PROGRESS IN GREENHOUSE CLIMATE MODELING

MATTHEW HUBER

Earth, Atmospheric and Planetary Sciences, Purdue University,
550 Stadium Mall Drive, West Lafayette, IN 47907 USA
huberm@purdue.edu

ABSTRACT.—This article discusses past greenhouse climates with an emphasis on recent progress in the comparison of models and data in the Eocene. These past climates—about 10°C warmer than modern—provide valuable lessons in the climate dynamics of a world with higher-than-modern greenhouse-gas concentrations, reduced temperature gradients, and a lack of polar ice. Some key points that emerge from this analysis are: 1) Past greenhouse climates are characterized by warmer-than-modern global mean temperatures and winter season warmth, as well as strong polar amplification of warming that generates weak meridional temperature gradients. 2) A trade-off exists between heat transport in a warmer world and a world with smaller temperature gradients; a range exists in which a warmer world with weak temperature gradients can transport as much heat as a cooler world with stronger gradients. 3) These features can be reproduced in models if global mean and tropical temperatures are allowed to increase significantly over modern values. 4) Local radiative-convective feedbacks play an important role, perhaps dominating over transport, or acting in conjunction with transport to maintain relatively warm, weak temperature gradient climates. 5) Past greenhouse climates appear to have many of the modern modes of variability from orbital to interannual. 6) Modeling efforts are most useful when done independently of proxy data; model output can then be compared with proxies to evaluate potential weaknesses in the model. 7) Model-data mismatch might be due to incorrect boundary or initial conditions or resolution, or to missing or incomplete representation of relevant physics. 8) Refinements and new developments in proxies for temperature and other variables offer much potential for discriminating among the next generation of modeled scenarios, thereby allowing a better understanding of the conditions under which greenhouse climates exist.

INTRODUCTION AND MOTIVATION

PALEOCLIMATE MODELING is the application of analytical or numerical techniques to predict the behavior of the climate system's components (i.e. atmosphere, ocean, sea ice), and sometimes their biogeochemical interactions, with the goal of understanding past climate change. Paleoclimate modeling primarily is a tool for exploring the mechanistic functioning of the climate system and for making concrete predictions based on the laws of physics that can be tested against paleoenvironmental proxies. 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, atmosphere and ocean circulation, and heat transport. Only analytical or numerical processes for which accurate conceptual models exists can be modeled. Consequently, failure of analytical or numerical models to match proxy data often highlights gaps in conceptual models of climate and in so doing, provides opportunities to develop new physical understandings that may apply across time scales. Indeed, modeling greenhouse climates pushes the limits of current models, which is simultaneously surprising and expected. Modeling in the absence of data provides neither increased understanding nor testable predictions, so model-data comparison is absolutely necessary to meet the goals of modeling.

The challenge in modeling past greenhouse climate is surprising given that these intervals represent the majority of the past 540 million years. It could be argued that a climate state with temperatures much warmer than modern, without substantial ice at sea level at one or both poles, is the normal climatic mode and that the glaciated, cool state, such as has existed over the past sev-

In *Reconstructing Earth's Deep-Time Climate—The State of the Art in 2012, Paleontological Society Short Course, November 3, 2012. The Paleontological Society Papers, Volume 18, Linda C. Ivany and Brian T. Huber (eds.), pp. 213–262.*

eral million years, is unusual. On the other hand, since we were born into this glaciated climate state, theories of near-modern climates are well developed and powerful, whereas greenhouse climates pose challenges to our understanding. Modern climate-change science (i.e., studies that aim to predict climatic evolution over the next century) focuses on the minutia of short time scales and global temperature variations of tenths of a degree. In contrast, past greenhouse intervals provide climate modelers a means to test fundamental assumptions regarding first-order controls on Earth's temperature, as well as the strengths and weaknesses of current climate models when subjected to large magnitude forcings. This kind of challenge allows models to 'stretch their legs,' which is important given the potentially large climate changes we may expect in the future. A review of what we know of past greenhouse worlds reveals that our capacity to predict and understand key features of such intervals is improving, but still limited. Recent progress has been significant, and past greenhouse climates may provide keys to understanding climate over a dynamic range substantially beyond what can be studied in the modern era.

Background

The primary factors driving regional, latitudinal (meridional), and global temperature distributions have been explored over the course of climate modeling history, and paleoclimate proxy records continue to be a wellspring of innovative theories and models. The two best-documented periods with greenhouse climates are the Cretaceous (145–65.5 Ma) and the Eocene (55.8–33.9 Ma), on which this article will concentrate. 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 Ma) were 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 mid-latitude continental interiors. 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:

(1) Global warmth: global mean surface temperatures much warmer (>10°C warmer) than the modern global mean temperature (15°C) and strangely, these climates experience significant variability;
(2) Equable climates: reduced seasonality in continental interiors compared to modern values, with winter temperatures above freezing; and
(3) Weak, 'low' temperature gradients: a significant reduction in equator-to-pole and vertical ocean temperature gradients (i.e., to ~20°C). Current annual mean surface air temperatures in the Arctic Ocean are about −15°C, and in the Antarctic are about −50°C, whereas tropical temperatures 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 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). However, evidence for reduced meridional and vertical ocean temperature gradients is much more controversial largely due to the nature of the rock and fossil records. Substantially reduced temperature gradients are reasonably well established only for the mid-Cretaceous and the early Eocene.

These warm climates were not lacking in variability; on the contrary, where high resolution records exist, multiple modes of variability, including orbital, inter-annual and sharp transitions, have been found. Superimposed on gradually evolving trends in greenhouse climates is a pervasive variability apparently driven or paced by orbital variations. Interannual variability also has been discovered. In addition, there exist abrupt climate shifts 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. All of these patterns may be studied from a climate modeling perspective to gain a better grasp of processes.

A variety of paleoclimate proxies, including oxygen isotopic paleotemperature estimates as well as newer proxies (e.g., Mg/Ca) and ones that might still be considered experimental, such as TEX_{86}, have been used to reconstruct greenhouse climates. New proxies and long proxy records of atmospheric carbon dioxide concentration (pCO_2) are becoming available. These allow us to constrain the potential sensitivity of climate change to

215greenhouse gas forcing. 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.

The reader is referred to prior reviews of greenhouse paleoclimate modeling as background (Valdes, 2000a,b; Deconto et al., 2000); this review is intended to cover some of the innovations that have occurred in the past decade. Arguably the most important innovation in paleoclimate modeling has been the coupling of dynamic atmospheric, oceanic, and sea-ice component models to achieve a greater level of realism and testability than is possible in an uncoupled framework in which climate components cannot interact with each other. Coupling is crucial because the advancement of our understanding of climate requires theories and models that are independent of the data to avoid circularity, and to make predictions testable against those proxy data. The cutting edge of modeling consists in taking these models and adding interactive terrestrial and ocean biogeochemical components, and incorporating fully dynamic, interactive ice sheets. The special challenges of paleoclimate modeling are balanced by unique opportunities to study the evolution of climate and the Earth system over its full dynamic range. If chosen carefully (i.e., only when the environmental signal provided by proxies is much greater than the noise inherent in the uncertain physics of models), paleoclimate modeling answers questions that are impossible to answer within near-modern settings.

This paper reviews what has been learned from various approaches using uncoupled atmospheric and oceanic models, and then discusses the more powerful methodology of coupled modeling. I focus on the Eocene because most coupled modeling has been focused on that interval, although I will also briefly describe the basic paleoclimate context of the Cretaceous. First, the paleoclimate of these intervals and the means by which we know them are discussed, then the types of models and the methods used in applying them are covered. Significant progress in modeling as measured by model-data comparison is discussed, then I conclude with a brief overview of the horizon of this rapidly evolving field.

This paper is intended to be read in a variety of ways. To get a quick idea of the pattern of each of the past greenhouse climates considered in this review, read the brief, unreferenced Overview below, then move on to the modeling section. More detailed, fully referenced descriptions of each interval are found in the corresponding 'Details' subsection. For those familiar with models and proxy data but interested in learning about progress in the field, skip ahead.

PALEOCLIMATE CONTEXT

Overview

Some general features of Cretaceous climate can be summarized as follows. Continental interior temperatures probably were above freezing year-round, and terrestrial ice at sea level was largely absent. The Cretaceous as a whole was a greenhouse world: most temperature records indicate high-latitude and deep-ocean warmth, and pCO_2 was high. Global mean surface temperatures were more than 10°C above modern, and the equator-to-pole temperature gradient was approximately 20°C. Peak Cretaceous warmth probably lies in the mid-Cretaceous (120–80 Ma), when tropical temperatures were probably between 30 and 35°C, mean annual polar temperatures were above 14°C, and deep ocean temperatures were 10–14°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, and a global thermal maximum does not necessarily denote a maximum everywhere at the same time. However, the accumulated data support an evolution of temperature from a cooler (although still warm compared to modern) early Cretaceous to peak warming by the Turonian (~90–94 Ma), returning to cooler temperatures by the latest Cretaceous. Some multiproxy evidence indicates the presence of (perhaps short-lived) below-freezing conditions at high latitudes and in continental interiors, and the apparent buildup of small-to-moderate terrestrial ice sheets potentially in the early Cretaceous Aptian (125–112Ma) and in the late Cretaceous Maastrichtian (70.6–65.5 Ma).

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 generally were much warmer than modern. Evidence exists for at least brief intervals of cooler-than Cretaceous/Eocene conditions in the deep sea and polar regions. Overall, in the first 10 million years after the asteroid impact, ecosystems recovered while the world followed a warming trend, reaching maximum temperatures between 56–49 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) at ~55 Ma, which warmed the world by ~5°C. The PETM is also associated with a severe shoaling (~2km) of the ocean calcite compensation depth (CCD) and a ≥3.0‰ negative stable carbon-isotope excursion (CIE) that is reflected in marine and soil carbonates. The confluence of ocean acidification, the CIE, and temperature-proxy records leads to the conclusion that the PETM and other hyperthermals represent a massive release of ^{13}C-depleted carbon and CO_2-induced global warming. One explanation for the CIE is the release of 2000–2500 Gt of isotopically light (60 per mil) 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, explosive Caribbean volcanism, or North Atlantic basaltic volcanism. There is some indication of an increase in the vigor of the hydrologic cycle, 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 turnover, while deep-sea benthic foraminifera suffered extinction (30–50% of species).

After the PETM, benthic records and terrestrial records indicate a several-million-year period of cooling followed by a return to extraordinarily warm conditions. The interval of ~52–49 Ma is known as the early Eocene Climatic Optimum (EECO), during which temperatures reached levels unparalleled in the Cenozoic (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 present-day, 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 ~5°C warmer than today, still much less than the extreme polar warming. It is a major challenge to explain how the high latitudes were kept as warm as 15–23°C with tropical temperatures only at ~35°C. 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 and numerical models of climate.

The greenhouse world is not fully understood, nor is the transition into the present cool world ('icehouse world') or the cause of global cooling. Our current understanding of these processes are as follows. Beginning in the middle Eocene (at ~48–46 Ma), the poles and deep oceans began to cool, and pCO_2 may have declined. The causes of this cooling and apparent carbon drawdown are unknown. Recently, a significant warming event was discovered during the late middle Eocene that indicates an interruption in the long-term Eocene cooling trend. Subsequently, the cooling continued and the diversity of planktic and benthic oceanic life-forms declined in the late Eocene to the earliest Oligocene (37.2–33.9Ma). Antarctic ice sheets achieved significant volume and reached sea level by about 33.9Ma, 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 kyr) increase in benthic foraminiferal $\delta^{18}O$ values in the earliest Oligocene (called the 'Oi1 event') has been interpreted as reflecting the establishment of an Antarctic ice sheet. Oxygen isotope records, however, reflect a combination of changes in ice volume and temperature, and it is still debated whether the Oi1 event was primarily due to an increase in ice volume or to cooling.

Cretaceous

Terrestrial data from the North Slope of Alaska (Spicer and Parrish, 1986; Parrish and Spicer, 1988) and northeastern Russia (Spicer et al., 2002), with modern mean-annual temperatures (MAT) of −14°C (Spicer et al., 2008), indicate that high Arctic MAT fell from ~13°C in the mid-Cretaceous (Herman and Spicer, 1996; Spicer et al., 2002) to 2 to 8°C by the Maastrichtian. Warm polar regions with winter temperatures above freezing are also inferred by the occurrence of fossil crocodilians and other aquatic vertebrates at mid-to-high latitudes during the Cretaceous (Markwick, 1998, 2007; Tarduno et al., 1998). The recent discovery of Late Cretaceous frost-intolerant floras in the remote interior of Siberia, far from ameliorating effects of the ocean, allows for a MAT reconstruction of ~12°C, which is 10°C warmer than model simulations (Spicer et al., 2008). While tropical temperatures were initially reconstructed to be quite cool (Barron, 1983), this is now recognized to be an artifact of diagenesis (D'Hondt and Arthur, 1996; Pearson et al., 2001). There is now more isotopic and elemental ratio evidence for substantial low-latitude warming during the Cretaceous (Norris and Wilson, 1998; Wilson et al., 2002; Bice et al., 2003; 2006). Temperature estimates from $\delta^{18}O$ and Mg/Ca values of extremely well-preserved surface-dwelling foraminifera have raised tropical to subtropical temperatures for the middle- (Norris and Wilson, 1998; Clarke and Jenkyns, 1999; Wilson and Norris, 2001; Norris et al., 2002; Wilson et al., 2002; Bornemann et al., 2008) to late Cretaceous (Wilson and Opdyke, 1996; Pearson et al., 2001). TEX_{86} SST data suggest early- (Dumitrescu et al., 2006; Littler et al., 2011; Jenkyns et al., 2012) to middle Cretaceous tropical conditions (Schouten et al., 2003; Forster et al., 2007a, b; Norris et al., 2002) of 32–36°C to over 45°C, depending on which version of the TEX_{86} temperature calibration (Kim et al., 2008, 2010) is applied (see Tierney, this volume). Mid- to high-latitude temperatures during the early (Littler et al., 2011) and late Cretaceous (Jenkyns et al., 2004) appear much warmer viewed through this proxy lens, and are hard to reconcile with the interpretations by some of moderate ice-sheet growth.

Intriguing evidence for seasonally freezing temperatures and/or permanent polar ice includes ice-rafted debris and lonestone-bearing mudstones together with glendonite pseudomorphs in Australia during the early Cretaceous Valanginian through the Albian stages (Alley and Frakes, 1993; Frakes et al., 1995; Price, 1999; Price and Nunn, 2010). Of these deposits, a distinct, two-meter-thick Cretaceous diamictite found in the Flinders Range of Australia represents evidence for glacial erosion (Alley and Frakes, 2003). Although these observations are supportive of cool, higher latitude/altitude temperatures, they are not considered unambiguous evidence of continental ice sheets (Bennett and Doyle, 1996; Hay, 2008). Single erratics and clasts often can be discounted as anomalous dispersal by kelp or storm deposits (as discussed in Markwick, 1998), but glendonite pseudomorphs are more difficult to discount.

Three interpretations of these sedimentological records can be made. First, the appearance of high-latitude warmth from nearly all the other proxies is incorrect and these climates were perennially cooler than most reconstructions. While this would certainly reduce the model-data discrepancies that have pervaded the literature, the overwhelming bulk of proxy evidence leads to the conclusion that this argument is not valid. The second alternative is that these are records of short cold snaps (Price and Nunn, 2010), probably related to orbital configurations that enhanced polar cooling. Whether this hypothesis is valid is debatable (Jenkyns et al., 2012), but the argument is physically plausible and worthy of further investigation. A third alternative is that records of cool, high-latitude temperatures (e.g., glendonite pseudomoprhs and ikaite) represent filters of the seasonal variability that are especially sensitive to winter season temperatures. For example, the presence of ikaite in Svalbard sediments tends to be interpreted as indicating sub-seafloor temperatures of < 4° (Spielhagen and Tripati, 2009) or < 7° (Price and Nunn, 2010), both of which are compatible with the densest water temperatures that might occur in the winter season in the Arctic while still being consistent with other proxy records. In other words, the presence of cold-sensitive proxies, such as pseudomorphs, could simply be revealing preferentially recorded colder winter-season temperatures (4–7°C) because this water becomes dense and sinks, forming a perennial cold layer at depth in enclosed basins.

Records of significant sea-level variations in the Cretaceous also argue for intervals of cooler high-latitude temperatures and ice-sheet accumulation. Recent reviews by Hay (2008) and Miller et al. (2011) provided a comprehensive summary of the greenhouse glacioeustasy debate. Evidence for glaciation rests mainly on inferred eustatic fluctuations from stratigraphic sequences occurring at frequencies too high for tectonic drivers. Some of these sequences also are argued to be linked to stable isotope variability and astronomical pacing (e.g., Gale et al., 2008; Galeotti et al., 2010b; Boulila et al., 2011). Accordingly, opposition to greenhouse glacioeustasy falls into two broad camps: those who question eustatic reconstructions based on ancient stratigraphic successions (Miall, 1992), and those who argue that isotopic data do not support changes in ice volume large enough to account for glaciation (Moriya, et al., 2010; Ando et al., 2009). Initial estimates of eustatic sea-level and inferred ice-volume changes from stratigraphic sequences (Vail et al., 1977; Haq et al., 1987) failed to adequately consider the potential for isostatic adjustments, changes in sediment load, dynamic topography associated with mantle dynamics (Lambeck and Chappell, 2001; Mitrovica et al, 2011), and basin subsidence (Hay and Southam, 1977; Christie-Blick et al., 1990; Hay, 2008). As a consequence, those sea-level estimates appear unreasonably large, requiring volumes of ice equal to or greater than the current Antarctic ice sheet to wax and wane during extraordinarily warm periods

The most convincing records of large magnitude sea-level changes are high-frequency, inferred eustatic oscillations observed during the Early to Late Cretaceous, including some during the peak warming of the Middle Cretaceous from the Western Interior of North America (Gale et al., 2008; Koch and Brenner, 2009), Europe, India (Gale et al., 2002), and central Italy (Galeotti et al., 2010a). Putatively, eustatic sea-level estimates from New Jersey, USA, and the Russian platform for the Late Cretaceous to early Eocene (including some for the Middle Cretaceous), which are assumed correlative with positive changes in low-resolution benthic $\delta^{18}O$ values from other locations (Huber et al., 2002), are interpreted as ~25 meters of sea level change occurring in less than one million years (Sickel et al., 2004; Miller et al., 2005a,b). This issue remains an area of active research.

Additional features of particular interest in the Cretaceous (and to a lesser extent, the Jurassic, 199.6–145.5 Ma) oceans are large, global, carbon-cycle perturbations termed oceanic anoxic events (OAEs). These were periods of high carbon burial that led to draw down 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 was then preserved in the organic-rich sediments known as black shales. At least two Cretaceous OAEs were 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, and polar temperatures in excess of 20°C.

Multiple hypotheses exist for explaining OAEs (Leckie et al., 2002; Sageman et al., 2006; McElwain et al., 2005; Flogel et al., 2011), 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), increased atmospheric pCO_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 potential feedback mechanism is poorly constrained. The emplacement of LIPs was sporadic and rapid on geological timescales, but note that almost three times as much oceanic crust was produced in the Cretaceous (in LIPs and spreading centers) as in any comparable time 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 circulation within isolated abyssal basins might have been important.

As the Cretaceous came to a close, climate fluctuated substantially (Wilf et al., 2003; MacLeod et al., 2005; MacLeod et al., 2011), but it is impossible to know how Cretaceous climate would have continued to evolve because the Cretaceous ended with a bang—the bolide that hit the Yucatan Peninsula at 65.5Ma profoundly perturbed the ocean and terrestrial ecosystems. 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.

Paleocene

The Paleocene is understudied, and much remains to be established about the details of climate evolution during this interval. In mid-latitude continental interiors, annual mean temperatures were ~15°C (Wilf et al., 1998; Wilf, 2000; Markwick, 1998), and the polar regions had annual mean temperatures of ~12°C (Arctic: Bice et al., 1996; Tripati et al., 2001; Antarctic: Poole et al., 2005). Deposits with glendonites, single erratics, and conglomeratic clasts have been found in Paleocene–Eocene sections on Svalbard (Spielhagen and Tripati, 2009). Whereas the erratics and conglomerate casts have multiple explanations, the presence of glendonites implies the occasional presence of cool temperatures in the Arctic. As described previously, it is likely that, given the location of these sequences in isolated Arctic Ocean basins, these cool temperatures represent winter season SSTs of 4–7°C (not annual mean values) that are preferentially represented at depth because this cool water mass is dense and sinks to the bottom. Winter season temperatures in this range would cause no conflict with the perception of the Paleocene as a largely ice-free, greenhouse climate, albeit a cooler one than the Eocene.

Even less is known in the tropics. Temperatures in Tanzania and the central, near-equatorial Pacific (ODP Site 865) were ~30°C (Huber, 2008). Currently, the most exciting paleoclimate record from the Paleocene is from the Cerrejón Formation of Colombia, South America. Head et al. (2009) discovered 58 to 60 Ma fossil vertebrae, estimated to be from eight individuals of the largest species of snake ever found. What makes the study so intriguing is that the authors relate the animal's immense projected size—13 meters long and more than 1 ton in weight—to a minimum annual mean temperature of 31 to 33°C. They extrapolated this temperature by using an empirical relationship between temperature and size derived from modern organisms. This temperature range is now corroborated by nearby TEX_{86} records (Jaramillo et al., 2010). These results suggest an equator-to-pole temperature gradient in the Paleocene of ~20°C, which is weak, but approximately achievable by climate models (Huber, 2009; Huber and Caballero, 2011).

PETM

The record of this interval 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 under 500 years, with a recovery over >125kyr and a total duration of ~170kyr (Rohl et al., 2007). The negative CIE was at least 2.5–3.5 per mil in deep oceanic records and 5–6 per mil in terrestrial and shallow-marine records (Koch et al., 1992; Dickens, 2011). 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 (Kennett and Stott, 1991); about 4–5°C in the deep sea (Zachos et al., 2001), in equatorial surface waters (Zachos et al., 2005) and the Arctic Ocean (Sluijs et al., 2006; Weijers et al., 2007); and by about 5°C on land at mid-latitudes in continental interiors (Fricke, 2003; Fricke and Wing, 2004; Wing et al., 2005). Pagani et al. (2006b) derived stable hydrogen (δD) and carbon ($\delta^{13}C$) isotope records of terrestrial-plant and aquatic-derived n-alkanes from the Arctic Ocean that reflect changes in hydrology, including surface water salinity and precipitation. They found some equivocal evidence of an enhanced hydrological cycle during the PETM that has been confirmed in other work (Harding et al., 2011).

Given the probable range of climate sensitivities for modern conditions, the implied carbon dioxide release for the PETM and other hyperthermals had to be massive in order to cause the observed temperature changes in an already greenhouse-gas-dominated world. Alternatively, climate sensitivity during the early Eocene was higher than average values estimated for today (Dickens et al., 1995; Higgins and Schrag, 2006; Pagani et al., 2006a; Zeebe et al., 2009). The source of carbon for the PETM and subsequent hyperthermals (discussed below) theoretically is constrained by the size of the CIE, the carbon isotope value of the source carbon, and changes in the CCD (Zeebe et al., 2009; Panchuk et al., 2008; Dunkley Jones et al., 2010).

Two dominant hypotheses for extreme warming have emerged with distinctly different implications for the climate system during greenhouse climates. A common explanation invokes a massive release of methane gas precariously trapped in marine sediments as methane clathrates (Dickens et al., 1995). In this scenario, clathrates are destabilized, and methane is either oxidized in the water column and/or released to the atmosphere, where it is quickly converted to CO_2. The methane hypothesis requires the release of over 4,000 PgC during the PETM (Zeebe et al., 2009), driven by rapid, pre-event warming and/or threshold temperatures in very deep waters (given that methane clathrates would only be stable in deep pelagic sediments where high sedimentary pressures offset the enhanced warmth of the Eocene deep ocean) (Archer et al., 2004; Archer and Buffett, 2005). The cause and nature of this pre-event warming has not been defined or well identified in paleotemperature reconstructions (Sluijs et al., 2007), and is further constrained by the appearance of orbital controls on the occurrence of hyperthermals. Whether or not the necessary quantity of methane was even available during Greenhouse times is another important debate (Buffet and Archer, 2004; Archer et al., 2004; Dickens, 2011).

An alternate hypothesis for the appearance of hyperthermals calls on the irreversible degradation of biomass (Kurtz et al., 2003; Higgins and Schrag, 2006; DeConto et al., 2012) leading to direct release of CO_2 into the atmosphere. This has been proposed to occur as tropical ecosystems crossed thermal thresholds (Huber, 2008), or as high-latitude permafrosts warmed, dried, and subsequently experienced oxidation of accumulated soil organic carbon (DeConto et al., 2012). However, the apparent coincidence of hyperthermals with combined high eccentricity and high obliquity (Lourens et al., 2005) implicates changes in high-latitude seasonal insolation as the ultimate trigger (DeConto et al., 2012). In this latter scenario, high-latitude permafrost catastrophically melted when the region crossed a critical temperature threshold due to the combination of slow, CO_2-induced warming and warm orbital geometries. CO_2 was then released from the oxidization of permafrost organic carbon. Model estimates for the amount of available permafrost carbon (DeConto et al., 2012), as well as modern rates of permafrost carbon sequestration and release (Schuur et al., 2008; 2009), accommodate geochemical requirements dictated by carbon cycle modeling (Panchuk et al., 2008; Cui et al., 2011) and lead to much higher CO_2 input and lower estimates of Earth System climate sensitivity (Pagani et al., 2006a; Dunkley Jones et al., 2010).

Presently, carbon cycle models do not allow discrimination between various carbon-cycle perturbation hypotheses. In lieu of novel results from new geochemical proxies, determining which of these hypotheses represent the primary hyperthermal trigger will depend on an improved understanding of how ocean pH changed and affected the global CCD, as well as a clearer understanding of the limits of methane and terrestrial carbon reservoir magnitudes and accumulation rates during a much warmer world. Nonetheless, these hypotheses suggest that extreme climate variability is an aspect of greenhouse climates, particularly as the climate system evolves from the colder spectrum of the greenhouse world toward the warmest extremes and threshold conditions. If methane or biomass is the primary source of carbon responsible for extreme warming, it suggests that Earth System climate sensitivity to CO_2 (i.e., climate sensitivity that includes slow and fast feedbacks) was high during the Eocene (Higgins and Schrag, 2006; Pagani et al., 2006a; Zeebe et al., 2009), even though cryospheric effects were absent.

Early Eocene

The first few million years of the Eocene saw a return to somewhat cooler conditions than the late Paleocene. Temperatures in continental midlatitudes dropped from 15–20°C in the earliest Eocene to 10–15°C by 54Ma (Bao et al., 1999; Wing, et al., 2000; Wing and Harrington 2001) Conditions warmed again in the EECO interval between ~53–49Ma. Annual-mean and cold-season continental temperatures were substantially warmer than modern, while meridional temperature gradients were greatly reduced (Wolfe, 1995; Greenwood and Wing, 1995; Barron, 1987). Reconstructions of warm climates on land are confirmed in the marine realm, with bottom water temperatures 10°C higher than modern values (Miller et al., 1987; Zachos et al., 2001; Lear et al., 2000). This implies that early Eocene winter temperatures in deep-water formation regions, located at the surface in high latitudes, could not have dropped much below 10°C , consistent with the high-latitude occurrence of frost-intolerant flora and fauna (Greenwood and Wing, 1995; Spicer and Parrish, 1990; Hutchison, 1982; Wing and Greenwood, 1993; Markwick, 1994, 1998).

As summarized in Huber (2008), SSTs of ~35°C are now reconstructed in the early Eocene tropics (Pearson et al., 2001, 2007; Tripati et al., 2003; Tripati and Elderfield, 2004; Zachos et al., 2003). At high northern latitudes, TEX_{86} (and its minor variants such as TEX_{86}') records from the ACEX core in the Arctic Ocean reveal snapshots of warm temperatures. During subsequent hyperthermal events, SSTs are reconstructed to range from 25–15°C. In the early Middle Eocene, TEX_{86} SSTs dropped to 8–12°C from early Eocene highs of 15° to 25°C. At high southern latitudes, Hollis et al. (2009), Creech et al. (2010), and Bijl et al. (2009) found evidence of 30–35°C SSTs in the South West Pacific between 55–65°S in the early Eocene.

These hot temperatures have major implications for understanding past climate dynamics and, particulary, of the equable climate problem if they are quantitatively accurate (there is little doubt that their trends are correct). One outgrowth of the increasing study of paleotemperature proxies and improved understanding of the myriad processes and mechanisms that affect proxies has been the unfortunate realization that large and difficult-to-quantify uncertainties persist in proxy interpretations (Shah et al., 2008; Ingalls et al., 2006; Herfort et al., 2006; Kim et al., 2010; Liu et al., 2009; Pearson et al., 2001; Huguet et al., 2006, 2007;Wuchter et al., 2004, 2005; Huguet et al., 2009; Trommer et al., 2009; Turich et al., 2007; Eberle et al., 2010; and see other papers in this volume). Of particular concern is the need to extrapolate calibrations out to temperatures and environmental conditions far beyond modern values. This can be especially difficult in the tropics, where conditions likely were much warmer than the warmest range of core-top calibrations of 30°C. This either requires extrapolating beyond core-top calibration, or using mesocosm calibrations that extend up to 40°C. For the Tanzanian TEX_{86} records of Pearson et al. (2007), peak early Eocene temperatures are either 35°C when extrapolating from the core-top TEX_{86} (GDGT2-index) calibration, or 39.4°C using the mesocosm-based TEX_{86} (GDGT2-index) calibration (Kim et al., 2010). The warmest value recorded by $\delta^{18}O$ in planktonic foraminifera of the same age (49.5 Ma) is 31.5°C. So, from what some might consider the best records at one time and one locality, reasonable arguments might be made to interpret tropical near-surface temperatures to be 31.5–39.4°C.

On the other hand, sometimes different proxies in a region show a remarkable level of congruence and temporal consistency—for example, in the southwest Pacific Ocean (Bijl et al., 2009; Hollis et al., 2009; Liu et al., 2009). However, even the congruence of these records may not prove their accuracy given their arguable lack of consistency with other records. For example, the presence of 11°C South Atlantic temperatures (Ivany et al., 2008) in the same latitude band as 30°C temperatures in the South Pacific (Bijl et al., 2009; Hollis et al., 2009; Liu et al., 2009) raises questions about the proxy interpretations. The occurrence of 10–14°C deep-ocean temperatures (Zachos et al., 2001) requires that some regions see temperatures fall to this value at least in winter, in agreement with the results of Ivany et al. (2008), but South Pacific records seem to preclude temperatures this cold unless open-ocean seasonality was enormous (a 30°C seasonal range). As recognized in many studies, seasonality and regional variation due to ocean-current heat transport are important considerations that may help reconcile the different proxies at high latitudes (Hollis et al., 2009), but serious, unexplained discrepancies persist at low latitudes (Huber, 2008; Liu et al., 2009). Despite these uncertainties, the warmth of this interval is unequivocal and difficult to explain.

Intriguingly, this warm period was subject to strong variability. Within the EECO are alternating warm and very warm (hyperthermal) periods. To date, it appears that the Eocene Thermal Maximums (ETM) 2 and 3 exhibit many of the same characteristics of the PETM, including a transient warming, a carbon isotope excursion, benthic foraminiferal assemblage changes, and dissolution horizons (Zachos et al., 2004; Lourens et al., 2005), implying that similar processes were responsible for each event. However, the magnitude of the CIE for each subsequent warming event decreases (Sluijs et al. 2009; Stap et al., 2010). Hyperthermals also appear to correspond to similar orbital geometries (Lourens et al., 2005; DeConto et al., 2012), and support the notion that hyperthermals were driven by a common trigger. 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. Given the normally assumed range of climate sensitivities (Pagani et al., 2006a), the implied release of carbon dioxide would have to have been massive to cause temperature changes that large in an already greenhouse-gas-dominated world.

Middle to late Eocene transition to the Icehouse

On top of a long-term cooling trend beginning at ~49 Ma, a series of transitions are noted. A substantial cooling in the Arctic to ~10°C occurred in the so-called "Azolla interval" at 48.5 Ma (Brinkhuis et al., 2006), followed by the first appearance of Arctic sea ice at 47.5 Ma (Stickley et al., 2009). Temperatures in the Southwest Pacific were found to decrease into the middle Eocene (Burgess et al., 2008), but apparently remaining near the balmy value of 20–26°C by the end of the Eocene (Bijl et al., 2009; Liu et al., 2009).

The middle Eocene climatic optimum (MECO;

~41–42 Ma) was first observed in a deep-sea sediment core from the Southern Ocean (Bohaty and Zachos 2003), but subsequent studies have verified a warming trend during the same time period at other locations, including the Demerara Rise in the Atlantic, Seymour Island, the Antarctic Peninsula, and Italy (Sexton et al, 2006; Ivany, 2007; Jovane et al, 2007). The $\delta^{18}O$ and $\delta^{13}C$ records in these studies indicate that the MECO was one of the most rapid periods of warming during the Cenozoic. The short-term nature of the event and the records showing joint changes in temperature and carbon cycles are consistent with the theory that this event was not driven by paleogeographic changes, but was driven by fluctuations in the concentration of atmospheric CO_2 (Bohaty and Zachos, 2003; Bijl et al., 2009).

The Eocene–Oligocene transition (from ~42 to 30 Ma) is marked by: stepwise patterns of cooling (Zachos et al., 2001; Feary et al., 1991; Lear et al., 2000; Dallai et al., 2001, Wolfe, 1994), changes in ocean productivity (Diester-Haass and Zahn, 2001; Salamy and Zachos, 1999), and by profound biotic change (c.f., Berggren and Prothero, 1992, papers in Prothero et al., 2003). The earliest Oligocene is marked by an apparent slight increase in seasonality (Ivany et al., 2000; Eldrett et al., 2009), a rapid increase in Antarctic glacial volume to roughly modern total ice volume (Miller et al., 1987; Zachos et al., 1996; Barker et al., 1999; Lear et al., 2000; Billups and Schrag, 2003), a massive CO_2 drawdown (Pearson et al., 2009; Pagani et al., 2011), and an abrupt increase in productivity of the world oceans associated with a global drop in the carbonate compensation depth of >1 km (van Andel et al., 1975; Shipboard Scientific Party, 2001; Lyle et al., 2005).

Until recently, surprisingly little was known about how SST changed during this transition. Liu et al. (2009) utilized U^{K}_{37}' and TEX_{86} SST proxy records from multiple marine sites spanning high to low latitudes to document significant SST changes during the transition. Both proxy records suggest that high-latitude SSTs averaged ~20°C near the end of the Eocene, ~10°C higher than previous oxygen-isotope-derived reconstructions. Through the transition, average high latitude SSTs cooled by 4.8±2.5°C, with maximum changes up to 10°C. Low latitudes were much warmer than modern, and exhibited relatively little cooling (0.5–3°C) during the transition, which is in agreement with the results of Pearson et al. (2008) and Lear et al. (2008) in Tanzania.

There are several proposed causes of this final transition to a world with substantial glaciation, including a Southern Ocean gateway opening, decreases in greenhouse gas concentrations, orbital configuration changes, and ice-albedo feedbacks. As described subsequently, the preponderance of evidence suggests that the drop in greenhouse gas concentrations (DeConto and Pollard, 2003; Huber et al., 2004; Stickley et al., 2004; Huber and Nof, 2006; Pearson et al., 2009; Pagani et al., 2011) was the dominant cause of this event, but the jury is still very much out (Barker and Thomas, 2004).

MODELS AND THEIR APPLICATION

The role of models and of proxies

Generally speaking, the role of models in paleoclimate either is to provide information necessary to fill a gap in interpretations from proxies (this is usually if the model is considered accurate), or to provide a self-consistent, physically based framework for comparison against proxies. If proxies and models agree, then this suggests that the physics (or chemistry or biology) that is generally considered accurate is indeed accurate. This provides an indication that models developed for modern climate can be extended past the range for which they were initially intended and tuned.

It is much more interesting and informative when proxies and models are in disagreement with each other. Assuming that the proxy data are being interpreted correctly, this implies that some aspect of the model has been falsified. Sometimes this falsification indicates that a part of the model is incorrect, and sometimes the falsification indicates that the model is incomplete (Table 1). As should be clear from the role of models and proxy data in this process, the less that proxies are used as inputs to a modeling study, the easier it is to falsify the model. As much as possible, modeling efforts are independent from data, and instead, they generate pseudo-data (model output) for comparison with proxies.

Most model-data comparisons focus on temperature because this is the primary climate variable, but the list of comparisons can be much longer, and includes: precipitation, salinity, wind or current strength and direction, as well as floral, biome, or sedimentological distributions. The saying goes, “All models are wrong, but some are useful.” This is certainly true, and offers a useful guide to understanding the role of modeling. But, TABLE *1*.—Potential explanations for data-model mismatches.

Category 1: Limitations of models with perfect physics	Is resolution in the atmosphere and ocean models too coarse to get correct dynamics or capture the 'proxy scale'? What resolution is enough?
	Are paleogeographic/paleobathymetric/soil/vegetation/ozone boundary conditions or initial conditions correct enough?
	Could a slight tweaking of a poorly constrained free parameter in the model dramatically improve the match with data?
Category 2: Missing or incorrect physics	Are there missing processes: e.g. polar stratospheric clouds, enhanced ocean mixing due to tropical cyclones or tides; Or incorrect physics: e.g. cloud parameterizations are substantially incorrect (wrong sign of feedbacks)?

it might be helpful to recognize an important corollary to this statement that arises in the field of paleoclimate. When investigating past climates we do not find old thermometers buried in ocean sediments or fossilized rain gauges in terrestrial strata. On the contrary, proxies are used to estimate paleoclimate variables, and proxies are, at heart, models. Some proxies, such as $\delta^{18}O$ of carbonate, have a direct physical underpinning in quantum mechanics, but the presence of vital effects, variations in seawater $\delta^{18}O$, variations in depth of production and seasonality, and post-burial diagenesis, render the process of interpreting an SST a rather complex modeling exercise in which each of these variables is accounted for. Some proxies have weak theoretical foundations and additional complexities in their interpretation. In any event, it is important to recognize that all proxies are models, and consequently, one could say, "All proxy records are wrong, but some are useful." This fact has been recognized in various forward proxy-model efforts of the type pioneered by Sellwood and Valdes (2006) and Schmidt (1999). Such forward modeling of variables measured in the geological record, such as climate sensitive facies or the $\delta^{18}O$ of foraminifera, represents an explicit recognition that proxies are models and model-data comparison becomes a better posed problem if the variables being compared are consistent with each other. It is conceivable that much model-data mismatch may disappear if forward proxy-model predictions are compared directly with proxies without introducing the intermediate step of projecting proxy records through modern statistical calibrations to retrodict past climates. Models and proxy data form a continuum, and integrated approaches are necessary to make progress.

After a model-data comparison has been carried out and weaknesses in the model found, normally one is left with the two categories presented in Table 1. Either the model is correct but is either not being driven with correct boundary or initial conditions or the simulation was not carried out at an appropriate resolution, or the model is missing a correct or complete representation of relevant physics. It should be noted that collection of proxy data can improve the initial or boundary condition data sets to alleviate problems of the first kind, but the choice of model resolution is up to the modeler, the model, and computational resources. Given that it can take several years of continuous integration to complete a fully coupled simulation of several thousand (model) years, resolution is per force often on the coarse end of reasonable. Even an otherwise perfect model may not capture the interactions of fine-scale dynamical features (i.e., eddies) with the large-scale circulations properly and hence, produces a poor model-data fit. From a modeling perspective, none of the options within the first category of Table 1 are particularly interesting—one can always argue that the model is acceptable, but it was forced with poor boundary conditions, or it would produce correct results if run at a higher resolution.

Errors from the second category in Table 1 are much more interesting, and are the potential route for paleoclimate proxy data to improve climate models. It is an interesting matter of the sociology of modelers and proxy-dataset generators that the former are likely to assume that a model-data discrepancy is likely to be of category 1, whereas the latter are more likely to leap immediately to the conclusion that there is something fundamentally wrong with model (category 2 errors). While no definitive methodology exists to separate category 1 and 2 errors, the latter have several characteristics. First, vigorous attempts to rule out category 1 errors must be made. Resolution, initial, and boundary conditions should be varied across a wide range of reasonable values. Multiple models and multiple proxy data sets should be used. But, if at the end of extensive study, model-data discrepancies of a magnitude significantly larger than the model-data difference in modern control simulations continue to be found in multiple models, and at a range of model resolutions and in multi-proxy data sets, then perhaps a category 2 error has been identified. This is the real potential value of paleoclimate data in improving climate models, but the bar is very high. No proxy data are directly used to improve paleoclimate models, the process is indirect. It requires extensive multi-model and multi-proxy data comparisons, and a process of deciphering where models may be seriously limited to reach the conclusion that models are seriously deficient.

Overview of the hierarchy of models

Paleoclimate modeling, as in all modeling, is best approached as an iterative processes driven initially and ultimately by interesting data that pose a puzzle. It begins with comparison to the predictions of simple theories, then moves to models of intermediate complexity, and continues on to models of greater sophistication. To achieve the goal of greater understanding, though, this process of complexification also must be run in reverse. The balances in comprehensive models must be diagnosed and simplified to their essence in order to elucidate fundamental physical processes that can be understood, bounded, and tested in models of all levels of complexity. Complex models may fully represent all important interactions, but may be so complex that they cannot be understood. Simple models may be fully understood, but lacking in necessary processes. It is only by climbing up and down the hierarchy of models that the necessary physics can be represented and understood.

Three families of models: simple, intermediate, and full.

Simple models.—Simple models are either numerical or analytical treatments in which the assumed important physics of the system has been drastically reduced in complexity. Such models may either include some measure of atmospheric or ocean dynamics (i.e., using Ekman and Sverdrup theories to predict upper ocean currents or quasi-geostrophic theory to understand mid-latitude storms), or may abstract such features entirely and adopt an analogue approach as exemplified by so-called box models.

Intermediate complexity models.—Intermediate complexity models, often coupled, operate either on a reduced domain (i.e., the upper ocean), with reduced dimensionality (i.e., two dimensions), with simplified atmospheres, at very low resolution, or using lower order equations (i.e., the Large-Scale Geostrophic Model), and frequently, all these simplifications are made (Claussen et al., 2002). The main reasons that earth-system models of intermediate complexity (EMICS) are used are to reduce the computational burden and to decrease the complexity of the problem, thus making it more tractable to understand the relevant physics. These tools can be useful for understanding the functioning of physical or biogeochemical processes. EMICS of today bear a strong similarity to the full complexity models of the previous decade. Partially, this is due to the sliding-scale nature of models: all models are gross simplifications of the real world, and with ever-increasing computational power and more sophisticated equations being used in these models, one decade's state of the art model is the next decade's model of intermediate complexity. The limitations introduced by reducing the complexity of the system or studying it with low resolution have serious consequences that must be acknowledged. On the other hand, these models are usually computationally cheap and a wide range of parameter space can be explored, yielding a greater feel for the overall sensitivity of model results to uncertainty in model physics or initial and boundary conditions.

Full complexity models.—Full complexity climate models simulate the dynamics of three-dimensional ocean or atmosphere circulation using so-called primitive equations. These models typically are referred to as general circulation models (GCMs), with the name alluding to the fact that the models have sufficient sophistication to represent the general circulation of the major fluid envelopes that make up the climate system. Such a sophisticated approach is advantageous when the physics governing the important dynamics are unclear, in which case, it is best to include all potential processes for the most complete window into past dynamics. It also can be useful when trying to compare with proxies, since such models approximate the actual physical setting in which such proxies are found to a substantial degree. Nevertheless, issues of resolution in the model and completeness of model physics that arise in models of intermediate complexity are not solved by using full-complexity models, only alleviated somewhat. In addition, with the level of detail that is possible within these models and the computational burden entailed by this sophistication, uncertainties introduced by sensitivity to model boundary and initial conditions, as well as model parameters, become a serious limitation relative to the models of intermediate complexity. When ocean (O) and atmosphere (A) GCMs are coupled together, they are called fully-coupled models.

On the other hand, application of simple dynamical models has received relatively little attention in the paleoclimate literature in recent decades, which is surprising given that these techniques did and still do play a significant role in filling in gaps in our understanding between full GCMs and box models. These techniques include wind-driven circulation theory using the Ekman and Sverdrup relations or use of the linearized momentum equations to predict vertical and horizontal ocean currents. Lying somewhere between these techniques and intermediate complexity OGCMs are shallow-water equation treatments of atmospheric or oceanic dynamics that can be directly compared with analytic techniques, but are numerical and two or two-and-a-half dimensional in nature. Some discussion of these simple methods is covered below, but most of the remainder of this paper will focus on full climate models with an emphasis on coupled models.

Types of full climate models

Four types of full dynamical numerical models have been used to address the questions posed by past greenhouse climates: (1) atmospheric (A) GCMs driven by fixed SST distributions; (2) AGCMs coupled to mixed-layer, "slab" oceans, with specified mixed-layer thickness and meridional heat fluxes or diffusivities; (3) ocean (O)GCMs usually driven by SST, precipitation minus evaporation (P-E), and wind fields taken from the former types of experiments; and (4) fully coupled ocean-atmosphere-sea ice dynamical models (CGCMs).

Fixed SST AGCMs.—In fixed SST AGCM experiments, the SST distribution is specified, and the object is to explore what atmospheric features (e.g., rainfall, wind, outgoing longwave radiation, terrestrial temperature patterns) exist in equilibrium with such a distribution, and the sensitivity of these features to other factors (e.g., topography). The key physical assumption in these experiments is that the thermal inertia of the ocean is so much larger than that of the atmosphere that the ocean temperatures may be considered climatologically constant, although seasonally varying. By changing the SST distribution that drives the model, assumptions about the role of ocean heat transport can be tested because the ocean heat transport implied by a fixed SST distribution is calculable as a residual from an AGCM simulation. When SSTs are specified, $p\mathrm{CO_2}$ may be changed as a sensitivity parameter, although the results must be interpreted cautiously in light of the fact that the top-of-the-atmosphere radiative balance may not be consistent between experiments. In specified SST experiments, sea ice cannot change dynamically.

Slab Ocean AGCMs.—In slab ocean AGCM experiments, SSTs are predicted, so it is possible to test SST sensitivity to $p\mathrm{CO_2}$ or other parameter variations (e.g., orbital parameters). It is explicitly assumed the ocean can be represented as a mixed layer of specified depth, i.e., that the important ocean thermal inertia is in the upper 50 m or so, and ocean poleward heat transport is at specified levels (often close to modern or higher). Ideally, the pattern of mixed-layer depth and poleward heat transport is derived from that in a coupled model (in which case it has a physical basis), but in the vast majority of cases, it is determined in an ad-hoc fashion. Sea ice normally is interactive in such studies, although sometimes it is thermodynamic and sometimes fully dynamic. Studies that demonstrate substantial high-latitude amplification of temperature response to increases in $p\mathrm{CO_2}$ or other forcings often do so because of the nonlinearity introduced by crossing a threshold from extensive sea-ice cover to little or no sea-ice cover (e.g., Sloan and Rea, 1996; Otto-Bliesner and Upchurch, 1997; Peters and Sloan, 2000; Beerling et al., 2011). It is important to keep in mind that proxy data imply (but do not rule out) that it was unlikely that sea ice was present at all during past, warm greenhouse climates (e.g. Markwick, 1998), therefore, this sensitivity is probably not representative of actual greenhouse climate systems.

Ocean-only modeling.–Paleoceanographic modeling may be used to investigate specific questions such as: (1) What was the current structure and tracer distribution of a given past time interval? (2) How sensitive are major oceanographic features (e.g. the Gulf Stream or the meridional overturning circulation (MOC)), to boundary condition changes? (3) How might modes of variability (e.g., interannual, interdecadal, and millennial-scale variability) change under different conditions? It may also answer some more general questions, such as: (4) What is the role of the ocean in governing the Earth System's response to greenhouse gas (GHG) forcing, and what is the ocean's role in GHG storage and release through Earth's history? and (5) To what degree are ocean heat transport (OHT) changes important for understanding climate change, and what is OHT sensitive to (GHGs, gateways, multiple equilibria)?

Oceanic vertical diffusion is one of the most important and most contentious issues in modern oceanography, and it is likely that it was even more important in Earth's past. The absolute background value of the diffusivity, and the physical dependencies that increase it above this background value, are much debated. While there is a bit of spread for an estimate of this background value, a reasonable observed number is .1 cm^2/s, and values in this range should be the default in paleoceanographic applications. Values 10–100 times greater than this were once commonly used even in modern simulations, which highlights deficiencies in those models. The inclusion of state-of-the-art treatments of diapycnal and isopycnal eddy mixing are important, and currently, it appears from the point of view of the tracer circulation and tracer properties that a combination of the KPP and Gent-McWilliams treatments of these processes is best (Gent et al., 1998). There is substantial room for improvement in these parameterizations, and ideally, several vertical diffusion methods should be tested because the resulting ocean circulations can be fundamentally different (e.g., Nilsson et al., 2003). Furthermore, the important issue of how tidal dissipation may have varied in the past has been neglected.

Coupled GCMs.–In coupled GCMs (CGCMS) the atmosphere and ocean, as well as sea ice, interact according to prognostic, dynamic equations, and they inherently produce self-consistent energy and momentum transfers between components. In these models, one or both parameters can change, unlike in other modeling approaches, and the simulation is able to go in whatever direction physics, tuning, and boundary conditions take it. In this sense, the simulations are both the worst and best of all modeling configurations. Worst in the sense that, with so much freedom, the coupled simulations can differ strongly from the proxy data. In that same sense they are also the best, since this allows falsification against proxy data.

MODELING METHODS

Boundary conditions

Models require boundary conditions to run. Some of these boundary conditions are derived from geological observations or proxy-based reconstructions (i.e., topography); others are inferred or calculated based on physical, biological or geochemical arguments (e.g., ozone distributions). For a fixed SST AGCM simulation, these boundary conditions usually include: SSTs, topography, trace gas concentrations, solar constants, a vegetation distribution (may be specified based on reconstructions or predicted by a model), soil properties, and other parameters. For ocean models, these include heat, moisture, and momentum fluxes (or state variables) passed to the ocean model throughout the simulation, as well as static features such as bathymetry. Ocean bathymetry is a boundary condition that has been a subject of concern in the construction of paleoceanographic modeling experiments (Bice et al., 1998). Some studies have employed a flat-bottom configuration, eschewing bathymetric variation altogether (Bush and Philander, 1997a,b; Barron and Peterson, 1991), and dynamical arguments suggest that most general ocean-current patterns and transports are not sensitive to bathymetric variations (summarized in Bush, 1997a). Nevertheless, the details of some currents certainly involve bathymetry (Bice et al., 1998), and perhaps more importantly, the degree of upwelling from the abyss is governed to a large extent by the presence of abyssal bathymetric rises (e.g., geostrophically balanced meridional flows along ocean ridges), and thus some representation is probably important even for the gross features of the ocean circulation (e.g., Toggweiler and Samuels, 1995; Vallis, 2000). The freshwater (salt) forcing of the ocean potentially is a important variable, with potential parameters including the placement of continental runoff into marginal seas and in determining the overall meridional gradient of precipitation minus evaporation (P-E). The former issue is important for determining the exact location of deep-water formation (Bice et al., 1997), and the latter is a factor in determining the strength of the MOC. Neither of these issues proves fatal to the study of paleoclimates with ocean models.

The atmosphere is coupled to the ocean through the air-water interface: i.e., via sea surface temperatures (SSTs), and fluxes of sunlight, heat, water, and momentum. It is coupled to the biosphere biogeochemically through fluxes of sunlight, carbon dioxide, dust, and traces of organic and inorganic chemicals. Thus, to understand the ocean's dynamics and phenomena, and to predict its role in climate and biogeochemical cycles, it is necessary to constrain the state of, and fluxes through, the interface between ocean and atmosphere (Seager et al., 1995; Power and Kleeman, 1994). This has important consequences.

In ocean models, fluxes traditionally have been handled through the use of so-called Haney restoring (Haney, 1971) on temperature and the specification of net virtual salt fluxes (mixed boundary conditions). In general, the weaker the restoring, the more closely this boundary condition mimics atmospheric interaction, but mixed boundary conditions have been shown to be susceptible to spurious multiple equilibria (Saravanan and McWilliams, 1995; Weaver and Sarachik, 1991; Zhang et al., 1993; Power and Kleeman, 1993,1994). When the SST distribution is fixed in an OGCM experiment, the implicit assumption is that the atmosphere has an infinite heat capacity relative to the ocean, opposite to the assumption in fixed SST AGCM experiments. In a steady-state and zonal average sense, the surface temperature gradient must map into the vertical temperature gradient, thus, bottom water temperatures in such experiments cannot be cooler than the coldest surface value. In fixed SST AGCM experiments, neither the coldest temperatures (upper atmospheric), nor terrestrial temperatures are fixed, so winds, heat transport, and surface land temperatures can be verified independently. In a fixed SST OGCM experiment, however, it is not clear how to explore the factors maintaining temperature distributions and heat fluxes. This highlights one fundamental difference between the ocean and the atmosphere–the atmosphere has two free boundaries, one at the top and one at the bottom, whereas the ocean has the top as its only free boundary.

Allowing these SSTs to vary somewhat by applying a strong restoring boundary condition alleviates this problem only marginally. As this fixed assumption is relaxed by using long restoring times, so-called bulk forcing (Brady et al., 1998, Doney et al., 1998), or other methods (Nong et al., 2000), the top boundary specification better approximates a real atmosphere, and better estimates ocean heat transport. The use of bulk forcing (Brady et al., 1998, Doney et al., 1998), or inclusion of even a simple representation of atmospheric fluxes (as in Nong et al., 2000; Najjar et al.. 2002) is preferable in those cases in which a coupled model is not used. Surface wind stress patterns are normally specified based on modern conditions, a theoretical distribution, or produced by an AGCM. Since a non-interactive atmosphere (one in which winds, temperature, and freshwater fluxes cannot adapt to changing ocean conditions) is inherently unrealistic, in most cases, a coupled model is preferable to an uncoupled model, even when the focus is strictly oceanographic (Doney et al., 1998). This also allows for a more realistic pattern of surface state and flux variations, and allows for OHT to be a well-posed, prognostic variable, which it is not if surface temperature patterns are specified (Huber et al., 2003; Nong et al., 2000; Seager et al., 1995).

A minimum requirement for realism in any climate model simulation is that it be a possible solution in a coupled system, because the real atmosphere and ocean are coupled. Thus, the oceanic or atmospheric heat transport produced by the circulation should not be substantially out of equilibrium with the heat transport implied by the surface heat-flux divergence patterns. Studies of modern climate using uncoupled models avoid these difficulties because the forcing at the top or bottom boundary is well known, and the goal is to understand the flows in equilibrium with this distribution: these flows are also observationally constrained and thus, the models are testable. However, in the deep past, constraints on circulation rates, zonal temperature gradients, heat transport, precipitation, and evaporation are weak, and the state of the fluxes at the boundary between ocean and atmosphere are not well known. Consequently, there are few means to provide an independent check on the predictions of uncoupled GCMs, and no guarantee of energetic self-consistency in the simulations. An uncoupled model-simulated circulation pattern that would collapse if coupled cannot reveal much about any real past climate. Therefore, regardless of whether a fully coupled model is used, a minimum requirement of uncoupled GCM experiments should be careful attention to the compatibility of fluxes passed through the ocean surface between the ocean and atmosphere and enforcement of basic conservation laws in experimental design.

Initial conditions

The choice of initial conditions has little impact on AGCM simulations because of the short memory of the atmosphere, although soil initial temperatures and moisture values may take a decade or two to reach equilibrium. The long temperature memory of the ocean means that choice of initial conditions for that component is more critical to reduce computational time. Proxies that might be used as initial conditions for modeling experiments are sparsely distributed in time and space, have inherent uncertainties associated with them, and may not constrain important quantities (e.g., salinity). This challenge is usually overcome by specifying an initial sea surface temperature (SST) distribution derived from proxies or using output from an atmospheric general circulation model (AGCM), which in turn predicted SSTs assuming some distribution of oceanic mixed-layer depth and heat transport. Surface salinities are normally based on output from such AGCM simulations after assuming an empirical relationship between precipitation minus evaporation (P-E) distributions and surface salinity, or alternatively, a global constant value can be specified as an initial condition. In paleoceanographic simulations carried out with a coupled ocean-atmosphere model, these initial condition values may be strongly departed from because of the model's evolution. In simulations that do not employ coupling with an atmospheric model, these distributions are more or less fixed. Deep ocean temperatures are normally taken from benthic foraminifera-derived temperature estimates (where available), and salinity is frequently set at the modern global mean value or, sometimes, is set to a different global mean value designed to accommodate past variations in global mean salinity.

Equilibration and spin-up

A related issue in performing coupled modeling experiments has been integrating the entire system, including the deep ocean, to quasi-steady state (so-called spin-up). Nevertheless, ocean models—at least the non-eddy resolving models employed in climate and paleoclimate studies—are computationally inexpensive relative to their atmospheric cousins. In ocean-only GCM studies, it is now feasible to integrate the model to equilibrium without so-called deep ocean acceleration, although accelerating the deep-ocean tracer fields during spin-up has been shown to speed equilibration without unduly affecting the final climate state (Danabasoglu et al., 1996, 2004). However, as described above, for most climate purposes, a coupled model is preferred so that surface temperature and salinity fields and ocean heat and freshwater transport are fully prognostic. Many coupled models historically displayed a drift of climate from modern values even for modern conditions, which precluded them for use in paleoclimates, but this modeling challenge has been overcome in the recent generation of models. It is now computationally feasible to integrate coupled models to equilibrium (≥ 4000 years), and this is probably somewhat preferable to iterative techniques used previously (Huber and Sloan, 2001) because it involves less ad-hoc manipulations.

PROGRESS IN THE MAJOR QUESTIONS OF GREENHOUSE PALEOCLIMATES

Many important questions are raised by patterns shown by past greenhouse climates and much remains to be understood. This complexity can be reduced into the three most vexing climate problems that have the broadest implications for climate theory. These three problems are the focus of most greenhouse paleoclimate modeling.

Global warmth with strong climate sensitivity

The pattern is one of global mean surface temperatures much warmer (>10°C warmer) than the modern global mean temperature (15°C), and strangely, these climates experience significant variability. How was this warmth maintained, and how can a warm world be highly variable given that most positive feedbacks we know are strongest at cooler temperatures?

It is likely that global mean temperatures in the Eocene and Cretaceous were much warmer than previously thought, given new estimates emerging from improved data coverage and the introduction of new proxies (Huber, 2008; Sluijs et al., 2006; Weijer et al., 2007; Liu et al., 2009; Head et al., 2009; Kowalski and Dilcher, 2002). This new characterization is a product of a new understanding of the potential contamination of the older tropical SST records (Zachos et al., 1994) by diagenesis (Schrag, 1999; Huber and Sloan, 2000; Pearson et al., 2001, 2007); the development of new proxies, such as TEX_{86}, MBT, Mg/Ca; and revised calibrations of older proxy data, such as leaf-margin analysis using the Kowalski and Dilcher (2002) calibration. Consequently, the long-recognized and difficult-to-explain pattern of greenhouse climate warmth has always been underestimated. It is now clear that the global mean temperature change from Modern to Eocene was not ~4° (Covey et al., 1996) but >10°C, although the polar amplification of this global mean temperature increase remains extreme because newer polar temperature estimates are even warmer than corresponding tropical temperature increases. The Cretaceous warming may have been even larger, but fewer data exist to constrain it.

Atmospheric pCO_2 is generally considered to play either a secondary or primary role in explaining global warmth during the Cretaceous and Eocene because concentrations were above modern (Pearson and Palmer, 2000; Pagani et al., 2005; Lowenstein and Demicco, 2006; Pearson et al., 2009; Pagani et al., 2011). As a secondary forcing agent, CO_2 concentrations may have primed the climate system to be sensitive to forcing by alterations in other boundary conditions, such as ocean gateways or insolation, or it may have enhanced nonlinear sensitivity through feedbacks, e.g., due to water vapor. Alternatively, pCO_2 could have been the primary forcing factor responsible for Eocene greenhouse climate. Whether the latter or former perspective is correct has profound implications for estimates of global climate sensitivity to pCO_2.

The upper range of pCO_2 has not been explored adequately because modelers tend to choose the conservative end (520–2200 ppm) of paleo-pCO_2 proxy estimates, even though they extend up to 4400 ppm in the early Eocene (Fletcher et al., 2008) or Cretaceous (Royer et al., 2004). This, in part, has been motivated by the fact that, at pCO_2 values above 2000 ppm, models have produced tropical SSTs warmer than reconstructions. On the other hand, these reconstructions have been reinterpreted upward (Pearson et al., 2001; Wilson et al., 2002; Pearson et al., 2007; Forster et al., 2007a, b; Bice et al., 2006) and this issue needs to be reexamined. It should be noted that whereas tension exists between the competing interests of warming winters in the interiors and overheating in the tropics, there is no great conflict between the pCO_2 required by any given climate model and the paleo-pCO_2 proxies. This is for two reasons.

First, the global mean temperature response to greenhouse gases in climate models, most commonly represented as the equilibrium sensitivity of temperature to a doubling of pCO_2 from 280 ppm to 560 ppm (S), is a parameter that varies from model to model and has no known true value (Roe and Baker, 2007). Estimates for modern-day conditions range from 1–6°C, and the range implied by paleoclimate conditions is similar, although perhaps pushed to the higher end of the range (Pagani et al., 2006a). A specific value of S is an emergent property of a given climate model. It is not directly specified and it is not trivial to adjust because it emerges as the difference of many large and countervailing processes. Consequently, even if pCO_2 in the past were perfectly known (which it is not), it would be necessary to either utilize a climate model with the correct S value (which is also unknown), or to adjust the input value of pCO_2 so that the product of the sensitivity and the pCO_2 yielded the correct global MAT response. In other words, it is the product of pCO_2 and S that is the relevant parameter to explore given that there currently is irreducible uncertainty in both terms of the equation. If the true value of S is 6°C per pCO_2 doubling and a study is using a model with S=3°C per doubling, then twice the value of pCO_2 would be required for an accurate simulation (barring chemistry-dependent interactions).

Second, the approximately logarithmic dependence of radiative forcing due to pCO_2 as a function of concentration moderates the importance of real or apparent variations of pCO_2 provided that the background values are high. All climate-model simulations to date require pCO_2 of at least 1000ppm to begin to match records of high-latitude and deep-ocean warmth, or even to match the new, warmer interpretation of warmer-than-modern tropical SSTs (Pearson et al., 2007); in fact, they suggest higher values (Lunt et al., 2012). At this higher-than-modern pCO_2 range (e.g., 2000 ppm), a variation in pCO_2 of 1000 ppm would produce relatively small impacts on global mean temperature (i.e., 1.5°C for S=3).

Consequently, provided pCO_2 was generally higher than modern in the Eocene or Cretaceous, the specific value is not particularly important to evaluating the correctness of a solution to the equable climate problem at this point. If the situation arose that a climate-model simulation produced accurate climate results, then backing out the true value of S would be a potential reward. However, that would require an accurate knowledge of pCO_2, and we are still far from that level of accurate representation of early Eocene climate, although we are close in the late Eocene (Pearson et al., 2009; Pagani et al., 2011).

An additional part of this sensitivity puzzle is that the data reveal the Eocene and Cretaceous to be variable, with strong climate- and carbon-cycle variability on orbital periodicities all the way down to high frequency (interannual variability). Orbital-scale variability of the system is pervasive on land (Fischer and Roberts, 1991; Machlus et al., 2008; Galeotti et al., 2010b; Abels et al., 2012; Deconto et al., 2012) and in the oceans (Cramer et al., 2003; Wade et al., 2001; Wade and Kroon, 2002; Coxall et al., 2005; Pälike et al., 2006 a, b; Burgess et al., 2008; Warnaar et al., 2009; Westerhold et al., 2007), and this is apparently without substantial land-ice. Sub-orbital multi-millenial variability also is found in records, although they may only reflect regional signals (Machlus et al., 2008; Aziz et al., 2008; Lenz et al., 2011; Stickley et al., 2012). Good evidence exists for strong decadal to interannual variability from the Cretaceous through the Pliocene, including El Niño-Southern Oscillation cyclicity (Ripepe et al., 1991; Huber and Caballero, 2003; Garric and Huber, 2003; Galeotti et al., 2010a; Davies et al., 2011; Ivany et al., 2011; Davies et al., 2012; Watanabe et al., 2011). Most theories for strong variability are built on Quaternary conceptual models and rely on enhancement of variability by feedbacks with the cryosphere or a cold abyssal ocean, but this approach cannot easily explain the variability of climate in past greenhouse intervals.

The dynamic range of global mean temperature variation is estimated to be ~8°C on time scales of millions of years (Wing et al., 2000; Zachos et al, 2001; Jenkyns, 2003; Wilf et al., 2003; Fricke and Wing, 2004; Dumitrescu et al., 2006; Thomas et al., 2006; Pearson et al., 2007; Donders et al., 2009; Sinninghe Damsté et al., 2010; Wang et al., 2012). On shorter time scales, global mean temperature changes of ~5°C appear often, such as various hyperthermal events (Lourens et al., 2005; Nicolo et al., 2007; Galeotti et al., 2010b), and the early Eocene is characterized by at least three rapid hyperthermal events that seem to have ushered in the EECO: the early Eocene maximums, PETM, ETM 2, and ETM 3 (known as ELMO and X-event, respectively). One of the major paradoxes associated with the PETM (and mutatis mutandis for other hyperthermals) is that the amount of carbon released during the event, as estimated from carbon isotope perturbations, is much smaller than that required to produce the reconstructed ~5°C global warming unless a very large value of global climate sensitivity is assumed (Pagani et al, 2006a; Hendricks and Schrag, 2007). There are two resolutions to this paradox: either sensitivity is not a strong function of global temperature or boundary conditions and it has a uniformly high value (S >4°C), or sensitivity strongly increased in the climate regime that preceded the PETM and other hyperthermals. The relatively small changes in pCO_2 (Pearson et al., 2009; Pagani et al., 2011) associated with strong global cooling during the Eocene–Oligocene transition (Lear et al., 2008; Pearson et al., 2008; Liu et al., 2009; Eldrett et al., 2009, Wade et al., 2011) and the substantial warming with a small change in pCO_2 during the Middle Miocene Climatic Optimum (Kürschner et al., 2008; Herold et al., 2011b, 2012) also provide suggestive evidence of a higher or increasing climate sensitivity in these past warmer worlds.

It is difficult to imagine a mechanism to produce very high sensitivity at the high range in temperatures of the background greenhouse state because most well-understood processes that increase sensitivity (e.g., ice-albedo feedbacks) operate at relatively low CO_2 and temperature values. With the warm high-latitude temperatures indicated by proxy data, there is no land or sea-ice and no snow, so such feedbacks were nonexistent. Nevertheless, the range of pCO_2 above 2000 ppm is little explored and surprises, perhaps due to the vegetation, chemistry, water vapor or cloud feedbacks, might be found.

Equable climate problem

The pattern of equable climates is reduced seasonality in continental interiors compared to modern values with winter temperatures above freezing. How do we warm continental interiors, generating above-freezing conditions in winter but without overheating summers–especially subject to the other constraints described in the other problems?

Continental interior temperatures produced by Eocene and Cretaceous climate-model simulations have been too cold compared to paleoclimate proxies, especially for winter climates (Sloan, 1994). Evidence for very low seasonality and above-freezing conditions that characterize continental interiors far from the mediating force of a warm ocean (Sloan and Barron, 1990; Spicer and Parrish, 1990; Wolfe, 1994; Greenwood and Wing, 1995; Herman and Spicer, 1996; Upchurch et al., 1999; Markwick, 2007) has posed a major mystery. Unlike oceans, the interiors of large continents have low heat capacities with climates largely modulated by the seasonal solar cycle (Crowley et al., 1986). Even if high-latitude ocean temperatures were very warm, this would do little to ameliorate continental interiors (Huber and Sloan 1999; Sloan et al., 2000). This so-called equable climate problem persists at relatively high greenhouse gas concentrations up to ~2000 ppm CO_2, where the model-data mismatch typically is ~20°C for winter temperatures (Shellito et al., 2003), although the discrepancy for mean-annual temperatures is typically less except for polar regions.

This model-data discrepancy has proven fertile ground for innovative thinking in climate modeling. Efforts to achieve the character of warm winter continental interior and polar temperatures have been approached in various ways including: large lakes (Sloan, 1994; Morrill et al., 2001); polar stratospheric clouds (Sloan et al., 1992; 1999; Sloan and Pollard, 1998; Kirk-Davidoff et al., 2002; Kirk-Davidoff and Lamarque, 2008); increased ocean heat transport (Berry, 1922; Covey and Barron, 1988; Sloan et al., 1995); finer-resolution simulations of continental interiors (Sewall and Sloan, 2006; Thrasher and Sloan, 2009); a permanent, positive phase of the Arctic Oscillation (Sewall and Sloan, 2001); altered orbital parameters (Sloan and Morrill, 1998; Lawrence et al., 2003; Sewall et al., 2004); altered topography and ocean gateways (Sewall et al., 2000); radiative convective feedbacks (Abbot et al., 2008, 2009a, b); altered vegetation (Sewall et al., 2000; Shellito and Sloan, 2006a ,b); and changes in sea surface temperature (SST) distributions (Huber and Sloan, 1999; Sloan et al., 2001; Sewall and Sloan, 2004). Certainly part of a solution to this problem lies in changes in Earth's albedo arising from feedbacks such as the lack of polar ice and enhanced vegetation at high latitudes (Otto-Bliesner and Upchurch, 1997; DeConto et al., 1999; Upchurch et al., 1999). Inclusion of these parameters has failed to reproduce the warm winter temperatures seen in proxies.

While some regions have proven sensitive to variations in these climate-forcing mechanisms, the general outcome of model investigations has been failure to provide a general solution to the equable climate problem. One mechanism might explain warmth at extreme high latitudes (e.g., cloud feedbacks), but leave temperatures within the western interior of North America unexplained (Abbot et al., 2009a) or vice versa (e.g., large lakes or altered topography). Ultimately, failure of these various hypothesized resolutions to the equable climate problem, whether tested individually or in concert, has suggested that they are not the leading order solution to the problem. Recently, progress has been made in solving this problem, and it has depended critically on new tropical SST constraints and consideration of strong global radiative forcing enhanced by local feedbacks (Pearson et al., 2001, 2007).

Low gradient problem

This climate problem can be divided into two sub-problems: the cool tropical paradox and polar amplification. The cool-tropics paradox, or tropical thermostat hypotheses, focus on apparent limits to tropical temperature increase in a warmer world. The process of polar amplification refers to the enhancement of warming in the extratropics compared with the global mean. These can be considered two sides of the same coin, but the mechanisms involved and the data on which they depend are different, and it is worthwhile to focus on each separately as well as in conjunction.

Cool tropics paradox or tropical thermostats.—This is an old and still-influential paradigm that holds that tropical temperatures varied little in the past. Pioneering work indicated that greenhouse intervals deeper in Earth's past were characterized by cooler-than-modern tropical temperatures (Shackleton and Boersma, 1981). Similarly groundbreaking efforts, i.e., the original CLIMAP reconstructions (CLIMAP Members, 1981), indicated little tropical cooling or even slight warming during the last glacial maximum. These data interpretations lend support to and gained support from a class of thermostats proposed by Newell (1979) and Ramanathan and Collins (1991), in which tropical SST is limited by either evaporation or cloud feedbacks that set in at a trigger temperature erroneously assumed to be a universal constant independent of atmospheric CO_2 concentration. Yet, there is a long history demonstrating the lack of basis in data, theory, or models of the trigger class of thermostats (Fu et al., 1992; Wallace, 1992; Hartmann and Michelsen, 1993; Pierrehumbert, 1995; Sud et al., 2008; van Hooidonk and Huber, 2009; Williams et al., 2009). Other thermostats (Clement et al., 1996; Miller, 1997; Kleypas et al., 2008) have met similar fates (Hartman and Michelsen, 2002; van Hooidonk and Huber, 2009). Nevertheless, the false notion that data and simple physical reasoning support a tropical thermostat persists in the paleoclimate community, often based on the observation of an apparent cut-off in modern and paleoclimate data.

Interpretations of the proxy data evolved in subsequent decades as new, independent proxy records, and improved understanding of biases in the proxies developed. A decade ago, it became well accepted that tropical temperatures cooled substantially (2–3°C) during glacials (Crowley, 2000), and tropical temperatures in greenhouse climates were acknowledged to have been at least as warm as, if not warmer than, modern values (Crowley and Zachos, 2000). A true picture of the degree to which previous reconstructions were biased to cool values then became evident (Huber and Sloan, 2000, Pearson et al., 2001). SSTs of ~35°C (with an error of ±5°) are now reconstructed in the tropics (Pearson et al., 2001; Tripati et al., 2003; Tripati and Elderfield, 2004; Zachos et al., 2004; Pearson et al., 2007; Huber, 2008; Lear et al., 2008) during the Eocene or Cretaceous (Norris et al., 2002; Schouten et al., 2003, 2007). Such extreme temperatures have major implications for our understanding of the dynamics of past climates and for the history the life.

Whether a tropical thermostat exists is fundamentally important for three reasons. The tropics, defined broadly (30°N–30°S), make up half of Earth's surface area and play an outsized role in determining past variations in global mean temperature and the sensitivity of this variable to forcing factors such as greenhouse-gas concentrations. The tropics also dominate global biodiversity, and have frequently been considered stable, safe havens for fauna and flora compared with the more variable high latitudes. Finally, because global atmosphere-ocean circulation is driven by temperature gradients, tropical temperatures provide the linchpin on which the rest of the general circulation depends.

For their part, paleoclimate modelers have concluded that hot tropical temperatures and the high concentrations of greenhouse gases that cause them are required to reproduce warm extratropics because standard models and dynamical theory do not produce equator-to-pole temperature gradients much weaker than they have been in modern times. Nevertheless, based on the supposition that the models are missing important physics, many hypotheses and novel mechanisms have been proposed that center on the existence of a 'thermostat' that maintains tropical temperatures at a fixed level. These attempts to include new feedbacks have illuminated many dark alleys of climate dynamics, but so far all have been dead ends (Williams et al., 2009).

If SSTs were truly ~35°C at times in Tanzania (19°S) or New Jersey (~30°N), some tropical regions must have been much hotter, especially on land. As summarized in Huber (2008), this has thought-provoking implications for paleoclimate, vegetation, and carbon-cycle evolution. First, tropical temperatures above 31°C offer no evidence for a climate thermostat—that is, a strict mechanism that maintains tropical SSTs in the modern range. Climate dynamicists who have been trying to explain thermostats for decades may have been chasing a chimera. Second, climate models might be able to reproduce warm poles and warm extratropical continental winters, given that these new tropical SSTs imply closer-to-modern temperature gradients. Third, during the warmest parts of the past 65 million years—i.e., the PETM and subsequent hyperthermal phases of the EECO—tropical vegetation in some regions may have been above the upper limits of its thermal tolerance. Most plants, especially the C3 plants that comprised Eocene floras, have physiological mechanisms that break down in the 35°–40°C range; in particular, they can die because photorespiration dominates over photosynthesis. Annual mean temperatures greater than 35°C can be plausibly reconstructed to have been widespread equatorward of 30° latitude (although not universally as evidenced by cooler temperatures in some regions by Keating-Bitoni et al. (2011) and Jaramillo et al. (2011)), so floras may have been thermally stressed, and perhaps undergoing water stress, in the warmest intervals. This point of view remains controversial and is the focus of ongoing work. It remains possible that Cretaceous and Eocene 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 proxies record tropical temperatures from below the mixed-layer, that is, greater than ~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 or other key biological limitations.

Polar amplification.—Climate models historically have failed to reproduce the small equator-to-pole temperature gradients indicated by Cretaceous and Eocene paleoclimate proxies (Barron et al., 1981; Barron, 1987; Barron et al., 1993; Sloan and Rea, 1995; Huber and Sloan 2000; Huber and

Sloan, 2001; Zhou et al., 2012), and given that the newer reconstructions have raised polar temperatures (Fricke and Wing, 2004; Jenkyns et al., 2004; Poulsen, 2004; Sluijs et al., 2006; Bijl et al., 2009; Hollis et al., 2009; Jenkyns et al., 2012) even more than tropical ones, a mystery still persists. This failure represents one of the greatest challenges in paleoclimate dynamics because it suggests that climate models fail to reproduce the leading order feedbacks in a warmer world (Valdes, 2011). Alternatively, a sampling bias toward warm-season values at high latitudes and cold-(or 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, but small (up to 5°C) biases are possibly explained by such depth, productivity, and seasonal biases.

A commonly advocated conjecture might appear to resolve the mysteries of greenhouse climates, and it relies on increased atmospheric or oceanic heat transport in conjunction with high greenhouse-gas concentrations. Attention has focused on the challenging problem of modeling paleo-ocean-atmosphere circulations and, especially, on placing bounds on the magnitude of heat transport in such climates. Increased poleward heat transport by the ocean or atmosphere has long been considered a solution to these problems (Berry, 1922; Barron, 1987; Covey and Barron, 1988; Rind and Chandler, 1991; Barron et al., 1993; Schmidt and Mysak, 1996). However, the supposition of low gradients propelled by enhanced heat transport is recognized as a climate conundrum: that is, enhanced poleward heat transport is necessary to maintain low meridional-temperature gradients, but small meridional temperature gradients reduce rates of poleward heat transport. Ocean-atmospheric models have produced meridional atmospheric and ocean heat transports close to modern (Huber and Sloan, 2001; Najjar et al., 2002; von der Heydt and Dijkstra, 2006). The missing required transport is inferred to be ~1 Petawatts at 45° latitude (Huber et al. 2003). Changes in atmospheric or oceanic heat transport have been always been the leading hypotheses explaining the low gradients of past greenhouse climates (Berry, 1922); consequently, below is a more detailed discussion of the role of heat transport in this important issue.

OVERVIEW OF THE ROLE OF THE ATMOSPHERE AND OCEANS

Atmospheric heat transport

While it is often assumed that increased latent-heat transport might account for a small meridional temperature gradient, models and theory have not produced this effect. This is surprising at first, given the exponential dependence on atmospheric saturation vapor pressure expected from the Clausius-Clapeyron relation. A warmer atmosphere could hold more water vapor provided that relative humidity distributions were not much different than modern. But, current theory and models have shown that latent heat transport increases with increasing temperature gradient (all things being equal), the opposite to what is required to maintain low gradients (Pierrehumbert, 2002). This is because increased latent heat transport requires a more intense hydrological cycle, which places important constraints on the dynamics. A more intense hydrological cycle as commonly defined in atmospheric dynamics (Pierrehumbert, 2002) refers to increased meridional water vapor transport from low to high latitudes. It does not mean more rain everywhere, as it is often used in the paleoclimate literature. Too often, pervasive records of wet Eocene and Cretaceous conditions have been used as evidence of an enhanced hydrological cycle and a wetter world everywhere.

In steady state, stronger divergence requires an intensification of evaporation (E) in net E zones and a counterbalancing increase in precipitation (P) in net P zones, leading to an increased meridional gradient of E-P (Held and Soden, 2006). Climate model simulations of the future show a general drying of the subtropics (and moistening of the deep tropics) in simulations of modern global warming, a fact that is relatively well-understood from basic theoretical arguments (Allen and Ingram, 2002; Held and Soden, 2006). Thus to first order, the general projected change in the net surface freshwater flux under increased greenhouse gas forcing is a simple amplification without a change in the spatial pattern. In steady state, provided that E>P equatorward of the subtropical margins (e.g., see the work of Ziegler et al. , 2003, suggesting this has been the case since the Permian) there must be net P averaged over the extratropics. Thus, we expect to find an apparent drying of the subtropical regions to be consistent with an increase in the vigor of the hydrological cycle, and a compensating moistening in high latitudes. This pattern appears to be playing itself out currently as revealed by trends in modern hydrological records (Zhang et al., 2007), and some evidence indicates the same in the Eocene (Pagani et al., 2006a; Harding et al., 2011).

On the other hand, global mean temperature introduces a nonlinearity to this simple picture. For this reason, it is crucial that most simulations that have been performed have not been driven by CO_2 concentrations (Pearson and Palmer, 2000; Lowenstein and Demicco, 2006) or tropical temperatures (Pearson et al., 2007) in the range that we now consider likely. Given that newer proxy records and interpretations push temperature estimates up almost across the board, the importance of nonlinear dependence of heat transport on the global mean climate must be considered. Thus, mechanisms may exist to achieve greater latent heat transport that have not been investigated for Eocene conditions based on thinking about modern global warming. Substantial focus has been applied to the issue of meridional circulation and transport changes in a modern global warming context. In general, two results are robust: global warming should intensify the hydrological cycle and increase the width of the Hadley Cell, displacing storm tracks poleward.

Recent dynamical work (Caballero and Langen, 2005; Frierson et al., 2007) has shown the limitation of Pierrehumbert's (2002) theoretical scaling, and it deserves renewed attention especially in light of new temperature proxy data. Because half of Earth's surface area lies between 30°N and 30°S, upward revision of tropical temperatures from ~26°C to ~35°C during the warmest phases has profound implications for the global mean temperature, which, in turn, has a nonlinear impact on features like Hadley Cell width and poleward latent heat transport. Caballero and Langen (2005) used an idealized atmospheric general circulation modeling framework to construct a regime diagram for the variations of atmospheric latent and sensible heat transport as a function of surface temperature and its meridional gradient. They found that poleward atmospheric latent-heat transport increased with increasing temperature gradient (ΔT), and increased with global mean surface temperature (Tm) over a wide range of values. For the combined range of Tm and ΔT characterizing the Eocene, latent and total heat transport did not increase with global warming for two reasons. First, eddy kinetic energy (EKE) of the storm tracks, which can be thought of as a qualitative measure of the thermal eddy diffusivity of the atmosphere, decreases at Tm > 15°C because of large increases in midlatitude static stability (which suppresses eddy activity). Second, as Tm increased, storm tracks migrated poleward in their simulations, resulting in storms becoming cut off from their supply of subtropical water vapor. Within the near-term global warming context, such a shift is only 2–3° latitude, but the storm track latitude should shift by 5–10° latitude over the range of potential Eocene climates. Extensions of this work by Frierson et al. (2007) showed that the Hadley Cell width increases ~0.2–0.3° latitude per °K warming. These results are corroborated by a series of studies of Hadley Cell width in modern and future global warming context that indicate similar sensitivities (Fu et al., 2006; Lu et al., 2007; Seager et al., 2007). For the Eocene, this would imply that the mean Hadley Cell margin was 2–3° poleward of its present position, with associated shifts in evaporation and precipitation zones and wind patterns expected accordingly.

Some results from studying modern global warming do not seem to be robust, and it is possible that studying the Eocene may provide unique insights given the magnitude of the forcing and the response. Theory, observations, and modeling indicate that future atmospheric specific humidity will increase at a rate of ~7% per °K of surface warming because of the exponential dependence of saturation vapor pressure on temperature (Soden et al., 2005). On the other hand, global precipitation rates have been proposed to increase at a much lower rate (~3% per °K) based on theoretical energy balance constraints and modeling results (Allen and Ingram, 2002; Trenberth et al., 2003). Despite the power and simplicity of these arguments the modern reality (as measured by satellites) indicates that the global precipitation sensitivity to warming is ~6% per °K of warming, or nearly identical to the expectation based on the Clausius-Clapeyron relation (Wentz et al., 2007). These results have been corroborated by analyses by Allan and Soden (2007) and Yu and Weller (2007). The discrepancy between theory, models and data for modern conditions may be directly relevant to the consistent failure of models under Eocene conditions. Future model data comparisons should focus on changes in precipitation as well as temperature.

Oceanic heat transport

One of paleoceanographic modeling's grandest challenges is to evaluate the validity of the hypothesis that changes in ocean heat transport (OHT) have occurred. So far, the jury is still very much out. A common misconception in paleoceanography is that evidence from proxies for warm sea-surface temperatures at high latitudes is unequivocal evidence for increased OHT, or more generally, that knowledge of ocean temperatures provides much information about OHT. It is always possible that ocean temperatures are warm because little heat is lost to the atmosphere. Yet, given the physical constraints on the atmosphere's ability to transport more heat under equable conditions, other mechanisms, including ocean heat transfer in the form of low-latitude halothermal circulation, were hypothesized. Justifications for enhanced ocean heat transport reach back to Chamberlin (1906), who called on the production of high-salinity waters sourced from highly evaporative, low-latitude regions during warm periods in Earth history.

Arguments regarding the source of deep waters and the direction of transport are now considered misguided. What matters for ocean heat transport is the amount of cooling and the vigor of ocean circulation. Considered by itself, it makes no difference where or at what temperature deep water forms; it is the temperature transformations along the circulation path that matter for OHT. If warm, salty, deep water with a temperature of 18°C forms in the tropics/subtropics and flows to the poles where it is maintained at this temperature, no heat is transported, regardless of the fact that the polar waters are quite warm. Weak temperature gradients lead to weak ocean heat transport regardless of its source (Huber et al., 2003). The production of warm saline bottom waters does not explain weak temperature gradient climates because ocean transport heat requires cooling somewhere at the surface.

A family of related problems that proxy interpretations have in providing insights into ocean heat transport explanations of greenhouse climates is one pathological case we can call the 'covered hot-tub paradox.' In an ocean covered with an insulating, opaque, and waterproof 'lid,' like the cover on a hot-tub, no heat is transported by the ocean because no heat is lost or gained through the surface, regardless of the details of the internal temperature and velocity distribution. In a more realistic example, imagine an atmosphere so thick with greenhouse gases that high-latitude outgoing longwave radiation is much diminished from modern values, in which case the atmosphere draws less heat from the ocean. In this extreme case, assume that the ocean currents transport warm, equatorial water all the way to the pole. Nevertheless, because of the strong greenhouse effect, no heat is lost from the ocean to be radiated by the atmosphere to space. In this idealized case, the ocean surface temperatures are as warm at the poles as in low latitudes and might be interpreted as an extreme example of increased OHT, when, in fact, the ocean has transported no heat. This is not merely a straw man argument, it is a very real hypothesis for past high-latitude warmth, i.e., that the ocean is warm at high latitudes because very little heat is lost in winter due to a strong greenhouse effect (Huber and Caballero, 2011; Lunt et al., 2012).

Thus, paleoenvironmental proxies must provide quantitative estimates of two quantities, temperature and velocity structure, to complete an estimate of OHT. Without a substantial improvement in our understanding of the velocity structure of the upper several kilometers of the ocean in the deep past, there is little that can be determined about OHT from existing proxy records that mostly constrain temperature. The relevant quantity is not where deep water formed, but how much circulation occurred. The rate at which heat is transported by circulation is a function of the velocities, diffusion, and temperature gradients, and analogous balances obtain for the transport of all tracers. Just as there are weak data-based constraints on ocean velocity, there are weak constraints on its integral properties, such as the meridional overturning circulation (MOC). One commonly applied constraint is the calculation of vertical and spatial gradients of $\delta^{13}C$, which is an isotopic tracer related to nutrient distribution and the carbon cycle (e.g., Corfield and Norris, 1996). Where sufficient $\delta^{13}C$ gradients exist, the location of regions of deep-water formation can be ascertained and bulk flow patterns may be estimated, although typically in much of the Eocene, these gradients are weak so establishing patterns is challenging. The use of other water-mass tracers, such as neodymium (e.g., Thomas, 2004), is a recent innovation that may add significant detail to the picture that has developed from $\delta^{13}C$ tracer distributions. In the Plio-Pliestocene, such hurdles have been partially overcome (e.g. Ravelo and Andresen, 2000) in large part due to the greater coverage afforded by working in more recent, observation-rich times. More geographic coverage and additional circulation proxies will be required to improve the situation in the Paleogene. Reconstructions of biogeographic, carbon-isotopic, benthic oxygenation, or trace-element patterns can help to unravel these difficult issues, and often can establish at least the directions of ocean currents (see Carter et al., 1996; McGowran et al., 1997; Huber et al., 2004; Thomas, 2004), but the magnitude of circulation is usually subject to very substantial uncertainty.

Horizontal and vertical velocities are prognosed by OGCMs as well as a host of simplified models, and are the quantities most directly tied to the well-understood and well-represented dynamics of the ocean. In general, an OGCM is not necessary to predict velocities. In modern oceanographic studies, the Sverdrup and Ekman relations provide a clear constraint on the horizontal and vertical circulations produced by an OGCM driven explicitly by AGCM wind fields, and a first approximation to the behavior of a coupled GCM (Danabasoglu, 1998; England et al., 1992). Such studies have demonstrated that theoretical tools can be used to unravel causal linkages between the atmosphere and ocean, even when an OGCM is used to model the ocean explicitly. The surface and upper ocean flows largely reflect the driving wind fields, and consequently, shallow water ocean models, or even analytic calculations (such as the 'island rule' or the Sverdrup transport), allow estimates of the strength, shape, and disposition of gyre circulations as well as the western boundary currents in equilibrium with them for a given wind field (Nof, 2000; Nof and Van Gorder, 2002; Huber and Nof, 2006). The Sverdrup transport (gyre circulations driven by the curl of the wind stress) corresponds to the vertically averaged (barotropic) mode of the ocean general circulation (Pedlosky, 1996). With further assumptions, these calculations allow testable predictions to be made about current strengths, circulation trajectories, and ocean heat transport (Huber and Sloan, 1999; Klinger and Marotzke, 2000; Held, 2001). Such predictions may be verified with records of paleo-circulation intensity and direction. In the deep ocean, OGCM results mainly have reproduced features that were adducible from Stommel-Arons theory (see results in Huber et al., 2003). Such predictions may be verified with records of paleo-circulation intensity and direction.

Unfortunately, current directions are weakly constrained, and current magnitudes are among the most poorly known features in deep-time paleoceanography because these features are not strongly constrained by direct proxies, and the indirect proxies are qualitative (Watkins and Kennett, 1972; Carter et al., 1996; McGowran, 1997). Nevertheless, a substantial amount of work has been done in predicting paleocean current directions and testing these against proxy interpretations. Some notable studies include those in the Permian (Winguth et al., 2002), the Cretaceous Tethyan (Barron and Peterson, 1991; Bush and Philander, 1997a, b) and Turonian seaways (Slingerland et al., 1996) and in the global ocean (Poulsen et al., 1998, 2001). In more recent periods, attempts have been made to understand changes in flow direction between the Atlantic and Pacific, i.e. during the Miocene (Nof and Van Gorder, 2002; Nisancioglu et al., 2003; Omta and Dijkstra, 2003), although explicit model-data comparison has been lacking in these studies. The more widespread use of sediment and biological transport and dispersal models in such investigations would substantially assist in the comparison to proxy records. Some first steps in that direction have been taken in Huber et al. (2004) and Ali and Huber (2009).

Upwelling is a feature more closely tied to the proxy record than ocean current direction because it is more directly linked to an observable quantity (productivity and burial). Understanding how upwelling patterns may have changed during the past can provide us with clues to the nature of circulation changes (Munk, 1966; England et al., 1992; Toggweiler and Samuels, 1995; Vallis, 2000), primary productivity shifts (Thomas, 1996; Thompson and Schmitz, 1997; Moore et al., 2004), and alterations in global cycling of chemical species (Broecker and Peng, 1982) that may have occurred. Upwelling predictions may be especially useful for placing constraints on water mass "aging" (Miller et al., 1987) and organic carbon burial rates (Handoh et al., 1999, Handoh, 2003) over these time intervals. For the purely wind-driven component of upwelling, an OGCM is not particularly necessary; a shallow-water model is adequate to the task (Handoh et al., 1999, 2003), or, calculation of Ekman pumping from the wind fields produces nearly identical results (Huber and Sloan 1999; Sloan et al., 1999; Sloan and Huber, 2000; Huber and Sloan, 2000; Sloan and Huber, 2001). Quantitatively tying the upwelling velocity to a sediment accumulation rate involves a deep understanding of many ecological and biogeochemical processes. Currently, our ability to make quantitative predictions on productivity, burial, and the carbon cycle (Pagani et al., 2011) in the past is more limited by our understanding of past biogeochemical cycles and paleoecology than by insights into past physical oceanography.

Rates of deep-water formation and deep-water flow are not currently predictable from proxy methods, which is unfortunate because these are the quantities that bear directly on the critical paleoceanographic questions, whereas the location of deep-water formation reveals little by itself about the ability of the circulation to transport heat or nutrients. Nevertheless, some physically justified conjectures on the evolution of ocean heat transport during these intervals have been made.

A leading paradigm of paleoceanography is that the opening and closing of ocean "gateways" has led to significant changes in poleward heat transport by the ocean, thereby dramatically altering climate and driving a large part of the evolution of climate through time. A number of modeling studies have demonstrated that the ocean's overturning circulation, heat transport, and resulting temperature distribution are somewhat sensitive to changes in ocean gateways (Mikolajewicz et al., 1993, 1997). Furthermore, there is a strong line of theoretical and modeling evidence that backs up the theory that wind-driven upwelling of deep water in the Southern Ocean plays a dominant role in modern ocean circulation (Toggweiler and Samuels, 1995; Nof, 2000), and that past changes in this upwelling may have been especially important in the high latitudes (Toggweiler and Bjornsson, 2000) and at low latitudes (Hotinski and Toggweiler, 2003). The most frequently cited example of gateway changes on OHT is due to the creation of an Antarctic Circumpolar Current (ACC). As noted in Toggweiler and Bjornsson (2000), creation of the ACC leads to a cooling of the Southern Hemisphere and a warming of the Northern Hemisphere because the flows associated with the ACC 'steal' heat across the equator (Toggweiler and Lea, 2010).

Gateways have been hypothesized to answer the following question: 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 ~100,000 years during the Oi1? A popular theory is thermal isolation of the Antarctic continent, where 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 (Kennett et al., 1975; Kennett, 1977). Data on the opening of Drake Passage to deep flow are equivocal, because even recent estimates range from middle Eocene to early Miocene (a range of 20 Ma). The leading evidence for the ocean gateway-heat transport mechanism is the apparent synchroneity of initiation of widespread glaciation and opening of an ocean gateway between Australia and Antarctica (the Tasman Gateway) near the Eocene–Oligocene boundary. The 'Tasman Gateway (TG) hypothesis' suggests that the onset of Antarctic glaciation and associated global cooling across the Eocene–Oligocene (E/O) transition resulted from thermal isolation of continental Antarctica from relatively warm equatorial-sourced limbs of surface current gyres via TG opening (Murphy and Kennett, 1986; Exon et al., 2001). The Tasman Gateway opened several million years before Oi1, which seems to make a close connection between Southern Ocean gateway changes and glaciation unsupportable (Stickley et al., 2004). 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 (Sijp et al., 2011). That does not mean that Southern Ocean gateway changes had no affect on climate. Evidence exists that a large part of the transition of thermal structure of the Atlantic Ocean from its Eocene configuration to its modern configuration occurred in the early Oligocene, after ACC development (Katz et al., 2011). The oceanographic changes associated with opening of Drake and Tasman gateways (the '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 0.6PW decrease in southward heat transport in the southern subtropical gyre (out of a peak of ~1.5PW), 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°C) change in the Antarctic polar ocean region (Huber et al., 2003; Huber et al., 2004; Sijp et al., 2011; Cristini et al., 2012). Climate and glaciological modeling implies that a leading role of this transition was a decrease in atmospheric greenhouse gas concentrations because Antarctic climate and ice sheets have been shown to be more sensitive to pCO_2 than to other parameters (Deconto and Pollard, 2003; Huber

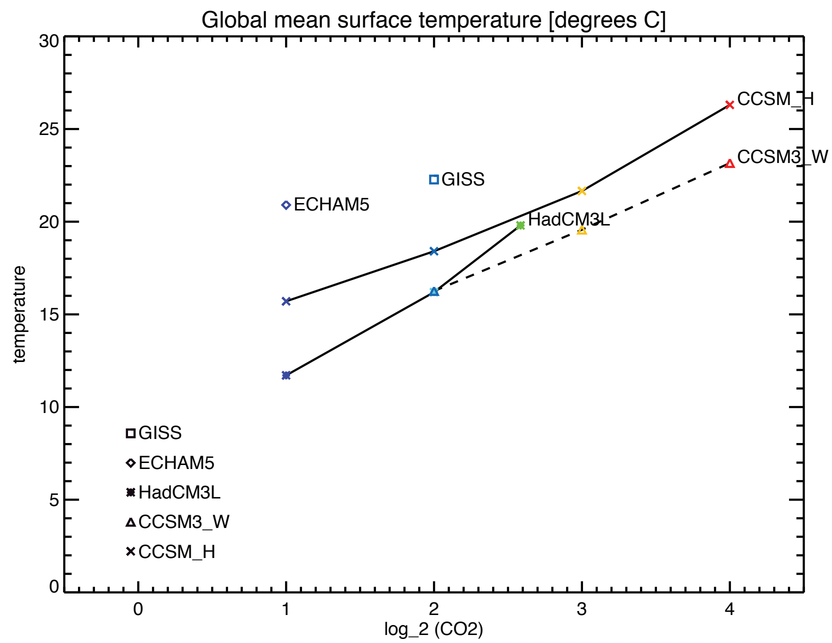

FIGURE *1*.—The global annual mean terrestrial near-surface atmospheric temperature as a function of the logarithm of pCO_2 for all simulations. The different symbols represent different modeling efforts as described in Table 2 and labeled on the figure. The values at log=0 are pre-Industrial control simulations carried out at ~280 ppm CO_2. The same symbols are used in succeeding figures. Figure adapted from Lunt et al. (2012).

and Nof, 2006). Consequently, Cenozoic cooling of Antarctica is no longer generally accepted as having been primarily caused by changes in oceanic circulation. 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. 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.

At the moment, it appears that the closed Southern Ocean gateways of the Eocene were not predominately responsible for high-latitude warmth, but that does not mean necessarily that changes in ocean heat transport were negligible contributors to Cenozoic climate deterioration. All ocean modeling studies have similar predictions with respect to the physical oceanographic response to gateway changes, but the surface temperature response, i.e., the climatic impact of this change, show a wide spread. In studies in which a simple representation of coupling of ocean-atmosphere exchanges have been included, opening and closing of Southern Ocean gateways results in Southern Ocean surface temperature changes from a low value of 0.8°C (Mikolajewicz et al., 1993; Cristini et al. 2012) a middle range of ~1.5°C (Nong et al., 2000) to a high value of ~3.5°C (Toggweiler and Bjornsson, 2000; Najjar et al., 2002). The spread of values represents differences in the treatment of the top boundary condition, including sea-ice feedbacks and damping by the atmosphere, rather than representing an important difference in the predicted ocean dynamics behavior. It is conceivable that changes in the ocean's nutrient distribution associated with gateway changes drove changes in atmospheric greenhouse gas concentrations (Heinze and Crowley, 1997; Huber and Nof, 2006; Scher and Martin, 2006; Pagani et al., 2011) that appear to have been responsible for the bulk of the EOT climate change.

More generally, models that simultaneously predict the behavior of the atmosphere and ocean have only produced total meridional atmospheric and ocean heat transports close to modern (Huber and Sloan, 2001; Najjar et al., 2002; von der Heydt and Dijkstra, 2006; Sijp et al., 2011; Winguth et al., 2011; Zhang et al., 2011; Lund et al.,

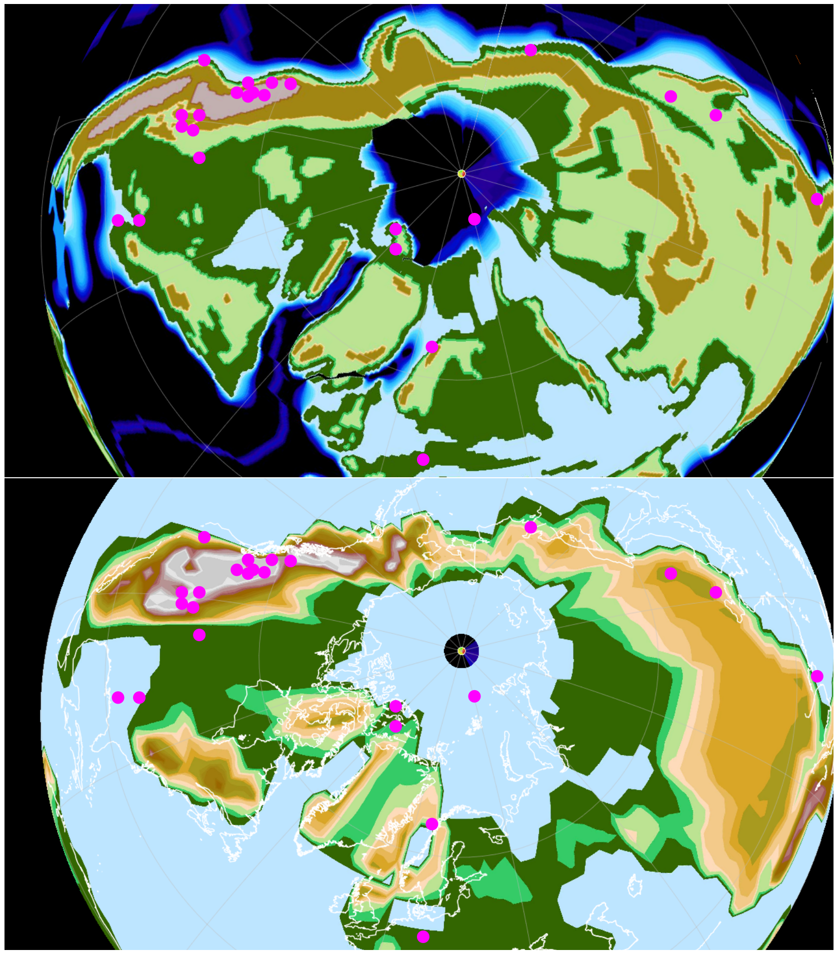

FIGURE 2.—Comparison of paleogeographic information used in the simulation with independent reconstructions from Mueller et al. (2008) via the freely available GPLATES software and via personal communication from Paul Markwick. In (a), a plate tectonic reconstruction for 55 mya from GPLATES including sea floor age is shown. Terrestrial paleoclimate proxy localities are indicated on this map with pink circles. The modern continental boundaries in their 55 mya positions as reconstructed by GPLATES compared with Markwick's paleogeography are shown in this panel. The topography used in CCSM3 model studies (bottom) is compared with the independently derived high resolution early Eocene topography of Markwick (based on Markwick (1998), top). In these figures, the paleoclimate proxy localities are indicated with magenta circles. The GPLATES-derived plate configuration is a reasonable match to the land-sea distribution developed in Sewall et al. (2000) and used in the simulations here. It should be noted that in some cases GPLATES-derived paleolocations must be adjusted slightly to fall on land or to be in the right location with respect to topography, as described further in Huber and Caballero (2011).

2012). How much ocean heat transport is required to explain low gradient climates? Climate models have been coupled to a 'slab' mixed-layer ocean model, which assumes that the important ocean thermal inertia is in the upper 50m and that ocean poleward heat transport is at specified levels, in order to predict equator-to-pole surface temperature gradients. With appropriate (Eocene or Cretaceous) boundary conditions, near-modern $p\mathrm{CO}_2$, 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. With higher $p\mathrm{CO}_2$, tropical SSTs increase without changing meridional gradients substantially, and continental interiors remain well below freezing in winter. Substantial high-latitude amplification of temperature response to increases in $p\mathrm{CO}_2$ or other forcings is generally obtained because of the nonlinearity introduced by crossing a threshold from extensive sea-ice cover to little or no sea-ice cover. Proxy data imply that it was unlikely that sea ice was present during the warmest of the greenhouse climates; therefore, this CO_2 sensitivity is probably not representative of the true past greenhouse intervals.

With heat transport approximately 3× modern in a slab ocean model and pCO_2 much higher than modern (1-2000 ppm, depending on the model), 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 traditionally have not been considered, such as those due to tropical cyclones (Emanuel, 2002; Sriver and Huber, 2007; Korty et al., 2007; Fedorov et al., 2010). Well-established physical mechanisms (Bony et al., 2006) cannot provide for this kind of increase, although the idea has been explored in simple models (Lyle, 1997; Emanuel, 2002). Further, there currently is no evidence from ocean circulation proxies for such large increases in ocean circulation rates (Hague et al., in press) and the associated decrease in ocean residence time it implies.

It is important to recognize that details matter when evaluating whether or not reconstructed temperature patterns and inferred ocean heat transport reflect some sort of paradox. For example, there is a range 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 two to three times as large (Huber et al., 2003). In other words, warm polar temperatures of 10°C and tropical temperatures of 34°C, like those observed in the early Eocene, can exist in equilibrium with near-modern ocean transport.

RECENT PROGRESS

Recent paleoclimate modeling of greenhouse climates has been significant. Here, rather than summarize all of these efforts, for the sake of brevity and clarity, I present work completed for the informal Eocene Model Intercomparison Project. By examining the features that are robust across models and comparing them with the robust results from proxy data compilations, a deeper understanding of the ability of the current generation of models to reproduce greenhouse climates can be gained.

Eocene Model Informal Intercomparison Project

Recently, an informal Eocene coupled model intercomparison project was carried out (EOMIP). This comparison is fully described in Lunt et al. (2012), from which I draw here. That study presented simulations that were published in the peer-reviewed literature, and carried out with fully dynamic atmosphere-ocean general circulation models (GCMs) with primitive-equation atmospheres and oceans. This comparison comprises a total of four models: 1) HadCM3L (Lunt et al., 2010), 2) ECHAM5/MPI-OM (Heinemann et al., 2009), 3) CCSM3 (Hollis et al., 2009; Liu et al., 2009; Shellito et al., 2009; Winguth et al., 2010, 2012; Huber and Caballero, 2011), and 4) GISS ModelE-R (Roberts et al., 2009). Models that are most similar to those used in future climate change projection were chosen for this project. Models with energy balance or non-dynamical atmospheres, such as GENIE (Panchuk et al., 2008) or the UVIC model (Weaver et al., 2001; Sijp et al., 2011), or those with non-dynamical oceans (Deconto and Pollard, 2003, were excluded.

There are two sets of CCSM3 simulations, one set by Winguth et al. (2010, 2012) and another by Huber (Ali and Huber, 2009; Huber et al., 2009; Liu et al., 2009; Huber and Caballero, 2011). All the models and simulations are summarized in Table 2.

EOMIP differs from more formal model intercomparisons, such as those carried out under the auspices of PMIP, in that the groups have carried out their own experimental design and simulations in isolation, and the comparison is being carried out post-hoc, rather than being planned from the outset. As such, the groups have used different paleogeographical boundary conditions and CO_2 levels to simulate Eocene climates. This has advantages and disadvantages compared to the more formal approach with a single experimental design. The main disadvantage is that a direct comparison between models is impossible due to even subtle differences in imposed boundary conditions; the main advantage is that, in addition to uncertainties in the models themselves, the model ensemble also represents the uncertainties in the paleoenvironmental conditions, and therefore more fully represents the uncertainty in climatic predictions for that time period. Most of these studies had previously carried out some

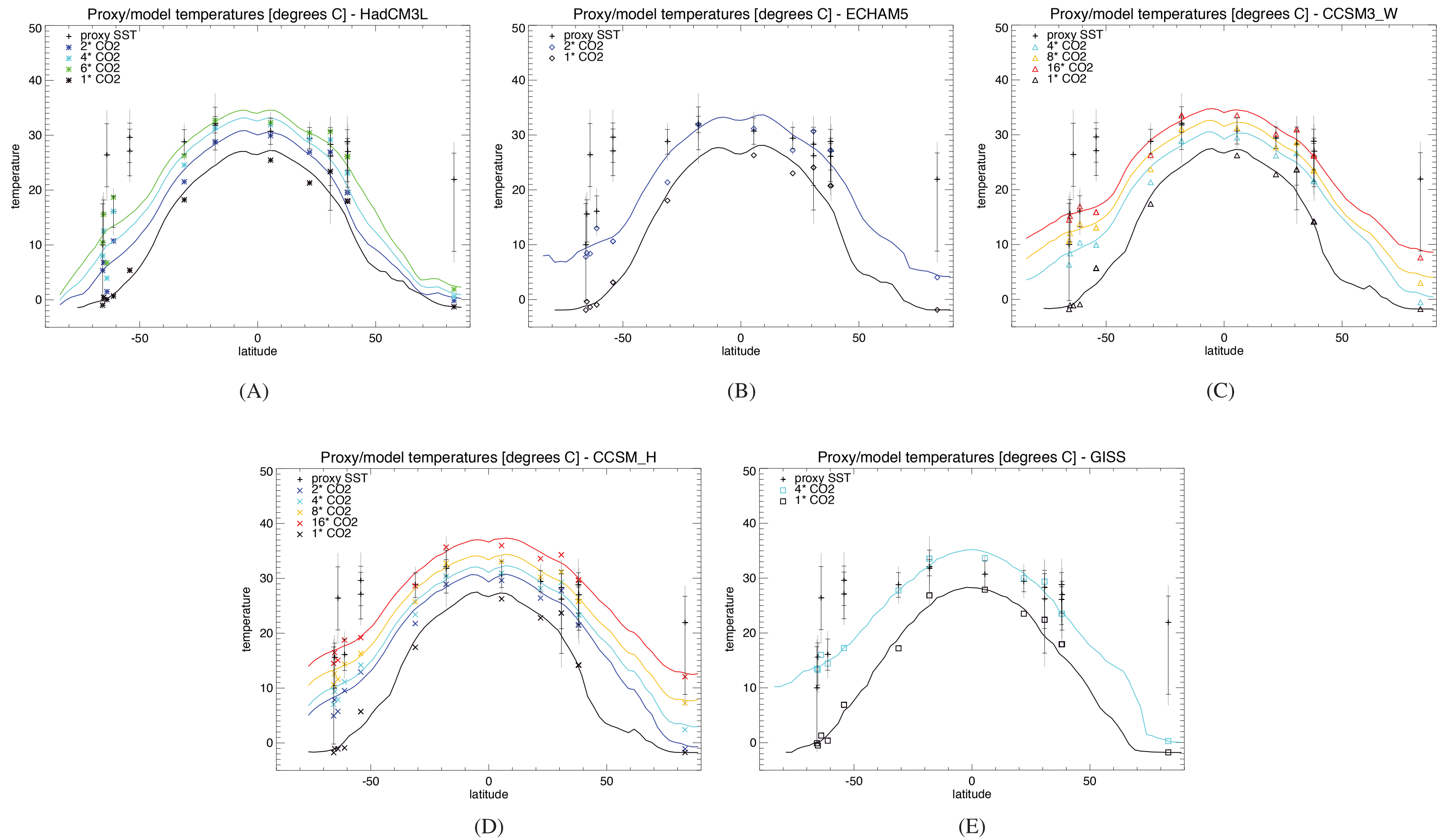

FIGURE 3.—Comparison of modeled SST with proxy-derived temperatures across latitude. The simulations at ×1 CO_2 are pre-industrial reference simulations. For the model results, the continuous lines represent the zonal mean, and the symbols represent the modeled temperature at the same location (longitude, latitude) as the proxy data. For the proxy data, the symbols represent the proxy temperature, and the error bars represent the range. The range is made up of two components: calibration uncertainty (black bar) and temporal uncertainty (gray bar). Adapted from Lunt et al. (2012).

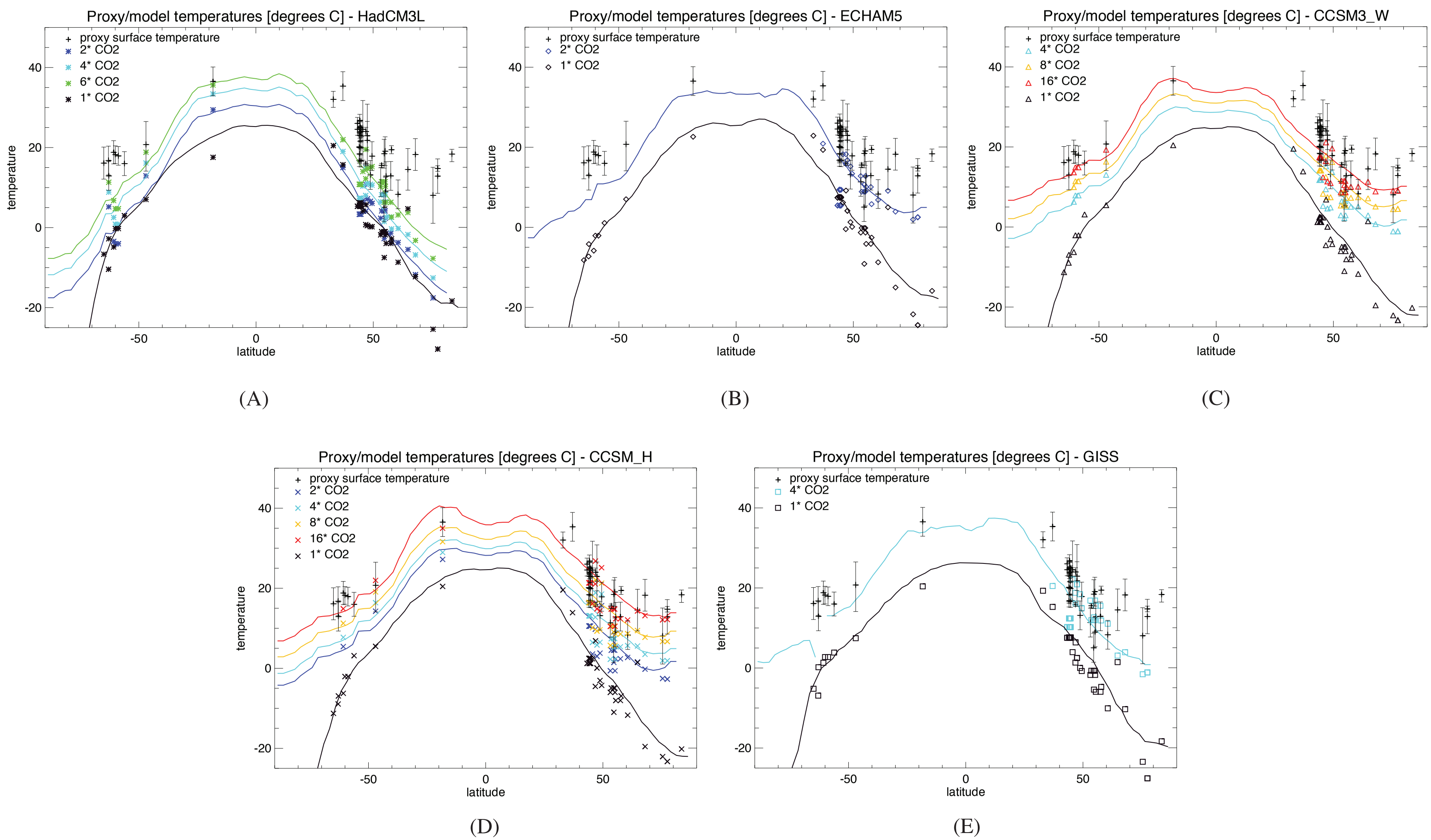

FIGURE 4.—Comparison of terrestrial modeled surface air temperature (SAT) with proxy-derived temperatures across latitude. The simulations at ×1 CO_2 are pre-industrial reference simulations. For the model results, the continuous lines represent the zonal mean, and the symbols represent the modeled temperature at the same location (longitude, latitude) as the proxy data. For the proxy data, the symbols represent the proxy temperature, and the error bars represent the range, as given by Huber and Caballero (2011). Adapted from Lunt et al. (2012).

TABLE 2.—Summary of model simulations in EoMIP. Some models have irregular grids in the atmosphere and/or ocean, or have spectral atmospheres. The atmospheric and ocean resolutions are given in number of gridboxes, X × Y × Z where X is the effective number of gridboxes in the zonal, Y in the meridional, and Z in the vertical. See the original references for more details.

Name	Reference	Model Name and Reference	Atmospheric grid spacing (longitude x latitude x height)	Ocean grid spacing (longitude x latitude x height)	CO_2	Vegetation	Paleogeography
HadCM	Lunt et al. (2010)	HadCM3L, Cox et al. (2001)	96x73x19	96x73x20	×2,4,6	homogenous shrubland	proprietary
ECHAM	Heineman et al. (2009)	ECHAM5/MPI-OM, Roeckner et al. (2003)	96x48x19	142x82x40	×2	homogenous woody savanna	Bice and Marotzke (2001)
CCSM_W	Winguth et al. (2010, 2012)	CCSM3, Collins et al. (2006); Yeager et al. (2006)	96x48x26	100x116x25	×4,8,16	adapted from Shellito and Sloan, 2006a.	Sewall et al. (2000)
CCSM_H	Liu et al. (2009); Huber and Caballero (2011)	same	96x48x26	100x122x25	×2,4,8, 16	adapted from Sewall et al, 2000	Sewall et al. (2000)
GISS	Roberts et al. (2009)	GISS ModelE-R, Schmidt et al. (2006)	72x45x20	72x45x13	×2	adapted from Sewall et al, 2000	Bice and Marotzke (2001)

form of model-data comparison; however, the models have not been formally inter-compared in a consistent framework. New data now allow a more robust and extensive evaluation of the models. Individual simulations are described further below.

Lunt et al. (2010) investigated the potential role of hydrate destabilization as a mechanism for the Paleocene-Eocene Thermal maximum (PETM, 55Ma) using the HadCM3L model. They found a switch in modeled ocean circulation that occurred between 2–4× pre-industrial concentrations of atmospheric CO_2, which resulted in a non-linear warming of intermediate ocean depths. They hypothesized that this could be a triggering mechanism for hydrate release. For the three Eocene simulations carried out (2, 4, and 6×), vegetation was set globally to a shrub plant functional type. The paleogeography is proprietary, but is illustrated in Lunt et al. (2010, supplementary information). An additional simulation at 3× CO_2 was carried out with the same model by Tindall et al. (2010), which incorporated oxygen isotopes into the hydrological cycle. The $\delta^{18}O$ of seawater from the Tindall et al. (2010) simulation is used in the SST compilation to inform the uncertainty range of the proxies based on measurements.

Heinemann et al. (2009) presented an ECHAM5/MPI-OM Eocene simulation and compared it to a pre-industrial simulation, diagnosing the reasons for Eocene warmth by making use of a simple 1-D energy-balance model. They reported a larger polar warming than many previous studies, which they attributed to local radiative forcing changes rather than modified poleward heat transport. The Eocene simulation was carried out under 2× CO_2 levels, and a globally homogeneous vegetation was prescribed with characteristics similar to present-day woody savannas.

Huber and Caballero (2011) presented a set of Eocene CCSM3 simulations, originally published by Liu et al. (2009), with the main aim of comparing these with a new compilation of proxy terrestrial temperature data. At high CO_2 (16×), they

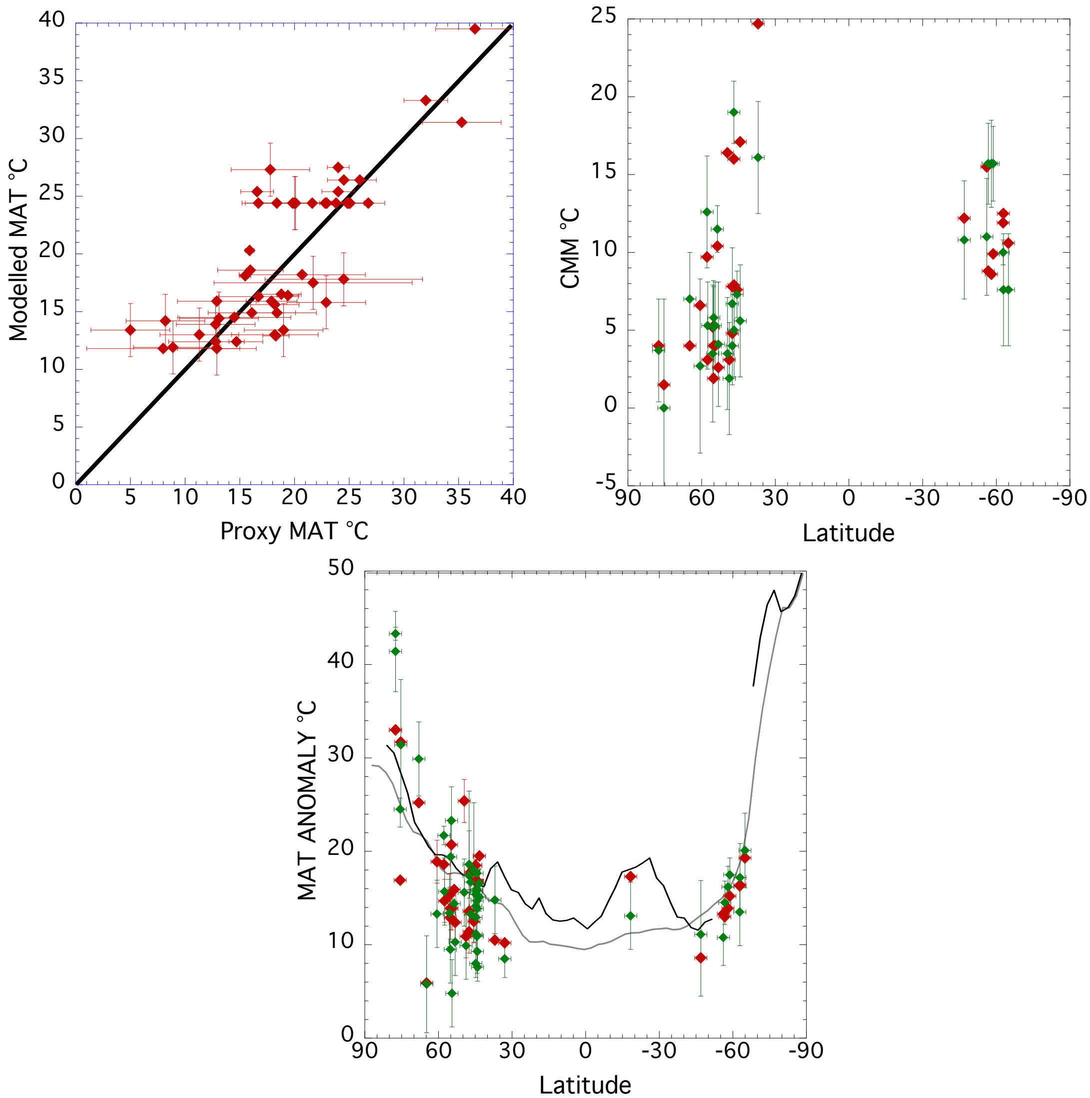

FIGURE 5.—(Upper left) A pointwise comparison of modeled annual mean terrestrial temperatures plotted from Huber's CCSM3 simulation at 4480 ppmv CO_2 against those reconstructed from terrestrial proxies are plotted, and the 1-to-1 line is plotted in black. (Upper right) Cold-month-mean temperature from same simulations (red) and proxy data estimates (green) versus latitude. (bottom) A pointwise comparison of the anomaly of annual mean terrestrial surface temperature from the simulation (red) and proxy data estimates (green) minus the modern temperature value at those locations versus latitude. The zonal mean anomaly of the model case minus the modern model climatology is shown in gray on this figure for reference. The land-only zonal mean anomaly is also shown in black. This is more fully described in Huber and Caballero, 2011.

obtained good agreement with data from mid and high latitudes. Winguth et al. (2010, 2012) carried out an independent set of CCSM3 simulations motivated by investigating the role of hydrates as a possible cause of the PETM. They found evidence of nonlinear ocean warming and enhanced stratification in response to increasing atmospheric CO_2 concentrations, and a shift of deep water formation from northern and southern sources to a predominately southern source.

The CCSM3 simulations done by Huber and Winguth differ mainly in the treatment of aerosols. In the CCSM3 simulation by Winguth, a high aerosol load is applied, whereas the CCSM simulation by Huber considers a lower-than-present-day aerosol distribution consistent with the approach by Kump and Pollard (2008), possibly justified by reduced ocean productivity and thus reduced DMS emissions. A globally reduced productivity is supported by the recent study of

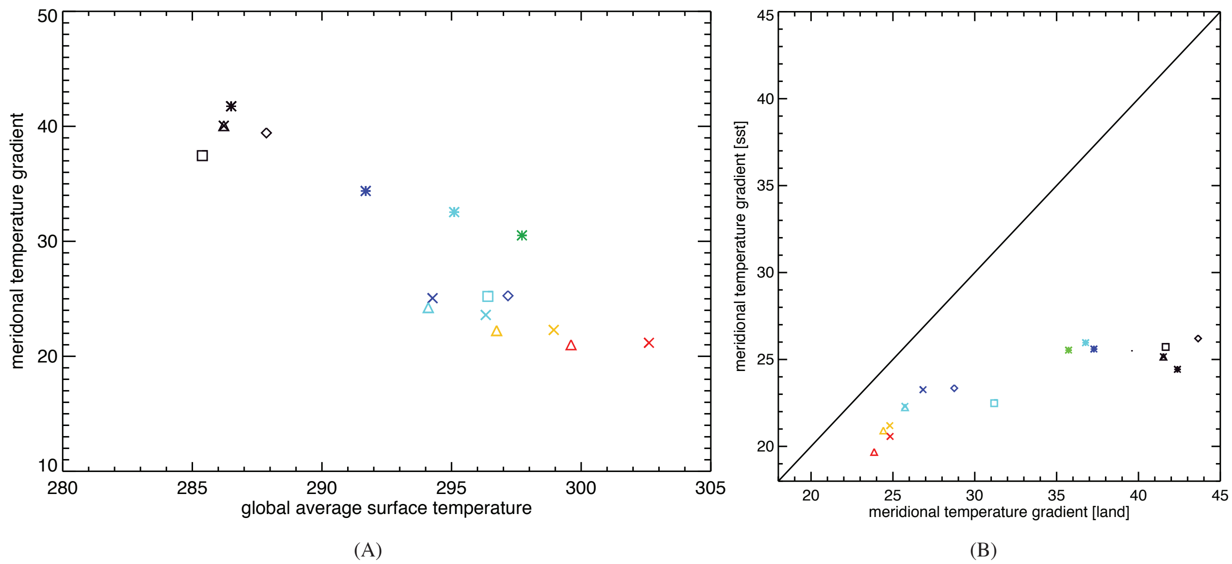

FIGURE 6.—A) Meridional sea-surface temperature gradient as a function of global mean surface temperature in ° K. B) Meridional surface temperature gradient over land versus ocean. Symbols and colors represent different models as in Figure 1. Adapted from Lunt et al. (2012).

Winguth et al. (2012). However, it remains uncertain to what extent intensified volcanism near the PETM might have increased the aerosol load (Storey et al., 2007). Work by Goldner et al. (2012) using a fully interactive, prognostic framework for predicting aerosol distributions suggests that the aerosol assumptions made in the Huber CCSM3 studies are more accurate than those made by Winguth.

Roberts et al. (2009) carried out an investigation into the role of the geometry of Arctic gateways in determining Eocene climate with the GISS ModelE-R. They found that restricting Arctic gateways led to freshening of the Arctic Ocean, similar to data associated with the Azolla event (Brinkhuis et al., 2006). They incorporated oxygen isotopes into the hydrological cycle in their model, and used the predicted isotopic concentrations of seawater to more directly compare with proxy temperature estimates.

To evaluate the various climate model simulations, both terrestrial and marine temperature datasets were used. The marine dataset was compiled for and described in Lunt et al. (2012), and the terrestrial data are identical to those presented in Huber and Caballero (2011). The global mean temperature produced by each model and simulation is shown in Figure 1. Figure 2 shows an example of how a pointwise model comparison is done with an example from the Huber CCSM3 simulations described in Huber and Caballero (2011). Figure 3 shows a zonal SST model-data comparison for each model. Each model is capable of simulating Eocene SSTs that are within the uncertainty estimates of the majority of data points. The data points that lie furthest from the model simulations are the ACEX TEX_{86}' Arctic SST estimate (Sluijs et al., 2006), and the $\delta^{18}O$ and TEX_{86} estimates from the southwest Pacific (Bijl et al., 2009; Hollis et al., 2009). The Arctic temperature reconstructions have uncertainty estimates such that, at high CO_2 (8–16×), the CCSM3 model simulations are just within agreement. At this CO_2 level, these models are also consistent with the tropical temperature estimates. It is likely that other models could also obtain similarly high Arctic temperatures if they were run at sufficiently high CO_2 or low aerosol forcing. Also, given that some of these models (e.g., HadCM) have a higher climate sensitivity than CCSM3, this model-data consistency could be potentially obtained at a lower CO_2 than in CCSM3.

Figure 4 shows the terrestrial temperature model-data comparison for each model. Those models that have been run at high CO_2 (i.e., CCSM3 simulations) show good agreement with the data across all latitudes. For example, focusing on CCSM3, as shown in Figure 5 and described further in Huber and Caballero (2011), at a high CO_2 value (4480), it is possible to produce terrestrial annual mean temperatures that are in

TABLE 3.—Major remaining questions and proxies that might help answer them.

Key Climate Parameter	Proxy	Important Questions
Tropical temperatures	$\delta^{18}O$ Mg/Ca TEX_{86} MBT/CBT Clumped Isotopes	•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, summer production? •Are productivity and calcification biased toward relatively cool conditions? •How much do preservation and diagenesis alter the proxy signals? •Can we only trust foraminifera with 'glassy' preservation? •Under what conditions does Mg/Ca produce valid temperatures? •What is the Mg/Ca of seawater as a function of time? •Are compound specific isotope analyses also subject to alteration? Seasonal biases? •What techniques can be used to constrain the upper range of terrestrial temperatures?
Extratropical temperatures	Leaf Physiognomy Palynofloral transfer functions Sea ice diatoms $\delta^{18}O$ Mg/Ca TEX_{86} MBT/CBT Clumped Isotopes	•Is there sea ice? Sea ice introduces a fundamental nonlinearity into climate. When did it first form? •Are high latitude proxies recording only the warm events or warm season? Is polar warmth a fiction of aliasing 'hyperthermal' events? •What is the oxygen isotopic composition of polar sea water? Do hydrological cycle and ocean circulation changes alter this?
Greenhouse gases	Boron Alkenones Stomata Nahcolite	•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? •Most proxies have decreasing accuracy at high CO_2 values, how do we estimate CO_2 accurately when it is high? •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?

TABLE 3.—Continued.→

general agreement with the data everywhere and winter season temperatures in similar agreement, and to generate a reduction of the meridional temperature gradient in substantive agreement with proxies. The other models do not simulate such high temperatures, but, as with SST, it does appear that if they had been run at higher CO_2, the model-data agreement would have been better. The HadCM model appears to be somewhat of an outlier in the Northern Hemisphere high latitudes, as it shows less polar amplification than the other models, an effect also seen in SST.

Lunt et al. (2012) provided a first-order estimate of the CO_2 level for each model that could give the best agreement with the proxy estimates. For HadCM, CCSM3 (Huber), and CCSM3 (Winguth), using SST, this is 2100 ppmv, 4100 ppmv, and 5400 ppmv, respectively, and using terrestrial annual mean temperature, this is 2800 ppmv, 4500 ppmv, and 6300 ppmv, respectively. These estimates come with many caveats; however, they do indicate the magnitude of the range of CO_2 values that could be considered consistent with model results. These values are significantly higher than those presented for this time period in the compilation of Beerling and Royer (2011). This implies that this generation of models may still be too insensitive to radiative forcing, or that missing radiative feedbacks still need to be included as boundary conditions in the models.

Several conclusions and implications were made in the Lunt et al. (2012) comparison project. The model results show a large spread in global mean temperatures, for example a 9°C range in surface air temperature under a single CO_2 value, and are characterized by warming preferentially in different regions. This highlights the importance of the inherent spread of values of climate sensitivity of the models and their different degrees of polar amplification. The models that have been run at sufficiently high CO_2 show very good agreement with the terrestrial data. No models show evidence for tropical thermostats nor support for a low value of climate sensitivity, i.e. $S < 1$ seems ruled out by the model-data comparison, and even $S < 2$ also seems incompatible given that it would require unexplainable amounts of CO_2 to match proxy temperatures. It is also intriguing that several models (CCSM3, HADCM3L, ECHAM5) seem to have higher paleoclimate sensitivity values or sensitivity values that increase with increasing global mean temperature (Figure 1), in seeming agreement with proxy implications.

The changes in meridional temperature gradient that bear on the weak gradient problem are summarized in Figure 6, which shows the surface temperature difference between the low latitudes (< 30°) and the high latitudes (> 60°) as a function of global mean temperature (Figure 6A) and how this is partitioned between land and ocean warming (Figure 6B). All of the Eocene simulations have a reduced meridional surface temperature gradient compared with the pre-industrial, and the gradient reduces further as CO_2 increases, i.e., polar amplification increases (Figure 6A). However, there is a high degree of inter-model variability in the absolute Eocene gradient, with HadCM appearing to be an outlier with a relatively high Eocene gradient. There is some indica-

Key Climate-Parameter	Proxy	Important Questions
Tropical temperatures	$\delta^{18}O$ Mg/Ca TEX_{86} MBT/CBT Clumped Isotopes	•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, summer production? •Are productivity and calcification biased toward relatively cool conditions? •How much do preservation and diagenesis alter the proxy signals? •Can we only trust foraminifera with 'glassy' preservation? •Under what conditions does Mg/Ca produce valid temperatures? •What is the Mg/Ca of seawater as a function of time? •Are compound specific isotope analyses also subject to alteration? Seasonal biases? •What techniques can be used to constrain the upper range of terrestrial temperatures?

tion that the models are asymptoting towards a minimum gradient of about 20°C. This is in agreement with the previous work of Huber et al. (2003) that implied that meridional temperature gradients of the order 20°C were physically realistic, even without large changes to meridional heat transport. Compared with preindustrial, the meridional surface temperature gradient reduces more on land than over ocean (Figure 6B). For HadCM, this applies also to the Eocene simulations as CO_2 increases. However, for the CCSM3 simulations, the meridional temperature gradient is reduced by a similar amount over land and ocean as a function of CO_2, with some indication, at maximum (16×) CO_2, that the SST gradient starts reducing more over ocean than over land. This implies that, when considering changes relative to the modern, it is possible to have substantially different temperature changes over land compared with over ocean at the same latitude. This shows the importance of differentiating terrestrial and oceanic signals when considering the consistency between different proxy data and between data and models.

Some of the differences between the model results can be attributed to differences in the experimental design. In particular, some models apply a very generic Eocene vegetation, which is not particularly realistic. A slightly more coordinated study could provide guidelines for ways to better represent Eocene vegetation (e.g., using palynological data or dynamic vegetation models where available). This would provide an ensemble of model results that better represent the true uncertainty in the model simulations. Other inconsistencies between model simulations should not necessarily be eliminated. For example, different models using different paleogeographical reconstructions may be more representative of the true spread of model results than if all groups used a single paleogeography.

On the data side, better understanding of the temperature proxies and their associated uncertainties—in particular, seasonal effects—is a clear goal for future work, as is greater geographical and finer temporal coverage. Perhaps most crucial of all, better CO_2 constraints from proxies would be of huge benefit to model-data comparison exercises such as this. Recently, much work is being undertaken in this area, but this should be intensified wherever possible. Note that at high CO_2, due to the logarithmic nature of the CO_2 forcing, proxies that may have relatively coarse precision at low CO_2 can actually provide very strong constraints on the CO_2 forcing itself. Such constraints on CO_2, combined with proxy temperature reconstructions with well defined uncertainty ranges, could provide a strong constraint on model simulations, providing quantitative metrics for assessing model performance, and could ultimately provide relative weightings for model simulations of future climates.

OUTLOOK

Significant progress has been made toward demonstrating that, with the right forcing, and with improved proxy data, it is possible to bring the models and proxies for past greenhouse climates into reasonably close agreement on some of the fundamental climate parameters such as global mean temperature, winter season terrestrial temperature, and meridional temperature gradient. This leaves much exciting work to be done.

New modeling tools enable the community to tackle a new generation of paleoclimate and paleoenvironmental questions, such as changes in the hydrological cycle (White et al., 2001; Poulsen et al., 2007; Speelman et al., 2010; Tindall et al., 2010; Clementz and Sewall, 2011; Bowen, 2011) and monsoons (Fricke et al., 2009; Herold et al., 2011a; Huber and Goldner, 2012) in past greenhouse climates. We are also now in a position to explore other interesting areas such as: fundamental transitions in atmospheric (Brielerly et al., 2009; Caballero and Huber, 2009) or oceanic (von der Heydt and Dijkstra, 2008; Hague et al., in press;) circulation modes; alterations in major modes of climate variability (Garric and Huber, 2003; Huber and Caballero, 2003; Lyle et al., 2008; Galeotti et al., 2010a; Lenz et al., 2010); and a wealth of interactions between atmospheric (Lamarque et al., 2006, 2007; Kump and Pollard, 2008; Beerling et al. 2009), terrestrial (Pound et al., 2011; Herold et al., 2011a; Deconto et al., 2012; Zhou et al., 2012) and oceanic biogeochemical cycles (Pagani et al., 2011; Winguth et al., 2011). The role of ice and sea-level changes through greenhouse climates is an area of particular interest (DeConto et al., 2008; De Boer et al., 2010; Lunt et al., 2010; Gasson et al., 2012). Consequently, areas on the cutting edge of greenhouse paleoclimate modeling currently include the addition of interactive isotopes, biogeochemistry, vegetation, and ice sheets into paleoclimate models, and, of course, the search for new, interesting lessons about climate dynamics that can be learned from studying the past.

New proxy records are necessary to drive this endeavor forward. Some areas of particular note are shown in Table 3, which aims to summarize some of the major remaining questions and the proxies that might assist in answering them.

ACKNOWLEDGEMENTS

MH would like to thank the organizers, Linda Ivany and Brian Huber, for putting this volume and short course together and for their forbearance. Reviews by Linda Ivany and Dan Horton greatly improved the manuscript. MH also would like to acknowledge that this contribution relies heavily on some prior collaborative studies (Huber and Caballero, 2011; Thomas and Huber, 2006; Thomas et al., 2006; Pagani et al., 2012, and Lunt et al, 2012). MH is funded by the NSF, specifically P2C2 grants 0902882 and 0902780.

REFERENCES

ABBOT, D. S., M. HUBER, G. BOUSQUET, AND C. C. WALKER. 2009a. High-CO_2 cloud radiative forcing feedback over both land and ocean in a global climate model. Geophysical Research Letters, L05702. doi:10.1029/2008GL036703.

ABBOT, D. S., AND E. TZIPERMAN. 2008. A high latitude convective cloud feedback and equable climates. Quarterly Journal of the Royal Meteorological Society, doi:10.1002/qj.211.

ABBOT, D. S., C. C. WALKER, AND E. TZIPERMAN. 2009b. Can a convective cloud feedback help to eliminate winter and spring sea ice at high CO_2 concentrations? Journal of Climatology, doi:/2009JCLI2854.

ALLEN, M. R., AND W. J. INGRAM. 2002. Constraints on future changes in the hydrological cycle. Nature, 419:224–228.

ALLEY, N. F., AND L. A. FRAKES. 2003. First known Cretaceous glaciation: Livingston Tillite Member of the Cadnaowie Formation, South Australia. Australian Journal of Earth Sciences, 50:139–144.

ANDO, A., B. T. HUBER, K. G. MACLEOD, T. OHTA, AND B.-K. KHIM. 2009. Blake Nose stable isotopic evidence against the mid-Cenomanian glaciation hypothesis. Geology, 37:451-454

AZIZ, H., F. J. HILGEN, G. M. VAN LUIJK, A. SLUIJS, M. J. KRAUS, J. M. PARES, AND P. D. GINGERICH. 2008. Astronomical climate control on paleosol stacking patterns in the upper Paleocene–lower Eocene Willwood Formation, Bighorn Basin, Wyoming. Geology, 36:531–534.

BAO, H., P. L. KOCH, AND D. RUMBLE, III. 1999. PaleoceneEocene climatic variation in western North America; evidence from the $\delta^{18}O$ of pedogenic hematite. Geological Society of America Bulletin, 111:1405–1415.

BARKER, P. F., BARRETT, P. J., COOPER, A. K., AND P. HUYBRECHTS. 1999. Antarctic glacial history from numerical models and continental margin sediments. Palaeogeography, Palaeoclimatology, Palaeoecology, 150:247–267.

BARKER, F., AND E. THOMAS. 2004. Origin, signature and palaeoclimate influence of the Antarctic Circumpolar Current. Earth Science Reviews, 66:143–162.

BARRON, E. J. 1983. A warm, equable Cretaceous: the nature of the problem. Earth-Science Reviews, 19:305-338.

BARRON, E. J. 1987. Eocene equator-to-pole surface ocean temperatures: A significant climate problem? Paleoceanography, 2:729–732.

BARRON, E. J., P. J. FAWCETT, W. H. PETERSON, D. POLLARD, AND S. L. THOMPSON. 1995. A "simulation" of mid-Cretaceous climate. Paleoceanography, 10:953.

BARRON, E .J., AND W. H. PETERSON. 1991. The Cenozoic ocean circulation based on ocean general circulation model results, Palaeogeography Palaeoclimatology Palaeoecology, 83:1–28.

BARRON, E., W. PETERSON, D. POLLARD, AND S. THOMPSON. 1993. Past climate and the role of ocean heat transport: Model simulations for the Cretaceous. Paleoceanography, 8:785–798.

BARRON, E. J., S. L. THOMPSON, AND S.H. SCHNEIDER. 1981. An ice-free Cretaceous? Results from climate model. Science, 212:501–508.

BECKMANN, B, S. FLÖGEL, P. HOFMANN, M. SCHULZ AND T. WAGNER. 2005. Orbital forcing of Cretaceous river discharge in tropical Africa and ocean response. Nature, 437:241–244.

BEERLING, D., R. A. BERNER, F. T. MACKENZIE, M. B. HARFOOT, AND J. A. PYLE. 2009: Methane and the CH4-related greenhouse effect over the past 400 million years. American Journal of Science, 309:97–113.

BEERLING, D., AND D. ROYER. 2011. Convergent Cenozoic CO_2 history, Nature Geosciences, 4:418–420.

BEERLING, D. J., A. FOX, D. S. STEVENSON, AND P. J. VALDES. 2011. Enhanced chemistry-climate feedbacks in past greenhouse worlds. Proceedings of the National Academy of Science USA, 108:9770–9775.

BERGGREN, W. A. AND D. R. PROTHERO. 1992. Eocene–Oligocene climatic and biotic evolution: An overview, p. 1–28. *In* D. R. PROTHERO, AND W. A. BERGGREN (eds.). Eocene–Oligocene Climatic and Biotic Evolution. Princeton University Press, Princeton, NJ.

BERRY, E. W. 1922. A possible explanation of upper Eocene climates. Proceedings of the American

Philosophical Society, 61:1–14.

BICE, K. L., E. J. BARRON, AND W. H. PETERSON. 1997. Continental runoff and early Cenozoic bottom water sources. Geology, 25:51–954.

BICE, K. L., E. J. BARRON, AND W. H. PETERSON. 1998. Reconstruction of realistic Early Eocene paleobathymetry and ocean GCM sensitivity to specified ocean basin configuration, p. 212–226. *In* T. J. CROWLEY, AND K. C. BURKE (eds.). Tectonic Boundary Conditions for Climate Reconstructions. Oxford University Press, New York.

BICE, K. L., D. BIRGEL, P. A. MEYERS, K. A. DAHL, K.-U. HINRICHS, AND R. D. NORRIS. 2006. A multiple proxy and model study of Cretaceous upper ocean temperatures and atmospheric CO_2 concentrations. Paleoceanography, 21:PA2002. doi:10.1029/2005PA001203.

BICE, K. L., B. T. HUBER, AND R. D. NORRIS. 2003. Extreme polar warmth during the Cretaceous greenhouse? Paradox of the late Turonian $\delta^{18}O$ record at Deep Sea Drilling Project Site 511. Paleoceanography, 18. doi:10.1029/2002PA000848.

BICE, K. L., AND J. MAROTZKE. 2001. Numerical evidence against reversed thermohaline circulation in the warm Paleocene/Eocene ocean. Journal of Geophysical Research, 106:11529-11542.

BIJL, P., S. SCHOUTEN, A. SLUIJS, G. REICHART, J. ZACHOS, AND H. BRINKHUIS. 2009. Early Palaeogene temperature evolution of the southwest pacific ocean. Nature, 461:776–779.

BILLUPS, K. AND D. SCHRAG. 2003. Application of benthic foraminiferal Mg/Ca ratios to questions of Cenozoic climate change. Earth and Planetary Science Letters, 209:181–195.

BOHATY, S.M., AND J. ZACHOS. 2003. Significant southern ocean warming event in the late middle Eocene. Geology, 31:1017–1020.

BONY, S., R. COLMAN, V. M. KATTSOV, R. P. ALLAN, C. S. BRETHERTON, J.-L. DUFRESNE, A. HALL, S. HALLEGATTE, M. M. HOLLAND, W. INGRAM, D. A. RANDALL, B. J. SODEN, G. TSELIOUDIS, AND M. J. WEBB. 2006. How well do we understand and evaluate climate change feedback processes? Journal of Climatology, 19:3445–3482.

BOULILA, S., B. GALBRUN, J. LASKAR, AND H. PÄLIKE. 2012. A ~9 myr cycle in Cenozoic δ13C record and long-term orbital eccentricity modulation: is there a link? Earth and Planetary Science Letters, 317-318:273–281.

BRADY, E.C., R.M. DECONTO, AND S.L. THOMPSON. 1998. Deep water formation and poleward ocean heat transport in the warm climate extreme of the Cretaceous (80 Ma). Geophysical Research Letters, 25:4205–4208.

BRINKHUIS H, S. SCHOUTEN, M. E. COLLINSON, A. SLUIJS, J. S. SINNINGHE DAMSTÉ, G. R. DICKENS, M. HUBER, T. M. CRONIN, J. ONODERA, K. TAKAHASHI, J. P. BUJAK, R. STEIN, J. VAN DER BURGH, J. S. ELDRETT, I. C. HARDING, A. F. LOTTER, F. SANGIORGI, H.VAN KONIJNENBURG-VAN CITTERT, J. W. DE LEEUW, J. MATTHIESSEN, J. BACKMAN, K. MORAN, AND THE EXPEDITION 302 SCIENTISTS. 2006. Episodic fresh surface waters in the Eocene Arctic Ocean. Nature, 441:606–609.

BROECKER, W. S., AND T.-H. PENG. 1982. Tracers in the Sea. Lamont-Doherty Earth Observatory. Palisades, N. Y.

BUSH, A. B. G. 1997. Numerical simulation of the Cretaceous Tethys Circumglobal Current. Science, 275:807–8.

BUSH, A. B. G., AND S .G. H. PHILANDER. 1997. The Late Cretaceous: simulation with a coupled ocean–atmosphere general circulation model. Paleoceanography, 12:495–516.

CABALLERO, R., AND P. L. LANGEN. 2005: The dynamic range of poleward energy transport in an atmospheric general circulation model. Geophysical Research Letters., 32, L02705. doi:10.1029/2004GL021581.

CARTER, R. M., L. CARTER, AND I. N. MCCAVE. 1996. Current controlled sediment deposition from the shelf to the deep ocean: The Cenozoic evolution of circulation through the SW Pacific gateway. Geologische Rundschau, 87:438–451.

CHRISTIE-BLICK, N., MOUNTAIN, G. S. AND K. G. MILLER. 1990. Seismic stratigraphic record of sea level change, p. 116– 140. *In* Sea-level Change. National Academy Press, Washington, D.C.

CLARKE, L. J., AND H. C. JENKYNS. 1999. New oxygen isotope evidence for long-term Cretaceous climatic change in the Southern Hemisphere. Geology, 27:699–702.

CLAUSSEN, M., L. A. MYSAK, A. J. WEAVER, M. CRUCIFIX, T. FICHEFET, M.-F. LOUTRE, S. L. WEBER, J. ALCAMO, V. A. ALEXEEV, A. BERGER, R. CALOV, A. GANOPOLSKY, H. GOOSSE, G. LOHMAN, F. LUNKEIT, I. I. MOHKOV, V. PETOUKHOV, P. STONE, AND Z. WANG. 2002. Earth System Models of Intermediate Complexity: Closing the gap in the spectrum of climate system models. Climate Dynamics, 18:579–586.

CLEMENT, A. C., R. SEAGER, M. A. CANE, AND S. E. ZEBIAK. 1996. An ocean dynamical thermostat. J. Climate, 9:2190–2196.

CLIMAP PROJECT MEMBERS. 1976. The surface of the Ice-Age Earth, Science, 191:1131–1137.

CORFIELD, R., AND R. D. NORRIS. 1996. Deep water circulation in the Paleocene Ocean, p. 443–456. *In* R. W. O'B. KNOX, R. M. CORFIELD, AND R. E. DUNAY (eds.). Correlation of the Early Paleogene in Northwest Europe. Geological Society of London Special Publications 101, London.

COVEY, C., AND E. BARRON. 1988. The role of ocean heat-transport in climatic-change. Earth-Science Reviews, 24:429–445.

COVEY, C., L. C. SLOAN, AND M. I. HOFFERT. 1996.

Paleoclimate data constraints on climate sensitivity: the paleocalibration method. Climatic Change, 32:165–184.

COXALL, H. K., A. WILSON, H. PALIKE, C. H. LEAR, AND J. BACKMAN. 2005. Rapid stepwise onset of Antarctic glaciation and deeper calcite compensation in the Pacific Ocean. Nature, 433:53–57.

CRAMER, B. S., J. D. WRIGHT, D. V. KENT, AND M. P. AUBRY. 2003. Orbital climate forcing of 13C excursions in the late Paleocene early Eocene (chrons C24nC25n). Paleoceanography 18:PA1097.

CRISTINI, L., K. GROSFELD, M. BUTZIN, AND G. LOHMANN. 2012. Influence of the opening of the Drake Passage on the Cenozoic Antarctic Ice Sheet: A modeling approach. Palaeogeography, Palaeoclimatology, Palaeoecology, 339–341:66–73.

CROWLEY, T. J. 2000. CLIMAP SSTs re-revisited, Clim. Dyn., 16: 241–255.

CROWLEY, T. J., D. A. SHORT, J. G. MENGEL, AND G. R. NORTH. 1986. Role of seasonality in the evolution of climate during the last 100 million years. Science, 231:579–584.

CROWLEY, T. J., AND J. C. ZACHOS. 2000. Comparison of zonal temperature profiles for past warm time periods, p. 50–76. *In* B. T. HUBER, K. G. MACLEOD, AND S. L. WING (eds.). Warm Climates in Earth History, Cambridge University Press.

DANABASOGLU, G. A. 2004. Comparison of global ocean circulation model solutions obtained with synchronous and accelerated integration methods. Ocean Modeling, 7:323–341.

DANABASOGLU, G., J. C. MCWILLIAMS, AND W. G. LARGE. 1996. Approach to equilibrium in accelerated global oceanic models. Journal of Climate, 9:1092–1110.

DAVIES, A., A. E. S. KEMP, AND H. PÄLIKE. 2011. Tropical ocean atmosphere controls on inter-annual climate variability in the Cretaceous Arctic. Geophysical Research Letters, 38:L03706.

DAVIES, A., A. E. S. KEMP, G. P. WEEDON, AND J. A. BARRON. 2012. El Niño–Southern Oscillation variability from the Late Cretaceous Marca Shale of California. Geology, 40:15–18.

DE BOER, B., R. VAN DE WAL, R. BINTANJA, L. LOURENS, AND E. TUENTER. 2010. Cenozoic global ice-volume and temperature simulations with 1-D ice-sheet models forced by benthic $\delta^{18}O$ records. Annals of Glaciology, 5:23–33.

DECONTO, R.M., GALEOTTI, S., PAGANI, M., TRACY, D., SCHAEFER, K., ZHANG, T., POLLARD, D., AND D. J. BEERLING. 2012. Past extreme warming events linked to massive carbon release from thawing permafrost. Nature, 484:87–91.

DECONTO, R. M., W. W. HAY, S. L. THOMPSON, AND J. BERGENGREN. 1999. Late Cretaceous climate and vegetation interactions: The cold continental interior paradox, p. 391–406. *In* E. Barrera and C. Johnson (eds.). Evolution of the Cretaceous Ocean/Climate System, Geological Society of America Special Paper 332, Boulder, CO.

DECONTO, R. M., AND D. POLLARD. 2003. Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2. Nature, 421:245–249.

DECONTO, R. M., D. POLLARD, P. A. WILSON, H. PÄLIKE, C. H. LEAR, AND M. PAGANI. 2008. Thresholds for Cenozoic bipolar glaciation. Nature, 455:652–656.

DECONTO, R. M., S. L. THOMPSON, AND D. POLLARD. 2000. Recent advances in paleoclimate modeling: toward better simulations of warm paleoclimates, p. 21–49. *In* B. T. HUBER, K. G. MACLEOD, AND S. L. WING (eds.). Warm Climates in Earth History, Cambridge University Press.

D'HONDT, S., AND M. A. ARTHUR. 1996. Late Cretaceous oceans and the cool tropic paradox. Science, 271:1838–1841.

DICKENS, G. R. 2011. Down the rabbit hole: toward appropriate discussion of methane release from gas hydrate systems during the Paleocene–Eocene thermal maximum and other past hyperthermal events. Climate of the Past, 7:831–846.

DICKENS, G. R., J. R. O'NEIL, D. K. REA, AND R. M. OWEN. 1995. Dissociation of oceanic methane hydrate as a cause of the carbon isotope excursion at the end of Paleocene. Paleoceanography, 10:965–971.

DONEY, S. C., W. G. LARGE, AND F. O. BRYAN. 1998. Surface ocean fluxes and water-mass transformation rates in the coupled Climate System Model. Journal of Climate, 9:1420–1441.

DUMITRESCU, M., S. C. BRASSELL, S. SCHOUTEN, E. C. HOPMANS, AND J. S. SINNINGHE DAMSTE´. 2006. Instability in tropical Pacific sea-surface temperatures during the early Aptian. Geology, 34:833–836.

EBERLE, J. J., H. C FRICKE, J. D. HUMPHREY, L. HACKETT, M. G. NEWBREY, AND J. H. HUTCHISON. 2010 Seasonal variability in Arctic temperatures during early Eocene time, Earth and Planetary Sciences Letters, 296:481–486.

ELDRETT, J. S., D. R. GREENWOOD, D.R., HARDING, I.C., AND M. HUBER. 2009. Increased seasonality through the Eocene to Oligocene transition in Northern high latitudes. Nature, 459:969–974.

EMANUEL, K. A. 2002. A simple model of multiple climate regimes, Journal of Geophysical Research—Atmospheric, 107:4077. doi:10.1029/2001JD001002.

FISCHER, A. G., AND L. T. ROBERTS. 1991. Cyclicity in the Green River Formation (lacustrine Eocene) of Wyoming. Journal of Sedimentary Petrology, 61:1146–1154.

FLETCHER, B. J., S. J. BRENTNALL, C. W. ANDERSON, R. A. BERNER, AND D. J. BEERLING. 2008. Atmospheric carbon dioxide linked with Mesozoic and early Cenozoic climate. Nature Geosciences, 1:43–48.

FLÖGEL, S., K. WALLMANN, C. J. POULSEN, J. ZHOU, A. OSCHLIES, S. VOIGT, AND W. KUHNT. 2011. Simulating the biogeochemical effects of volcanic CO_2 degassing on the oxygen-state of the deep ocean during the Cenomanian/Turonian Anoxic Event (OAE2). Earth and Planetary Science Letters, 305:371–384.

FORSTER, A., S. SCHOUTEN, M. BAAS, AND J. S. SINNINGHE DAMSTÉ. 2007a. Mid-Cretaceous (Albian–Santonian) sea surface temperature record of the tropical Atlantic Ocean. Geology, 7:919–922.

FORSTER, A., S. SCHOUTEN, K. MORIYA, P. A. WILSON, AND J. S. SINNINGHE DAMSTÉ. 2007b. Tropical warming and intermittent cooling during the Cenomanian/Turonian oceanic anoxic event 2: sea surface temperature records from the equatorial Atlantic. Paleoceanography, 22. doi:10.1029/2006PA001349.

FRAKES, L. A., N. F. ALLEY, AND M. DEYNOUX. 1995. Early Cretaceous ice rafting and climate Zonation in Australia. International Geology Review, 37:567–583.

FRICKE, H. C. 2003. Investigation of early Eocene water-vapor transport and paleoelevation using oxygen isotope data from geographically widespread mammal remains. Geological Society of America Bulletin, 115:1088–1096.

FRICKE, H. C., B. Z. FOREMAN, AND J. O. SEWALL. 2009. Integrated climate model-oxygen isotope evidence for a North America monsoon during the Late Cretaceous. Earth and Planetary Science Letters, 29:11–21.

FRICKE, H. C., AND S. L. WING. 2004. Oxygen isotope and paleobotanical estimates of temperature and δ^{18}O-Latitude gradients over North America during the early Eocene. American Journal of Science, 304: 612–635.

FU, R., A. DEL GENIO, W. B. ROSSOW, AND W. T. LIU. 1992. Cirrus-cloud thermostat for tropical sea-surface temperatures tested using satellite data. Nature, 358:394–397.

FU, Q., C. M. JOHANSON, J. M. WALLACE, AND T. REICHLER. 2006. Enhanced mid-latitude tropospheric warming in satellite measurements. Science, 312:1179.

GARRIC, G., AND M. HUBER. 2003. Quasi-decadal variability in paleoclimate records: Sunspot cycles or intrinsic oscillations? Paleoceanography, 18. doi:10.1029/2002PA000869.

GALE, A. S., J. HARDENBOL, B. HATHAWAY, W. J. KENNEDY, J. R. YOUNG, AND V. PHANSALKAR. 2002. Global correlation of Cenomanian (Upper Cretaceous) sequences: evidence for Milankovitch control on sea level. Geology, 30:291–294.

GALE, A. S., S. VOIGT, B. B. SAGEMAN, AND W. J. KENNEDY. 2008. Eustatic sea-level record for the Cenomanian (Late Cretaceous)—extension to the Western Interior Basin, USA. Geology, 36:859–862.

GALEOTTI, S., A. VON DER HEYDT, M. HUBER, D. BICE, H. DIJKSTRA, T. JILBERT, L. LANCI, AND G.-J. REICHART. 2010a. Evidence for active El Niño Southern Oscillation variability in the late Miocene greenhouse climate. Geology, 38:419–422.

GALEOTTI, S., S. KRISHNAN, M. PAGANI, L. LANCI, A. GAUDIO, J. C. ZACHOS, S. MONECHI, G. MORELLI, AND L. LOURENS. 2010b. Orbital chronology of Early Eocene hyperthermals from the Contessa Road section, central Italy. 2010. Earth and Planetary Sciences Letters, 290:192–200.

GASSON, E., M. SIDDALL, D. J. LUNT, O. J. L. RACKHAM, C. H. LEAR, AND D. POLLARD. 2012. Exploring uncertainties in the relationship between temperature, ice volume, and sea level over the past 50 million years. Review of Geophysics, 50, RG1005. doi:10.1029/2011RG000358.

GENT, P. R., F. O. BRYAN, G. DANABASOGLU, S. C. DONEY, W.R. HOLLAND, W. G. LARGE, AND J. C. MCWILLIAMS. 1998. The NCAR Climate System Model global ocean component. Journal of Climate, 11:1287–1306.

GOLDNER, A., M. HUBER, AND R. CABALLERO. Does Antarctica glaciation cool the world? Climate of the Past Discussions, submitted.

GREENWOOD, D. R., J. R. BASINGER, AND R. Y. SMITH. 2010. How wet was the Arctic Eocene rain forest? Estimates of precipitation from Paleogene Arctic macrofloras. Geology, 38:15–18.

GREENWOOD, D. R., AND S. L. WING. 1995. Eocene continental climates and latitudinal temperature gradients. Geology, 23:1044–1048.

HANDOH, I. C., G. R. BIGG, E. J. W. JONES, AND M. INOUE. 1999. An ocean modeling study of the Cenomanian Atlantic: equatorial paleo-upwelling, organic-rich sediments and the consequences for a connection between the proto-North and South Atlantic. Geophysical Research Letters, 26:223–226.

HANDOH, I. C., G. R. BIGG, AND E. J. W. JONES. 2003. Evolution of upwelling in the Atlantic Ocean Basin, Palaeogeography Palaeoclimatology Palaeoecology, 202:31–58.

HANEY, R. L. 1971. Surface thermal boundary condition for ocean circulation models. J. Phys. Oceanogr., 1:241–248.

HARDING, I. C., A. J. CHARLES, J. E. A. MARSHALL, H. PALIKE, A. P. ROBERTS, P. A. WILSON, E. JARVIS, R. THORNE, E. MORRIS, R. MOREMON, R. B. PEARCE, AND S. AKBARI. 2011. Sea-level and salinity fluctuations during the Paleocene–Eocene thermal maximum in Arctic Spitsbergen. Earth and Planetary Sciences Letters, 303:97–107.

HARTMANN, D. L., AND M. L. MICHELSEN. 1993.

Large-scale effects on the regulation of tropical sea-surface temperature. Journal of Climatology, 6:2049–2062.

HARTMANN, D. L., AND M. L. MICHELSEN. 2002. No evidence for iris. Bulletin of the American Meteorological Society, 83: 49– 254.

HAY, W. W. 2008. Evolving ideas about the Cretaceous climate and ocean circulation. Cretaceous Research, 29:725–753. HAQ, B. U., J. HARDENBOL, AND P. R.VAIL. 1987. Chronology of fluctuating sea levels since the Triassic. Science, 235:1156–1167.

HAY, W. W., AND J. R. SOUTHAM. 1977. Modulation of marine sedimentation by the continental shelves, p. 569–604. *In* N. R. ANDERSON, AND A. MALAHOFF (eds.). The Fate of Fossil Fuel CO_2 in the Oceans. Marine Science Series, 6. Plenum Press, New York.

HEAD, J. J., J. I. BLOCH, A. K. HASTINGS, J. R. BOURQUE, E. A. CADENA, F. HERRERA, P. D. POLLY, AND C. A. JARAMILLO. 2009. Giant boid snake from the Palaeocene neotropics reveals hotter past equatorial temperatures. Nature, 457:715–718.

HEINEMANN, M., J. H. JUNGCLAUS, AND J. MAROTZKE. 2009. Warm Paleocene/Eocene climate as simulated in ECHAM5/MPI-OM. Climate of the Past, 5:785–802.

HEINZE, C. AND T. J. CROWLEY. 1997. Sedimentary response to gateway circulation changes. Paleoceanography, 12:742–754.

HERMAN, A.B., AND R. A. SPICER. 1996. Palaeobotanical evidence for a warm Cretaceous Arctic Ocean. Nature, 380:330–333.

HEROLD, N. K., M. HUBER, D. R. GREENWOOD, R. D. MÜLLER, AND M. SETON. 2011a. Early to middle Miocene monsoon climate in Australia. Geology, 39:3–6.

HEROLD, N.K., D. MÜLLER, AND M. HUBER. 2011b. Insights into Miocene ocean circulation from a coupled climate model: Part I Atmospheric Circulation, Journal of Climate, 24:6353–6372.

HEROLD, N., HUBER, M., MÜLLER, R.D. AND M. SETON. Modelling the Miocene climatic optimum: ocean circulation. 2012. Paleoceanography, 27:PA1209. doi:10.1029/2010PA002041.

HIGGINS, J. A., AND D. P. SCHRAG. 2006. Beyond methane: towards a theory for the Paleocene–Eocene thermal maximum. Earth and Planetary Sciences Letters, 245:523–537.

HOLLIS, C. J., L. HANDLEY, E. M. CROUCH, H. E. G. MORGANS,J. A. BAKER, J. CREECH, K. S. COLLINS, S. J. GIBBS, M. HUBER, S. SCHOUTEN, J. C. ZACHOS, AND R. D. PANCOST. 2009. Tropical sea temperatures in the high-latitude South Pacific during the Eocene. Geology, 37:99–102.

HOTINSKI, R. M., AND J. R. TOGGWEILER. 2003. Impact of a Tethyan circumglobal passage on ocean heat transport and "equable" climates, Paleoceanography, 18. doi: 10.10.1029 /2001PA000730.

HUBER, M. 2008. A hotter greenhouse? Science, 321:353–354.

HUBER, M. 2009. Snakes tell a torrid tale. Nature, 457:669–671.

HUBER, M., H. BRINKHUIS, C. E. STICKLEY, K. DOOS, A SLUIJS, J. WARNAAR, S. A. SCHELLENBERG, AND G. L. WILLIAMS. 2004. Eocene circulation of the Southern Ocean: was Antarctica kept warm by subtropical waters: Paleoceanography, 19, PA4026.

HUBER, M., AND R. CABALLERO. 2003. Eocene El Niño: Evidence for robust tropical dynamics in the "hothouse." Science, 299:877–88.

HUBER, M., AND A. GOLDNER. 2011. Eocene monsoons. Journal of Asian Earth Sciences, 44:3–23.

HUBER, M., AND D. NOF. 2006. The ocean circulation in the southern hemisphere and its climatic impacts in the Eocene. Palaeogeography Palaeoclimatology Palaeoecology, 231:9–28.

HUBER, B. T., R. D. NORRIS, AND K. G. MACLEOD. 2002. Deep sea paleotemperature record of extreme warmth during the Cretaceous Period. Geology, 30:123–126.

HUBER, M., AND L. C. SLOAN. 1999. Warm climate transitions: a general circulation modeling study of the Late Paleocene thermal maximum (~56 Ma). Journal of Geophysical Research, 104:16633–16655.

HUBER, M., AND L. C. SLOAN. 2000. Climatic responses to tropical sea surface temperature changes on a "greenhouse" Earth. Paleoceanography, 15:443–450.

HUBER, M., AND L. C. SLOAN. 2001. Heat transport, deep waters, and thermal gradients. Geophysical Research Letters, 28:3481–3484.

HUBER, M., L. C. SLOAN, AND C. SHELLITO. 2003. Early Paleogene oceans and climate: A fully coupled modelling approach using NCAR's CSM, p. 25–47. *In* S. L. WING, P. D. GINGERICH, B. SCHMITZ, AND E. THOMAS (eds.). Causes and Consequences of Globally Warm Climates in the Early Paleogene. Geological Society of America Special Paper 369, Boulder, CO.

HUTCHINSON, J. H. 1982. Turtle, crocodilian, and champsosaur diversity changes in the Cenozoic of the north-central region of western United States. Palaeogeography, Palaeoclimatology, Palaeoecology, 37:149–164.

IVANY, L. C., T. BREY, M. HUBER, D. P. BUICK, AND B. R. SCHÖNE. 2011. El Niño in the Eocene greenhouse recorded by fossil bivalves and wood from Antarctica, Geophysical Research Letters, 38, L16709, doi:10.1029/2011GL048635.

IVANY, L. C., K. C. LOHMANN, D. B. BLAKE, F. HASIUK, R. B. ARONSON, A. GLASS, AND R. MOODY. 2008. Eocene climate record of a high southern latitude continental shelf: Seymour Island, Antarctica, Geological Society of America Bulletin, 120:659–678.

JARAMILLO, C. A., D. OCHOA, L. CONTRERAS, M. PAGANI, H. CARVAJAL-ORTIZ, L. M. PRATT, S. KRISHNAN, A. CARDONA, M. ROMERO, L. QUIROZ, G. RODRIGUEZ, M. J. RUEDA, F. DE LA PARRA, S. MORÓN, W. GREEN, G. BAYONA, C. MONTES, O. QUINTERO, R. RAMIREZ, G. MORA, S. SCHOUTEN, H. BERMUDEZ, R. NAVARRETE, F. PARRA, M. ALVARÁN, J. OSORNO, J. L. CROWLEY, V. VALENCIA, AND J. VERVOORT. 2010. Effects of rapid global warming at the Paleocene-Eocene boundary on neotropical vegetation. Science, 330:957–961. doi:10.1126/science.1193833

JENKYNS, H .C. 2003. Evidence for rapid climate change in the Mesozoic–Palaeogene greenhouse world: Philosophical Transactions of the Royal Society, 361:1885–1916.

JENKYNS, H. C., A. FORSTER, S. SCHOUTEN, J. S. SINNINGE DAMSTÉ. 2004. Higher temperatures in the Late Cretaceous Arctic Ocean. Nature, 432:888–892.

JENKYNS, H., L. SCHOUTEN-HUIBERS, S. SCHOUTEN, AND J. SINNINGHE DAMSTÉ. 2012. Warm Middle Jurassic–Early Cretaceous high-latitude sea-surface temperatures from the Southern Ocean. Climate of the Past, 8:215–226.

KATZ, M. E., B. S. CRAMER, J. R. TOGGWEILER, G. ESMAY, C. LIU, K. G. MILLER, Y. ROSENTHAL, B. S. WADE, AND J. D. WRIGHT. 2011. Impact of Antarctic circumpolar current development on late Paleogene ocean structure. Science, 332:1076–1079.

KENNETT, J. P. 1977. Cenozoic evolution of Antarctic glaciations, the circum-Antarctic ocean and their impact on global paleoceanography. Journal of Geophysical Research, 82:3843–3860.

KENNETT, J. P., R. E. HOUTZ, P. B. ANDREW, A. R. EDWARDS, V. A. GOSTIN, M. HAJOS, M. HAMPTON, D. G. JENKINS, S. V. MARGOLIS, A. F. OVENSHINE, AND K. PERCH-NIELSEN. 1975. Cenozoic paleoceanography in the southwest Pacific Ocean, Antarctic glaciation and the development of the Circum-Antarctic current. Initial Reports of the Deep Sea Drilling Project, 29:1155–1169.

KENNETT, J. P., AND L. D. STOTT. 1991. Abrupt deep-sea warming, palaeoceanographic changes and benthic extinctions at the end of the Palaeocene. Nature, 353:225–229.

KIM, J.-H., S. SCHOUTEN, E. C. HOPMANS, B. DONNER, AND J. S. SINNINGHE DAMSTÉ. 2008. Global sediment core-top calibration of the TEX_{86} paleothermometer in the ocean. Geochimica et Cosmochimica Acta, 72:1154–1173.

KIM, J.-H., J. VAN DER MEER, S. SCHOUTEN, P. HELMKE, V. WILLMOTT, F. SANGIORGI, N. KOC, E. C. HOPMANS, AND J. S. SINNINGHE DAMSTÉ. 2010. New indices and calibrations derived from the distribution of crenarchaeal isopreoid tetraether lipids: Implications for past sea surface temperature reconstructions. Geochimica et Cosmochimica Acta, 74: 4639–4654.

KIRK-DAVIDOFF, D. B., AND J.-F. LAMARQUE. 2008. Maintenance of polar stratospheric clouds in a moist stratosphere. Climate of the Past, 4:69–78.

KIRK-DAVIDOFF, D. B., D. P. SCHRAG, AND J. G. ANDERSON. 2002. On the feedback of stratospheric clouds on polar climate. Geophysical Research Letters, 29.

KLEYPAS, J. A., G. DANABASOGLU, AND J. M. LOUGH. 2008. Potential role of the ocean thermostat in determining regional differences in coral reef bleaching events, Geophysical Research Letters, 35, L03613. doi:10.1029/2007GL032257.

KOCH, J. T., AND R. L. BRENNER. 2009. Evidence for glacioeustatic control of large, rapid sea-level fluctuations during the Albian-Cenomanian: Dakota formation, eastern margin of western interior seaway, USA. Cretaceous Research, 30:411-423.

KOCH, P. L., J. C. ZACHOS, AND P. D. GINGERICH. 1992. Correlation between isotope records in marine and continental carbon reservoirs near the Palaeocene/Eocene boundary. Nature, 358:319–322.

KOWALSKI, E. A., AND D. L. DILCHER. 2002. Warmer temperatures for terrestrial ecosystems. Proceedings of the National Academy of Sciences USA, 100:167–170.

KÜRSCHNER, W. M., Z. KVACEK, AND D. L. DILCHER. 2008. The impact of Miocene atmospheric carbon dioxide fluctuations on climate and the evolution of terrestrial ecosystems. Proceedings of the National Academy of Sciences USA, 105:449–453.

KURTZ, A. C., L. R. KUMP, M. A. ARTHUR, J. C. ZACHOS, AND A. PAYTAN. 2003. Early Cenozoic decoupling of the global carbon and sulfur cycles. Paleoceanography, 18.

LAMBECK, K. AND J. CHAPPELL. 2001. Sea level change through the last glacial cycle. Science, 292:679–686.

LEAR, C. H., T. R. BAILEY, P. N. PEARSON, H. K. COXALL, AND Y. ROSENTHAL. 2008. Cooling and ice growth across the Eocene–Oligocene transition. Geology, 36:251–254.

LEAR, C. H., H. ELDERFIELD, AND P. A. WILSON. 2000. Cenozoic deep-sea temperature sand global ice volumes from Mg/Ca in benthic foraminiferal calcite. Science, 287:269–272.

LECKIE, R. M., T. J. BRALOWER, AND R. CASHMAN. 2002. Oceanic anoxic events and plankton evolution: Biotic response to tectonic forcing during the mid-Cretaceous. Paleoceanography, 17. doi:10.1029/2001PA000623,.

LENZ, O. K., V. WILDE, AND W. RIEGEL. 2010. A 600 k.y. record of El Niño-Southern Oscillation (ENSO): evidence for persisting teleconnections during the middle Eocene greenhouse climate of Central Europe. Geology, 38:627–630.

LENZ, O. K., V. WILDE, AND W. RIEGEL. 2011. Short-term fluctuations in vegetation and phytoplankton during the middle Eocene greenhouse climate: a 640-kyr record from the Messel oil shale (Germany). International Journal of Earth Sciences, 100:1851–1879.

LITTLER, K., S. A. ROBINSON, P. R. BOWN, A. J. NEDERBRAGT, AND R. D. PANCOST. 2011. High sea-surface temperatures during the Early Cretaceous Epoch. Nature Geoscience, 4:169-172.

LIU, Z., M. PAGANI, D. ZINNIKER, R. DECONTO, M. HUBER, H. BRINKHUIS, S. SHAH, R. LECKIE, AND A. PEARSON. 2009. Global cooling during the Eocene–Oligocene climate transition, Science, 323:1187–1190.

LOURENS, L. J., A. SLUIJS, D. KROON, J. C. ZACHOS, E. THOMAS, U. RÖHL, J. BOWLES, AND I. RAFFI. 2005. Astronomical pacing of late Palaeocene to early Eocene hyperthermal events. Nature, 435:1083–1087.

LOWENSTEIN, T. K., AND R. V. DEMICCO. 2006. Elevated Eocene atmospheric CO_2 and its subsequent decline. Science, 313:1928.

LUNT, D. J., T. DUNKLEY-JONES, M. HEINEMANN, M. HUBER, A. LEGRANDE, A. WINGUTH, C. LOPTSON, J. MAROTZKE, J. TINDALL, P. VALDES, AND C. WINGUTH. 2012. A model-data comparison for a multi-model ensemble of early Eocene atmosphere-ocean simulations: EoMIP. Climate of the Past Discussions, 8:1229–1273.

LUNT, D. J., P. J. VALDES, T. DUNKLEY-JONES, A. RIDGWELL, A. M. HAYWOOD, D. N. SCHMIDT, R. MARSH, AND M. MASLIN. 2010. CO_2-driven ocean circulation changes as an amplifier of Paleocene–Eocene thermal maximum hydrate destabilization. Geology, 38:875–878.

LYLE, M. 1997. Could early Cenozoic thermohaline circulation have warmed the poles? Paleoceanography, 12:161–167.

LYLE, M., J. BARRON, T. J. BRALOWER, M. HUBER, A. OLIVAREZ-LYLE, A. C. RAVELO, D. K. REA, AND P. A. WILSON. 2008. The Pacific Ocean and the Cenozoic evolution of climate. Review of Geophysics. doi:10.1029/2005RG000190.

LYLE, M., A. OLIVAREZ-LYLE, J. BACKMAN, AND A. TRIPATI. 2005. Biogenic sedimentation in the Eocene equatorial Pacific—the stuttering greenhouse and Eocene carbonate compensation depth, p. 1–35. *In* P. A. WILSON, M. LYLE, AND J. V. FIRTH (eds.), Proceedings of the Ocean Drilling Program, Scientific Results.

MACHLS, M. L., P. E. OLSEN, N. CHRISTIE-BLICK, AND S. R. HEMMING. 2008. Spectral analysis of the lower Eocene Wilkins Peak Member, Green River Formation, Wyoming: Support for Milankovitch cyclicity, Earth and Planetary Science Letters, 268:64–75.

MACLEOD, K. G., B. T. HUBER, AND C. ISAZA-LONDOÑO. 2005. North Atlantic warming during "global" cooling at the end of the Cretaceous. Geology, 33:437–440.

MACLEOD, K. G., B. T. HUBER, T. PLETSCH, U. RÖHL, AND M. KUCERA. 2001. Maastrichtian foraminiferal and paleoceanographic changes on Milankovitch time scales. Paleoceanography, 16:133–154.

MACLEOD, K. G., C. ISAZA-LONDOÑO, E. E. MARTIN, A. JIMENEZ BERROCOSO, AND C. BASAK. 2011. Changes in North Atlantic circulation at the end of the Cretaceous greenhouse interval. Nature Geoscienceence, 779–782.

MACLEOD, K. G., C. I. LONDOÑO, E. E. MARTIN, A. JIMENEZ BERROCOSO, AND C. BASAK. 2011. Changes in North Atlantic circulation at the end of the Cretaceous greenhouse interval. Nature Geosciences, 4:779–782. MARKWICK, P. J. 1994. "Equability," continentality, and Tertiary "climate": The crocodilian perspective. Geology, 22: 613–616.

MARKWICK, P. J. 1998. Fossil crocodilians as indicators of Late Cretaceous and Cenozoic climates: Implications for using palaeontological data in reconstructing palaeoclimate, Palaeogeography Palaeoclimatology Palaeoecology, 137: 205–271.

MARKWICK, P. J. 2007. The palaeogeographic and palaeoclimatic significance of climate proxies for data-model comparisons, p. 251–312. *In* M. Williams, A. M. Haywood, F. J. Gregory, and D. N. Schmidt (eds.). Deep-Time Perspectives on Climate Change: Marrying the Signal from Computer Models and Biological Proxies. The Micropalaeontological Society, Special Publications. The Geological Society of London.

MCELWAIN, J. C., J.WADE-MURPHY, AND S. P. HESSELBO. 2005. Changes in carbon dioxide during an oceanic anoxic event linked to intrusion into Gondwana coals. Nature, 435:479:482.

MCGOWRAN, B., Q. LI, J. CANN, D. PADLEY, D. M. MCKIRDY, AND S. SHAFIK. 1997. Biogeographic impact of the Leeuwin Current in southern Australia since the late middle Eocene. Palaeogeography Palaeoclimatology Palaeoecology, 136:19–40.

MIALL, A. D. 1992. Exxon global cycle chart: An event for every occasion? Geology, 20: 787–790.

MIKOLAJEWICZ, U., E. MAIERREIMER, T. J. CROWLEY, AND K. Y. KIM. 1993. Effect of Drake and Panamanian gateways on the circulation of an ocean model. Paleoceanography, 8:409–426.

MIKOLAJEWICZ, U., AND T. J. CROWLEY. 1997. Response of a coupled ocean/energy balance model to restricted flow through the Central American isthmus. Paleoceanography, 12:429–441.

MILLER, K. G., T. R. JANACEK, M. E. KATZ, AND D. J. KEIL. 1987. Abyssal circulation and benthic foraminiferal changes near the Paleocene/Eocene Boundary. Paleoceanography, 2:741–761.

MILLER, K. G., M. A. KOMINZ, J. V. BROWNING, J. D.

WRIGHT, G. S. MOUNTAIN, M. E. KATZ, P. J. SUGARMAN, B. S. CRAMER, N. CHRISTIE-BLICK, AND S. F. PEKAR. 2005a. The Phanerozoic record of sea-level change. Science, 310: 1293-1298.

MILLER, K. G., G. S. MOUNTAIN, J. D. WRIGHT, AND J. V. BROWNING. 2011. A 180-million-year record of sea level and ice volume variations from continental margin and deep-sea isotopic records. Oceanography, 24:40–53.

MILLER, K. G., J. D. WRIGHT, AND J. V. BROWNING. 2005. Visions of ice sheets in a greenhouse world, p. 215–231. *In* A. PAYTAN, AND C. DE LA ROCHA, (eds.). Ocean Chemistry Throughout the Phanerozoic. Marine Geology Special Issue 217.

MILLER, R. L. 1997. Tropical thermostats and low cloud cover. Journal of Climate, 409–440.

MITROVICA, J. X., N. GOMEZ, E. MORROW, C. HAY, K. LATYCHEV, AND M. E. TAMISIEA. 2011. On the robustness of predictions of sea level fingerprints. Geophysical Journal International, 187:729–742.

MOORE, T. C., JR., J. BACKMAN, I. RAFFI, C. NIGRINI, A. SANFILIPPO, H. PALIKE, AND M. LYLE. 2004. Paleogene tropical Pacific: Clues to circulation, productivity, and plate motion. Paleoceanography, 19.

MORIYA, K., P. A. WILSON, O. FRIEDRICH, J. ERBACHER, H. KAWAHATA. 2010. Testing for ice sheets during the mid-Cretaceous greenhouse using glassy foraminiferal calcite from the mid-Cenomanian tropics on Demerara Rise. Geology, 7:615–618.

NAJJAR, R. G., G. T. NONG, D. SEIDOV, AND W. H. PETERSON. 2002. Modeling geographic impacts on early Eocene ocean temperature. Geophysical Research Letters, 29, doi:10.1029/2001GL014–438.

NEWELL, R. E. 1979. Climate and the ocean. American Journal of Science, 67:405–416.

NICOLO, M. J., G. R. DICKENS, C. J. HOLLIS, AND J. ZACHOS. 2007. Multiple early Eocene hyperthermals: their sedimentary expression on the New Zealand continental margin and in the deep sea. Geology, 35:699–702.

NILSSON, J., G. BROSTRÖM AND G. WALIN. 2003. The thermohaline circulation and vertical mixing: does weaker density stratification give stronger overturning? Journal of Physical Oceanography, 33:2781–2795.

NISANCIOGLU, K. H., M. E. RAYMO, AND P. H. STONE. 2003. Reorganization of Miocene deep water circulation in response to shoaling of the Central American Seaway. Paleoceanography, 18. doi:10.1029/2002PA000767, 2003.

NOF, D. 2000. Does the wind control the import and export of the South Atlantic? Journal of Physical Oceanography, 30:2650–2667.

NOF, D., AND S. VAN GORDER. 2002. Did an open Panama Isthmus correspond to an invasion of Pacific water into the Atlantic? Journal of Physical Oceanography, 33:1324–1336.

NONG, G. T., R. G. NAJJAR, D. SEIDOV, AND W. H. PETERSON. 2000. Simulation of ocean temperature change due to the opening of the Drake Passage. Geophysical Research Letters, 27:2689–2692.

NORRIS, R. D., K. L. BICE, E. A. MAGNO, AND P. A. WILSON. 2002. Jiggling the tropical thermostat in the Cretaceous hothouse. Geology, 30:299–302.

NORRIS, R. D. AND P. A. WILSON. 1998. Low-latitude sea-surface temperatures for the mid-Cretaceous and the evolution of planktic foraminifera. Geology, 26:823-826.

OMTA, A. W. AND H. A. DIJKSTRA. 2003. A physical mechanism for the Atlantic-Pacific flow reversal in the early Miocene. Global and Planetary Change, 788:1–12.

OTTO-BLIESNER, B. L., AND G. R. UPCHURCH. 1997. Vegetation-induced warming of high-latitude regions during the late Cretaceous period. Nature, 385:804–807.

PAGANI, M., K. CALDEIRA, D. ARCHER, AND J. C. ZACHOS. 2006a. An ancient carbon mystery. Science, 314:1556–1557.

PAGANI, M., M. HUBER, LIU, Z., S. M. BOHATY, J. HENDERIKS, W. SIJP, S. KRISHNAN, AND R. M. DECONTO. 2011. The Role of Carbon Dioxide during the Onset of Antarctic Glaciation. Science, 334:1261–1264.

PAGANI, M., N. PEDENTCHOUK, M. HUBER, A. SLUIJS, S. SCHOUTEN, H. BRINKHUIS, J. S. SINNINGHE DAMSTÉ, G. R. DICKENS, AND THE IODP EXPEDITION 302 SCIENTISTS. 2006b. Arctic hydrology during global warming at the Palaeocene/Eocene thermal maximum. Nature, 442:671–675.

PAGANI, M., J. ZACHOS, K. H. FREEMAN, S. BOHATY, AND B. TIPPLE. 2005. Marked change in atmospheric carbon dioxide concentrations during the Oligocene. Science, 309:600–603.

PÄLIKE, H., J. FRAZIER, AND J. C. ZACHOS. 2006a. Extended orbitally forced palaeoclimatic records from the equatorial Atlantic Ceara Rise. Quaternary Science Reviews, 25:3138–3149.

PÄLIKE, H., R. D. NORRIS, J. O. HERRLE, P. A. WILSON, H. K. COXALL, C. H. LEAR, N. J. SHACKLETON, A. K. TRIPATI, AND B. S. WADE. 2006b. The heartbeat of the Oligocene climate system. Science, 314:1894–1898.

PANCHUK, K., A. RIDGWELL, AND L. R. KUMP. 2008. Sedimentary response to Paleocene–Eocene thermal maximum carbon release: a model-data comparison. Geology, 36:315–318.

PARRISH, J. T. 1998. Interpreting Pre-Quaternary Climate from the Geologic Record, Columbia University Press, 348pp.

PEARSON, P., P. W. DITCHFIELD, J. SINGANO, K. G. HARCOURT-BROWN, C. J. NICHOLAS, R. K. OLSSON, N. J. SHACKLETON, AND M. A. HALL. 2001. Warm tropical sea surface temperatures in the Late

Cretaceous and Eocene epochs. Nature, 413:481–487.

PEARSON, P. N., G. L. FOSTER, AND B. S. WADE. 2009. Atmospheric carbon dioxide through the Eocene–Oligocene climate transition. Nature, 461:1110–1113.

PEARSON, P. N., I. K. MCMILLAN, B. S. WADE, T. D. JONES, H. K. COXALL, P. R. BOWN, AND C. H. LEAR. 2008. Extinction and environmental change across the Eocene–Oligocene boundary in Tanzania. Geology, 36:179–182.

PEARSON, P. N., AND M. R. PALMER. 2000. Atmospheric carbon dioxide concentrations over the past 60 million years. Nature, 406:695–699.

PEARSON, P. N., B. E. VAN DONGEN, C. J. NICHOLAS, R. D. PANCOST, S. SCHOUTEN, J. M. SINGANO, AND B. S. WADE. 2007. Stable warm tropical climate through the Eocene Epoch. Geology, 35:211–214.

PETERS, R. B., AND L. C. SLOAN. 2000. High concentrations of greenhouse gases and polar stratospheric clouds: A possible solution to high-latitude faunal migration at the latest Paleocene thermal maximum. Geology, 28:979–982.

PIERREHUMBERT, R. T. 1995. Thermostats, radiator fins, and the local runaway greenhouse. Journal of Atmospheric Sciences, 1784–1806.

PIERREHUMBERT, R. T. 2002. The hydrologic cycle in deep time climate problems. Nature, 419:191–198.

POOLE, I., D. CANTRILL, AND T. UTESCHER. 2005. A multi-proxy approach to determine Antarctic terrestrial palaeoclimate during the Late Cretaceous and Early Tertiary. Palaeogeography Palaeoclimatology, Palaeoecology, 222:95–121.

POULSEN, C. J., E. J. BARRON, M. A. ARTHUR, AND W. H. PETERSON. 2001. Response of the mid-Cretaceous global oceanic circulation to tectonic and CO_2 forcings. Paleoceanography, 16.

POULSEN, C. J., E. J. BARRON, W. H. PETERSON, AND P. A. WILSON. 1999. A reinterpretation of mid-Cretaceous shallow marine temperatures through model-data comparison. Paleoceanography, 14:679–697.

POULSEN, C. J., D. POLLARD, AND T. S. WHITE. 2007. GCM simulation of the $\delta^{18}O$ content of continental precipitation in the middle Cretaceous: a model-proxy comparison. Geology, 35:199–202.

POULSEN, C. J., D. SEIDOV, E. J. BARRON, AND W. H. PETERSON. 1998. The impact of paleogeographic evolution on the surface oceanic simulation and the marine environment within the mid-Cretaceous Tethys. Paleoceanography, 13:546–559.

POWER, S. B., AND R. KLEEMAN. 1994. Surface flux parameterisation and the response of OGCMs to high latitude freshening. Tellus, 46:86–95.

PRICE, G. D. 1999. The evidence and implications of polar ice during the Mesozoic. Earth Science Reviews, 48:183–210.

PRICE, G. D., AND E. V. NUNN. 2010. Valanginian isotope variation in glendonites and belemnites from Arctic Svalbard: transient glacial temperatures during the Cretaceous greenhouse. Geology, 38:251–254.

PROTHERO, D. R., L. C. IVANY, AND E. A. NESBITT (EDS.). 2003. From Greenhouse to Icehouse: the Marine Eocene–Oligocene Transition. Columbia University Press, New York.

RAMANATHAN, V., AND W. COLLINS. 1991. Thermodynamic Regulation of Ocean Warming by Cirrus Clouds Deduced from Observations of the 1987 E1 Nino. Nature, 27–32.

RIND, D., AND M. CHANDLER. 1991. Increased ocean heat transports and warmer climate. Journal of Geophysical Research, 96:7437–7461.

RIPEPE, M., L. T. ROBERTS, AND A. G. FISCHER. 1991. ENSO and sunspot cycles in varved Eocene oil shales from image analysis. Journal of Sedimentary Petrology, 61:1155–1163.

ROBERTS, C. D., A. N. LEGRANDE, AND A. K. TRIPATI. 2009. Climate sensitivity to Arctic seaway restriction during the early Paleogene. Earth and Planetary Science Letters, 286:576–585.

ROHL, U., T. WESTERHOLD, T. J., BRALOWER, AND J. C. ZACHOS. 2007. On the duration of thePaleocene-Eocene thermal maximum (PETM). Geochemistry Geophysics Geosystems, 8:Q12002.

ROYER, D. L., R. A. BERNER, I. P. MONTANEZ, N. J.TABOR, AND D. J. BEERLING. 2004. CO_2 as a primary driver of Phanerozoic climate. GSA Today, 14:4–10.

SAGEMAN, B. B., S. R. MEYERS, AND M. A. ARTHUR. 2006. Orbital time scale and new C-isotope record for Cenomanian–Turonian stratotype. Geology, 34:1251128.

SCHER, H. D., AND E. E. MARTIN. 2006. Timing and climatic consequences of the opening of Drake Passage. Science, 312:428–430.

SCHMIDT, G. A. 1999. Forward modeling of carbonate proxy data from planktonic foraminifera using oxygen isotope tracers in a global ocean model. Paleoceanography, 14:482–497.

SCHOUTEN S., A. FORSTER, F. E. PANOTO, AND J. S. SINNINGHE DAMSTÉ. 2007. Towards the calibration of the TEX_{86} palaeothermometer for tropical sea surface temperatures in ancient greenhouse worlds. Organic Geochemistry 38:1537–1546.

SCHOUTEN S., E. HOPMANS, A. FORSTER, Y. VAN BREUGEL, M. M. M. KUYPERS, AND J. S. SINNINGHE DAMSTÉ. 2003. Extremely high sea-surface temperatures at low latitudes during the middle Cretaceous as revealed by archael membrane lipids. Geology, 31:1069–1072.

SCHRAG, D. P. 1999. Effects of diagenesis on the isotopic record of late Paleogene tropical sea surface temperatures, Chemical Geology, 161:215–224.

SCHUUR, E. A., G., J. BOCKHEIM, J. G. CANADELL, E. EUSKIRCHEN, C. B. FIELD, S. V. GORYACHKIN, S.

HAGEMANN, P. KUHRY, P. M. LAFLEUR, H. LEE, G. MAZHITOVA, F. E. NELSON, A. RINKE, V. E. ROMANOVSKY, N. SHIKLOMANOV, C. TARNOCAI, S. VENEVSKY, J. G. VOGEL, AND S. A. ZIMOV. 2008. Vulnerability of permafrost carbon to climate change: implications for the global carbon cycle. Bioscience, 58:701–714.

SCHUUR, E. A. G., J. G. VOGEL, K. G. CRUMMER, H. LEE, J. O. SICKMAN, AND T. E. OSTERKANP. 2009. The effect of permafrost thaw on old carbon release and net carbon exchange from tundra. Nature, 459:556–559.

SEAGER, R., Y. KUSHNIR, AND M. A. CANE. 1995. On heat flux boundary conditions for ocean models. Journal of Physical Oceanography, 25:3219–3230.

SEAGER, R., M. F. TING, I. M. HELD, Y. KUSHNIR, J. LU, G. VECCHI, H. P. HUANG, N. HARNIK, A. LEETMAA, N.-C. LAU, C. LI, J. VELEZ, AND N. NAIK. 2007. Model projections of an imminent transition to a more arid climate in southwestern North America. Science, 316:1181–1184.

SEAGER, R., AND G. A. VECCHI. 2010: Greenhouse warming and the 21st Century hydroclimate of southwestern North America. Proceedings of the National Academy of Sciences USA. 107:21277–21282.

SELLWOOD, B.W., AND P. J. VALDES. 2006. Mesozoic climates: General circulation models and the rock record. Sedimentary Geology, 190:269–287.

SEWALL, J. O., AND L. C. SLOAN. 2001. Equable Paleogene climates: the result of a stable, positive Arctic Oscillation? Geophysical Research Letters, 28:3693–3695.

SEWALL, J. O., AND L. C. SLOAN. 2004. Less ice, less tilt, less chill: the influence of a seasonally ice-free Arctic Ocean and changing obliquity on early Paleogene climate. Geology, 32:477–480.

SEWALL, J. O., AND L. C. SLOAN. 2006. Come a little bit closer: a high-resolution climate study of the early Paleogene Laramide foreland. Geology, 34:81–84.

SEWALL, J. O., L .C. SLOAN, M. HUBER, AND S. WING. 2000. Climate sensitivity to changes in land surface characteristics: Global and Planetary Change, 26:445–465.

SEXTON, P. F., P. A. WILSON, AND P. N. PEARSON. 2006. Microstructural and geochemical perspectives on planktic foraminiferal preservation: "glassy" versus "frosty." Geochemistry Geophysics Geosystems 7. doi:10.1029/2006GC001291.

SHACKLETON, N. J., AND A. BOERSMA. 1981. The climate of the Eocene ocean. Journal of the Geological Society of London, 138:153–157.

SHELLITO, C. J., J.-F. LAMARQUE, AND L. C. SLOAN. 2009. Early Eocene Arctic climate sensitivity to pCO_2 and basin geography. Geophysical Research Letters, 36:L09707.

SHELLITO, C., AND L. C. SLOAN. 2006a. Reconstructing a lost Eocene Paradise, Part I. Simulating the change in global floral distribution at the initial Eocene thermal maximum. Global and Planetary Change, 50:1–17.

SHELLITO, C., AND L. C. SLOAN. 2006b. Reconstructing a lost Eocene Paradise, Part II: On the utility of dynamic global vegetation models in pre-Quaternary climate studies. Global and Planetary Change, 50:18–32.

SHELLITO, C. J., L. C. SLOAN, AND M. HUBER. 2003. Climate model sensitivity to atmospheric CO_2 levels in the Early–Middle Paleogene. Palaeogeography Palaeoclimatology Palaeoecology 193:113–123.

SICKEL, W. A. V., M. A. KOMINZ, K. G. MILLER, AND J. V. BROWNING. 2004. Late Cretaceous and Cenozoic sea-level estimates: backstripping analysis of borehole data, onshore New Jersey. Basin Research, 16:451–465.

SINNINGHE DAMSTÉ, J. S., E. C. VAN BENTUM, G. J. REICHART, J. PROSS, AND S. SCHOUTEN. 2010. A CO_2 decrease-driven cooling and increased latitudinal temperature gradient during the mid-Cretaceous Oceanic Anoxic Event 2. Earth and Planetary Science Letters, 293:97–103.

SLOAN, L. C. 1994. Equable climates during the early Eocene: Significance of regional paleogeography for North American climate. Geology, 22:881–884.

SLOAN, L. C., AND E. J. BARRON. 1992. A comparison of Eocene climate model results to quantified paleoclimatic interpretations: Palaeogeography, Palaeoclimatology, Palaeoecology, 93:183–202.

SLOAN, L. C., AND M. HUBER. 2001. Eocene oceanic responses to orbital forcing on precessional time scale. Paleoceanography, 16:101–111.

SLOAN, L. C., M. HUBER, T. J. CROWLEY, J. O. SEWALL, AND S. BAUM. 2001. Effect of sea surface temperature configuration on model simulations of "equable" climate in the Early Eocene. Palaeogeography Palaeoclimatology Palaeoecology, 167:321–335.

SLOAN, L. C., M. HUBER, AND A. EWING. 1999. Polar stratospheric cloud forcing in a greenhouse world, p. 273–293. *In*: F. ABRANTES AND A. MIX (EDS.). Reconstructing Ocean History: A Window into the Future. Kluwer Academic/Plenum Publishers.

SLOAN, L. C., AND D. POLLARD. 1998. Polar stratospheric clouds: A high latitude warming mechanism in an ancient greenhouse world. Geophysical Research Letters, 25:3517–3520.

SLOAN, L .C., AND D. K. REA. 1996. Atmospheric CO_2 of the early Eocene: A general circulation modeling sensitivity study, Palaeogeography, Palaeoclimatology, Palaeoecology, 119:275–292.

SLOAN, L. C., J. C. G. WALKER, AND T. C. MOORE. 1995. The role of oceanic heat transport in early Eocene climate. Paleoceanography, 10:347–356.

SLOAN, L. C., J. C. G. WALKER, T. C. MOORE, D. K. REA, AND J. C. ZACHOS. 1992. Possible methane induced polar warming in the early Eocene. Nature, 357:320–322.

SLUIJS, A., H. BRINKHUIS, S. SCHOUTEN, J. C. ZACHOS, C. M. JOHN, S. BOHATY, J. S. SINNINGHE DAMSTÉ, E. M. CROUCH, AND G. R. DICKENS. 2007. Environmental precursors to rapid carbon injection at the Paleocene Eocene boundary. Nature, 450:1218–1221.

SLUIJS, A., S. SCHOUTEN, T. H. DONDERS, P. L. SCHOON, U. RÖHL, G.-J. REICHART, F. SANGIORGI, J.-H. KIM, J. S. SINNINGHE DAMSTÉ AND H. BRINKHUIS. 2009. Warm and wet conditions in the Arctic region during Eocene Thermal Maximum 2. Nature Geoscience, 2:777–780.

SLUIJS, A., S. SCHOUTEN, M. PAGANI, M. WOLTERING, H. BRINKHUIS, J. S. SINNINGHE DAMSTÉ, G. R. DICKENS, M. HUBER, G.-J. REICHART, R. STEIN, J. MATTHIESSEN, L. J. LOUREN, N. PEDENTCHOUK, J. BACKMAN, K. MORAN, AND THE EXPEDITION 302 SCIENTISTS. 2006. Subtropical Arctic Ocean conditions during the Palaeocene Eocene thermal maximum. Nature, 441:610–613.

SPICER, R. A., A. AHLBERG, A. B. HERMAN, C. C. HOFMANN, M. RAIKEVICH, P. J. VALDES, AND P. J. MARKWICK. 2008. The Late Cretaceous continental interior of Siberia: A challenge for climate models. Earth and Planetary Science Letters, 267:228–235.

SPICER, R. A., A. AHLBERG, A. B. HERMAN, S. P. KELLEY, M. I. RAIKEVICH, AND P. M. REES. 2002. Palaeoenvironment and ecology of the middle Cretaceous Grebenka flora of northeastern Asia, Palaeogeography Palaeoclimatology Palaeoecology, 184:65–105.

SPICER, R. A., AND J. T. PARRISH. 1986. Paleobotanical evidence for cool north polar climates in middle Cretaceous (Albian–Cenomanian) time. Geology, 14:703–706.

SPICER, R. A., AND J. T. PARRISH. 1990. Late Cretaceous–early Tertiary palaeoclimates of northern high latitudes: a quantitative view, Journal of the Geological Society of London, 147:329–341.

SPIELHAGEN, R. F., AND A. TRIPATI. 2009. Evidence from Svalbard for near-freezing temperaturesand climate oscillations in the Arctic during the Paleocene and Eocene. Palaeogeography Palaeoclimatology Palaeoecology, 278: 48–56.

STICKLEY, C. E., H. BRINKHUIS, S. A. SCHELLENBERG, A. SLUIJS, U. RÖHL, M. FULLER, M. GRAUERT, M. HUBER, J. WARNAAR, AND G. L. WILLIAMS. 2004. Timing and nature of the deepening of the Tasmanian Gateway, Paleoceanography, PA4027. doi:10.1029/2004PA001022.

STICKLEY, C. E. , N. KOÇ, , R. B. PEARCE, A. E. S. KEMP, R. W. JORDAN , F. SANGIORGI, AND K. ST. JOHN. 2012. Variability in the length of the sea ice season in the Middle Eocene Arctic. Geology, 40:727–730. doi:10.1130/G32976.1.

STICKLEY, C. E., K. ST. JOHN, N. KOÇ, R. W. JORDAN, S. PASSCHIER, R. B. PEARCE, AND L. E. KEARNS. 2009. Evidence for middle Eocene Arctic sea ice from diatoms and ice-rafted debris. Nature, 460:376–379.

STOREY, M., R. A. DUNCAN, AND C. C. SWISHER. 2007. Paleocene–Eocene Thermal Maximum and the opening of the Northeast Atlantic. Science, 316:587–589.

SUD, Y. C., G. K. WALKER, Y. P. ZHOU, G. A. SCHMIDT, K.-M. LAU, AND R. F. CAHALAN. 2008. Effects of doubled CO_2 on tropical sea surface temperatures (SSTs) for onset of deep convection and maximum SST: Simulations based inferences, Geophysical Research Letters, 35:L12707. doi:10.1029/2008GL033872.

TARDUNO, J. A., D. B. BRINKMAN, P. R. RENNE, R. D. COTTRELL, H. SCHER, AND P. CASTILLO. 1998. Evidence for extreme climatic warmth from Late Cretaceous Arctic vertebrates. Science, 282:2241–2244.

THOMAS, D. J. 2004. Evidence for production of North Pacific deep waters during the early Cenozoic Greenhouse. Nature, 430:65–68.

THOMAS, E., H. BRINKHUIS, M. HUBER, AND U. ROEHL. 2006. An ocean view of the early Cenozoic Greenhouse World. Oceanography, 19:63–72.

THOMPSON, E. I., AND B. SCHMITZ. 1997. Barium and the late Paleocene delta C-13 maximum: Evidence of increased marine surface productivity. Paleoceanography, 12:239–254.

THRASHER, B. L., AND L. C. SLOAN. 2009. Carbon dioxide and the early Eocene climate of western North America. Geology, 37:807–810.

TINDALL, J., R. FLECKER, P. J. VALDES, D. N. SCHMIDT, P. MARKWICK, AND J. HARRIS. 2010. Modelling oxygen isotope distribution of ancient seawater using a coupled ocean-atmosphere GCM: implications for reconstructing early Eocene climate. Earth and Planetary Sciences Letters, 292:265–273.

TOGGWEILER, J. R., AND D. W. LEA. 2010. Temperature differences between the hemispheres and ice age climate variability. Paleoceanography, 25:PA2212. doi:10.1029/2009PA001758.

TOGGWEILER, J. R., AND H. BJORNSSON. 2000. Drake Passage and paleoclimate, Journal of Quaternary Sciences, 15:319–238.

TOGGWEILER, J. R., AND B. SAMUELS. 1995. Effect of Drake Passage on the global thermohaline circulation. Deep-Sea Research, 42:477–500.

TRENBERTH, K. E., A. DAI, R. M. RASMUSSEN, AND D. B. PARSONS. 2003. The changing character of precipitation. Bulletin of the American Meteorological Society, 84:1205–1217.

TRIPATI, A., M. L. DELANEY, J. ZACHOS, L. D. ANDER-

SON, D. C. KELLY, AND H. ELDERFIELD. 2003. Tropical sea surface temperature reconstruction for the early Paleogene using Mg/Ca ratios of planktonic foraminifera. Paleoceanography, 18. doi:10.1029/2003PA000937.

TRIPATI, A. K., AND H. ELDERFIELD. 2004. Abrupt hydrographic changes in the equatorial Pacific and subtropical Atlantic from foraminiferal Mg/ Ca indicate greenhouse origin for the thermal maximum at the Paleocene–Eocene Boundary. Geochemistry Geophysics Geosystems 5:Q02006.

TRIPATI, A., J. ZACHOS, L. MARINCOVICH JR., AND K. BICE. 2001. Late Paleocene Arctic coastal climate inferred from molluscan stable and radiogenic isotope ratios, Palaeogeography Palaeoclimatology Palaeoecology, 170:101–113.

UPCHURCH, G. R., JR., B. L. OTTO-BLIESNER, AND C. R. SCOTESE. 1999. Terrestrial vegetation and its effects on climate during the latest Cretaceous. *In* E. BARRERA, AND C. C. JOHNSON (eds.). Evolution of the Cretaceous Ocean-Climate System: Boulder, Colorado, Geological Society of America Special Paper 332.

VAIL, P. R., R. M. MITCHUM, JR., AND S. THOMPSON III. 1977. Global cycles of relative changes of sea level. In: C. E. PAYTON (ed.). Seismic Stratigraphy and Global Changes of Sea Level. American Association of Petroleum Geologists Memoir 26, 83–97.

VALDES, P. J. 2000a. Warm climate forcing mechanisms, p. 3–20. *In* B. T. HUBER, K. G. MACLEOD, AND S. L. WING (eds.). Warm Climates in Earth History. Cambridge University Press.

VALDES, P. J. 2000b. Paleoclimate modeling. *In* P. MOTE AND A. O'NEILL (eds.). Numerical Modeling of the Global Atmosphere in the Climate System, Kluwer Academic Publishers.

VALLIS, G. K. 2000. Large-scale circulation and production of stratification: Effects of wind, geometry, and diffusion, Journal of Physical Oceanography, 30:933–954.

VAN HOOIDONK, R., AND M. HUBER. 2009. Equivocal evidence for a thermostat and unusually low levels of coral bleaching in the Western Pacific Warm Pool. Geophysical Research Letters, 36:L06705. doi:10.1029/2008GL036288.

VON DER HEYDT, A., AND H. A. DIJKSTRA. 2006. Effect of ocean gateways on the global ocean circulation in the late Oligocene and early Miocene. Paleoceanography, 21:1–18.

VON DER HEYDT, A., AND H. A. DIJKSTRA. 2008. The effect of ocean gateways on ocean circulation patterns in the Cenozoic. Global and Planetary Change, 62:132–146.

WADE, B. S., A. J. P. HOUBEN, W. QUAIJTAAL, S. SCHOUTEN, Y. ROSENTHAL, K. G. MILLER, M. E. KATZ, J. D. WRIGHT, AND H. BRINKHUIS. 2012. Multiproxy record of abrupt sea surface temperature cooling across the Eocene–Oligocene transition in the Gulf of Mexico. Geology, 40:159–162.

WADE, B. S., AND D. KROON. 2002. Middle Eocene regional climate instability: Evidence from the western North Pacific. Geology, 30:1011–1014.

WADE, B. S., D. KROON, AND R. D. NORRIS. 2001. Orbitally forced climate change in late mid- Eocene time at Blake Nose (Leg 171B): Evidence from stable isotopes in foraminifera. *In* D. KROON, R. D. NORRIS, AND A. KLAUS (eds.). Western North Atlantic Palaeogene and Cretaceous palaeoceanography: Geological Society of London Special Publication, 183: 273–291.

WALLACE, J. M. 1992. Effect of deep convection on the regulation of tropical sea-surface temperature. Nature, 357:230–231.

WARNAAR, J., P. K. BIJL, M. HUBER, L. SLOAN, H. BRINKHUIS, U. RÖHL, R. SRIVER, AND H. VISSCHER. 2009. Orbitally forced climate changes in the Tasman sector during the Middle Eocene. Palaeogeography Palaeoclimate Palaeoecology, 280:361–370.

WATANABE, T., A. SUZUKI, S. MINOBE, T. KAWASHIMA, K. KAMEO, K. MINOSHIMA, Y. M. AGUILAR, R. WANI, H. KAWAHATA, T. NAGAI, AND T. KASE. 2011. Permanent El Niño during the Pliocene warm period not supported by coral evidence. Nature, 471:209–211.

WATKINS, N. D., AND J. P. KENNETT. 1972. Regional sedimentary disconformities and upper Cenozoic changes in bottom water velocities between Australia and Antarctica. Antarctic Research, 19:317–334.

WEAVER, A. J., M. EBY, E. C. WIEBE, C. M. BITZ, P. B. DUFFY, T. L. EWEN, A. F. FANNING, M. M. HOLLAND, A. MACFADYEN, H. D. MATTHEWS, K. J. MEISSNER, O. SAENKO, A. SCHMITTNER, H. WANG, AND M. YOSHIMORI. 2001. The UVic Earth System Climate Model: Model description, climatology and application to past, present and future climates. Atmosphere-Ocean, 39:361–428.

WEAVER, A. J, AND E. S. SARACHIK. 1991. The role of mixed boundary condition in numerical models of ocean's climate. Journal of Physical Oceanography, 21:1470–1493.

WEIJERS, J. W. H., S. SCHOUTEN, A. SLUIJS, H. BRINKHUIS, AND J. S. SINNINGHE DAMSTÉ. 2007. Warm Arctic continents during the Palaeocene–Eocene thermal maximum. Earth and Planetary Sciences Letters, 261:230–238.

WESTERHOLD, T., U. ROHL, J. LASKAR, I. RAFFI, J. BOWLES, L. J. LOURENS, AND J. C. ZACHOS. 2007. On the duration of magnetochrons C24r and C25n and the timing of early Eocene global warming events: Implications from the Ocean Drilling Program Leg 208 Walvis Ridge depth transect. Paleoceanography 22: PA2201. doi:10.1029/2006PA001322.

WHITE, T. S., L. A. GONZALEZ, G. A. LUDVIGSON, AND C. POULSEN. 2001. Middle Cretaceous greenhouse hydrologic cycle of North America. Geology, 29:363–366.

WILF, P. 2000. Late Paleocene–early Eocene climate changes in southwestern Wyoming: Paleobotanical analysis: Geological Society of America Bulletin, 112:292–307.

WILF, P., K. C. BEARD, K. S. DAVIES-VOLLUM, AND J. W. NOREJKO. 1998. Portrait of a late Paleocene (early Clarkforkian) terrestrial ecosystem: Big Multi Quarry and associated strata, Washakie Basin, southwestern Wyoming, Palaios, 13:514–532.

WILF, P., K. R. JOHNSON, AND B. T. HUBER. 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 USA, 100: 599–604.

WILLIAMS, I. N., R. T. PIERREHUMBERT, AND M. HUBER. 2009. Global warming, convective threshold and false thermostats. Geophysical Research Letters, 36: L21805. doi:10.1029/2009GL039849

WILSON, P. A., AND R. D. NORRIS. 2001. Warm tropical ocean surface and global anoxia during the mid-Cretaceous period. Nature, 412:425-429.

WILSON, P. A., R. D. NORRIS, AND M. J. COOPER. 2002. Testing the Cretaceous greenhouse hypothesis using glassy foraminiferal calcite from the core of the Turonian tropics on Demerara Rise. Geology, 30:607–610.

WILSON, P. A., AND B. N. OPDYKE. 1996. Equatorial sea-surface temperatures for the Maastrichtian revealed through remarkable preservation of metastable carbonate. Geology, 24:555-558.

WING, S. L., H. BAO, AND K. L. KOCH. 2000. An early Eocene cool period? Evidence for continental cooling during the warmest part of the Cenozoic, p. 197–237. *In* B. T. HUBER, K. G. MACLEOD, AND S. L. WING (eds.). Warm Climates in Earth History. Cambridge University Press.

WING, S. L., AND D. R. GREENWOOD. 1993. Fossils and fossil climate: the case for equable continental interiors in the Eocene. Philosophical Transactions of the Royal Society of London B, 341:243–252.d.

WING, S. L., AND G. J. HARRINGTON. 2001. Floral response to rapid warming in the earliest Eocene and implications for concurrent faunal change. Paleobiology, 27:539–563.

WING, S. L., G. J. HARRINGTON, F. A. SMITH, J. I. BLOCH, D. M. BOYER, AND K. H. FREEMAN. 2005. Transient floral change and rapid global warming at the Paleocene–Eocene boundary. Science, 310:993–996.

WINGUTH, A. M. E., C. HEINZE, J. E. KUTZBACH, E. MAIER-REIMER, U. MIKOLAJEWICZ, D. B. ROWLEY, P. M. REES, AND A. M. ZIEGLER. 2002. Simulated warm polar currents during the Middle Permian. Paleoceanography, 17:1057.

WINGUTH, A., C. SHELLITO, C. SHIELDS, AND C. WINGUTH. 2010. Climate Response at the Paleocene–Eocene thermal maximum to greenhouse gas forcing—a model study with CCSM3. Journal of Climate, 23:2562–2584.

WINGUTH, A. M. E., E. THOMAS, AND C. WINGUTH. 2012. Global decline in ocean ventilation, oxygenation, and productivity during the Paleocene–Eocene Thermal Maximum: Implications for the benthic extinction. Geology, 40:263–266.

WOLFE, J. A. 1994. Tertiary climatic changes at middle latitudes of western North America. Palaeogeography Palaeoclimatology Palaeoecology 108:195–205.

WOLFE, J. A. 1995. Paleoclimatic estimates from Tertiary leaf assemblages. Annual Review of Earth and Planetary Sciences, 23:119–142.

ZACHOS, J.C., G. R. DICKENS, AND R. E. ZEEBE. 2008. An early Cenozoic perspective on greenhouse warming and carbon-cycle dynamics. Nature, 451:279–283.

ZACHOS, J., M. PAGANI, L. SLOAN, E. THOMAS, AND K. BILLUPS. 2001. Trends, rhythms, and aberrations in global climate 65 Ma to present. Science, 292:686–694.

ZACHOS, J. C., T. M. QUINN, AND K. A. SALAMY. 1996. High-resolution (104 years) deep-sea foraminiferal stable isotope records of the Eocene-Oligocene climate transition. Paleoceanography, 11:251–266.

ZACHOS, J. C., S. SCHOUTEN, S. BOHATY, T. QUATTLEBAUM, A. SLUIJS, H. BRINKHUIS, S. GIBBS, AND T. J. BRALOWER. 2006. Extreme warming of mid-latitude coastal ocean during the Paleocene–Eocene Thermal Maximum: Inferences from TEX_{86} and isotope data. Geology, 34:737–740.

ZACHOS, J. C., L. D. STOTT, AND K. C. LOHMANN. 1994. Evolution of early Cenozoic marine temperatures: Paleoceanography, 9:353–387.

ZACHOS, J. C., M. W. WARA, S. BOHATY, M. L. DELANEY, M. R. PETRIZZO, A. BRILL, T. J. BRALOWER, AND I. PREMOLI-SILVA. 2003. A transient rise in tropical sea surface temperature during the Paleocene–Eocene Thermal Maximum. Science, 302:1551–1554.

ZEEBE, R. E., J. ZACHOS, AND G. R. DICKENS. 2009. Carbon dioxide forcing alone insufficient to explain Palaeocene–Eocene Thermal Maximum warming. Nature Geoscience, 2:576–580.

ZHANG, S., R. J. GREATBACH, AND C. A. LIN. 1993. A reexamination of the polar halocline catastrophe and implications for coupled ocean-atmosphere modeling. Journal of Physical Oceanography, 23:287–299.

ZHANG, Y., J. M. WALLACE, AND D. S. BATTISTI. 1997. ENSO-like decade-to-century scale variability:1900–93. Journal of Climate, 10:1004–1020.

ZHANG, Z., K. H. NISANCIOGLU, F. FLATØY, M. BENTSEN, I. BETHKE, AND H. WANG. 2011. Tropical seaways played a more important role than high latitude seaways in Cenozoic cooling. Climate of the Past, 7:801–813.

ZHOU, J., C. J. POULSEN, N. ROSENBLOOM, C. SHIELDS, AND B. BRIEGLEB. 2012. Vegetation-climate interactions in the warm mid-Cretaceous, Climate of the Past, 8:565–576.

ZIEGLER, A. M., G. ESHEL, P. M. REES, T. A ROTHFUS, D. B. ROWLEY, AND D. SUNDERLIN. 2003. Tracing the tropics across land and sea: Permian to present. Lethaia, 36:227–254.