PALEOCEANOGRAPHY: THE GREENHOUSE WORLD

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Introduction

Understanding and modeling a world very different from that of today, for example, a much warmer world with a so-called 'greenhouse' climate, requires a thorough grasp of such processes as the carbon cycle, ocean circulation, and heat transport in the oceans – indeed it pushes the limits of our current conceptual and numerical models. A review of what we know of past greenhouse worlds reveals that our capacity to predict and understand key features of such intervals is still limited.

Greenhouse climates represent most of the past 540 My, called the Phanerozoic, the part of Earth history for which an extensive fossil record is available. One could thus argue that a climate state with temperatures much warmer than modern, without substantial ice at sea level at one or both Poles, is the 'normal' mode, and the glaciated, cool state – such as has existed over the past several million years – is unusual. Since we were born into this glaciated climate state, however, our theories of near-modern climates are well developed and powerful, whereas greenhouse climates with their different boundary conditions challenge our understanding.

The two best-documented periods with greenhouse climates are the Cretaceous ($\sim$145–65.5 Ma) and the Eocene ($\sim$55.8–33.9 Ma), on which this article will concentrate. But, other intervals, both earlier (e.g., Silurian, 443.7–416 Ma) and more recent (e.g., early–middle Miocene, 23.03–11.61 Ma, mid-Pliocene, 3.5 Ma) are characterized by climates clearly warmer than today, although less torrid than the Eocene or Cretaceous. Furthermore, within long-term greenhouse intervals, there may lurk short-term periods of an icier state.

The alien first impression one derives from looking at the paleoclimate records of past greenhouse climates is the combination of polar temperatures too high for ice formation and warm winters within continental interiors at mid-latitudes. In fact, there are three main characteristics of past greenhouse climates that are remarkable and puzzling, and hence are the focus of most research. These three key features are:

- Global warmth: a global mean surface temperatures much warmer ($>10\,°C$) than the modern global mean temperature ($15\,°C$);
- Equable climates: reduced seasonality in continental interiors compared to modern, with winter temperatures above freezing; and
- Low temperature gradients: a significant reduction in equator-to-pole and vertical ocean temperature gradients (i.e., to $<20\,°C$). Annual mean surface air temperatures today in the Arctic Ocean are about $-15\,°C$, in the Antarctic about $-50\,°C$, whereas tropical temperatures are on average $26\,°C$, hence the modern gradient is 40–75 $°C$.

Not all greenhouse climates are the same: some have one or two of these key features, and only a few have all the three. For example, there is ample and unequivocal evidence for global warmth and equable climates throughout large portions of the Mesozoic (251–65.5 Ma) and the Paleogene (65.5–23.03 Ma) ([Figure 1](#)). But, evidence for reduced meridional and vertical ocean temperature gradients is much more controversial, however, and (largely due to the nature of the rock and fossil records) only reasonably well established for the mid-Cretaceous and the early Eocene (*see* Cenozoic Climate – Oxygen Isotope Evidence and Paleoceanography).

Superimposed on these fundamental aspects of greenhouse climate modes is a pervasive variability apparently driven by orbital variations (*see* Plio-Pleistocene Glacial Cycles and Milankovitch Variability), and abrupt climate shifts (*see* Abrupt Climate Change), indicative either of crossing of thresholds or of sudden changes in climatic forcing factors, and notable changes in climate due to shifting paleogeographies, paleotopography, and ocean circulation changes (*see* Heat Transport and Climate).

A variety of paleoclimate proxies, including oxygen isotopic paleotemperature estimates (*see* Cenozoic Climate – Oxygen Isotope Evidence), as well as newer proxies such as Mg/Ca (*see* Determination of Past Sea Surface Temperatures), and ones that might still be considered experimental, such as TEX_{86} (*see* Paleoceanography), have been used to reconstruct greenhouse climates. New proxies and long proxy records of atmospheric carbon dioxide concentrations (pCO_2)
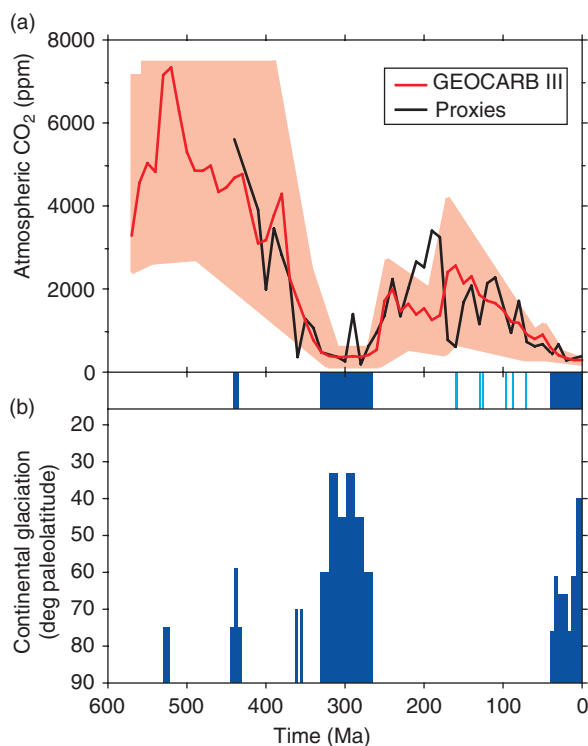
Figure 1 (a) Atmospheric carbon dioxide concentrations produced by geochemical models and proxies for the Phanerozoic. (b) Major intervals of continental glaciation and the latitude to which they extend. The major periods in Earth's history with little continental ice correspond to those periods with high greenhouse gas concentrations at this gross level of comparison. From Royer DL, Berner RA, Montanez IP, Tabor NJ, and Beerling DJ (2004) CO_2 as a primary driver of Phanerozoic climate. *GSA Today* 14: 4–12.

are becoming available and allow us to constrain the potential sensitivity of climate change to greenhouse gas forcing (Figure 1). The investigation of past greenhouse climates is making exciting and unprecedented progress, driven by massive innovations in multiproxy paleoclimate and paleoenvironmental reconstruction techniques in both the marine and terrestrial realms, and by significant developments in paleoclimate modeling (*see* Paleoceanography and Paleoceanography, Climate Models in).

Cretaceous

The Cretaceous as a whole was a greenhouse world (Figure 1): most temperature records indicate high latitude and deep ocean warmth, and $p\mathrm{CO_2}$ was high. Nevertheless, substantial multiproxy evidence indicates the presence of, perhaps short-lived, below-freezing conditions at high latitudes and in continental interiors, and the apparent buildup of moderate terrestrial ice sheets (potentially in the

early Cretaceous Aptian, 125–112 Ma, and in the late Cretaceous Maastrichtian, 70.6–65.5 Ma). Peak Cretaceous warmth and the best example of a greenhouse climate lies in the mid-Cretaceous (∼120–80 Ma), when tropical temperatures were between 30 and 35 °C, mean annual polar temperatures above 14 °C, and deep ocean temperatures were *c*. 12 °C. Continental interior temperatures were probably above freezing year-round, and terrestrial ice at sea level was probably absent. Global mean surface temperatures were much more than 10 °C above modern, and the equator-to-pole temperature gradient was approximately 20 °C.

It is not clear exactly when the thermal maximum occurred during this overall warm long period (e.g., in the Cenomanian, 99.6–93.5 Ma, or in the Turonian, 93.5–89.3 Ma), because regional factors such as paleogeography and ocean heat transport might substantially alter the expression of global warmth (Figure 2), and a global thermal maximum does not necessarily denote a global maximum everywhere at the same time.

A feature of particular interest in the Cretaceous (and to a lesser extent in the Jurassic, 199.6–145.5 Ma) oceans are the large global carbon cycle perturbations called 'oceanic anoxic events' (OAEs). These were periods of high carbon burial that led to drawdown of atmospheric carbon dioxide, lowering of bottom-water oxygen concentrations, and, in many cases, significant biological extinction. Most OAEs may have been caused by high productivity and export of organic carbon from surface waters, which has then preserved in the organic-rich sediments known as black shales. At least two Cretaceous OAEs are probably global, and indicative of ocean-wide anoxia at least at intermediate water depths, the Selli (late early Aptian, ∼120 Ma) and Bonarelli events (Cenomanian–Turonian, ∼93.5 Ma). During these events, global sea surface temperatures (SSTs) were extremely high, with equatorial temperatures of 32–36 °C, polar temperatures in excess of 20 °C. Multiple hypotheses exist for explaining OAEs, and different events may bear the imprint of one mechanism more than another, but one factor may have been particularly important.

Increased volcanic and hydrothermal activity, perhaps associated with large igneous provinces (LIPs) (*see* Igneous Provinces), changed increased atmospheric $p\mathrm{CO_2}$, and altered ocean chemistry in ways to promote upper ocean export productivity leading to oxygen depletion in the deeper ocean. These changes may have been exacerbated by changes in ocean circulation and increased thermohaline stratification which might have affected benthic oxygenation, but the direction and magnitude of this

(a) Temperature difference
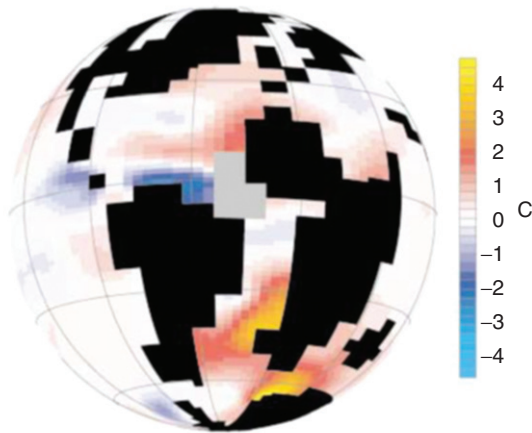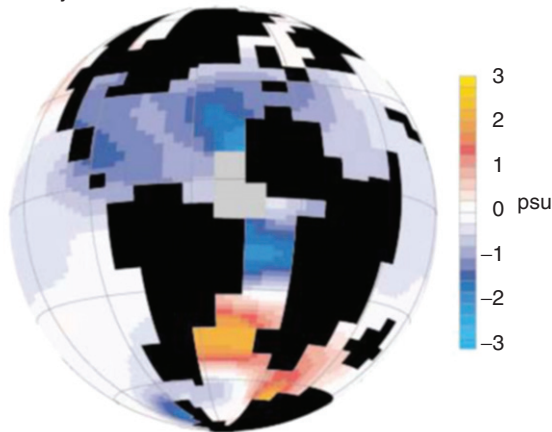

(b) Salinity difference

Figure 2 In these coupled ocean–atmospheric climate model results from Poulsen *et al.* (2003), major changes in proto-Atlantic Ocean temperature and salinity occur by deepening of the gateway between the North and South Atlantic. This study concluded that part of the pattern of warming, anoxia, and collapse of reefs in the mid-Cretaceous organic-rich sediments known as black might have been driven by ocean circulation changes engendered by rifting of the Atlantic basin.

potential feedback are poorly constrained. The emplacement of LIPs was sporadic and rapid on geological timescales, but *in toto* almost 3 times as much oceanic crust was produced in the Cretaceous (in LIPs and spreading centers) as in any comparable period. The resulting fluxes of greenhouse gases and other chemical constituents, including nutrients, were unusually large compared to the remainder of the fossil record. The direct impact of this volcanism on the ocean circulation through changes in the geothermal heat flux was probably small as compared to the ocean's large-scale circulation, but perturbations to the circulation within isolated abyssal basins might have been important.

As the Cretaceous came to a close, climate fluctuated substantially (Figure 3), but it is impossible to know how Cretaceous climate would have

continued to evolve, because the Cretaceous ended with a bang – the bolide which hit the Yucatan Peninsula at 65.5 Ma profoundly perturbed the ocean and terrestrial ecosystems (*see* Paleoceanography). Much of the evolution of greenhouse climates is related to the carbon cycle and hence to biology and ecology, so it is difficult to know whether the processes that maintained warmth in the Cretaceous with such different biota are the same as in the Cenozoic.

Early Paleogene

After the asteroid impact and subsequent mass extinction of many groups of surface-dwelling oceanic life-forms, the world may have cooled for a few millennia, but in the Paleocene (65.5–55.8 Ma) conditions were generally much warmer than modern, although evidence for at least intervals of deep sea and polar temperatures cooler than peak Cretaceous or Eocene warmth exists. Overall, in the first 10 My after the asteroid impact, ecosystems recovered while the world followed a warming trend, reaching maximum temperatures between ~ 56 and ~ 50 Ma (latest Paleocene to early Eocene).

Within the overall greenhouse climate of the Paleocene to early Eocene lies a profoundly important, but still perplexing, abrupt climatic maximum event, the Paleocene–Eocene Thermal Maximum (PETM). The record of this time period is characterized by global negative anomalies in oxygen and carbon isotope values in surface and bottom-dwelling foraminifera and bulk carbonate. The PETM may have started in fewer than 500 years, with a recovery over ~ 170 ky. The negative carbon isotope excursion (CIE) was at least 2.5–3.5‰ in deep oceanic records and 5–6‰ in terrestrial and shallow marine records (Figure 4). These joint isotope anomalies, backed by independent temperature proxy records, indicate that rapid emission of isotopically light carbon caused severe greenhouse warming, analogous to modern anthropogenic fossil-fuel burning.

During the PETM, temperatures increased by 5–8 °C in southern high-latitude sea surface waters; *c.* 4–5 °C in the deep sea, equatorial surface waters, and the Arctic Ocean; and *c.* 5 °C on land at mid-latitudes in continental interiors. There is some indication of an increase in the vigor of the hydrologic cycle (Figure 4), but the true timing and spatial dependence of the change in hydrology are not clear. Diversity and distribution of surface marine and terrestrial biota shifted, with migration of thermophilic biota to high latitudes and evolutionary
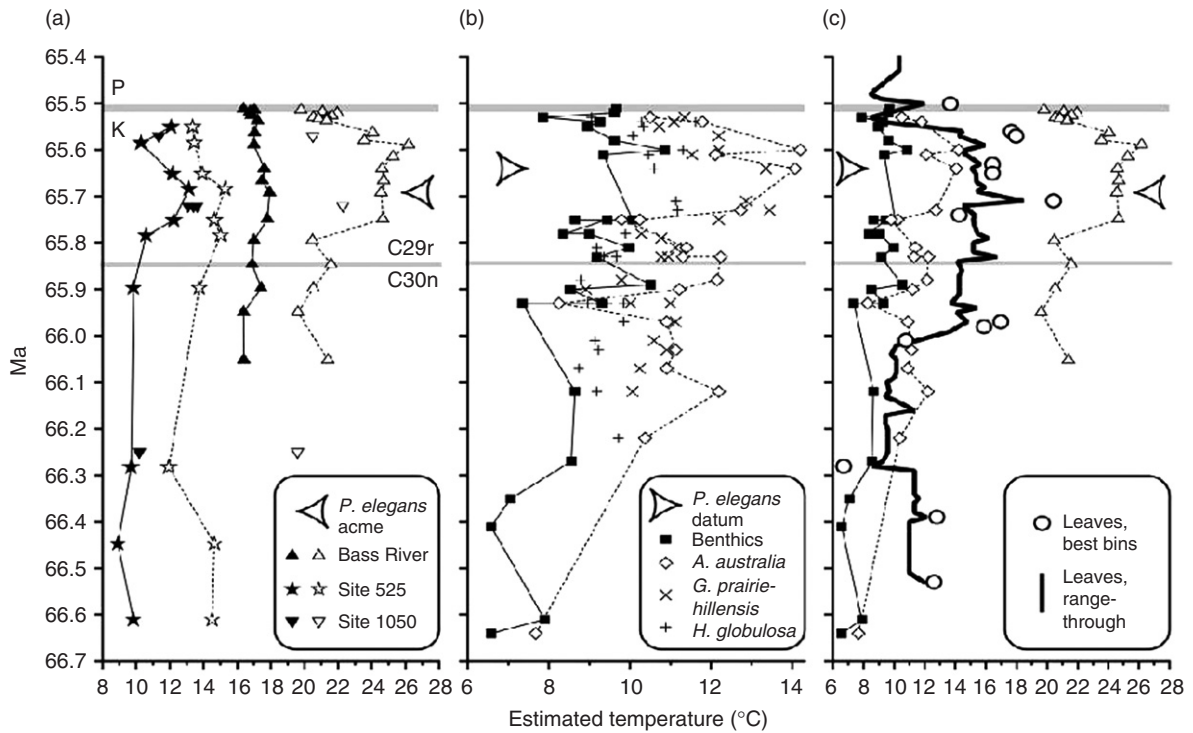
Figure 3 A summary of climate change records near the end of the Cretaceous (Maastrichtian). Paleotemperatures estimated from oxygen isotope data from benthic (filled symbols) and planktonic (open symbols) foraminiferea from middle (a) and high (b) latitudes. (c) Combined representative data (with leaf data) from (a) and (b) with terrestrial paleotemperature estimates derived from macrofloral records for North Dakota. The end of Cretaceous was clearly a time of very variable climate. From Wilf P, Johnson KR, and Huber BT (2003) Correlated terrestrial and marine evidence for global climate changes before mass extinction at the Cretaceous–Paleogene boundary. *Proceedings of the National Academy of Sciences of the United States of America* 100: 599–604.
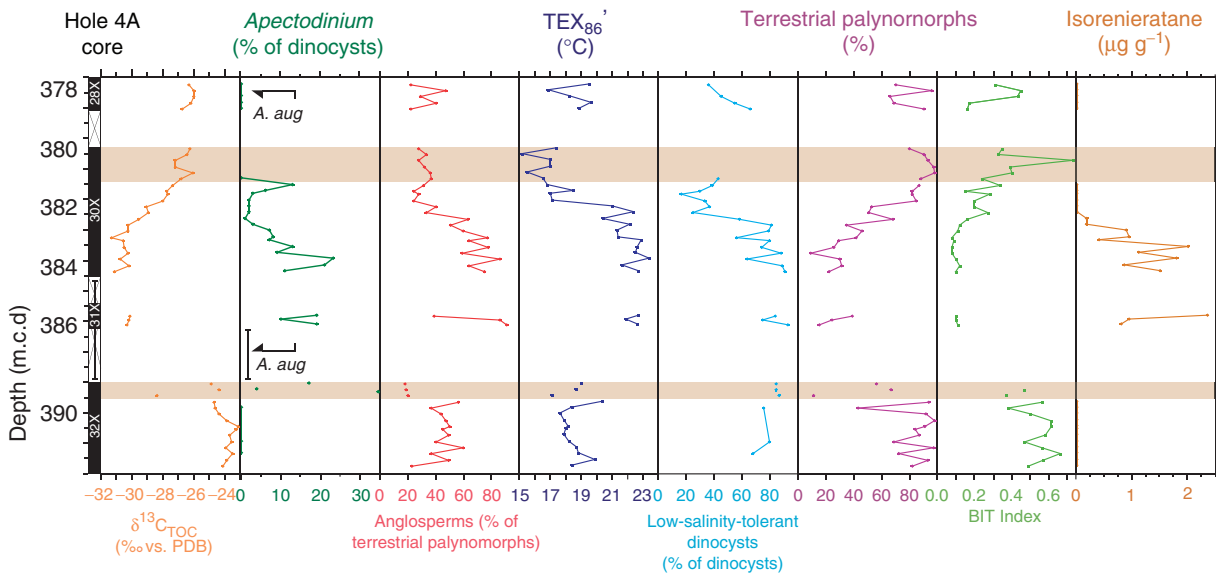

Figure 4 IODP Site 302 enabled the recovery of PETM palynological and geochemical records from the Arctic Ocean. The carbon isotope records the PETM negative excursion (far left). The dinocyst *Apectodinium* (second from left) is a warm-water species, and a nearly global indicator of the PETM. TEX$_{86}$ is a new paleotemperature proxy (middle) which reveals extreme polar warmth even previous to the PETM, and a 5 °C warming with the event. Other indicators show an apparent decrease in salinity in the Arctic ocean. From Sluijs A, Schouten S, Pagani M, *et al.* (2006) Subtropical Arctic Ocean conditions during the Palaeocene Eocene thermal maximum. *Nature* 441: 610–613 (doi:10.1038/nature/04668).

turnover, while deep-sea benthic foraminifera suffered extinction (30–50% of species). There was widespread oceanic carbonate dissolution: the calcium carbonate compensation depth (CCD) rose by more than 2 km in the southeastern Atlantic.

One explanation for the CIE is the release of ~ 2000–2500 Gt of isotopically light ($\sim -60\%$) carbon from methane clathrates in oceanic reservoirs (see Methane Hydrates and Climatic Effects). 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 a 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, explosive Caribbean volcanism, or North Atlantic basaltic volcanism.

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 >2 km rise in the CCD is more than that estimated for a release of 2000–2500 Gt C. In addition, pre-PETM atmospheric pCO_2 levels of greater than 1000 ppm require much larger amounts than 2500 Gt C to raise global temperatures by 5 °C at estimated climate sensitivities of 1.5–4.5 °C for a doubling of CO_2. Alternate sources of carbon include a large range of options: 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 methane release from extensive marshlands; and mantle plume-induced lithospheric gas explosions.

The interval of ~ 52–49 Ma is known as the early Eocene Climatic Optimum (EECO; Figure 5), during which temperatures reached levels unparalleled in the Cenozoic (the last 65.5 My) with the brief exception of the PETM. 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 temperatures were >30 °C warmer than modern, global deep water temperatures were ~ 10–12 °C warmer than today, and polar ice sheets probably did not reach sea level – if they existed at all. Interestingly, EECO tropical ocean temperatures, once thought to be at modern or even cooler values, are now considered to have been ~ 8 °C warmer than today, still much less than the extreme polar warming.

We do not know how the high latitudes were kept as warm as 15–23 °C with tropical temperatures only at ~ 35 °C (Figure 6). The associated small latitudinal temperature gradients make it difficult to explain either high heat transport through the atmosphere or through the ocean given the well-proven conceptual
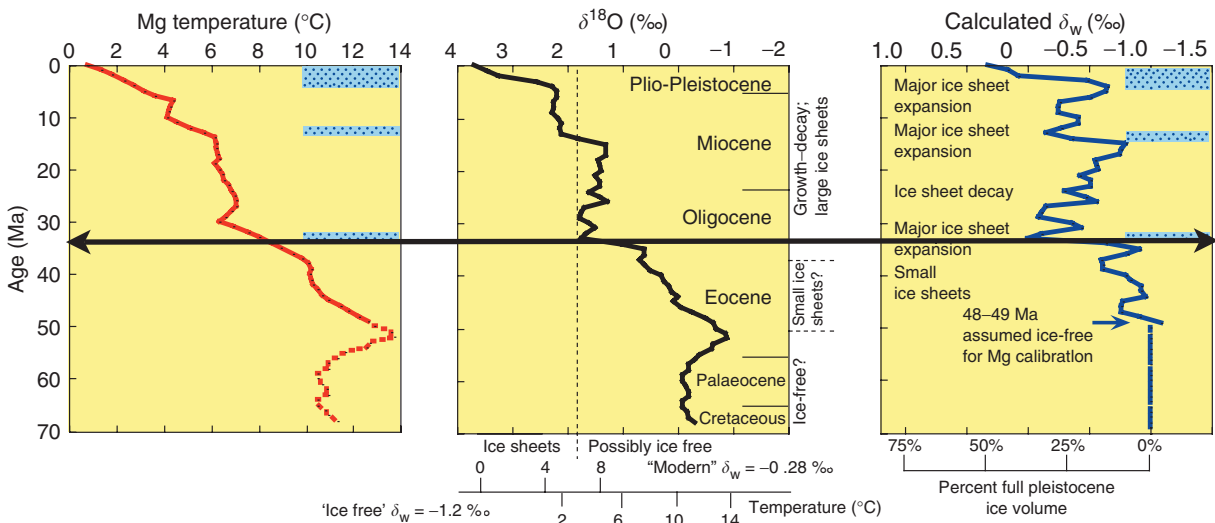

Figure 5 Lear *et al.* (2000) utilized Mg/Ca ratios from benthic foraminifera to create a paleotemperature proxy record (left). When compared against the benthic oxygen isotope record (middle) there is a general congruence in pattern which is expected, given that a large part of the oxygen isotope record also reflects temperature change. When taken in combination, the two records can be used to infer changes in the oxygen isotopic composition (right) of seawater, a proxy for terrestrial ice volume. The major ice buildup at the beginning of the Oligocene (Oi1) is apparent in the substantial change in oxygen isotopic composition in the absence of major Mg/Ca change (solid horizontal line).
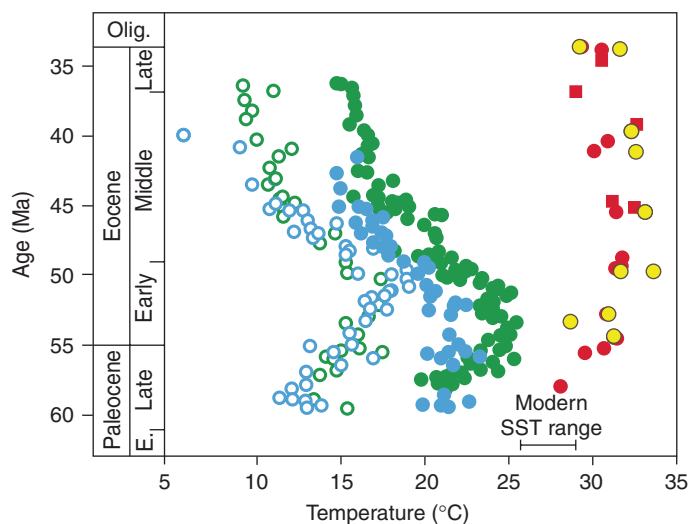
Figure 6 Pearson *et al.* (2007) paleotemperature estimates from Tanzania (red) derived from oxygen isotope ratios of planktonic foraminiferal tests and from TEX$_{86}$ (yellow). The other data points show benthic and polar surface paleotemperature estimates. The Tanzanian data reveal tropical temperatures were significantly warmer than modern but that they were stable compared to benthic and polar temperatures through the Cenozoic climate deterioration. The authors' interpretation is that previous tropical temperature estimates (filled blue and green circles) were spuriously tracking deep ocean temperatures (open blue and green circles) and temperature trends because of diagenesis. If correct, this implies that tropical temperatures were at least 5 °C warmer than previously thought in greenhouse intervals.

and numerical models of climate (Figure 7). Despite more than two decades of work on this paradox of high polar temperatures and low heat transport, climate models consistently compute temperatures for high latitudes and mid-latitude continental winters that are lower than those indicated by biotic and chemical temperature proxies. Even this paradigm of greenhouse climate displayed profound variability. Within the EECO are alternating warm and very warm (hyperthermal) periods. These hypothermals are characterized by dissolution horizons associated with isotope anomalies and benthic foraminiferal assemblage changes and have been identified in upper Paleocene–lower Eocene sediments worldwide, reflecting events similar in nature to, but less severe than, the PETM. It remains an open question whether these hyperthermals directly reflect greenhouse gas inputs, or cumulative effects of changing ocean chemistry and circulation, perhaps driven by orbital forcing. While pCO$_2$ levels may have been high (1000–4000 ppm) to maintain the rise of the global mean temperatures of this greenhouse world, it does not simultaneously explain the small temperature gradients (as low as 15 °C), warm poles, and warm continental interiors. Furthermore, we are still far from explaining the amplitude of orbital, suborbital, and nonorbital variability exhibited by these climates. The mystery is compounded by the rapidity by which the Eocene greenhouse world came to an end with the sudden emplacement of a massive Antarctic ice sheet
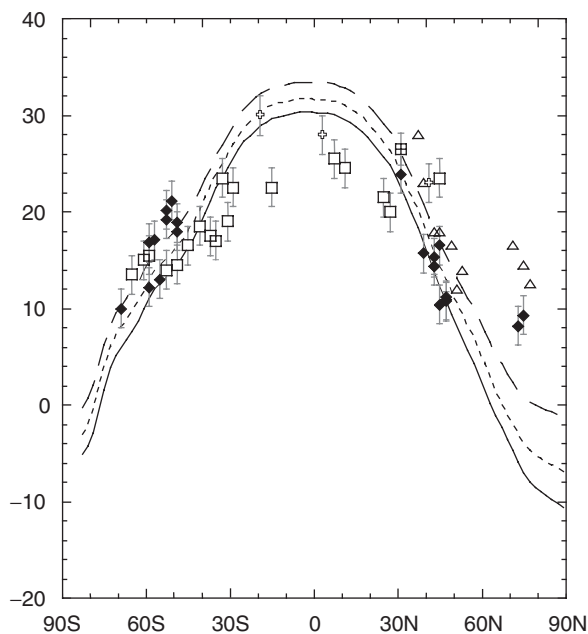

Figure 7 As described in Shellito *et al.* (2003), model-predicted zonally averaged mean annual temperatures compared with paleotemperature proxy records for the Eocene. These simulations were carried out with Eocene boundary conditions and with pCO$_2$ specified at three levels: 500 (solid), 1000 (dotted), and 2000 (dashed lines) ppm. The other symbols represent terrestrial and marine surface temperature proxy records. With increasing greenhouse gas concentrations, temperatures gradually come into better agreement with proxies at high latitudes, but the mismatch persists until extremely high values are reached, at which point tropical temperatures are too warm.

near the epoch's end, but the transition provides critical clues to understanding the functioning of a greenhouse climate.

The End of the Greenhouse

We do not fully understand the greenhouse world, and neither do we understand the transition into the present cool world ('icehouse world') and the cause of global cooling. This much we think we do know. Beginning in the middle Eocene (at ~ 48.6 Ma), the Poles and deep oceans began to cool, and $p\text{CO}_2$ may have declined. The cause of this cooling and apparent carbon draw-down are unknown. As this cooling continued, the diversity of planktic and benthic oceanic life-forms 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. Small, wet-based ice sheets may have existed through the late Eocene, but a rapid (~ 100 ky) increase in benthic foraminiferal $\delta^{18}\text{O}$ values in the earliest Oligocene (called the 'Oi1 event') has been interpreted as reflecting the establishment of the Antarctic ice sheet. Oxygen isotope records, however, reflect a combination of changes in ice volume and temperature (*see* Cenozoic Climate – Oxygen Isotope Evidence), and it is still debated whether the Oi1 event was primarily due to an increase in ice volume or to cooling.

Paleotemperature proxy data (Mg/Ca) suggest that the event was due mainly to ice volume increase, although this record might have been affected by changes in position of the CCD in the oceans at the time. 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. Whatever the Oi1 event was in detail, it is generally seen as the time of change from a largely unglaciated to a largely glaciated world, and the end of the last greenhouse world. There are several proposed causes of this final transition to a world with substantial glaciation, including Southern Ocean gateway opening, decreases in greenhouse gas concentrations, orbital configuration changes, and ice albedo feedbacks.

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 ~ 100 Ky during the Oi1? A popular theory is 'thermal isolation' of the Antarctic continent: opening of the Tasman Gateway and Drake Passage 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 cannot constrain the validity of this hypothesis by data on the opening of Drake Passage, because even recent estimates range from middle Eocene to early Miocene (a range of 20 My). The Tasman Gateway opened several million years before Oi1 which seems to make a close connection between Southern Ocean gateway changes and glaciation unsupportable. Data on microfossil distribution indicate that there probably was no warm current flowing southward along eastern Australia, because a counterclockwise gyre in the southern Pacific prevented warm waters from reaching Antarctica. Furthermore, ocean–atmosphere climate modeling indicates that the change in meridional heat transport associated with ACC onset was insignificant at high latitudes. That does not mean that Southern Ocean gateway changes had no impact on climate. The oceanographic changes associated with opening of Drake and Tasman gateways (the so-called 'Drake Passage effect') has been reproduced by climate model investigations of this time interval, and the magnitude of the change anticipated from idealized model results has been reproduced. Opening of Drake and Tasman gateways produces a 0.6 PW decrease in southward heat transport in the southern subtropical gyre (out of a peak of ~ 1.5 PW), and shifts that heat into the Northern Hemisphere. The climatic effect of this shift is felt primarily as a small temperature change in the subtropical oceans and a nearly negligible ($< 3\,^\circ\text{C}$) change in the Antarctic polar ocean region.

Climate and glaciological modeling implicates a leading role of this transition to decreases in atmospheric greenhouse gas concentrations, because Antarctic climate and ice sheets have been shown to be more sensitive to $p\text{CO}_2$ than to other parameters. Consequently, Cenozoic cooling of Antarctica is no longer generally accepted as having been primarily caused by changes in oceanic circulation only. Instead, decreasing CO_2 levels with subsequent processes such as ice albedo and weathering feedbacks (possibly modulated by orbital variations) are seen as significant long-term climate-forcing factors.

The difficulty with this explanation is that existing $p\text{CO}_2$ reconstructions do not show a decrease near the Oi1 event; instead, the primary decreases were much earlier or much later. This suggests that our understanding remains incomplete. Either $p\text{CO}_2$ proxies or climate models could be incorrect, or, more interestingly, they may not contain key feedbacks and processes that might have transformed the long-term greenhouse gas decline into a sudden ice buildup. It is likely, both from a proxy and modeling

perspective, that changes in Earth's orbital configuration played a role in setting the exact timing of major Antarctic ice accumulation, but this area remains a subject of active research.

Problems Posed by the Greenhouse Climate

A strong line of geologic evidence links higher greenhouse gas concentrations with low temperature gradients and warm global mean and deep ocean temperatures through the Cretaceous and early Paleogene. This correlation is fairly strong, but not as strong, for equable continental climates. The concentrations necessary to sufficiently warm high latitudes and continental interiors, however, are so high that keeping tropical SSTs relatively cool becomes a problem. Less obvious and more successful mechanisms have been put forward to solve this problem, such as forcing by polar stratospheric clouds and tropical cyclone-induced ocean mixing. So far, all attempts to accurately reproduce warm Eocene continental interior temperatures lead to unrealistically hot tropical temperatures in view of biotic and geochemical proxy estimates (*see* Determination of Past Sea Surface Temperatures). As summarized in Table 1 and discussed below, many opportunities exist for continued progress to solve these problems.

Cretaceous and early Paleogene tropical SST reconstructions might be biased toward cool values, and other variables need to be constrained to more precisely define the data–model mismatch in the Tropics. For instance, if planktonic foraminifera record tropical temperatures from below the mixed-layer, that is greater than *c.* 40 m depth, rather than 6 m, the low-gradient problem is ameliorated, but such a habitat may be difficult to reconcile with evidence for the presence of photosymbionts. Alternatively, a sampling bias toward warm season values at high latitudes and cold (upwelling) season temperatures in low latitudes could partially resolve the low-gradient problem. Warm terrestrial extratropical winter temperatures tend to rule out a large high-latitude bias in planktonic foraminiferal SST estimates.

A commonly advocated conjecture might appear to resolve the mysteries of greenhouse climates, and it relies on oceanic heat transport in conjunction with high greenhouse gas concentrations. Attention has focused on the challenging problem of modeling paleo-ocean circulations and, especially, on placing bounds on the magnitude of ocean heat transport in such climates.

Climate models have been coupled to a 'slab' mixed-layer ocean model, that is, assuming that the important ocean thermal inertia is in the upper 50 m and that ocean poleward heat transport is at specified levels, to predict equator-to-pole surface temperature gradients. With appropriate (Eocene or Cretaceous) boundary conditions, near-modern pCO_2, and ocean heat transport specified at near-modern values, an equator-to-pole temperature gradient (and continental winter temperatures) is very close to the modern result (Figure 7). With higher pCO_2, tropical SSTs increase (to $\sim 32\,°C$ for 1000 ppm) without changing meridional gradients substantially, and continental interiors remain well below freezing in winter. Substantial high-latitude amplification of temperature response to increases in pCO_2 or other forcings is generally obtained because of the non-linearity introduced by crossing a threshold from extensive sea ice cover to little or no sea ice cover.

Proxy data imply that it was unlikely – but not impossible – that sea ice was present during the warmest of the greenhouse climates; therefore, this sensitivity is probably not representative for the true past greenhouse intervals. Nevertheless, evidence is growing of the possibility of the presence of sea ice and potentially ice sheets during some of the cooler parts of greenhouse climates. With heat transport approximately 3 times to modern in a slab ocean model, and pCO_2 much higher than modern (> 2000 ppm), some of the main characteristics of greenhouse climates are reproduced, but we do not know how to reach such a high heat transport level without invoking feedbacks in the climate system that have not traditionally been considered, such as those due to tropical cyclones.

There is a range, however, of smaller-than-modern temperature gradients that may not imply an increase in ocean heat transport over modern values. Near-modern values of ocean heat transport could support an equator-to-pole surface temperature gradient of $24\,°C$, whereas a $15\,°C$ gradient requires ocean heat transport roughly 2–3 times as large. In other words, warm polar temperatures of $10\,°C$ and tropical temperatures of $34\,°C$ can exist in equilibrium with near-modern ocean transport values, but climates with smaller gradients are paradoxical. Thus even small errors or biases in tropical or polar temperatures affect our interpretation of the magnitude of the climate mystery posed by past greenhouse climates.

In conclusion, two main hypotheses potentially operating in conjunction might explain greenhouse climates. The first is increased greenhouse gas concentration, but we need better understanding of the coupled carbon cycle and climate dynamics on short

Table 1   Greenhouse climate research: key areas to resolve the low gradient and equable climate problems

Tropical proxies may be biased to cool values	What are the true depth habits of planktonic foraminifera? Are 'mixed-layer' dwellers calcifying below the mixed layer?	Could proxies reflect seasonal biases due to upwelling? Are productivity and calcification limited to relatively cool conditions?	How much do preservation and diagenesis alter the proxy signals? Can we only trust foraminifera with 'glassy' preservation? Under what conditions do Mg/Ca produce valid temperatures? Are compound-specific isotope analyses also subject to alteration?
Polar temperature proxies might be biased to warm values	Is there sea ice? Sea ice introduces a fundamental nonlinearity into climate. When did it first form?	Are proxies recording only the warm events or warm season? Is polar warmth a fiction of aliasing?	What is the oxygen isotopic composition of polar seawater? Do hydrological cycle and ocean circulation changes alter this?
Models missing key features	Is resolution in the atmosphere and ocean models too coarse to get correct dynamics or capture the 'proxy scale'? What resolution is enough?	Is there missing physics: e.g., polar stratospheric clouds, enhanced ocean mixing; or incorrect physics: e.g., cloud parametrizations?	Are paleogeographic/ paleobathymetric/ soil/vegetation/ozone boundary conditions correct enough?
Greenhouse gases	How accurate are existing pCO_2 proxies? How can they be improved? No proxy for methane or other greenhouse gases. How did other radiatively important constituents vary?	The sensitivity of climate models to greenhouse gas forcing varies from model to model and the 'correct' values are an unknown, what can we learn from forcing models with greenhouse gases?	Carbon cycle feedbacks. What caused long-term and short-term carbon perturbations? How did climate and the carbon cycle feed back onto each other?
Circulation proxies	Although ocean circulation tracers (e.g., Nd) provide information on the sources and directions of flow, there are currently no tracers that directly reflect ocean flow rates. Can quantitative proxies for rates of deep water formation and flow be developed? Was the ocean circulation weaker or stronger?	Density-driven flows in the ocean are function of temperature and salinity distributions. Most proxies focus on temperature. Can existing techniques for estimating paleosalinity (especially surface) be refined and extended? Can new, quantitative proxies be developed? Was the ocean circulation of past greenhouse climates driven primarily by salinity gradients?	Oceanic carbon storage and oxygenation are largely controlled by ocean circulation. Can integrated carbon–oxygen-isotope-dynamical models reveal unique and important linkages between proxies and ocean dynamics?

and long timescales. The second is a physical mechanism that increases poleward heat transport strongly as the conditions warm.

See also

Abrupt Climate Change. Cenozoic Climate – Oxygen Isotope Evidence. Determination of Past Sea Surface Temperatures. Heat Transport and Climate. Methane Hydrates and Climatic Effects. Paleoceanography. Paleoceanography, Climate Models in. Plio-Pleistocene Glacial Cycles and Milankovitch Variability.

Further Reading

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Biographical Sketch

Matthew Huber is a climate modeler who aims to understand the dynamics and phenomenology of past climates – especially the Eocene. He also enjoys long walks on the beach.

Ellen Thomas' research interests revolve around benthic foraminifera, which she studies in order to investigate the impact of environmental and climate changes on living organisms in the oceans on timescales from millions of years to decades. She studied earth sciences at the University of Utrecht (Netherlands), obtaining a PhD in 1979. She moved to the US to become a postdoctoral researcher at the Department of Chemistry, Arizona State University (Tempe, AZ), and moved on to serve as a staff scientist at the Deep-Sea Drilling Project, first based at Scripps Institution of Oceanography (La Jolla, CA), later at the Lamont-Doherty Geological Observatory (Palisades, NY). From there she went to Wesleyan University (Middletown, CT) as a researcher in the Department of Earth and Environmental Sciences, spent time as a lecturer in marine micropaleontology at the Departments of Earth Sciences and Sub-department of Quaternary Research in Cambridge (UK), then returned to the US as a research professor at Wesleyan University and a senior researcher at Yale University (Department of Geology and Geophysics; New Haven, CT).