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Ellen Thomas  
*Yale University*

Henk Brinkhuis  
*Utrecht University*

Matthew Huber  
*Purdue University, mhuber@purdue.edu*

Ursula Röhl  
*Bremen University*

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# AN OCEAN VIEW OF THE EARLY CENOZOIC GREENHOUSE WORLD

BY ELLEN THOMAS, HENK BRINKHUIS, MATTHEW HUBER, AND URSULA RÖHL

The Deep Sea Drilling Project (DSDP; 1966–1983) and the Ocean Drilling Program (ODP; 1983–2003) have supplied an amazing amount of information used in reconstruction of past climates, and the present Integrated Ocean Drilling Program (IODP) continues to do so. Here we compare current thinking on early Cenozoic climate (“The Greenhouse World,” ~ 65.5–33.5 million years ago [Ma]) (Figure 1) with those published 25 years ago, to highlight what we have learned in the last 25 years of drilling the ocean floor and where we face continuing challenges.

The Cenozoic began with an asteroid

impact on the Yucatan Peninsula and subsequent mass-extinction of many groups of surface-dwelling oceanic life forms. The world may have cooled for a few millennia, but during the next 10 million years (Myrs), ecosystems recovered while the world warmed, reaching maximum temperatures between ~ 56 and ~ 50 Ma (latest Paleocene to early Eocene). At that time, crocodiles, tapir-like mammals, and palm trees flourished around an Arctic Ocean with warm, sometimes brackish surface waters. Temperatures did not reach freezing even in continental interiors at mid to high latitudes, polar surface tempera-

tures and global deep water temperatures were more than 10–12°C warmer than today, and polar ice sheets probably did not reach sea level—if they existed at all. From the middle Eocene on (~ 48.6 Ma), global deep waters and high-latitude surface waters cooled. The diversity of planktic and benthic oceanic forms of life declined in the late Eocene to the earliest Oligocene (~ 37.2–33.9 Ma). Antarctic ice sheets achieved significant volume and reached sea level by about 33.9 Ma, while sea ice might have covered parts of the Arctic Ocean by that time.

We do not fully understand the causes of the warm climate, the amplitude of its (possibly orbitally driven) variability, or the processes by which high latitudes could have been kept so warm. Atmospheric CO<sub>2</sub> levels may have been high (1000–4000 ppm) and more important than ocean circulation in maintaining the warm temperatures of the Greenhouse World. Understanding Earth’s past warm climates, their temporal variability, mechanisms of latitudinal heat

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**Ellen Thomas** ([ellen.thomas@yale.edu](mailto:ellen.thomas@yale.edu)) is Senior Research Scientist, Marine Micropaleontology, Department of Geology and Geophysics, Yale University, New Haven, CT, USA and Research Professor, Department of Earth and Environmental Sciences, Wesleyan University, Middletown, CT, USA. **Henk Brinkhuis** is Associate Professor, Palaeoecology, Institute of Environmental Biology, Utrecht University, Laboratory of Palaeobotany and Palynology, Utrecht, Netherlands. **Matthew Huber** is Assistant Professor, Earth and Atmospheric Sciences Department, Purdue University, West Lafayette, IN, USA. **Ursula Röhl** is Senior Research Scientist, Center for Marine Environmental Research (MARUM) and DFG Research Center for Ocean Margins (RCOM), Bremen University, Bremen, Germany.

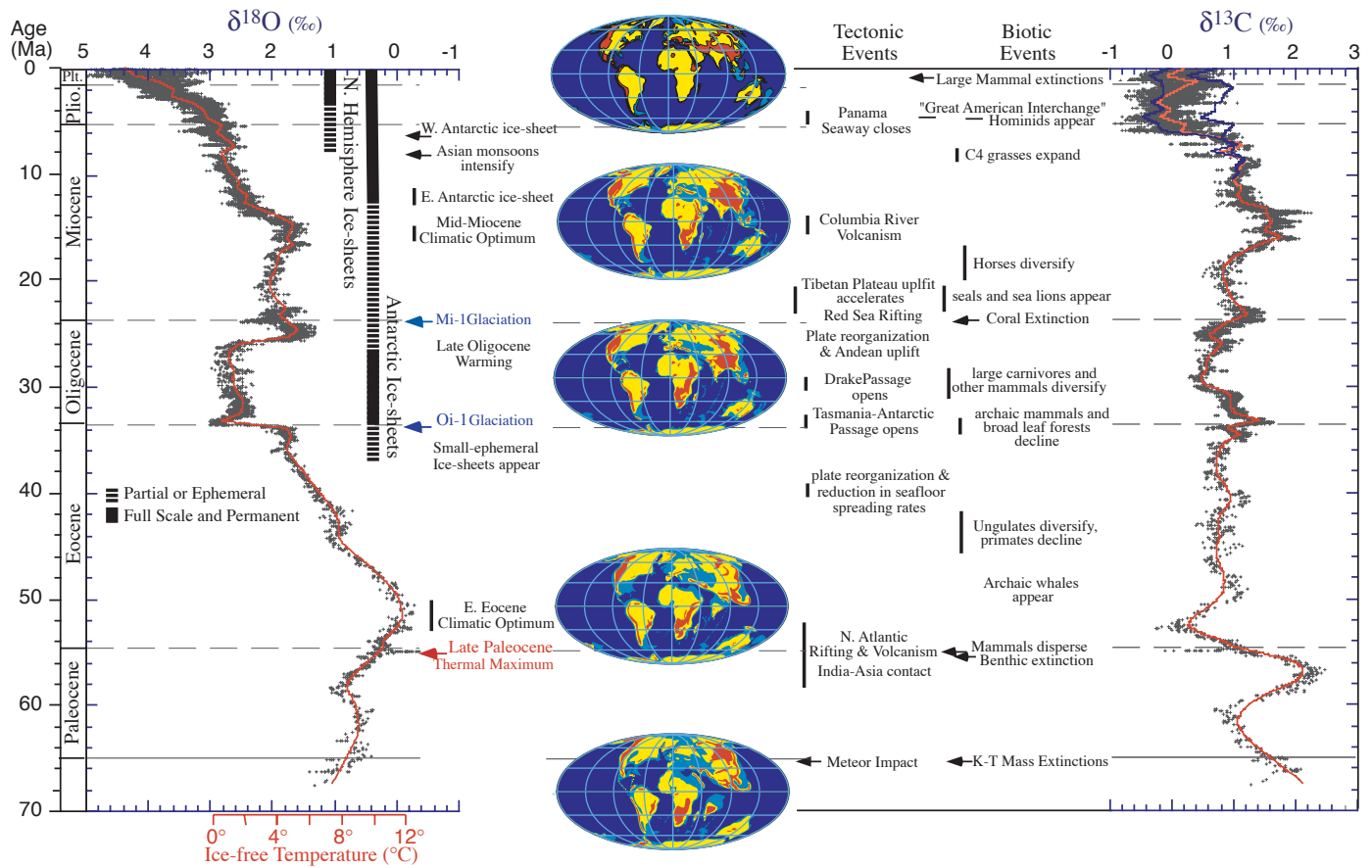


Figure 1. Global deep-sea oxygen and carbon isotope records based on data compiled from more than 40 DSDP and ODP Sites (Zachos et al., 2001). The sedimentary sections from which these data were generated are all pelagic, i.e., deposited at depths of > ~ 1000 m, and consist of fine-grained, carbonate rich (> 50% CaCO<sub>3</sub>) oozes and chalks. Most data are from the analysis of two long-lived benthic foraminiferal genera (*Cibicidoides* and *Nuttallides*); the data were corrected for genus-specific isotope effects by adding 0.64‰ to the values for *Cibicidoides*, 0.4‰ to values for *Nuttallides*. The numerical ages are relative to the geomagnetic polarity timescale of Berggren et al. (1995). Raw data were smoothed using a 5-point running mean, curve-fitted with a locally weighted mean. For carbon isotope values two different curves were derived for times later than the middle Miocene, with the blue curve representing Atlantic Ocean values, the red curves Pacific Ocean values; prior to 15 Ma inter-basin gradients are insignificant or absent. The δ<sup>18</sup>O temperature scale was computed for an ice-free ocean (~ 1.2‰ standard mean ocean water [SMOW]) and thus only applies to the time before the onset of large scale glaciation of the Antarctic continent. From the early Oligocene on, and possibly for some time before that, much of the variability in the δ<sup>18</sup>O values may reflect volume changes in polar ice sheets. The vertical bars provide a rough estimate of the ice volume in each hemisphere relative to the ice volume during the last glacial maximum. The dashed bar represents ice volume < 50%, full bar > 50%. Some key tectonic and biotic events are listed.

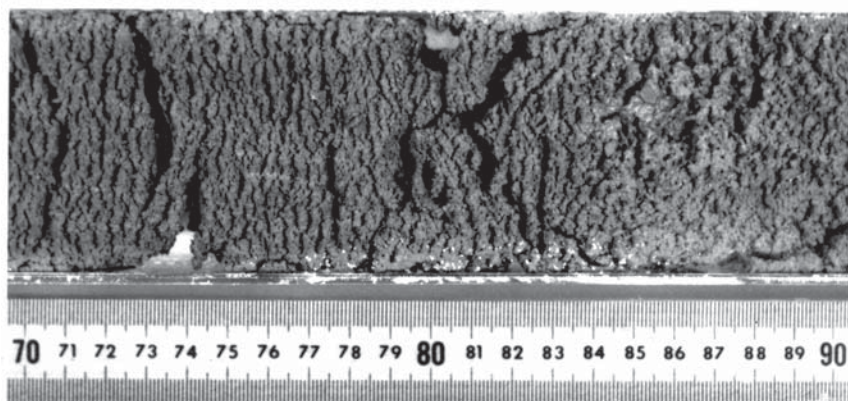
transport, the hydrological cycle, and the role of greenhouse-gas concentration and ocean circulation are of crucial importance to predicting the transition to a future greenhouse world. In this paper, we document the importance of ocean drilling for the reconstruction of past climates and a different planet Earth.

### CORING TECHNIQUES, MEASURING TIME, AND PROXIES: THE LAST 25 YEARS

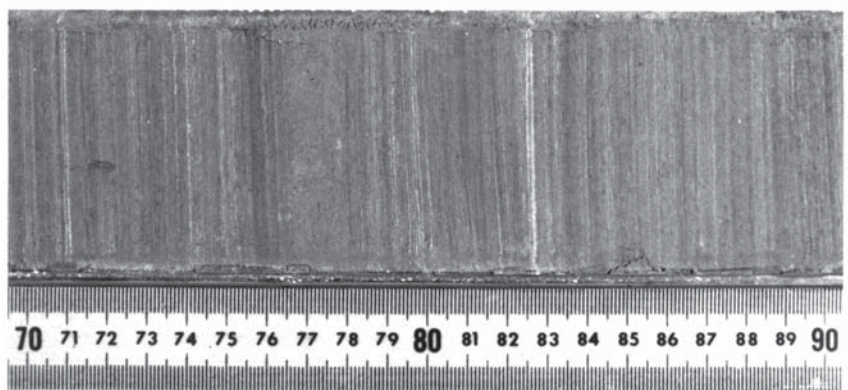
Technologies now taken for granted after publication of a seminal paper on Cenozoic climate (Haq, 1981). The Hydraulic Piston Corer (HPC), so critical to

obtaining long sections of undisturbed sediment, was first used on the drilling vessel (DV) *Glomar Challenger* in 1979 (Leg 64). Cores recovered by the HPC are in stark contrast to the disturbed sediments recovered using rotary coring (Figure 2). Coring in only one hole at one site, however, is not optimal in the case

## COMPARISON PHOTOS ROTARY CORING VS HYDRAULIC PISTON CORING



STANDARD ROTARY CORE  
HOLE: 479 CORED INTERVAL: 90.0–107.5 m



HYDRAULIC PISTON CORE  
HOLE: 480 CORED INTERVAL: 95–99.5 m

Figure 2. Examples of rotary drilling and hydraulic piston coring in the Gulf of California, Sites 479 and 480, DSDP Leg 64 (first leg to use hydraulic piston coring).

of sediment drilling for detailed climate studies. Material may be lost between each 9.5-m core, parts of the hole may be double-cored, or sediments can be disturbed between the cores. Such problems inspired the development of the multi-sensor track (MST) on the DV *JOIDES Resolution* (1989), the ODP drillship. On the MST, magnetic susceptibility of

whole core sections is measured, and the “wiggly lines” of magnetic susceptibility (and other parameters) are evaluated to plan recovery of offset cores in several holes at one site. In this way, a stratigraphically complete, composite section (splice) can be recovered (Figure 3). Such sections are critical in the development of high-resolution climate records.

Piston cores permitted major advances to be made in the development of a precise and accurate geobiomagnetic polarity timescale, which is absolutely necessary to correlate data globally (last updated by Gradstein et al., 2004). Orbitally tuned timescales (see Moore and Pälike, this issue), giving time resolution in  $10^4$  years, are available for the Neogene (last 23 Myrs), and the Oligocene to late Eocene. Timescale tuning for the middle Eocene and older epochs faces fundamental issues because the precision of the orbital solutions is still limited, and there are relatively large uncertainties in radiometric age constraints. For an astronomically calibrated Paleogene (65–23 Ma) timescale, only the stable 405-kyr-long eccentricity period should be used for tuning (Laskar et al., 2004). Given that stability, one can obtain a best fit and derive numerical ages for magnetochron boundaries and short-lived events (Westerhold et al., in press).

Many analytical developments focused on new proxies to elucidate past environmental conditions. For example, Haq (1981) did not show a record of bulk or benthic foraminiferal  $\delta^{13}\text{C}$  values, which are now used extensively to reconstruct oceanic productivity, the biological pump of carbon transfer to the deep ocean, deep ocean circulation, and (indirectly) atmospheric  $\text{CO}_2$  levels (e.g., Shackleton, 1987). In 1981, paleo- $\text{CO}_2$  levels were hardly discussed, probably because neither estimates from carbon-cycle and climate modeling (e.g., Berner, 1994; DeConto and Pollard, 2003), nor estimates from such proxies as alkenone carbon isotopes (Pagani et al., 2005) or boron isotopes (Pearson and Palmer, 2000) existed. In addition, in 1981 there

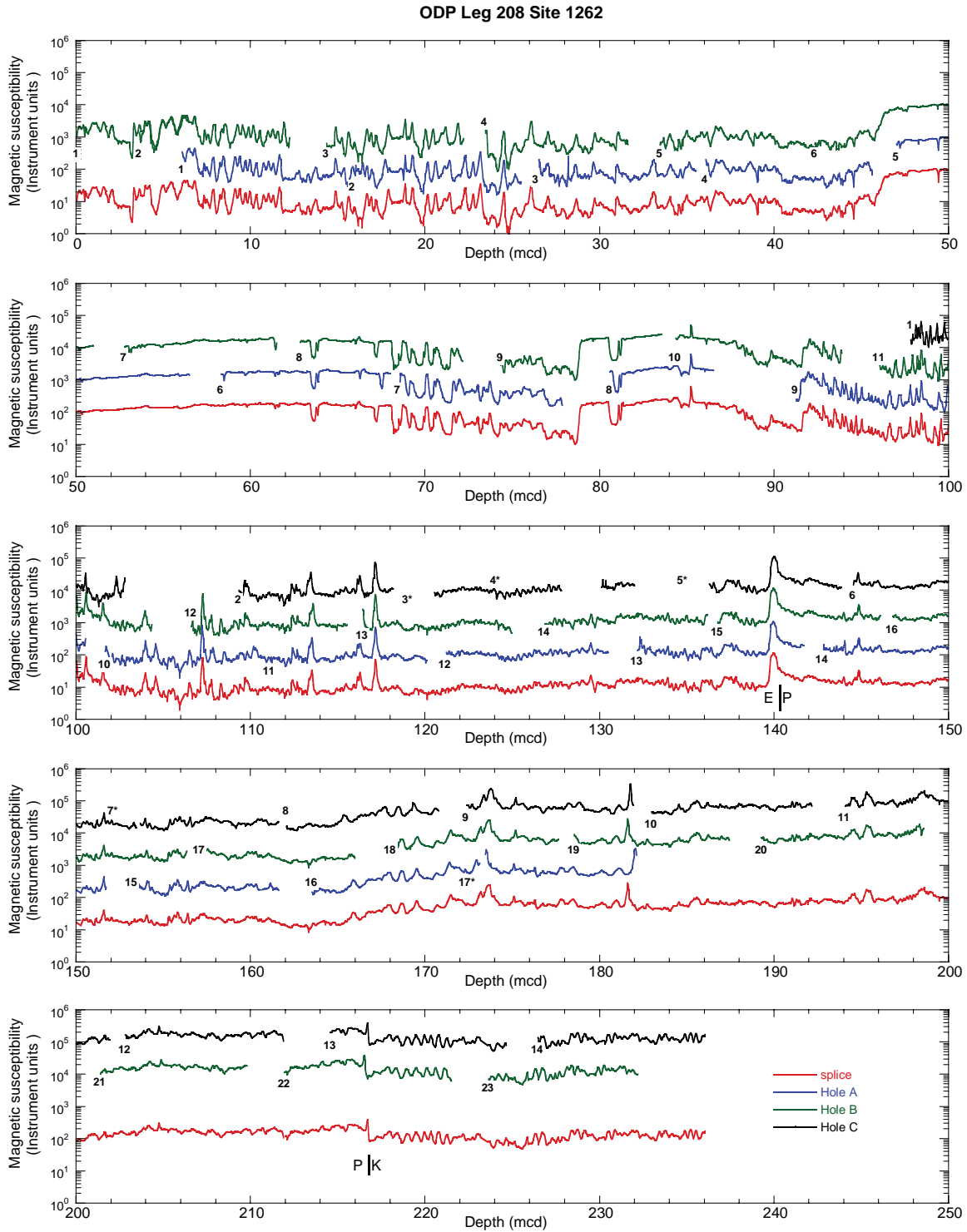


Figure 3. Magnetic susceptibility data for ODP Site 1262 plotted versus composite depth below seafloor, exhibiting orbital cyclicity (mainly eccentricity cycles) in Paleogene sediments. A higher magnetic susceptibility value corresponds to a lower weight percent  $\text{CaCO}_3$ ; note, for example, the sharp drop in  $\text{CaCO}_3$  percentage at the Paleocene/Eocene boundary and the long-term decrease in  $\text{CaCO}_3$  percentage at the Cretaceous/Paleogene boundary. Data from Holes 1262A (blue), 1262B (green), and 1262C (black) are offset from the spliced record (red) by 10, 100, and 1000 times their values, respectively. Numbers near the top of the individual core records are the core numbers. Source: Zachos et al. (2004).

was little discussion of the possible role that the changing energy received from the Sun at specific latitudes, as modulated by variability in orbital configuration of the Earth (Milankovitch forcing), played on climate prior to the development of large Northern Hemisphere ice sheets in the Plio-Pleistocene (last 3.0–2.5 Myrs), possibly at least in part because of a lack of high-resolution sediment records.

### PALEOGENE OCEAN BASINS AND CIRCULATION

Continents were in different positions in the Paleogene (Figure 4) than they are now, limiting ocean currents and heat transport. Past continental positions have long been known approximately (Haq, 1981), but the exact timing of gateway opening and closing remains a point of major debate. India and Asia started to collide by the late Paleocene–

early Eocene or earlier. West of Greenland, seafloor spreading in the Labrador Sea started in the middle Paleocene; to the east of Greenland, only shallow connections between the Arctic and the North Atlantic existed in the Paleocene, and a shallow connection existed between the Arctic and Tethys Oceans (Figure 5). Full-scale seafloor spreading in the Norwegian Sea started close to the end of the Paleocene. The Tasman Gate-

Eocene Annual Average temperature + currents 1120 ppm

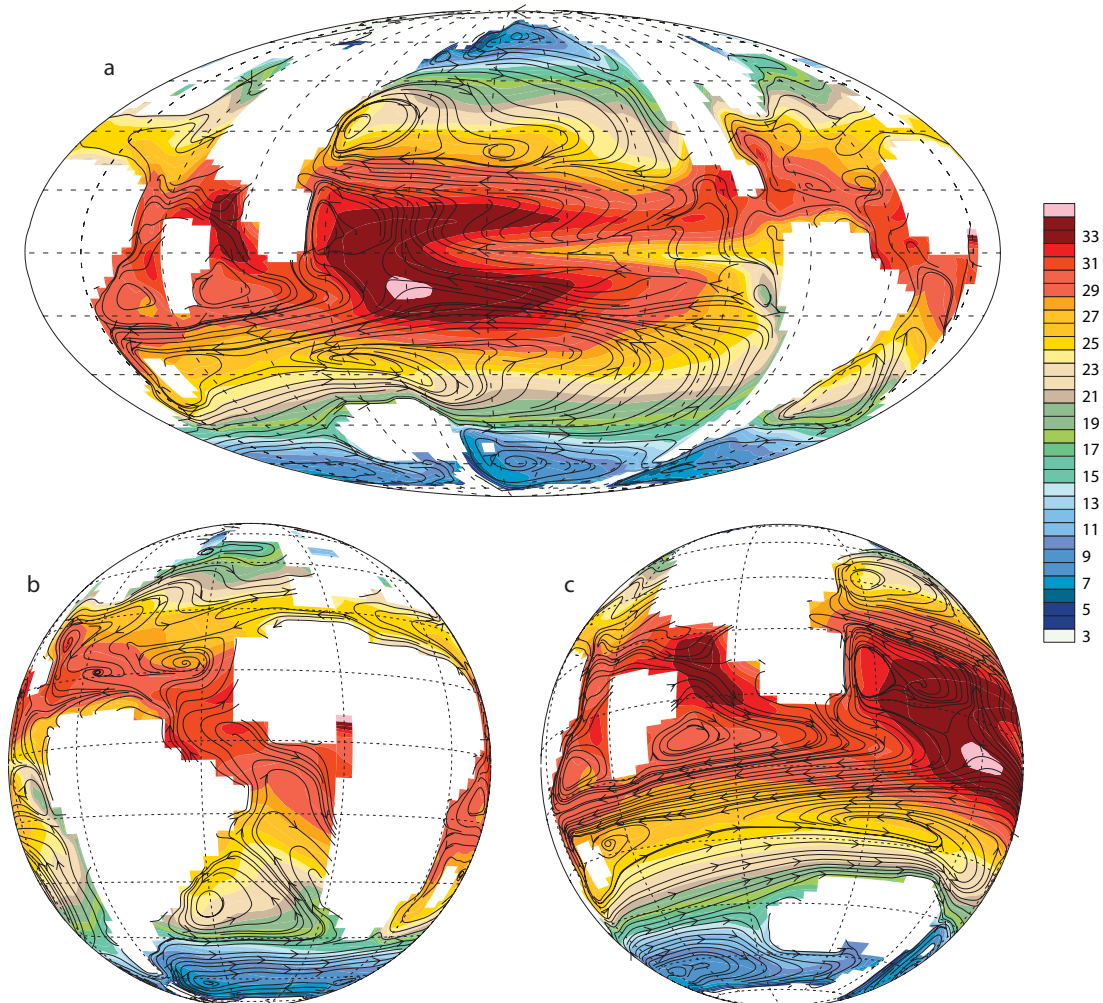


Figure 4. Eocene continental positions, temperature, and currents at 38-m depth in the ocean model; atmospheric CO<sub>2</sub> level at 1120 ppm. This depth is chosen to emphasize the gyre circulations and because it probably is representative of the conditions recorded by sea-surface-temperature proxies. The climate simulation used to produce the figures is described in Huber et al. (2004), but these figures were not published.

way (Australia - Antarctica) deepened during the late Eocene through the earliest Oligocene (~ 36.5–30.2 Ma; Stickley et al., 2004), but there is a large discrepancy in estimates of the opening time of Drake Passage, from the late middle Eocene (~ 41 Ma; Scher and Martin, 2006) to the early Miocene (~ 20 Ma; Anderson and Delaney, 2005).

In 1981, many workers favored the theory that Paleocene-Eocene interme-

diate to deep-water masses formed by sinking of warm, salty waters produced by excess evaporation in low-latitude marginal seas. Polar regions were so warm, however, that water masses that would by today's standards be subtropical (~ 12°C) (Brinkhuis et al., 2006) were still dense enough to fill the abyss by deep convection at high latitudes, as they do today (Huber et al., 2004). In climate models, subtropical deep waters are not

easily produced, and a Paleogene “thermospheric” ocean is no longer generally accepted. Many see the Southern Ocean as the dominant deep-water source region through the Cenozoic, but there is still little agreement on Paleogene deep-sea circulation patterns, for example, whether/when there was a deep-water source in the northern Atlantic or Pacific (Via and Thomas, 2006). Problems in reconstructing flow patterns may in

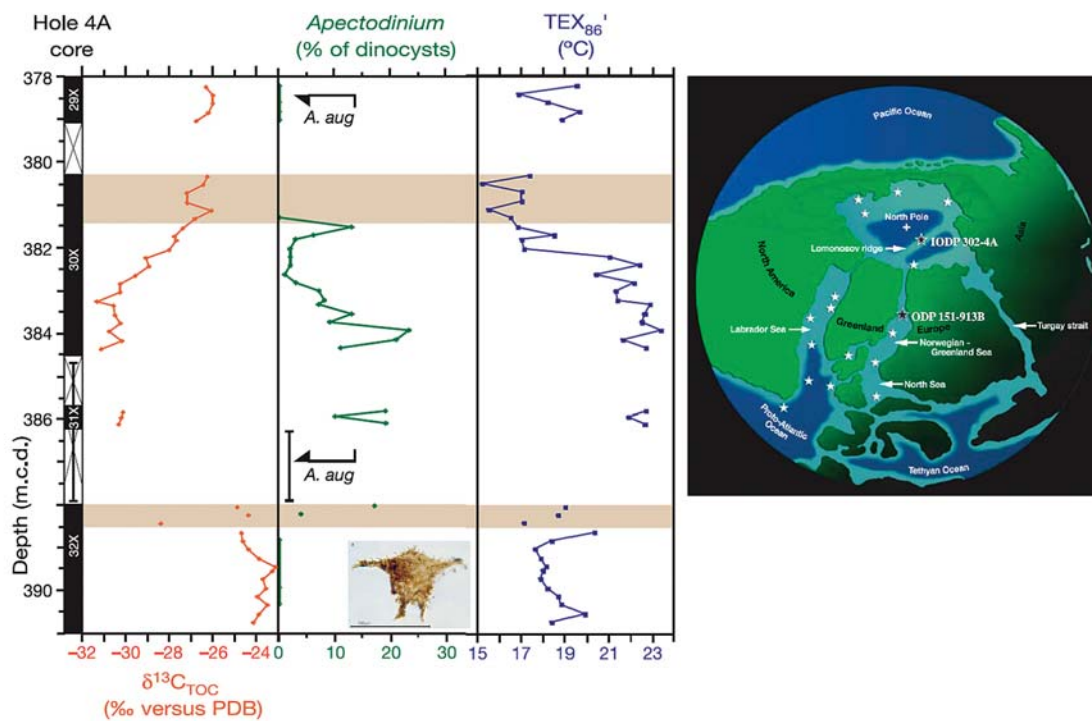


Figure 5. Paleogeographic map of the Arctic Ocean (Brinkhuis et al., 2006). To the left: carbon isotopic composition of total organic carbon at IODP Site 302-4A, showing the occurrence of the negative carbon isotope excursion (CIE) indicating the PETM. The relative abundance of *Apectodinium* (inset: *Apectodinium augustum*) shows that the abundance of these warm water dinoflagellate cysts increased even in the Arctic Ocean, indicating a worldwide distribution during the PETM. Arctic Ocean surface water temperatures before, during, and after the PETM are reconstructed using the  $TEX_{86}'$  proxy (Sluijs et al., 2006), used in non-carbonate sediments. The map indicates locations (white stars) where microfossils of the free-floating freshwater fern *Azolla* have been found during an ~ 800 kyr interval close to the end of the early Eocene (~ 50 Ma), and the abundant occurrence in coeval sediments outside the Arctic (Labrador, Norwegian, Greenland Seas) indicates spill over.

part be due to weak Paleogene gradients in bottom-water properties among the world's oceans, possibly because circulation was fundamentally different from the present (e.g., Emanuel, 2002).

### THE CATASTROPHIC EVENT AT THE CRETACEOUS/TERTIARY BOUNDARY

Knowledge of the asteroid impact at the end of the Cretaceous with its crater on the Yucatan Peninsula (Mexico) and the subsequent mass extinction, one of the largest of the Phanerozoic, has been strongly influenced by scientific drilling. There is still vigorous debate about the processes by which an impact could cause extinction. Proposed mechanisms include darkness caused by dust or sulfate particles, global wildfires, continental margin collapse, and mega-tsunamis and/or gas hydrate dissociation, severe acid rain and acidification of the oceans, and global cooling and/or warming. Debate also continues about the patterns of propagation of extinction through the world's ecosystems and about the recovery of biotic diversity. Cyst-forming photosynthesizers (e.g., diatoms and dinoflagellates) and the siliceous heterotrophic radiolarians survived, as did deep-sea bottom-dwelling foraminifera, but extinction was severe (> 90 percent of all species) among surface-dwelling planktic foraminifera and calcareous nannofossils (e.g., d'Hondt, 2005).

What do these extinction patterns tell us about the effect of the asteroid impact on oceanic primary productivity and the "biological pump" (i.e., the transfer of organic matter from the sea surface to the ocean floor)? Both were thought to have collapsed for millions

of years ("Strangelove Ocean"; Hsü and Mackenzie, 1985), as shown by the collapse of the gradient between oceanic deep and surface carbon isotope values. Productivity in terms of biomass, however, could have recovered quickly while diversity remained low: surviving phytoplankton would bloom as soon as light conditions allowed. Low gradients of benthic-planktic carbon-isotope values are thought to have persisted for millions of years because of a collapse of the biological pump (rather than primary productivity), possibly due to extinction of fecal pellet producers or a shift to smaller-celled primary producers (d'Hondt, 2005).

Regional occurrence of dysoxic to anoxic bottom waters directly after the extinction, however, suggests that the biological pump may have recovered rapidly (e.g., by coagulation by sticky diatoms or cyanobacteria, and ballasting with biogenic silica or terrigenous dust) at least regionally, as supported by the lack of extinction of deep-sea benthic foraminifera (Thomas, in press). The persistent low benthic-planktic carbon-isotope gradients must then be explained and may have had a more complex origin than earlier envisaged. One possible origin of these low isotope gradients is some combination of diagenetic "vital effects" because the surface-dwelling carriers of the isotope record underwent severe extinction; post-extinction records are derived from different species, and those records may have been influenced by input of light carbon to the surface ocean-atmosphere system by biomass burning or by methane from dissociation of gas hydrates due to continental margin slumping.

### EARLY EOCENE CLIMATIC OPTIMUM

Earth may have cooled for a few millennia after the end Cretaceous bolide impact (Galeotti et al., 2004). That cooling was followed by warming that continued through the end of the Paleocene (Figures 1 and 4), making the early Eocene the warmest period of the Cenozoic (e.g., Haq, 1981; Zachos et al., 2001). We do not know how the high latitudes were kept at temperatures as great as 15°C or more; latitudinal temperature gradients were low and thus would not permit high heat transport through the atmosphere and ocean (Huber et al., 2004). Climate models consistently compute temperatures for high latitudes that are lower than those suggested by biotic and chemical temperature proxies.

The early Eocene started with an extreme, short-lived global warming event, the Paleocene-Eocene Thermal Maximum (PETM). The time period was characterized by negative oxygen and carbon isotope values in surface- and bottom-dwelling foraminifera and bulk carbonate globally. The PETM may have started in fewer than 1000 years, with recovery over ~ 170 kyr (Röhl et al., 2006). The negative carbon isotope excursion (CIE) was at least 2.5 ‰ in deep oceanic records, 5–6 ‰ in terrestrial and shallow marine records. These joint isotope anomalies indicate that rapid emission of isotopically light carbon caused severe greenhouse warming, similar to fossil-fuel burning (review in Bowen et al., 2006).

During the PETM, temperatures increased by up to 8°C in southern high-latitude sea surface waters; about 4–5°C in the deep sea, in equatorial surface



waters, and the Arctic Ocean (Figure 5); and by about 5°C on land at mid latitudes in continental interiors. Humidity and precipitation were high, especially at middle to high latitudes (Pagani et al., 2006). Diversity and distribution of surface marine and terrestrial biota shifted, with migration of thermophilic biota to high latitudes and evolutionary turnover, while deep-sea benthic foraminifera suffered extinction (30–50 percent of species). There was widespread oceanic carbonate dissolution: the calcium carbonate compensation depth (CCD) rose by greater than 2 km in the southeastern Atlantic.

Since 1995 (Dickens et al., 1995; Matsumoto, 1995), the CIE has been explained by the release of ~ 2000–2500 gigatons (Gt) of isotopically light (~ -60‰) carbon from methane clathrates in oceanic reservoirs. Oxidation of methane in the oceans would have stripped oxygen from the deep waters, leading to hypoxia, and the shallowing of the CCD, leading to widespread dissolution of carbonates. Proposed triggers of gas hydrate dissociation include warming of the oceans by a change in oceanic circulation, continental slope failure, sea-level lowering, a comet impact, explosive Caribbean volcanism, or North Atlantic basaltic volcanism (review in Thomas, in press).

Arguments against gas hydrate dissociation as the cause of the PETM include low estimates (500–3000 Gt C) for the size of the oceanic gas hydrate reservoir in the recent oceans, implying even smaller ones in the warm Paleocene oceans. The observed greater than 2-km rise in the CCD is greater than estimated for a release of 2000–2500 Gt of carbon.

In addition, pre-PETM atmospheric pCO<sub>2</sub> levels of greater than 1000 ppm require much larger amounts than 2500 Gt of carbon to raise global temperatures by 5°C at estimated climate sensitivities of 1.5°–4.5°C for a doubling of CO<sub>2</sub>. Alternate sources of carbon include a large range of options: a comet; organic matter heated by igneous intrusions in the North Atlantic, by subduction in Alaska, or by continental collision in the Himalayas; peat burning; oxidation of organic matter after desiccation of inland seas; and mantle-plume-induced lithospheric gas explosions (reviews in Bowen et al., 2006; Thomas, in press).

Was the PETM unique? Dissolution horizons associated with isotope anomalies and benthic foraminiferal assemblage changes (called “hyperthermals”) have been identified in upper Paleocene–lower Eocene sediments, but there is as yet no agreement on a link of the PETM to a specific orbital configuration (Westerhold et al., in press). Do hyperthermals reflect greenhouse gas inputs, or cumulative effects of changing ocean chemistry and circulation attributed to Milankovitch forcing? If the PETM was one of a series of events occurring at orbital periodicities through the late Paleocene–early Eocene, its cause probably was not singular (i.e., a large volcanic eruption or a comet impact), but intrinsic to Earth’s climate system (i.e., orbital forced changes in insolation, influencing oceanic circulation and chemistry through positive feedback mechanisms). The Early Eocene Climatic Optimum has been seen as a prolonged period with continuously high temperatures, but it could have been a time of alternating warm and very warm (hyperthermal)

periods, comparable to alternation of glacial and interglacial periods during overall colder times in Earth’s history.

Global cooling started at the end of the early Eocene to the early middle Eocene (Figure 1). What triggered that cooling remains an unsolved question. Arctic Ocean surface waters had low salinities in the early Eocene (Figure 5), which were the lowest during the latest early Eocene. The Arctic Ocean basin opened to the North Atlantic in the earliest middle Eocene. Could fluctuations in freshwater supply via the new connection between the Arctic Ocean and North Atlantic Ocean have influenced global climate so that global cooling began with that connection through changing deep-water circulation patterns when densities declined in areas of freshwater outflow? Freshwater spills from the Arctic had been speculated to have caused the terminal Eocene cooling event (~ 33.7 Ma; Thierstein and Berger, 1978). More recent drilling results might support a come-back of this hypothesis, but modified, with spillover possibly involved in triggering the cooling starting the early middle Eocene (~ 49 Ma).

## LATE EOCENE COOLING/ GLACIATION

The net effect of climate change during the Cenozoic was drastic cooling. Ocean drilling results refined the cooling patterns (Figure 1) and through analysis of these data over the last 25 years, the initiation of glaciation has been defined at increasingly earlier ages. In the 1970s, Antarctica was thought not to be cold enough to support continental ice sheets until the late Miocene. In 1981 the main features of the record were commonly

discussed in terms of cooling rather than polar ice volume (Hag, 1981): the “traditional interpretation of the  $\delta^{18}\text{O}$  curve” (Shackleton and Kennett, 1975) invoked cooling Antarctic surface waters (thus global deep waters) in the earliest Oligocene, followed by further cooling and buildup of the Antarctic ice sheet in the middle Miocene. From the early 1990s on a rapid ( $\sim 100$  kyr) increase in benthic foraminiferal  $\delta^{18}\text{O}$  values in the earliest Oligocene (Oi-1 event; Figure 1) have been interpreted as reflecting at least in part the establishment of the Antarctic ice sheet, although small, wet-based ice sheets may have existed through the late Eocene (Coxall et al., 2004; Miller et al., 2005). Whether the Oi-1 event was due to an increase in ice volume or to cooling is still debated. Paleotemperature data (Mg/Ca) suggest that the event was due to ice-volume increase, but values might have been affected by changes in the CCD. Ice-sheet modelers argue that the event contains both an ice volume and a cooling component, and several planktonic microfossil records suggest surface water cooling at the time.

The Antarctic Continent had been in a polar position for tens of millions of years before glaciation started, so why did a continental ice sheet form in the earliest Oligocene in  $\sim 100,000$  years (Figure 1)? A popular theory is “thermal isolation” of the Antarctic Continent (Kennett, 1977). Opening of the Tasman Gateway and Drake Passage (Figure 4) triggered the initiation of the Antarctic Circumpolar Current (ACC), reducing meridional heat transport to Antarctica by isolation of the continent within a ring of cold water. We can not constrain the validity of this hypothesis by data on

the opening of Drake Passage, because estimates range from middle Eocene to early Miocene. Climate modeling, however, indicates that the change in meridional heat transport associated with ACC onset was insignificant (Huber et al., 2004); probably no warm current flowed southwards along eastern Australia (Figure 4) because a counterclockwise gyre in the southern Pacific (Haq, 1981) prevented warm waters from reaching Antarctica. The concentration of greenhouse gases in the atmosphere may have been the main forcing for the development of the Antarctic ice sheet (e.g., De Conto and Pollard, 2003). Cenozoic cooling of Antarctica is thus no longer generally accepted as having been caused by changes in oceanic circulation only, while the role of decreasing  $\text{CO}_2$  levels with subsequent processes such as ice albedo and weathering feedbacks (possibly modulated by orbital variations) is seen as a significant long-term climate-forcing factor.

#### EPILOGUE: HOW HAVE OUR IDEAS CHANGED IN 25 YEARS?

- An asteroid impacted Earth at the end of the Cretaceous. Debate is still vigorous on the processes that caused extinction and the patterns of recovery of the biota.
- Timescales have become much more refined; an early Paleogene, orbitally tuned timescale may become available in the near future.
- Extreme, short global warming at the beginning of the Eocene was probably caused by rapid emission of greenhouse gases, but we do not know the source or process of emission.
- Climates of the late Paleocene-early Eocene Greenhouse World may have

been punctuated by hyperthermal events, either caused by greenhouse gas emissions or by orbitally driven fluctuations of a different nature.

- Opening of the Arctic to the world oceans might have been a factor in middle Eocene global cooling.
- The Tasman Gateway was open (shallow) in the middle Eocene and deepened in the late Eocene, but we do not yet know exactly when Drake Passage opened.
- There are doubts whether the initiation of the ACC could have been a major causal factor in the glaciation of the Antarctic continent.

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