapid rates is glacioeustasy.

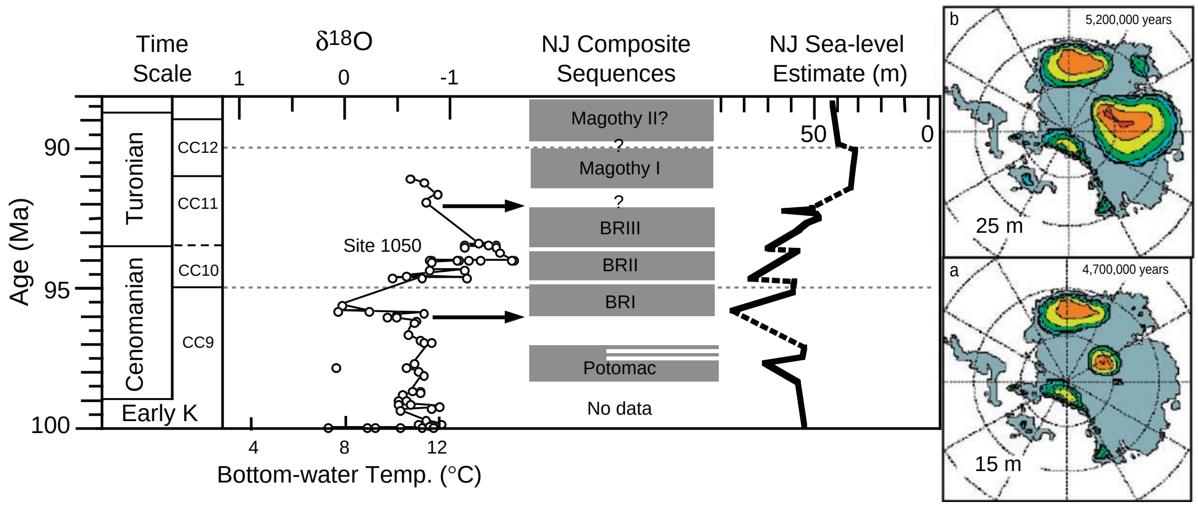


Fig. 5. Enlargement of Fig. 4 focusing on the peak warmth interval of the Cenomanian to Turonian. Shown at right are maps of 15 and 25 m equivalent ice volumes after DeConto and Pollard (2003).

The timing and amplitude of New Jersey sea-level record is very similar to the Russian platform backstripped record (Fig. 4; Sahagian et al., 1996), strengthening our interpretation of the New Jersey curve as a eustatic record. Six early Late Cretaceous eustatic changes from the Russian platform (Sahagian et al., 1996) are similar to New Jersey; two events are not discernable in the Russian platform record which also does not extend younger than 83 Ma. Fourteen of 16 eustatic lowerings reported by EPR (Haq et al., 1987) show correlative events (within ± 0.5 my) in New Jersey (Fig. 4), although the EPR amplitudes of change are much higher. Five to six later Late Cretaceous sequences in Northwest Europe (Hancock, 1993) appear to correlate with sequences in New Jersey, though limited age control, time scale problems, and the lack of backstripping in the NW Europe data precludes closer comparison. The correspondence amongst these records indicates a eustatic control on Late Cretaceous sequence boundaries; the two backstripped estimates (NJ and the Russian Platform) requires rates far in excess of 10 m/my.

High rates of sea-level change (>10 m/my) can only be explained by the growth and decay of continental ice sheets. The alternative to invoking Late Cretaceous ice sheets is that global sea-level changes were paced by as yet undefined mechanisms, because none of the other hypothesized mechanisms (steric effects, storage in lakes, deep-water changes, groundwater, or sea ice) can explain the observed 15–40 m changes in <1 my (Miller et al., 2004). Temperature change can explain rapid sea-level changes, but the effect is small (i.e., a 10 °C global warming would only cause a 10 m sea-level rise; Pitman and Golovchenko, 1983). Lakes can only explain at most 5 m of sea-level change (Sahagian et al., 1996). Opening and closing of Mediterranean basins could explain ~10+ m virtually instantaneously, but it is impossible to explain the number of Late Cretaceous to early Eocene sea-level events with this mechanism. Only ice-volume changes can explain the large (tens of meters) sea-level changes of the Cretaceous to early Eocene. These changes are robust and cannot be attributed to uncertainties in backstripping. The major uncertainty in backstripping derives from paleowater depth estimates which indicate typically 40–60 m in water depth changes (see Miller et al., 2004 for full documentation); even if paleodepths are in error by a factor of two, this still requires $\gg 10$ m of eustatic change in less than 1 my.

Sea-level studies thus pose a fundamental enigma because they predict glacioeustatic changes during intervals when the world was presumed to be ice-free. Matthews and colleagues (Matthews and Poore, 1980; Matthews, 1984; Prentice and Matthews, 1988; Matthews and Frohlich, 2002) recognized this enigma and continually challenged the paleoceanographic and paleoclimatology communities to rethink their hypotheses.

3. Benthic foraminiferal δ^{18}O records

Oxygen isotopic composition of marine carbonates provides a proxy to infer ice-volume changes. The δ^{18}O values of marine carbonates is an imperfect ice-volume proxy because it reflects temperature changes, global ice volume, and local evaporation/precipitation effects. Shackleton and Kennett (1975) and Savin et al. (1975) first reasoned that deep-sea benthic foraminiferal δ^{18}O records (e.g., Fig. 1) should reflect high-latitude temperature and ice-volume variations because modern deep and bottom waters (>2000 m) form in high latitudes, though influenced by low-latitude, high salinity water masses (e.g., Mediterranean Outflow Water and the North Atlantic Current; Worthington, 1970; Reid, 1979). Their conjecture was supported by the Cenozoic parallelism between benthic and high latitude planktonic foraminiferal δ^{18}O records and by the fact that the highest δ^{18}O values are in the deep sea and high-latitude surface waters. There has been considerable discussion given to the notion that warm, salty deep waters (WSDW) formed in low latitudes may have played a more important role during past warm climates such as the Late Cretaceous–Eocene (e.g., Chamberlain, 1906; Brass et al., 1982). However, evidence for such WSDW is surprisingly lacking. In fact, seismic stratigraphy (Mountain and Miller, 1992), carbon isotopic reconstructions (Pak and Miller, 1992), and the distribution of hiatuses (Kennett, 1977; Wright and Miller, 1993) all indicate that the Southern Ocean was the predominant deep-water source during the early Paleogene greenhouse world and that WSDW played a minor role.[1] Though data on Cretaceous oceans are more limited, carbon isotopic reconstructions and comparison of Southern Ocean surface δ^{18}O values with deep-sea benthic foraminiferal δ^{18}O values indicate that southern latitudes were also the main source for later Cretaceous deep waters (Barrera and Savin, 1999).

Benthic foraminiferal δ^{18}O values can be used to indicate the presence of large ice sheets for the Oligocene–Recent (see below; Miller and Fairbanks, 1983; Miller et al., 1987, 1991), although the precise contribution of ice-volume versus temperature changes to δ^{18}O variations is still not known within 50% (Miller et al., 1987). Benthic foraminiferal δ^{18}O values reflect the very large deep-sea reservoir and are not influenced by local changes in δ^{18}O values that potentially overprint planktonic δ^{18}O records (see discussion in Miller et al., 1991). Miller and Fairbanks (1983) and Miller et al. (1987) noted that high deep-water δ^{18}O values (>1.8‰ in *Cibicidoides*) are incompatible with an ice-free world. The derivation of the "1.8‰ line" (equivalent to the 50 m line in Fig. 1) is straightforward. Modern Atlantic *Cibicidoides* values are ~2.7–2.8‰ (offset from equilibrium by −0.64‰; Shackleton and Hall, 1984) with ambient temperatures of ~2 °C; if the modern ice sheets were melted, mean ocean δ^{18}O values would be ~0.9‰ lower (Shackleton and Kennett, 1975). Thus, δ^{18}O values >1.8‰ in *Cibicidoides* require either deep-water temperatures colder than today's or the presence of large ice sheets. *Cibicidoides* δ^{18}O values regularly exceeded 1.8‰ beginning in the earliest Oligocene (Fig. 1), indicating an icehouse world (Miller et al., 1991). These isotopic arguments were validated by glaciomarine evidence (Miller et al., 1991; Zachos et al., 1992) that proved that large ice sheets have typified much of the Oligocene and younger icehouse. In the icehouse world, ice-volume, sea-level, and δ^{18}O variations occurred in lock step (Miller et al., 1996).

In the greenhouse world, low benthic foraminiferal δ^{18}O values indicate very warm deep warm deep-water temperatures and relatively small ice volumes. For example, minimum benthic foraminiferal δ^{18}O values of −1.0‰ and −1.3‰ occurred in the early Eocene and late Cenomanian respectively, corresponding to bottom-water temperatures of 14 and 15.5 °C.[2] If ice sheets were present at these times, even warmer deep-

[1] Thomas (2004) presented Nd evidence for a North Pacific deep-water source in the Paleogene, but this evidence is contradicted by the fact that Pacific deep-water has the lowest δ^{13}C values of any ocean basin and thus was probably furthest from the source (Pak and Miller, 1992).

[2] Temperatures were computed assuming an ice-free world, $\delta_w = -1.2$‰ PDB; Shackleton and Kennett (1975), with *Nuttallides* values of −0.76‰ relative to equilibrium (Pak and Miller, 1992), and the paleotemperature equation in Barrera and Savin (1999).

water temperatures are required. We assume that the world was ice-free when benthic foraminiferal δ^{18}O values were extremely low (approximately -0.6‰, 13 °C;=the 0 ice volume line in Fig. 1); otherwise benthic foraminiferal δ^{18}O values place few constraints on the presence of ice sheets.

Warmth does not mean stable or uniform conditions as indicated by large (>1‰) benthic foraminiferal δ^{18}O variations during the Late Cretaceous to early Eocene greenhouse world (Figs. 1, 3–5), with some remarkably high δ^{18}O values (exceeding 1.0‰, Fig. 1). Oxygen isotopic values less than 1.8‰ in *Cibicidoides* do not require the presence of large sheets, though the highest δ^{18}O values are strongly suggestive of ice sheets during the coolest period (~7 °C deep waters) of the Late Cretaceous, the earliest Maastrichtian. Still, the low δ^{18}O values for most of the Late Cretaceous to early Eocene requires deep-water temperatures greater than 8 °C. Such warm temperatures in deep-water source regions indicate that Antarctic ice sheets, if present, remained well inland, having little to no influence on coastal regions. Nevertheless, there appears to be a link between δ^{18}O increases and sequence boundaries during the greenhouse Late Cretaceous–Eocene, just as there is in the icehouse of the Oligocene–Recent.

Eocene δ^{18}O records are coarsely sampled (Fig. 3), though it is clear that fairly large (>0.5‰) δ^{18}O increases occurred as three steps in the early middle Eocene (Chron C21r starting at 49 Ma, C21n starting at 48 Ma, and Chron C20 starting at 46.2 Ma) and two in the late middle Eocene (42.6 and 40.2 Ma). All five of these increases correlate with hiatuses/sequence boundaries on the New Jersey coastal plain (Fig. 3), suggesting that sea-level lowerings were caused by glacioeustasy (Browning et al., 1996, 1997). The early Eocene δ^{18}O record is less clear. Several increases in planktonic δ^{18}O records from Site 865 (Fig. 3) appear to correlate with New Jersey sequence boundaries (Fig. 3). The benthic foraminiferal record is too coarsely sampled and/or the changes apparently too small to be detected (Fig. 3). Herein lies a fundamental problem with trying to detect glacioeustatic changes in a greenhouse world: a small (15 m) to moderate (25 m) ice sheet would only affect global δ^{18}O values by 0.15–0.25‰ (Table 1; see below for discussion) unless the δ^{18}O signal was amplified by attendant cooling.

Isotopic studies of the Late Cretaceous are still in their infancy because there are few suitable Upper Cretaceous sections due to problems in recovery and diagenesis of deeply buried sections (i.e., isotopic studies of sections with >400 m burial are suspect; Miller et al., 1987). Recent studies provided the first moderately detailed sections that adequately sampled the Cenomanian to lower Turonian (Site 1050, 100–91 Ma; Huber et al., 2002) and the upper Campanian–Maastrichtian (Site 690 and 463; 74–65 Ma; Barrera and Savin, 1999) (Figs. 1, 4, 5). However, much of the late Turonian to late Campanian (91–74 Ma) is still poorly sampled and few meaningful comparisons can be made for this part of the Late Cretaceous.

The Campanian/Maastrichtian boundary is remarkable in the context of isotopic records of the last 100 my (Fig. 1). Benthic and planktonic foraminiferal δ^{18}O values record increases of 1‰ and 0.6‰, respectively, across this boundary, and benthic foraminiferal δ^{18}O values exceed 1.0‰ in the Maastrichtian (Barrera and Savin, 1999). Such high values were not reached again until the Oligocene and require deep-water temperatures less than 6 °C (assuming an ice-free world) to 8 °C (assuming an ice sheet ~40% of modern). Miller et al. (1999, 2003, 2004) noted that

Table 1

Calibration					
δ^{18}Oseawater‰/10 m	0.0110	0.0100	0.0080	0.0075	0.0055
Mean ice	−44‰	−40‰	−32‰	−30‰	−22‰
Sea level	δ^{18}Oseawater	δ^{18}Oseawater	δ^{18}Oseawater	δ^{18}Oseawater	δ^{18}Oseawater
15 m	0.17‰	0.15	0.12	0.11	0.08
20 m	0.22	0.20	0.16	0.15	0.11
25 m	0.28	0.25	0.20	0.19	0.14
30 m	0.33	0.30	0.24	0.23	0.17
35 m	0.39	0.35	0.28	0.26	0.19

the Campanian/Maastrichtian δ^{18}O increase correlated with the most dramatic of the Late Cretaceous sequence boundaries in New Jersey (=basal Navesink sequence) and was most likely associated with growth of a fairly large ice sheet.

Mid-Cenomanian and mid-Turonian δ^{18}O increases are intriguing because they bracket the warmest interval of the past 200 my, the late Cenomanian to early Turonian. A 1‰ mid-Cenomanian benthic foraminiferal δ^{18}O increase is well defined at Site 1050 (Fig. 5) and correlates remarkably well with a major sequence boundary in New Jersey separating Bass River from Potomac sequences (Fig. 5). The mid-Turonian δ^{18}O increase is less well defined due to a coring gap at Site 1050, though it is in excess of 0.75‰ (Fig. 5). It also appears to correlate with a major sequence boundary/hiatus in New Jersey separating the Bass River and Magothy sequences (Fig. 5). The ~1‰ δ^{18}O increases at 92–93 and 96 Ma cannot be entirely attributed to ice-volume changes because this would require ice sheets larger than those of modern times, and in fact much of the δ^{18}O signal must be attributed to deep-sea (hence inferred high latitude) temperature change. The challenge is to put constraints on how much of the signal is due to temperature.

Comparison of global sea-level estimates from New Jersey with δ^{18}O records allows us to estimate partitioning between ice volume and temperature in foraminiferal records, albeit tempered by the relatively coarse sampling resolution of Late Cretaceous δ^{18}O records. In the greenhouse world, warm deep-water temperatures and relatively small ice volumes preclude any unequivocal statement about ice volume based solely on the δ^{18}O values. However, assumptions about thermal history can be made and an ice-volume estimate can be obtained from δ^{18}O, and these estimates can be tested against sea-level/ice volume estimates provided by continental margin records. For the Campanian/Maastrichtian sequence boundary in New Jersey, Miller et al. (1999, 2004) inferred a eustatic lowering of ~40 m.[3] For the mid-Turonian (ca. 92 Ma) lowering, Miller et al. (2004) estimated 25–30 m of lowering (Fig. 5), whereas Sahagian et al. (1996) estimated ~45 m for this event in the Russian platform. The mid-Cenomanian event has a minimal eustatic lowering of ~25–30 m at Ancora (Fig. 5; Miller et al., 2004) and 25 m in Sahagian et al. (1996).

Calibration of sea level to δ^{18}O variations introduces another source of uncertainty because the late Pleistocene δ^{18}O-sea level calibration of 0.11‰/10 m (Fairbanks and Matthews, 1978) is probably not appropriate for a warmer planet with less Raleigh fractionation. While the average isotopic composition of past ice sheets is unknown, ice sheets during greenhouse intervals must have had higher δ^{18}O values than the late Pleistocene ice sheets. Miller et al. (1987) noted that permanent ice sheets today are less than −20‰, suggesting a minimum limit of approximately 0.055‰/10 m. Pekar et al. (2002) obtained an Oligocene calibration of 0.1‰/10 m, similar to the calibration assumed by DeConto and Pollard (2003). Greenhouse ice sheets would have had higher δ^{18}O values. Ice sheets with higher mean δ^{18}O values have less influence on the mean ocean δ^{18}O value per unit sea level change than their counterparts with much lower δ^{18}O values (Table 1). We derive relationships between the glacioeustatic δ^{18}O and sea-level changes by dividing the possible mean isotopic composition of ice sheets by the mean depth of the ocean (Table 1) and tabulate global changes in seawater δ^{18}O values for several different glacioeustatic lowerings (Table 1).

For the Late Cretaceous, if we assume mean ice sheet δ^{18}O values were approximately −30‰, then a calibration of 0.075‰/10 m yields a maximum mean ocean δ^{18}O value of ~0.3‰ for the 40 m Campanian/Maastrichtian glacioeustatic lowering (Table 1). The Campanian/Maastrichtian δ^{18}O increases would then correspond to a deep-water cooling of 3 °C and a tropical surface-water cooling of ~1 °C. Similarly, the ~1‰ mid-Cenomanian δ^{18}O increase is associated with a minimal eustatic lowering of ~25–30 m and a deep-water cooling of 3 °C. In both cases, the δ^{18}O signal consists of approximately 25–33% ice volume and 66–75% temperature (vs. the Pleistocene where the signal is 66% ice volume; Fairbanks, 1989).

[3] Lowstands were generally not sampled and thus amplitude estimates are minima. For example, the ca. 71.5 Ma lowering was estimated as 40 m based on: 1) a ~20 m eustatic drop occurred during deposition of the previous sequence from ca. 74 to 71 Ma at both Ancora and Bass River; 2) an unknown amount of lowering occurred during the hiatus ca. 71–69 Ma hiatus; and 3) sea-level rose by >20 m by the time deposition resumed.

4. Comparison with modeling results

DeConto and Pollard (2003) provided a coupled Global Climate Model (the Genesis 2.1 GCM) with an ice sheet model to evaluate the influence of atmospheric CO_2 opening on the development of a large Antarctic ice sheet. Their ice sheet model accounts for bedrock loading, lapse rate effects of topography, surface mass balance, basal melting and ice flow and computes both the sea-level and δ^{18}O effects of ice growth over 10 my of model run. Though computed for Oligocene continental configurations, the modeling results are applicable to the older record as long as Antarctica was a polar continent (i.e., for the entire interval considered here). They used an empirical calibration of 0.1‰/10 m for δ^{18}O/sea level change, a reasonable estimate for Oligocene ice sheets, but probably too high for greenhouse ice sheets considering the discussion above (Table 1). They concluded that the Antarctic ice sheet formed in response to critical levels of CO_2 (DeConto and Pollard, 2003). Four map views of Antarctic ice volume (Fig. 1) of DeConto and Pollard (2003) are shown that correspond to glacioeustatic lowerings of 15, 25, 40, and 50 m and illustrate the following.

1) At greater than 3 times modern CO_2 (840 ppmv), small isolated ice caps form in the highest elevation of Dronning Maud land, the Gamburtsev Plateau, and the Transantarctic Mountains (e.g., the ice configuration corresponding to ~15 m lowerings; Fig. 1; DeConto and Pollard, 2003). DeConto and Pollard (2003) provide sea-level calibrations for each of their maps; the map for 15 m (Fig. 1) illustrates the computed distribution of ice sheets that are equivalent to 15 m sea-level falls estimated from backstripping in Fig. 4 (e.g., sequence boundaries at the bases of BRII and BRIII of the late Cenomanian; Fig. 5).
2) At ~3 times modern CO_2, a small ice sheet threshold is reached owing to height/mass balance feedback; the 3 ice sheets continue to grow, but not coalesce. All three ice sheets were isolated from the coast (DeConto and Pollard, 2003). This configuration is equivalent to ~25 m eustatic lowerings (DeConto and Pollard, 2003). Thus, maps of this ice-sheet configuration (Figs. 1 and 5) are appropriate for the mid-Cenomanian and mid-Turonian eustatic lowerings that we compute to be ~25 m. We suggest that such sea-level falls were typical for the larger Late Cretaceous to Eocene sea-level lowerings (Figs. 3 and 4).
3) The ice caps begin to coalesce with about 40 m eustatic lowering (DeConto and Pollard, 2003). We suggest that this configuration was achieved only once during the Late Cretaceous with the Campanian/Maastrichtian event (Fig. 1).
4) At <2.8 times modern CO_2, a continental scale ice cap forms as the 3 nodes unite, lowering sea level by over 50 m (DeConto and Pollard, 2003). We suggest that this configuration was not achieved until the early Oligocene. It was at this point that calving ice, IRD, and grounded diamictons become common in the record (Miller et al., 1991; Zachos et al., 1992, 1994) and the world had truly entered into the icehouse.

Under true icehouse conditions, the ice sheet reaches the coast, limiting its growth (DeConto and Pollard, 2003). At this point, the Antarctic ice sheet begins to be an influence on, not just a response to, global climate change. The ice sheet stimulates a strengthening of Antarctic Bottom Water Formation that is teleconnected throughout the deep ocean as observed in the early Oligocene (Kennett, 1977; Wright and Miller, 1993). Ice volume changes reflected as δ^{18}O variations also begin to respond strongly to the 41 ky tilt cycle (Zachos et al., 1997); prior to the later Eocene, the response of δ^{18}O variations and small- to medium-sized ice sheets under greenhouse conditions was mainly in the eccentricity band (e.g., the 2.4 my very long eccentricity cycle and the 405 ky long eccentricity cycle as noted by DeConto and Pollard, 2003).

5. A new vision of a Greenhouse world

We propose a fundamentally different vision of Earth's cryospheric evolution that reconciles warm, generally ice-free poles with cold snaps that resulted in glacioeustatic lowerings. Based on the models of DeConto and Pollard (2003), Late Cretaceous–early Eocene ice sheets reached maximum volumes of 8–

12×10^6 km^3 (20–30 m glacioeustatic equivalent), with the exception of the large Campanian/Maastrichtian event, which reached 17×10^6 km^3 (40 m glacioeustatic equivalent). With possibly the latter exception, ice sheets did not reach the Antarctic coast, which reconciles ice with relative warmth in coastal Antarctica (Fig. 1) including terrestrial floral data.

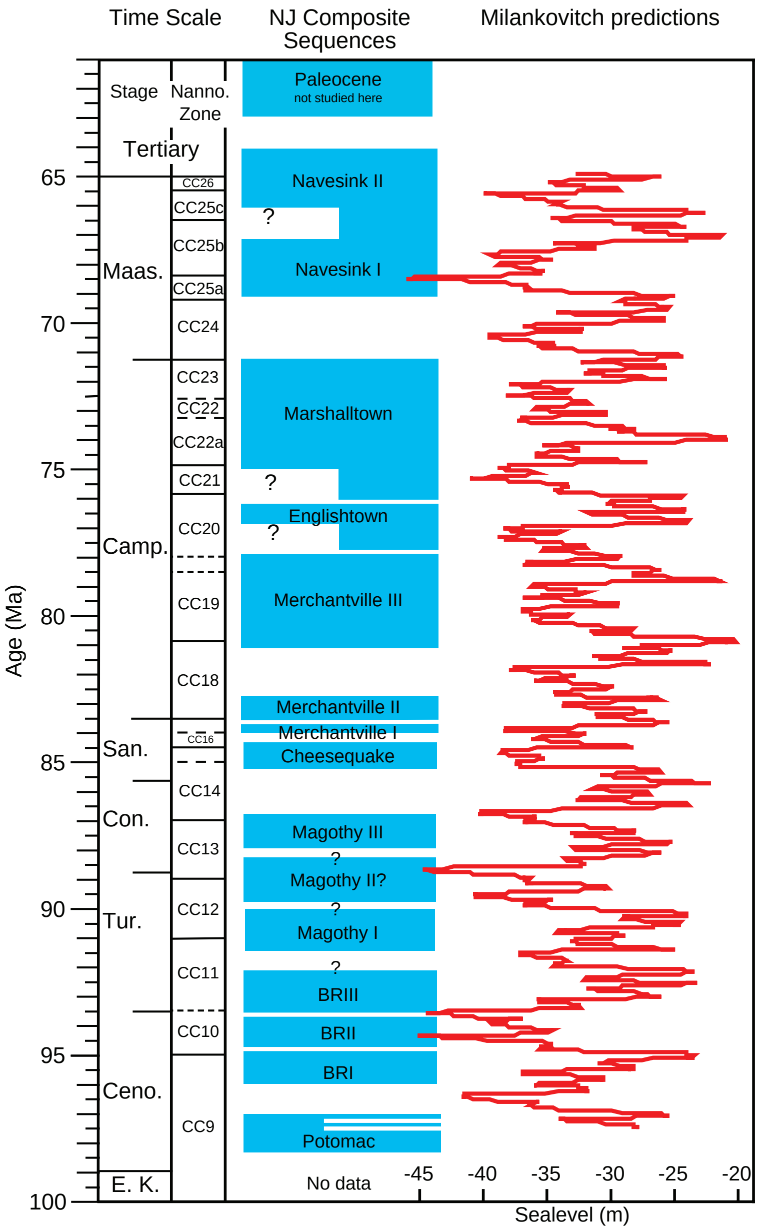


Fig. 6. Comparison of the New Jersey composite Late Cretaceous sequences (Fig. 4) with astronomically based glacioeustatic predictions of Matthews and Frohlich (2002). Note that the sea-level scale is inverted relative to Fig. 4, with lowerings to the left. This figure illustrates that every hiatus/sequence boundary is associated with a major predicted sea-level fall. The predominant beat in the predicted curve is the ca. 2.4 my eccentricity cycle.

Our vision explains terrestrial floral data that indicate warmth in Antarctica (e.g., Francis, 1999) and the presence of Cretaceous dinosaurs in Antarctica (e.g., Case et al., 2000). Terrestrial pollen indicates that Antarctica supported a *Nothofagus*-conifer fern-*Proteacea* forests during the greenhouse conditions Eocene, similar to Tasmania today (Askin, 2002). By the early Oligocene, vegetation was restricted to *Nothofagus* woodlands (Askin and Raine, 2000; Askin, 2002), similar to modern Patagonia. It has always been puzzling how such warm terrestrial indicators could be reconciled with a glaciated Antarctica. The maps in Fig. 1 illustrate that, unlike today when polar conditions are experienced throughout the continent, Antarctic climates in the greenhouse and early icehouse must have been more varied and complex. Medium-sized ice sheets existed ($8–12\times10^6$ km^3; 20–30 m glacioeustatic equivalent) at the same time pollen indicates that cool, temperate rainforests (e.g., similar to modern Tasmania) survived, if not thrived on the continent.

The early Oligocene provides an excellent example that illustrates the complexity of deciphering the early history of this continent. Most paleoceanographers concede that a large ice sheet existed on Antarctica during the earliest Oligocene (ca. 33 Ma). Pekar et al. (2001) estimated that growth of this ice sheet caused a 55 m lowering (i.e., equivalent to 88% of the modern East Antarctic ice sheet). However, this ice sheet collapsed about 1 my later, and sea level rose 45 m (Pekar et al., 2001). At the same time, *Nothofagus* leaves found in the Ross Sea (Barrett, 1999) indicate nearby southern birch forests. The ice sheets subsequently grew and decayed numerous times during the Oligocene (Miller et al., 1991; Pekar et al., 2001) and by the late Oligocene–middle Miocene the continent saw mostly a mossy tundra vegetation (Askin, 2002). By this time, ice sheets had triumphed over the forests, eliminating refugia that must be invoked to explain the early Oligocene reappearance of *Nothofagus.* Such refugia existed on a larger scale during the Late Cretaceous–Eocene as the glacial conditions (Fig. 5) alternated with interglacial conditions of comparative warmth. The Oligocene to middle Miocene was a transitional period, with a wet based EAIS (Marchant et al., 1993) that was near modern volume at times. By the middle Miocene, Antarctica was a polar desert with a permanent ice cap (Marchant et al., 1993; Fig. 1).

Various studies have documented that Milankovitch astronomical forcing controls not only icehouse ice-volume changes (Hays et al., 1976), but also greenhouse climate changes (e.g., Herbert, 1997; Cramer et al., 2003). DeConto and Pollard (2003) included Milankovitch-scale precessional (19/23 ky), obliquity (41 ky), and eccentricity (~100 ky) periodicities in their sea-level model for Antarctic cryospheric evolution. Matthews and Frohlich (2002) computed an estimated Cretaceous sea-level/ice-volume curve based on Milankovitch forcing (Fig. 6). Minima in this curve appear to correlate with sequence boundaries in New Jersey (Fig. 6), again supporting a glacioeustatic origin for these sequence boundaries. The dominant beats in the predicted curve are the ca. 2.4 my and 405 ky very long and long eccentricity cycles, with hints of a ca. 1 my tilt cycle. Sea level and ice volume undoubtedly changed on shorter time scales (e.g., the precessional and 41 ky tilt cycles), but the stratigraphic record in New Jersey has sedimentation rates that are only high enough to resolve the longer period changes.

We suggest that Late Cretaceous–Eocene ice sheets only existed during short intervals of peak Milankovitch insolation. Though we cannot yet unequivocally determine the tempo of greenhouse changes, comparison with the earliest Oligocene high resolution δ^{18}O records is illustrative: Zachos et al. (1997) showed that the my-scale earliest Oligocene Oi1 glacial event was a composite of numerous 41 ky increases; the Oi1 δ^{18}O peak actually occurred during approximately 2–4 tilt maxima, suggestion a duration of on the order of 100 ky. We suggest that Late Cretaceous–Eocene ice sheets were also equally emphemeral, lasting during successive insolation mimima on the order of 100 ky. Much of the Late Cretaceous–Eocene may in fact have been ice-free, but ice sheets were not completely absent.

We do not question that warm high-latitude climates dominated much of the Late Cretaceous to early Eocene and that warmth may have been caused by high CO_2. We do question the assumption that the poles were ice-free for hundreds of millions of years. Royer et al. (2004) said it well: the presence of “cool snaps” (e.g., our glacioeustatic lowerings) does not indicate that the planet was in a cool mode for the entire Triassic to Eocene. Nevertheless, the paradigm that these warm intervals precluded polar ice must be re-evaluated and the search for evidence of these ice sheets must be pursued.

Acknowledgements

We thank M.E. Katz, B. Cramer, and two anonymous reviewers for reviews, M.E. Katz for comments, A. Payton and C. De La Rocha for inviting us to submit to this issue, and members of the Coastal Plain Drilling Project who are not listed here for developing the stratigraphy and backstripping of the onshore New Jersey boreholes. Supported by NSF grants OCE 0084032 (Miller), EAR97-08664 (Miller), EAR99-09179 (Miller), the New Jersey Geological Survey, and the ODP. The views expressed here do not necessarily reflect those of the supporting agencies.

References

Askin, R.A., 2002. Oligocene and Neogene Vegetation and Climate History of The Ross Sea Region, Antarctica, GSA, Abstr. Programs, Paper No. 73-10.

Askin, R.A., Raine, J.J., 2000. Oligocene and Early Miocene terrestrial palynology of the Cape Roberts drillhole CRP-2/2A, Victoria Land Basin, Antarctica. Terra Antart. 7, 493–501.

Austin, J.A., Christie-Blick, N., Malone, M.J., et al., 1998. Proc. ODP, Init. Repots, vol. 174A. Ocean Drilling Program, College Station, TX. 324 pp.

Barrera, E., Savin, S.M., 1999. Evolution of late Campanian–Maastrichtian marine climates and oceans. Geol. Soc. Am., Special Paper 332, 245–282.

Barrett, P., 1999. Antarctic climate history over the last 100 million years. Terra Antart. Rep. 3, 53–72.

Bond, G.C., Kominz, M.A., 1984. Construction of tectonic subsidence curves for the early Paleozoic miogeocline, southern Canadian Rocky Mountains: implications for the subsidence mechanisms, age of breakup, and crustal thinning. Geol. Soc. Amer. Bull. 95, 155–173.

Brass, G.W., Southam, J.R., Peterson, W.H., 1982. Warm saline bottom water in the ancient ocean. Nature 296, 620–623.

Browning, J.V., Miller, K.G., Pak, D.K., 1996. Global implications of Eocene greenhouse and doubthouse sequences on the New Jersey coastal plain: the icehouse cometh. Geology 24, 639–642.

Browning, J.V., Miller, K.G., Van Fossen, M., Liu, C., Pak, D.K., Aubry, M.-P., Bybell, L.M., 1997. Early to middle Eocene sequences of the New Jersey coastal plain and their significance for global climate change. Proc. Ocean Drill. Program Sci. Results 150X, 229–242.

Cande, S.C., Kent, D.V., 1992. A new geomagnetic polarity time scale for the Late Cretaceous and Cenozoic. J. Geophys. Res. 97, 13917–13951.

Case, J.A., Martin, J.E., Chaney, D.S., Reguero, M., Marenssi, S.A., Santillana, S.M., Woodburne, M.O., 2000. The first duck-billed dinosaur (Family Hadrosauridae) From Antarctica. J. Vertebr. Paleontol. 20, 612–614.

Chamberlain, T.C., 1906. On a possible reversal of deep-sea circulation and its influence on geologic climates. J. Geol. 14, 371–372.

Cramer, B., Wright, J.D., Kent, D.V., Aubry, M.-P., 2003. Orbital climate forcing of $\delta^{13}C$ excursions in the late Paleocene–early Eocene (Chrons C24n–C25n). Paleoceanography 18, 1097.

DeConto, R.M., Pollard, D., 2003. Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2. Nature 421, 245–249.

Fairbanks, R.G., 1989. Glacio-eustatic record 0–16,000 years before present: influence of glacial melting rates on Younger Dryas event and deep ocean circulation. Nature 342, 637–642.

Fairbanks, R.G., Matthews, R.K., 1978. The marine oxygen isotopic record in Pleistocene coral, Barbados, West Indies. Quat. Res. 10, 181–196.

Fisher, A.G., 1984. Two Phanerozoic supercycles. In: Berggren, W.A., Van Couvering, J.A. (Eds.), Catastrophes and Earth History. Princeton University Press, Princeton, N.J., pp. 129–150.

Francis, J.E., 1999. Evidence from fossil plants for Antarctic paleoclimates over the past 100 million years. Terra Antart. Rep. 3, 43–52.

Hancock, J.M., 1993. Transatlantic correlations in the Campanian–Maastrichtian stages by eustatic changes of sea-level. Geol. Soc. London, Spec. Publ. 70, 241–256.

Haq, B.U., Hardenbol, J., Vail, P.R., 1987. Chronology of fluctuating sea levels since the Triassic (250 million years ago to present). Science 235, 1156–1167.

Hays, J.D., Imbrie, J., Shackleton, N.J., 1976. Variations in the Earth's orbit: pacemakers of the ice ages. Science 194, 1121–1132.

Herbert, T.D., 1997. A long marine history of carbon cycle modulation by orbital-climatic changes. PNAS 94, 8362–8369.

Huber, B.T., Hodell, D.A., Hamilton, C.P., 1995. Mid- to Late Cretaceous climate of the southern high latitudes: stable isotopic evidence for minimal equator-to-pole thermal gradients. Geol. Soc. Amer. Bull. 107, 392–417.

Huber, B.T., Norris, R.D., MacLeod, K.G., 2002. Deep sea paleotemperature record of extreme warmth during the Cretaceous. Geology 30, 123–126.

Kennett, J.P., 1977. Cenozoic evolution of Antarctic glaciation, the Circum-Antarctic Ocean, and their impact on global Paleoceanography. J. Geophys. Res. 82, 3843–3860.

Kominz, M.A., Miller, K.G., Browning, J.V., 1998. Long-term and short-term global Cenozoic sea-level estimates. Geology 26, 311–314.

Kominz, M.A., Van Sickel, W.A., Miller, K.G., Browning, J.V., 2003. Sea-Level Estimates for the Latest 100 Million Years: One-Dimensional Backstripping of Onshore New Jersey Boreholes. 22nd Annual GCSSEPM Foundation Bob F. Perkins Research Conference, Sequence Stratigraphic Models for Exploration and Production: Evolving Methodology, Emerging Models and Application Case Histories, pp. 303–315.

Marchant, D.R., Denton, G.H., Swisher, C.C., 1993. Miocene–Pliocene–Pleistocene glacial history of Arena Valley, Quartermain Mountains, Antarctica. Geogr. Ann. 75A, 269–302.

Matthews, R.K., 1984. Oxygen-isotope record of ice-volume history: 100 million years of glacio-eustatic sea-level fluctuation. AAPG Mem. 36, 97–107.

Matthews, R.K., Frohlich, C., 2002. Maximum flooding surfaces and sequence boundaries: comparisons between observations and orbital forcing in the Cretaceous and Jurassic (65–190 Ma). GeoArabia 7, 503–538.

Matthews, R.K., Poore, R.Z., 1980. Tertiary $\delta^{18}O$ record and glacio-eustatic sea-level fluctuations. Geology 8, 501–504.

Miall, A.D., 1991. Stratigraphic sequences and their chronostratigraphic correlation. J. Sediment. Petrol. 61, 497–505.

Miller, K.G., Fairbanks, R.G., 1983. Evidence for Oligocene–middle Miocene abyssal circulation changes in the western North Atlantic. Nature 306, 250–252.

Miller, K.G., Fairbanks, R.G., Mountain, G.S., 1987. Tertiary oxygen isotope synthesis, sea-level history, and continental margin erosion. Paleoceanography 2, 1–19.

Miller, K.G., Wright, J.D., Fairbanks, R.G., 1991. Unlocking the Ice House: Oligocene–Miocene oxygen isotopes, eustasy, and margin erosion. J. Geophys. Res. 96, 6829–6848.

Miller, K.G., Mountain, G.S., the Leg 150 Shipboard Party, and Members of the New Jersey coastal Plain Drilling Project, 1996. Drilling and dating New Jersey OligoceneMiocene sequences: ice volume, global sea level, and Exxon records. Science 271, 1092–1094.

Miller, K.G., Mountain, G.S., Browning, J.V., Kominz, M., Sugarman, P.J., Christie-Blick, N., Katz, M.E., Wright, J.D., 1998. Cenozoic global sea-level, sequences, and the New Jersey transect: results from coastal plain and slope drilling. Rev. Geophys. 36, 569–601.

Miller, K.G., Barrera, E., Olsson, R.K., Sugarman, P.J., Savin, S.M., 1999. Does ice drive early Maastrichtian eustasy? Global $\delta^{18}O$ and New Jersey sequences. Geology 27, 783–786.

Miller, K.G., Browning, J.V., Sugarman, P.J., et al., 2002. 174AS leg summary: sequences, sea level, tectonics, and aquifer resources: coastal plain drilling. In: Miller, K.G., Sugarman, P.J., Browning, J.V., et al., (Eds.), Proc. ODP, Init. Repts., 174AX (Suppl.). Ocean Drilling Program, College Station, TX, pp. 1–40.

Miller, K.G., Sugarman, P.J., Browning, J.V., Kominz, M.A., Hernandez, J.C., Olsson, R.K., Wright, J.D., Feigenson, M.D., Van Sickel, W., 2003. A chronology of Late Cretaceous sequences and sea-level history: glacioeustasy during the Greenhouse World. Geology 31, 585–588.

Miller, K.G., Sugarman, P.J., Browning, J.V., Kominz, M.A., Olsson, R.K., Feigenson, M.D., Hernandez, J.C., 2004. Upper Cretaceous sequences and sea-level history, New Jersey Coastal Plain. Geol. Soc. Amer. Bull. 116, 368–393.

Mountain, G.S., Miller, K.G., 1992. Seismic and geologic evidence for early Paleogene deep-water circulation in the western North Atlantic. Paleoceanography 7, 423–439.

Mountain, G.S., Miller, K.G., Blum, P., et al., 1994. Proc. Ocean Drill. Program, Initial Rep., vol. 150. Ocean Drilling Program, College Station, TX.

Pak, D.K., Miller, K.G., 1992. Late Paleocene to Early Eocene benthic foraminiferal stable isotopes and assemblages: implications for deepwater circulation. Paleoceanography 7, 405–422.

Pekar, S.F., Christie-Blick, N., Kominz, M.A., Miller, K.G., 2001. Evaluating the stratigraphic response to eustasy from Oligocene strata in New Jersey. Geology 29, 55–58.

Pekar, S.F., Christie-Blick, N., Kominz, M.A., Miller, K.G., 2002. Calibration between glacial eustasy and oxygen isotopic data for the early icehouse world of the Oligocene. Geology 30, 903–906.

Pitman, W.C. III, Golovchenko, X., 1983. The effect of sea-level change on the shelf edge and slope of passive margins. SEPM Spec. Publ. 33, 41–58.

Prentice, M.L., Matthews, R.K., 1988. Cenozoic ice-volume history: development of a composite oxygen isotope record. Geology 17, 963–966.

Reid, J.L., 1979. On the contribution of the Mediterranean Sea outflow to the Norwegian–Greenland Sea. Deep Sea Res. Part A, 1199–1223.

Royer, D.L., Berner, R.A., Montanez, I.P., Tabor, N.J., Beerling, D.J., 2004. CO_2 as a primary driver of Phanerozoic climate. GSA Today 14, 4–10.

Sahagian, D., Pinous, O., Olferiev, A., Zakaharov, V., Beisel, A., 1996. Eustatic curve for the Middle Jurassic–Cretaceous based on Russian platform and Siberian stratigraphy: zonal resolution. Am. Assoc. Pet. Geol. Bull. 80, 1433–1458.

Savin, S.M., 1977. The history of the Earth's surface temperature during the last 100 million years. Annu. Rev. Earth Planet. Sci. 5, 319–355.

Savin, S.M., Douglas, R.G., Stehli, F.G., 1975. Tertiary marine paleotemperatures. Geol. Soc. Amer. Bull. 86, 1499–1510.

Shackleton, N.J., Boersma, A., 1981. The climate of the Eocene ocean. J. Geol. Soc. (London) 138, 153–157.

Shackleton, N.J., Kennett, J.P., 1975. Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: oxygen and carbon isotope analysis in DSDP Sites 277, 279, and 281. Initial Rep. Deep Sea Drill. Proj. 29, 743–755.

Shackleton, N.J., Backman, J., Zimmerman, H., Kent, D.V., Hall, M.A., Roberts, D.G., Schnitker, D., Baldauf, J., 1984. Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region. Nature 307, 620–623.

Shackleton, N.J., Hall, M.A., 1984. Oxygen and carbon isotope stratigraphy of Deep Sea Drilling Project Hole 552A: Plio-Pleistocene glacial history. Initial Rep. Deep Sea Drill. Proj. 81, 599–609.

Steckler, M.S., 1981. Thermal and mechanical evolution of Atlantic-type margins (PhD thesis): New York, Columbia University. 261 pp.

Steckler, M.S., Watts, A.B., 1978. Subsidence of the Atlantic-type continental margins off New York. Earth Planet. Sci. Lett. 41, 1–13.

Stoll, H.M., Schrag, D.P., 1996. Evidence for glacial control of rapid sea level changes in the early Cretaceous. Science 272, 1771–1774.

Thomas, D.J., 2004. Evidence for deep-water production in the North Pacific Ocean during the early Cenozoic warm interval. Nature 430, 65–68.

Tucholke, B.E., Hollister, C.D., Weaver, F.M., Vennum, W.R., 1976. Continental rise and abyssal plain sedimentation in the southeast Pacific Basin-Leg 35 Deep Sea Drilling Project. Initial Rep. Deep Sea Drill. Proj. 35, 359–400.

Vail, P.R., Mitchum Jr., R.M., Todd, R.G., Widmier, J.M., Thompson III, S., Sangree, J.B., Bubb, J.N., Hatlelid, W.G., 1977. Seismic stratigraphy and global changes of sea level. Am. Assoc. Petrol. Geol., Mem. 26, 49–212.

Van Sickel, W.A., Kominz, M.A., Miller, K.G., Browning, J.V., 2004. Late Cretaceous and Cenozoic sea-level estimates: backstripping analysis of borehole data, onshore, New Jersey. Basin Res. 16, 451–465.

Watts, A.B., Steckler, M.S., 1979. Subsidence and eustasy at the continental margin of eastern North America. AGU Maurice Ewing Series 3, 218–234.

Wei, W., 1994. Age conversion table for different time scales. J. Nannoplankton Res. 2, 71–73.

Wise, S.W., Breza, J.R., Harwood, D.M., Wei, W., 1991. Palaeogene glacial history of Antarctica. In: Mãœller III, D.W., McKenzie, J.A., Weissert, H. (Eds.), Controversies in Modern Geology: Evolution of Geological Theories in Sedimentology, Earth History and Tectonics. Academic Press, London, pp. 133–171.

Worthington, L.V., 1970. The Norwegian Sea as a Mediterranean basin. Deep-Sea Res. 17, 77–84.

Wright, J.D., Miller, K.G., 1993. Southern Ocean influences on late Eocene to Miocene deepwater circulation. Ant. Res. Ser. 60, 1–25.

Wright, J.D., Miller, K.G., 1996. Control of North Atlantic deep water circulation by the Greenland-Scotland Ridge. Paleoceanography 11, 157–170.

Zachos, J.C., Breza, J., Wise, S.W., 1992. Earliest Oligocene ice-sheet Expansion on East Antarctica: stable isotope and sedimentological data from Kerguelen Plateau. Geology 20, 569–573.

Zachos, J.C., Stott, L.D., Lohmann, K.C., 1994. Evolution of marine temperatures during the paleogene. Paleoceanography 9, 353–387.

Zachos, J.C., Quinn, T.M., Salamy, K., 1996. High resolution (104 yr) deep-sea foraminiferal stable isotope time series. Paleoceanography 11, 251–266.

Zachos, J.C., Flower, B.P., Paul, H., 1997. Orbitally paced climate oscillations across the Oligocene/Miocene boundary. Nature 388, 567–571.