PALEOCLIMATE MODELING, PRE-QUATERNARY

Introduction

Pre-Quaternary paleoclimate modeling is the science of simulating climate change over the earliest 99.9% of Earth's history using numerical models of the climate system. The record of pre-Quaternary climate documents enormous climate swings with global ice-covered and ice-free end members. A number of climate factors shaped Quaternary and pre-Quaternary paleoclimate on timescales ranging from hundreds of thousands to several years. However, in contrast to Quaternary paleoclimate, tectonically-controlled processes (including continental drift, orogenies, and fluctuations in the long-term carbon cycle) and solar evolution were also important controls on pre-Quaternary climate. Understanding how these factors have influenced climate change through the Precambrian, Paleozoic, Mesozoic, and Cenozoic eras, and how they are expressed in the geological record is a principal objective of pre-Quaternary paleoclimate modeling.

The utility of pre-Quaternary paleoclimate modeling has been severalfold. Pre-Quaternary climate modeling has contributed to ocean, atmosphere, and geological sciences by: (a) identifying mechanisms of climate change; (b) quantifying the climate response to a variety of forcing factors; (c) recognizing climatic feedbacks that amplify/dampen climatic forcings; (d) identifying limitations of climate proxies from the geological record; and (e) demonstrating strengths/ shortcomings in numerical climate models. Pre-Quaternary paleoclimate modeling has made fundamental contributions to climate science. Among these, the recognition that the Earth's climate has experienced enormous variability on a multitude of temporal and spatial scales as a result of complex interactions between multiple climate factors is one of the most profound.

Methodology

Today, "paleoclimate models" are synonymous with numerical climate models implemented to study past climates. Prior to the wide availability of numerical climate models, paleoclimate models also referred to conceptual models constructed using physically-based rules about the general circulation of the atmosphere and ocean (e.g., Ziegler et al., 1977; Parrish et al., 1982). Numerical climate models are mathematical expressions of the theoretical laws that govern the climate system, approximated and written in computer code for fast, efficient computation. The model domain, frequently global in extent, is discretized into an array of horizontal and vertical grid cells. Not all models incorporate the full suite of governing equations in three dimensions. In fact, a hierarchy of numerical climate models has been used to study past climates. Numerical climate models can range in complexity from zero-dimensional energy balance models (EBMs) that simply express the conservation of energy, to three-dimensional general circulation models (GCMs) of the ocean and atmosphere that predict fluid flow on a rotating sphere heated by solar radiation. Many aspects of the climate system cannot be explicitly calculated, frequently because specific phenomena develop and act on sub-grid cell scale (i.e., over distances that are much smaller than represented by a model grid cell). Clouds provide an instructive example. Cloud motions vary over a horizontal scale of tens to hundreds of meters. Yet, the horizontal resolution of a GCM is typically tens to hundreds of kilometers. In order to represent clouds in GCMs, cloud processes are approximated or parameterized. Many of the differences between common types of numerical climate models (e.g., GCMs) arise from different parameterizations. To date, there are several different classes of GCMs, including atmosphere-only GCMs, ocean-only GCMs, coupled atmosphere-ocean GCMs, and GCMs coupled to models of the biosphere, cryosphere, and lithosphere. Climate models also differ in their horizontal and vertical resolution. As computational speed and efficiency has increased, so has the complexity and resolution of numerical paleoclimate models. For a more detailed description of numerical climate models, the reader is referred to one of several good texts on the subject (e.g., Washington and Parkinson, 1986; Trenberth, 1992; McGuffie and Henderson-Sellers, 1997).

Most models used in pre-Quaternary climate studies were initially developed and fine-tuned for the modern climate, and then modified for use in pre-Quaternary studies. The modifications for "paleo" use are generally limited to changes in the model's initial and boundary conditions. Table P1 lists the boundary conditions that have been explored in paleoclimate studies of the pre-Quaternary. A critical aspect of paleoclimate modeling is accurately defining the boundary conditions for a past age. For example, consider the difficulty of reconstructing the Cretaceous sea-surface temperatures (SSTs), continental vegetation, bathymetry, or atmospheric pCO_2. The task of reconstructing past boundary conditions is often made difficult or impossible because: (a) the geological record has been destroyed, incompletely sampled, or was never preserved; (b) the temporal resolution of the proxy is too coarse or too uncertain; (c) the spatial resolution (e.g., a single drill or field site) may be incompatible with the model resolution; or (d) the geological proxy may have lost its original signal through alteration or have an equivocal interpretation. In the absence of a detailed reconstruction of a particular boundary condition, a generalization or simplification has often been made (e.g., specification of globally uniform vegetation type or zonally-averaged SSTs). Numerous studies have demonstrated that differences within the uncertainty of a boundary condition can have regional or global climatic consequences (e.g., Kutzbach and Ziegler, 1993; Otto-Bliesner, 1998; Poulsen et al., 1998; Sewall et al., 2000; Huber and Sloan, 2000). In pre-Quaternary paleoclimate studies, initial conditions have usually been paid little attention for two reasons. First, the initial conditions are unknown. Second, very few pre-Quaternary proxy reconstructions have centennial or better resolution. Consequently, most paleoclimate modeling studies time-average the model results to mask interannual and interdecadal climate variability. Insofar as multiple climate states are possible, initial conditions may be important, and have been examined to a limited degree (e.g., Bice and Marotzke, 2001; Herrmann et al., 2003).

Table P1 Common paleoclimate boundary conditions

Continental distribution
Topography
Bathymetry
Land surface characteristics[a]
Vegetation properties
Soil properties
Land ice distribution and height[a]
Solar luminosity
Orbital parameters
Eccentricity
Obliquity
Precession
Atmospheric gases
Carbon dioxide
Methane
Nitrous oxide
CFCs
Ozone
Sea-surface temperatures[a]
Drainage basins or continental runoff

[a]In some studies, boundary condition may not be necessary if the parameter is explicitly calculated. For example, sea-surface temperatures are not necessary if a mixed-layer or fully dynamic ocean model is implemented.

Pre-Quaternary paleoclimate modeling studies have tended to focus on specific intervals or time slices of Earth's history (e.g., the mid-Cretaceous), rather than the continuum spanning all or a portion of Earth's 4.6 billion years. Practical limitations, including the computational costs incurred by long (i.e., greater than a few thousand years) simulations, the relative incompleteness of the paleoclimate proxy records, and the technical complexity of dynamically varying boundary conditions (specifically geography or orography), have motivated this time-slice approach. As these limitations diminish, long climate simulations may be possible, particularly for simpler or coarse-resolution models. The timescales of interest have varied tremendously, ranging from tens of millions of years to only a few years, and are mainly constrained by the resolution of paleoclimate proxy records.

Pre-Quaternary paleoclimate models have been utilized in several fashions. Due to the limitations of boundary condition reconstructions, many paleoclimate modeling studies fall within the category of sensitivity experiments – modeling experiments in which one parameter is varied at a time and compared to a control case. Alternatively, some studies have attempted to "simulate" a particular time interval by specifying the "best" boundary conditions and then comparing the simulation to proxy reconstructions. In practice, many studies mix these two approaches.

Paleoclimate modeling contributions

In this section, progress achieved on a number of pre-Quaternary paleoclimate problems through paleoclimate modeling has been summarized. As much as possible, the types of models used have been described in order to provide a sense of the evolution of the paleoclimate modeling field.

Faint Young Sun paradox

Stellar evolution models indicate that solar luminosity was 25–30% lower during the early Precambrian than today. Yet, the earliest one-dimensional EBMs and atmospheric GCMs predicted that much smaller decreases (2–5%) in solar luminosity would have triggered climate instability once sea ice had reached ~30° latitude, leading to a runaway ice-albedo feedback (Budyko, 1969; Sellers, 1969). Subsequent EBM experiments showed that the critical latitude for a runaway ice-albedo feedback was not entirely fixed; increasing the heat transport caused the instability to move to higher latitudes (Held and Suarez, 1974). The prediction of an early Precambrian ice-covered Earth is inconsistent with the presence of sedimentary rocks early in Earth history and evidence of primitive life at 3.5 Ga. Moreover, EBM calculations indicate that the formation of highly reflective CO_2 clouds would have made an ice-covered Earth irreversible (Caldiera and Kasting, 1992). This conflict between climate model predictions and the geological record has become known as the Faint Young Sun paradox.

Several climatic factors have been hypothesized to resolve the Faint Young Sun paradox, including the atmospheric concentration of greenhouse gases, the extent and configuration of continents, and the rotation rate and obliquity of the early Earth. Calculations using one-dimensional radiative-convective models indicate that high concentrations of NH_3 (Sagan and Mullen, 1972), CO_2 (Owen et al., 1979; Kuhn and Kasting, 1983;

Kasting et al., 1984; Kiehl and Dickinson, 1987) and/or CH_4 (Kiehl and Dickinson, 1987; Pavlov et al., 2000) in the atmosphere could have maintained surface temperatures above freezing by enhancing the Earth's greenhouse effect.

Theories of continental formation suggest that the extent of continental crust was considerably less than today. The small continental area may have affected the Precambrian climate by reducing the Earth's albedo (Henderson-Sellers and Henderson-Sellers, 1989; Kuhn et al., 1989; Gérard et al., 1992; Jenkins et al., 1993), increasing poleward heat transport (Endal and Schatten, 1982), and altering the global cloud fraction (Jenkins et al., 1993; Longdoz and François, 1997). These changes are hypothesized to compensate for the low solar luminosity at least partially, by impeding the equatorward growth of sea ice and increasing the solar radiation captured at the Earth's surface. However, the influence of continental configuration on Precambrian climate is highly model dependant. For example, in response to a global ocean, the cloud coverage decreased in the atmospheric GCM used by Jenkins et al. (1993) due to a shift from non-convective to convective cloud types, which have a smaller cloud fraction. In contrast, cloud coverage increased in the quasi-three-dimensional model used by Longdoz and François (1997) due to a more active hydrological cycle.

The Earth's rotation velocity has decreased through its history, causing day length to increase from 14-h days at 4.0 Ga to 24-h days (Walker and Zahnle, 1986). The dynamical consequences of an enhanced rotation rate during the Precambrian have been studied using EBMs and GCMs (Hunt, 1979; Kuhn et al., 1989; Jenkins et al., 1993). A faster rotation velocity reduces the scale size of synoptic disturbances, leading to a reduction in poleward heat transports, cold high latitude temperatures, and larger meridional temperature gradients (Hunt, 1979). Dynamical changes in eddy and mean motions may also reduce global cloud coverage, causing a 2 °C increase in global average air temperature that potentially compensated for the low Precambrian solar luminosity (Jenkins et al., 1993).

Snowball Earth hypothesis

The late Precambrian witnessed the most severe glaciations in Earth history. Paleomagnetic evidence from South Australia and Northwest Canada confirm the low-latitude ($<10°$) settings of late Precambrian glacial deposits. Paleoclimate models have been used extensively to evaluate the factors responsible for the low-latitude glaciation, and have examined the influences of paleogeography (Crowley and Baum, 1993; Chandler and Sohl, 2000; Poulsen et al., 2002), atmospheric CO_2 concentrations (Chandler and Sohl, 2000; Hyde et al., 2000; Baum and Crowley, 2001), continental surface characteristics (Baum and Crowley, 2001), ocean heat transport (Chandler and Sohl, 2000; Poulsen et al., 2001b), and ice-sheet dynamics (Hyde et al., 2000; Donnadieu et al., 2003). These modeling studies demonstrate that a combination of climate forcings can produce conditions for low-latitude glaciation (Chandler and Sohl, 2000; Hyde et al., 2000).

The low-latitude glacial deposits, and other sedimentary (deposition of banded iron formations and thick post-glacial carbonates) and geochemical evidence (carbon isotopes from carbonates), has been cited as confirmation that the entire Earth may have been completely ice-covered during the late Precambrian (Kirschvink, 1992; Hoffman et al., 1998). This idea, known as the Snowball Earth hypothesis, has gained support from some climate models. EBMs and atmospheric GCMs have simulated an ice-covered Earth under late Precambrian conditions (when the solar luminosity was about 6% less than present) (Jenkins and Smith, 1999; Baum and Crowley, 2001; Poulsen et al., 2001). However, this result is highly model dependent. Climate models that explicitly calculate ocean circulation and heat transport do not permit an ice-covered Earth under late Precambrian conditions. The energy released by convection at the sea-ice margin and the large heat capacity of the ocean work against the ice-albedo feedback (Poulsen et al., 2001; Bendtsen, 2002). The extent of late Precambrian ice is critical to the Earth's recovery – escape from an ice-covered state would have required ~300–1,000 × present atmospheric levels (PAL) of CO_2 (Caldiera and Kasting, 1992), while escape from a world with water at the equator would have required as little as 4 × PAL CO_2 (Crowley et al., 2001).

An alternative explanation for low-latitude glaciation is that the Earth's obliquity was substantially greater than at any time in the Phanerozoic (Williams, 1975, 1986). If the obliquity were higher than 54°, the poles would be the warmest region on Earth, and the equator would be the coldest. Atmospheric GCMs confirm that an extreme obliquity ($\geq 60°$) would produce freezing temperatures on low-latitude continents while maintaining above freezing conditions at high latitudes (Oglesby and Ogg, 1999; Jenkins, 2000; Donnadieu and Ramstein, 2002). The missing ingredient in this hypothesis is a reasonable physical mechanism for causing large changes in Earth's obliquity.

Ordovician glaciation

Geological evidence exists for a late Ordovician (~440 Ma) glaciation. This short-lived (~1 million year) glaciation (Brenchley et al., 1995, 2003) was remarkable because atmospheric CO_2 levels were high ($14 \pm 6 \times$ PAL) during the late Ordovician (Yapp and Poths, 1992). Numerical climate models of increasing complexity have been used to determine the conditions permitting glaciation at high CO_2 levels. Early studies using 2-D EBMs focused on the role of the late Ordovician paleogeography (Crowley et al., 1987; Crowley and Baum, 1991a), and specifically the orientation of Gondwanaland relative to the South Pole. With an edge of Gondwanaland near the South Pole, the thermal inertia of the ocean prevented continental summer temperatures from rising above freezing, thus allowing permanent snow cover (Crowley et al., 1987; Crowley and Baum, 1991a). Subsequent GCM experiments have confirmed the EBM result (Gibbs et al., 2000), but have also shown that the continental configuration of Gondwanaland is not a sufficient condition for glaciation. The influences of additional climatic factors on Ordovician glaciation have since been tested, including atmospheric CO_2, topography, ocean heat transport, orbital parameters, and snow/ice albedo (Crowley and Baum, 1995; Gibbs et al., 1997; Poussart et al., 1999; Herrmann et al., 2003). These studies generally conclude that glaciation is possible with high (8–14 × PAL) atmospheric CO_2 levels given favorable orbital parameters (i.e., a cold Southern Hemisphere summer configuration) and continental topography. With orbital forcing varying from cold-summer to warm-summer configurations, ice-sheet model calculations indicate that CO_2 levels must fall to 8 × PAL to grow a permanent ice sheet (Herrman et al., 2003).

Gondwanan glaciations

Ice sheets on Gondwana persisted for ~55 million year during the Permo-Carboniferous (275–330 Ma), and reached a size comparable to that of the Pleistocene ice sheets (Crowley and Baum, 1991b). The presence of ice sheets on a supercontinent is surprising since enhanced seasonality would have produced summer temperatures on the Gondwana ice sheet that may have been 15 °C greater than temperatures over the Laurentide Ice Sheet (Crowley, 1994). Two-dimensional EBM calculations indicate that a reduced solar luminosity (~3% less than modern) and favorable orbital parameters could compensate for the supercontinental effect, allowing freezing summer temperatures over the Gondwanan ice sheet (Crowley et al., 1991). Crowley and Baum (1992a) used a series of two-dimensional EBM experiments with a combination of evolving climatic factors (geography, geography + solar luminosity, geography + solar luminosity + CO_2) to simulate the estimated extent of the Gondwanan ice sheet. To simulate both the initiation and demise of the ice sheet, changes in geography, solar luminosity, and, most importantly, CO_2 were required (Crowley and Baum, 1992a).

EBM and GCM modeling of the Permo-Carboniferous glaciation has demonstrated that the onset and growth of the Gondwanan ice sheet may have been highly nonlinear due to a Small Ice Cap Instability (SICI) (Baum and Crowley, 1991; Crowley et al., 1994). A coupled climate-ice sheet model of the Gondwanan ice sheet also shows critical behavior with small changes in solar luminosity (0.0005%) leading to large differences (>10×) in the simulated ice volume (Hyde et al., 1999). The model also exhibits multiple equilibria. The melting of a large ice sheet due to an increase in CO_2 to 2 × PAL results in a small, stable Gondwanan ice sheet. Yet, no ice sheet is simulated given ice-free initial conditions and 2 × PAL CO_2 (Hyde et al., 1999).

Pangean climate

Paleogeography has been recognized as a first-order control on climate. During the late Paleozoic and early Mesozoic, the continents were agglomerated into the supercontinent Pangea. The supercontinental configuration had large consequences for Earth's climate. Typical features simulated by climate models include extreme continentality (i.e., a seasonal temperature range (>45 °C) that surpassed that of modern Eurasia) resulting from the small heat capacity of land (Crowley et al., 1989; Kutzbach and Gallimore, 1989; Kutzbach and Ziegler, 1993; Crowley and Baum, 1994; Wilson et al., 1994; Gibbs et al., 2002), strong monsoonal systems along the Tethyan coast (Kutzbach and Gallimore, 1989; Kutzbach and Ziegler, 1993; Wilson et al., 1994; Gibbs et al., 2002), and intense aridity in continental interiors due to the depletion of atmospheric moisture over the long continental trajectories (Kutzbach and Gallimore, 1989; Kutzbach and Ziegler, 1993; Wilson et al., 1994; Fawcett and Barron, 1998; Gibbs et al., 2002). Two-dimensional EBM calculations predict that Pangean surface temperatures would have been greatly influenced by orbital parameters with a maximum range of ~14–16 °C between maximum and minimum orbital insolation values (as determined from Pleistocene fluctuations) (Crowley and Baum, 1992b). In addition, atmospheric GCM experiments using an idealized Pangean paleogeography indicate that tropical and subtropical precipitation and runoff would have varied by 50% between the extreme phases of the precessional cycle with enhanced hydrology when perihelion (aphelion) occurred in summer (winter) (Kutzbach, 1994).

Mountains and plateaus may have played an important secondary role in controlling Pangean climate. The intensification of radiative heating and cooling over plateaus produced more extreme high and low pressure systems, which acted with the topography to guide the winds and focus precipitation (Kutzbach and Ziegler, 1993; Otto-Bliesner, 1993, 1998; Wilson et al., 1994; Hay and Wold, 1998). The effects of uplift were greatest in low latitude regions because of the influence of mountains on the position of the Inter-Tropical Convergence Zone (ITCZ) (Otto-Bliesner, 1998; Hay and Wold, 1998). Otto-Bliesner found that the presence of high (1,000–3,000 m) Central Pangean Mountains impeded the seasonal northward migration of the ITCZ, enhancing local precipitation by 68% in July and possibly explaining the extensive occurrence of tropical, late Carboniferous coals.

Considerable effort has been made to evaluate, through comparison with climate proxies, the ability of atmospheric GCMs to simulate elements of the Pangean climate (Kutzbach and Ziegler, 1993; Fawcett et al., 1994; Pollard and Schulz, 1994; Wilson et al., 1994; Rees et al., 1999, 2002; Gibbs et al., 2002). In general, atmospheric GCM simulations tend to do a "fair to good" job of simulating the requisite conditions for lithologic climatic indicators (Pollard and Schulz, 1994; Wilson et al., 1994; Gibbs et al., 2002) and sedimentary structures generated by severe weather (PSUCLIM, 1999a; PSUCLIM, 1999b). In comparison to climate inferred from Permian paleobotanical data, an atmospheric GCM predicts temperatures that are too cold in the high latitudes of the Southern Hemisphere (Rees et al., 1999, 2002). The absence of ocean dynamics (particularly the absence of warm polar currents, upwelling zones, and the explicit calculation of ocean heat transport) and the coarse model resolution (particularly the poor or nonexistent resolution of narrow mountain ranges, lakes, and coastlines) in the atmospheric GCMs have been implicated as potential reasons for the discrepancies between the model predictions and the proxy indicators (Pollard and Schulz, 1994; Rees et al., 1999, 2002; Gibbs et al., 2002).

Several atmospheric GCM studies have focused on the Jurassic (Chandler, 1994; Chandler et al., 1992; Moore et al., 1992; Valdes and Sellwood, 1992; Valdes, 1994), a geological interval that saw the rifting of Pangea. The large continental blocks shared many features with that of Pangea, including high continentality, intense continental aridity, and strong monsoonal systems (Moore et al., 1992). The latter conclude that the Jurassic warmth evidenced by climate proxies may be explained by elevated atmospheric CO_2. In contrast, Chandler et al. (1992) suggested that a Jurassic simulation with specified, warm SSTs was in energy balance without high atmospheric CO_2, implying that a warm Jurassic climate could have been the product of enhanced poleward heat transport through the ocean. However, in a Jurassic experiment with a specified, reduced meridional sea-surface gradient, the implied ocean heat transport was much smaller than in a present-day simulation, suggesting that enhanced ocean heat transport may not viable for the Jurassic (Valdes, 1994). A comparison of Jurassic climates simulated using different GCMs has confirmed that the treatment of the ocean is an important variable that likely explains model differences in high latitude climate prediction. Furthermore, mixed-layer ocean models without poleward heat transport exaggerate the equator to pole temperature gradient, while specified SSTs may not be sustainable (Valdes, 1994).

Cretaceous greenhouse climate

The Cretaceous was a period of global warmth with ice-free continents and globally averaged surface temperatures 6–14 °C higher than present (Barron, 1983). The focus of the earliest Cretaceous modeling studies was to understand the factors that led to a warm Cretaceous period. Using a simple planetary albedo model, Thompson and Barron (1981) concluded that reductions in albedo resulting from reduced land area, snow cover, and sea ice, could account for all of the Cretaceous warmth. Subsequent calculations using a one-dimensional EBM (Barron et al., 1981), a mean-annual atmospheric GCM coupled to an energy balance ocean model with no thermal inertia (Barron and Washington, 1982), and an atmospheric GCM with seasonally varying solar insolation and a mixed-layer ocean (Barron et al., 1993a) indicated that the role of continental geography on global-average Cretaceous surface temperature was minor. On the other hand, an increase in atmospheric CO_2 levels (2–10 × PAL) caused substantial global-average warming (Barron and Washington, 1985; Barron et al., 1993) due to a decrease in global albedo resulting from melting of snow/ice and an increase in water vapor. In fact, an increase to 4 × PAL CO_2 levels caused a global-average warming of 3.6–5.5 °C. An additional CO_2 feedback, enhanced high-latitude cloud cover, may further contribute to Cretaceous high-latitude warming (Sellwood and Valdes, 1997).

Ironically, the greenhouse gas solution to Cretaceous global warmth introduced a new climate problem: model-predicted Cretaceous tropical temperatures exceeded the estimates from climate proxies and approached the thermal tolerance of some tropical organisms while high-latitude regions were still too cold (Barron and Washington, 1985; Barron et al., 1993). Efficient poleward heat transport was suggested as a way to reduce the Cretaceous surface thermal gradient (Barron et al., 1981; Barron, 1983), but this flew in the face of classical theories that suggested a reduced thermal gradient would displace large-scale circulation features towards the poles and cause sluggish atmospheric and oceanic circulation. However, early atmospheric and coupled ocean-atmosphere GCM simulations demonstrated that a reduction in the surface thermal gradient did not cause sluggish circulation (Barron and Washington, 1982; Manabe and Bryan, 1985; Bush and Philander, 1997). Rather, the vertically integrated meridional temperature gradient was maintained or slightly increased through compensation by differential latent heating, leading to slightly increased zonal-average wind speeds at many latitudes (Barron and Washington, 1982).

Because of its enormous heat capacity, the ocean was identified as a possible source of enhanced poleward heat transport (Barron et al., 1981; Barron and Washington, 1982, 1985; Schneider et al., 1985; Covey and Barron, 1988; Covey and Thompson, 1989; Rind and Chandler, 1991). The idea of enhanced ocean heat transport was strengthened by atmospheric GCM experiments that demonstrate that an increase in ocean heat transport is only partially compensated by a reduction in atmospheric heat transport (Covey and Thompson, 1989) and could produce substantial global climate warming (Covey and Thompson, 1989; Rind and Chandler, 1991; Barron et al., 1993). Cretaceous atmospheric GCM simulations with enhanced oceanic heat transports predict high-latitude warming and low-latitude cooling, reducing the meridional surface temperature gradient (Barron et al., 1993; Barron et al., 1995; Poulsen et al., 1999). Yet, to date, a mechanism for sustaining high ocean heat transport has not been identified, though some modeling studies suggest the possibility of intensified surface circulation (Bush and Philander, 1997; Hotinski and Toggweiller, 2003).

Despite the specification of elevated atmospheric CO_2 and enhanced ocean heat transport in atmospheric GCMs of the Cretaceous, mean-annual temperatures in the continental interiors remained subfreezing, and at odds with the paleobotanical and sedimentological evidence (Schneider et al., 1985; Barron et al., 1995). This problem has been markedly ameliorated by including non-uniform vegetation, either by direct specification or by prediction with a vegetation-ecology model, in atmospheric GCM experiments (Otto-Bliesner and Upchurch, 1997; DeConto et al., 1999; Upchurch et al., 1999). The inclusion of high-latitude forest, in particular, produced a large warming effect that contributed to a 2.2 °C increase in global average temperature. These low-albedo forests warmed the high-latitude continents, which then transferred more heat to the high-latitude oceans, impeding sea-ice formation and warming coastal regions (Otto-Bliesner and Upchurch, 1997).

Ocean circulation in a warm Cretaceous climate has received considerable attention, because of the possibility that the thermohaline circulation may have reversed, resulting in subtropical deepwater formation (i.e., warm saline deepwater) and sluggish meridional ocean circulation (Chamberlin, 1906; Brass et al., 1982). Ocean GCM studies have largely undermined the hypothesis that the global thermohaline circulation was completely reversed. Using a coarse 5° × 5° ocean GCM with mean-annual forcing, Barron and Peterson (1990) reported significant subtropical deepwater formation in eastern Tethys with elevated (4 × PAL) CO_2 levels. However, subsequent finer-resolution ocean GCM simulations with seasonal forcing have predicted only limited convection at subtropical sites with most convection occurring at high-latitude Southern Hemisphere locations (Brady et al., 1998; Poulsen et al., 2001). Cretaceous GCM studies also do not support the notion of a sluggish, global meridional circulation; meridional overturning circulation is similar in experiments with different meridional surface temperature gradients (Manabe and Bryan, 1985; Brady et al., 1998).

On the basis of biogeographic and paleogeographic reconstructions, a Tethys circumglobal current has been inferred for the Cretaceous. Ocean GCM results indicate that a Tethys circumglobal current may have driven high rates of upwelling, cooling tropical temperatures and warming northern high latitudes, thereby reducing the Cretaceous meridional thermal gradient (Hotinski and Toggweiller, 2003). However, ocean GCMs have had mixed success simulating a Tethys circumglobal current (Seidov, 1986; Barron and Peterson, 1989; Bush, 1997; Bush and Philander, 1997), because the current is very sensitive to continental geometry (Poulsen et al., 1998). Regional ocean circulation and water properties may also have been highly sensitive to opening/closing of Cretaceous gateways (Poulsen et al., 2001; 2003). In addition to global models of ocean circulation, regional circulation models have been implemented to predict detailed Cretaceous circulation in the Western Interior Seaway of North America (Ericksen and Slingerland, 1990; Kump and Slingerland, 1999).

Warm "equable" Paleocene-Eocene climate

The early Eocene was the warmest interval in the Cenozoic, with global-average surface temperatures 2–4 °C warmer than present (Barron, 1987). Floral and faunal proxies suggest that continental temperatures were warm with a small annual range. Yet, on the basis of perpetual January and July atmospheric GCM simulations with specified SSTs, Sloan and Barron (1990) reported that warm, equable continental interiors could not be maintained in light of the small thermal inertia of land surfaces. Subsequently, a number of studies have attempted to reconcile the GCM predictions with early Eocene proxy evidence from the continental interiors (Sloan and Barron, 1990, 1992; Sloan and Cirbus, 1994; Sloan and Morrill, 1998; Sloan and Pollard, 1998; Sloan et al., 2001). Several factors have been identified that ameliorate the model-predicted high-latitude continental temperature range, including the presence of a large interior lake (Sloan and Cirbus, 1994), enhanced atmospheric CO_2 levels (Sloan and Cirbus, 1994; Shellito et al., 2003), cold summer-warm winter orbital parameters (Sloan and Morrill, 1998), and the specification of seasonally-varying, warm SSTs (Sloan et al., 2001).

Atmospheric GCM results indicate that elevated atmospheric CO_2 levels (at least $3 \times$ PAL) could explain the Eocene global warmth, but introduce a problem familiar to warm climates – tropical overheating (Sloan and Rea, 1995; Shellito et al., 2003). As in the Cretaceous studies, enhanced oceanic heat transport has been proposed as a possible mechanism for reducing the Eocene meridional surface temperature gradient (Barron, 1987; Covey and Barron, 1988; Rind and Chandler, 1991). Again, a mechanism for enhancing ocean heat transport has not been identified. Using an uncoupled ocean GCM forced by atmospheric GCM fields, Bice et al. (2000) suggest that the distribution of ocean heat transports between the Northern and Southern Hemisphere may be sensitive to the basin configuration. However, a coupled ocean-atmosphere model of the Eocene demonstrated reduced heat transports relative to the modern (Huber and Sloan, 2001). Polar stratospheric clouds, frozen water vapor clouds that form in polar regions, may be a partial solution for producing warm climates with low surface temperature gradients (Sloan and Pollard, 1998). By absorbing outgoing, long-wave radiation, prescribed polar stratospheric clouds in an atmospheric GCM caused warming of the troposphere and melting of sea ice, leading to significant (up to 20 °C) high-latitude warming (Sloan and Pollard, 1998).

Paleoclimate modeling studies have been conducted to determine how the ocean circulated in a warm Eocene climate. Atmospheric GCM experiments with specified SSTs displayed an intensification of the Hadley circulation and associated extremes in equatorial precipitation and subtropical evaporation, leading O'Connell et al. (1996) to suggest that conditions might have been ripe for warm, saline deepwater formation in the eastern Tethys Ocean. Indeed, Eocene ocean GCM experiments support the possibility of limited convection of warm, saline subtropical water (Barron and Peterson, 1991), provided the moisture flux from the atmosphere is favorable (Bice et al., 1997; Bice and Marotzke, 2001). In an attempt to simulate a "haline" mode circulation, Bice and Marotzke (2001) show that large perturbations to the atmospheric moisture flux (evaporation-precipitation) enhance subduction of warm, saline water to the depths, but do not increase subtropical convective mixing (the mechanism controlling deepwater formation). A coupled ocean-atmosphere simulation of the Eocene exhibits sluggish meridional circulation and warm, saline deep water (Huber and Sloan, 2001); however, the published results do not indicate whether the warm, saline water was formed by subduction or convection.

Late Cenozoic climate deterioration

After the warm, ice-free climate of the Cretaceous and early Cenozoic, a long-term cooling trend ensued, punctuated by the development of the Antarctic Ice Sheet near the Eocene/Oligocene boundary, and culminating in the Pleistocene ice age. Several mechanisms have been cited as instigators of the Cenozoic cooling, including changes in continental distribution, plateau uplift, oceanic gateways, and atmospheric CO_2. Barron (1985) used an annual-average atmospheric GCM to test the hypothesis that the evolution of the distribution and size of continental land masses caused the Cenozoic cooling trend. This model exhibited minor sensitivity to the Cenozoic paleogeographic changes, including changes in topography and continental position, and did not demonstrate a systematic decrease in global-average surface temperature.

Geologic evidence supports a significant increase in uplift rates and absolute elevation in southern Asia (the Tibetan Plateau and Himalayan Mountains) and the American West during the Cenozoic (Ruddiman et al., 1989), though the details of the uplift history remain uncertain. In a series of atmospheric GCM sensitivity experiments testing the influence of mountain elevation on climate, Kutzbach et al. (1989, 1993) showed that progressive uplift in Southern Asia and the American West resulted in largely linear changes in heating rates, vertical motion patterns, and low-level winds over the plateaus. In a similar series of experiments, Manabe and Broccoli (1990) and Broccoli and Manabe (1992) observed that large-amplitude stationary waves occur in response to the plateaus; general subsidence and infrequent storm development upstream of the troughs of these waves contribute to continental aridity. Changes in circulation associated with uplift led to patterns of regional climatic changes (colder winters over Northern Hemisphere continents; drier summers along the American Pacific coast and in the interior of Eurasia; winter drying of the American northern plains and the interior of Asia; and maintenance of warm/wet conditions along the southeast coasts of Asia and the United States), which are consistent with those in the Northern Hemisphere during the last 10 or 15 million years (Ruddiman and Kutzbach, 1989; Manabe and Broccoli, 1990). Atmospheric GCM experiments have also shown that Tibetan uplift and increased summer radiation through orbital changes are also primary controls on monsoon strength (Prell and Kutzbach, 1992). Despite the large regional climate response, plateau uplift contributes little to global cooling, indicating the need for additional climatic forcing to explain the Cenozoic cooling (Ruddiman and Kutzbach, 1989).

Ocean gateways have also been implicated in Cenozoic cooling. Kennett (1977) hypothesized that the opening of the Drake Passage near the Eocene/Oligocene boundary created an Antarctic Circumpolar Current (ACC), which thermally isolated Antarctica, leading to growth of the Antarctic Ice Sheet. Ocean GCMs that test the effect of the Drake Passage predict the formation of an ACC-reduced poleward heat transport in the high-latitude Southern Hemisphere, cooling high-latitude surface temperatures up to several degrees (ranging from 0.8–4 °C), depending on the type of sea-surface boundary conditions (restoring vs. non-restoring) (Mikolajewicz et al., 1993; Toggweiler and Samuels, 1995; Bice et al., 2000; Toggweiler and Bjornsson, 2000). In contrast, atmospheric GCM results indicated that warmer SSTs favor snowfall in the continental interior, promoting Antarctic Ice Sheet growth (Oglesby, 1989). Ocean GCMs also exhibit large circulation changes in response to an open Central American Isthmus, which existed prior to 3–4 Ma (Maier-Reimer et al., 1990; Mikolajewicz et al., 1993). In the open isthmus case, North Atlantic surface water is diluted by low salinity Pacific water, collapsing North Atlantic deepwater production. Because the North Atlantic circulation is an important source of heat in the high latitudes of the North Atlantic, Maier-Reimer et al. (1990) speculate that a compensating climatic factor must have warmed the region prior to the initiation of Northern Hemisphere Pleistocene glaciation.

Results from an asynchronously coupled ice sheet-atmosphere GCM suggest that Antarctic glaciation was induced primarily by declining atmospheric CO_2 (DeConto and Pollard, 2003). The ice sheet model exhibits highly nonlinear behavior; once an atmospheric CO_2 threshold (between ~3 and 2 × PAL) is crossed, Antarctic ice caps expand rapidly, with large orbital variations. A parameterization of the Drake Passage opening and ACC formation has a modest (and secondary) effect on ice sheet mass balance (DeConto and Pollard, 2003). Vegetation changes may also have contributed to Cenozoic cooling. Using an atmospheric GCM, Dutton and Barron (1997) propose that the evolution of grasslands and tundra in the Miocene may have caused a 1.9 °C global cooling, mainly due to the higher albedos of these vegetation types.

Warm Pliocene climate

The middle Pliocene was the most recent period in Earth history that was significantly warmer than the present. In contrast to other pre-Quaternary intervals, the boundary conditions for the Pliocene are fairly well-known as a result of the concerted efforts of the U.S. Geological Survey's PRISM (Pliocene Research, Interpretations, and Synoptic Mapping) Group. Several modeling groups have utilized the PRISM data sets to simulate the middle Pliocene climate (Chandler et al., 1994; Sloan et al., 1996; Haywood et al., 2000a,b). Despite incremental enhancements to the PRISM data set and the use of different atmospheric GCMs with varying resolution, the Pliocene simulations agree in several respects. As compared with the present, the Pliocene simulations exhibit global warming, a reduction in the equator-to-pole surface temperature gradient and zonal wind strength, and enhanced Northern Hemisphere high-latitude precipitation (Chandler et al., 1994; Sloan et al., 1996; Haywood et al., 2000b). The GCM results support the possibility that enhanced thermohaline circulation and concomitant increases in ocean heat transport could explain the middle Pliocene warmth (Chandler et al., 1994; Sloan et al., 1996). Alternatively, regional intensification of atmospheric and oceanic circulations may have induced greater heat transports from the equatorial region, warming Europe and the Mediterranean (Haywood et al., 2000a). Modeling results indicate that warmer SSTs and reduced ice cover in the Northern Hemisphere gave way to intensification of the Icelandic low-pressure and Azores high-pressure systems in the North Atlantic. The resulting surface pressure gradient increased the annual westerly wind velocity and wind stress, ultimately enhancing the flow of the Gulf Stream and North Atlantic Current as well (Haywood et al., 2000a).

Summary

Pre-Quaternary paleoclimate modeling is the science of simulating Earth's climate prior to the Quaternary using numerical climate models. The pre-Quaternary witnessed climate states that were fundamentally different to those of the Quaternary and the modern day. Since the earliest paleoclimate modeling studies using EBMs, considerable progress has been achieved in understanding the factors that have controlled Earth's past climates. Paleoclimate modeling studies have directly contributed to reshaping and debunking climate hypotheses, recognizing new climatic processes, and quantifying the climate response to various climatic factors. Much of the scientific progress has proceeded hand-in-hand with the development of increasingly sophisticated climate models. Yet, many outstanding questions remain about the evolution of Earth's climate. Future progress will be realized through further improvement and enhancement of paleoclimate models, the continued documentation of Earth's environmental and climatic history, and the persistent ingenuity of paleoclimate scientists.

Christopher J. Poulsen

Bibliography

Barron, E.J., 1983. A warm, equable Cretaceous: The Nature of the problem. *Earth Sci. Rev.*, **19**, 305–338.
Barron, E.J., 1985. Explanations of the Tertiary global cooling trend. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **50**, 45–61.
Barron, E.J., 1987. Eocene equator-to-pole surface ocean temperatures: A significant climate problem? *Paleoceanography*, **2**, 729–739.
Barron, E.J., and Peterson, W.H., 1989. Model simulation of the Cretaceous ocean circulation. *Science*, **244**, 684–686.
Barron, E.J., and Peterson, W.H., 1990. Mid-Cretaceous ocean circulation: Results from model sensitivity studies. *Paleoceanography*, **5**, 319–337.
Barron, E.J., and Peterson, W.H., 1991. The Cenozoic ocean circulation based on ocean general circulation model results. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **83**, 1–28.
Barron, E.J., and Washington, W.M., 1982. Cretaceous climate: A comparison of atmospheric simulations with the geologic record. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **40**, 103–133.
Barron, E.J., and Washington, W.M., 1985. Warm Cretaceous climates: High atmospheric CO_2 as a plausible mechanism. In Sundquist, E.T., and Broecker, W.S. (eds.), *The Carbon Cycle and Atmospheric CO_2: Natural Variations, Archean to Present.* Washington, DC: American Geophysical Union Geophysical Monograph 32, pp. 546–553.
Barron, E.J., Thompson, S.L., and Schneider, S.H., 1981. An ice-free Cretaceous? Results from climate model simulations. *Science*, **212**, 501–508.
Barron, E.J., Fawcett, P.J., Pollard, D., and Thompson, S., 1993a. Model simulations of Cretaceous climates: The role of geography and carbon dioxide. *Phil. Trans. R. Soc. Lond.*, **341**, 307–316.
Barron, E.J., Peterson, W.H., Pollard, D., and Thompson, S., 1993b. Past climate and the role of ocean heat transport: Model simulations for the Cretaceous. *Paleoceanography*, **8**, 785–798.
Barron, E.J., Fawcett, P.J., Peterson, W.H., Pollard, D., and Thompson, S.L., 1995. A "simulation" of mid-Cretaceous climate. *Paleoceanography*, **10**, 953–962.
Baum, S.K., and Crowley, T.J., 2001. GCM response to late Precambrian (~590 Ma) ice-covered continents. *Geophys. Res. Lett.*, **28**, 583–586.
Baum, S.K., and Crowley, T.J., 1991, Seasonal snowline instability in a climate model with realistic geography: Application to Carboniferous (~300 MA) glaciation, Geophys. *Res. Lett.*, **18**, 1719–1722.
Bendtsen, J., 2002. Climate sensitivity to changes in solar insolation in a simple coupled climate model. *Clim. Dyn.*, **18**, 595–609.
Bice, K.L., and Marotzke, J., 2001. Numerical evidence against reversed thermohaline circulation in the warm Paleocene/Eocene ocean. *J. Geophys. Res.*, **106**, 11529–11542.
Bice, K.L., Barron, E.J., and Peterson, W.H., 1997. Continental runoff and early Cenozoic bottom-water sources, *Geology*, **25**, 951–954.
Bice, K.L., Scotese, C.R., Seidov, D., and Barron, E.J., 2000. Quantifying the role of geographic change in Cenozoic ocean heat transport using uncoupled atmosphere and ocean models. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **161**, 295–310.
Brady, E.C., DeConto, R.M., and Thompson, S.L., 1998. Deep water formation and poleward ocean heat transport in the warm climate extreme of the Cretaceous (80 Ma). *Geophys. Res. Lett.*, **25**, 4205–4208.
Brass, G.W., Southam, J.R., and Peterson, W.H., 1982. Warm saline bottom water in the ancient ocean. *Nature*, **296**, 620–623.
Brenchley, P.J., Carden, G.A.F., and Marshall, J.D., 1995. Environmental changes associated with the "first strike" of the Late Ordovician mass extinction. *Mod. Geol.*, **22**, 69–82.
Brenchley, P.J., Carden, G.A., Hints, L., Kalijo, D., Marshall, J.D., Martma, T., Meidla, T., and Nolvak, J., 2003. High-resolution stable isotope stratigraphy of Upper Ordovician sequences: Constraints on the timing of bioevents and environmental changes associated with mass extinction and glaciation. *Geol. Soc. Am. Bull.*, **96**, 597–610.

Broccoli, A.J., and Manabe, S., 1992. The effects of orography on middle latitude northern hemisphere dry climates. *J. Clim.*, **5**, 1181–1201.
Budyko, M.I., 1969. The effect of solar radiation variations on the climate of the earth. *Tellus*, **21**, 611–619.
Bush, A.B.G., 1997. Numerical simulation of the Cretaceous Tethys circumglobal current. *Science*, **275**, 807–810.
Bush, A.B.G., and Philander, S.G.H., 1997. The late Cretaceous: Simulations with a coupled ocean-atmosphere general circulation model. *Paleoceanography*, **12**, 495–516.
Caldiera, K., and Kasting, J.F., 1992. Susceptibility of the early Earth to irreversible glaciation caused by carbon dioxide clouds. *Nature*, **359**, 226–228.
Chamberlin, T., 1906. On a possible reversal of deep-sea circulation and its influence on geologic climates. *J. Geol.*, **14**, 363–373.
Chandler, M.A., 1994. Depiction of modern and Pangean deserts: Evaluation of GCM hydrological diagnostics for paleoclimate studies. In Klein G.D. (ed.), *Pangea: Paleoclimate, Tectonics, and Sedimentation During Accretion, Zenith, and Breakup of a Supercontinent.* Boulder, CO: Geological Society of America Special Paper 288, 117–138.
Chandler, M.A., Rind, D., and Ruedy, R., 1992. Pangaean climate during the Early Jurassic: GCM simulations and the sedimentary record of paleoclimate. *Geol. Soc. Am. Bull.*, **104**, 543–559.
Chandler, M.A., Rind, D., and Thompson, R., 1994. Joint investigations of the middle Pliocene climate. II. GISS GCM northern hemisphere results. *Global Planet. Change*, **9**, 197–219.
Chandler, M.A., and Sohl, L.E., 2000. Climate forcings and the initiation of low-latitude ice sheets during the Neoproterozoic Varanger glacial interval. *J. Geophys. Res.*, **105**, 20737–20756.
Covey, C., and Barron, E., 1988. The role of ocean heat transport in climatic change. *Earth Sci. Rev.*, **24**, 429–445.
Covey, C., and Thompson, S.L., 1989. Testing the effects of ocean heat transport on climate. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **75**, 331–341.
Crowley, T.J., 1994. Pangean climates. In Klein G.D. (ed.), *Pangea: Paleoclimate, Tectonics, and Sedimentation During Accretion, Zenith, and Breakup of a Supercontinent.* Boulder, CO: Geological Society of America Special Paper 288, 25–39.
Crowley, T.J., and Baum, S.K., 1991a. Toward reconciliation of Late Ordovician (~440 Ma) glaciation with very high CO_2 levels. *J. Geophys. Res.*, **96**, 22597–22610.
Crowley, T.J., and Baum, S.K., 1991b. Seasonal snowline instability in a climate model with realistic geography: Application to Carboniferous (~300 Ma) glaciation. *Geophys. Res. Lett.*, **18**, 1719–1722.
Crowley, T.J., and Baum, S.K., 1992a. Modeling late Paleozoic glaciation. *Geology*, **20**, 507–510.
Crowley, T.J., and Baum, S.K., 1992b. Milankovitch fluctuations on the supercontinents. *Geophys. Res. Lett.*, **19**, 793–796.
Crowley, T.J., and Baum, S.K., 1993. Effect of decreased solar luminosity on late Precambrian ice extent. *J. Geophys. Res.*, **98**, 16723–16732.
Crowley, T.J., and Baum, S.K., 1994. General circulation model study of Late Carboniferous interglacial climates. *Palaeoclimates*, **1**, 3–21.
Crowley, T.J., and Baum, S.K., 1995. Reconciling Late Ordovician (440 Ma) glaciation with very high (14X) CO_2 levels. *J. Geophys. Res.*, **100**, 1093–1101.
Crowley, T.J., Mengel, J.G., and Short, D.A., 1987. Gondwanaland's seasonal cycle. *Nature*, **329**, 803–807.
Crowley, T.J., Hyde, W.T., and Short, D.A., 1989. Seasonal cycle variations on the supercontinent of Pangea. *Geology*, **17**, 457–460.
Crowley, T.J., Baum, S.K., and Hyde, W.T., 1991. Climate model comparisons of Gondwana and Laurentide glaciations. *J. Geophys. Res.*, **96**, 9217–9226.
Crowley, T.J., Yip, K.J. J., and Baum, S.K., 1994. Snowline instability in a general circulation model: application to Carboniferous glaciation. *Clim. Dyn.*, **10**, 363–376.
Crowley, T.J., Hyde, W.T., and Peltier, W.R., 2001. CO_2 levels required for deglaciation of a "Near-Snowball" Earth. *Geophys. Res. Lett.*, **28**, 283–286.
DeConto, R.M., Hay, W.W., Thompson, S.L., and Bergengren, J., 1999. Late Cretaceous climate and vegetation interactions: Cold continental interior paradox. In Barrera, E., and Johnson, C.C. (eds.), *Evolution of the Cretaceous Ocean-Climate System.* Boulder, CO: Geological Society of America Special Paper 332, 391–406.
DeConto, R.M., and Pollard, D., 2003. Rapid Cenozoic glaciation of Antarctic induced by declining atmospheric CO_2. *Nature*, **421**, 245–249.
Donnadieu, Y., Fluteau, F., Ramstein, G., Ritz, C., and Besse, J., 2003. Is there a conflict between the Neoproterozoic glacial deposits and the snowball Earth interpretation: An improved understanding with numerical modeling. *Earth Planet. Sci. Lett.*, **208**, 101–112.
Donnadieu, Y., and Ramstein, G., 2002. Is high obliquity a plausible cause for Neoproterozoic glaciations? *Geophys. Res. Lett.*, doi: 10.1029/2002GL015902.
Dutton, J.F., and Barron, E.J., 1997. Miocene to present vegetation changes: A possible piece of the Cenozoic cooling puzzle. *Geology*, **25**, 39–41.
Endal, A.S., and Schatten, K.H., 1982. The faint young-sun climate paradox: Continental influences. *J. Geophys. Res.*, **87**, 7295–7302.
Ericksen, M.C., and Slingerland, R., 1990. Numerical simulations of tidal and wind-driven circulation in the Cretaceous Interior Seaway of North America. *Geol. Soc. Am. Bull.*, **102**, 1499–1516.
Fawcett, P.J., and Barron, E.J., 1998. The role of geography and atmospheric CO_2 in long-term climate change: Results from model simulations for the Late Permian to present. In Crowley, T.J., and Burke, K. (eds.), *Tectonic Boundary Conditions for Climate Reconstructions.* New York: Oxford University Press, pp. 21–38.
Fawcett, P.J., Barron, E.J., Robison, V.D., and Katz, B.J., 1994. The climatic evolution of India and Australia from the Late Permian to mid-Jurassic: A comparison of climate model results with the geological record. In Klein, G.D. (ed.), *Pangea: Paleoclimate, Tectonics, and Sedimentation During Accretion, Zenith, and Breakup of a Supercontinent.* Boulder, CO: Geological Society of America Special Paper 288, 139–157.
Gérard, J.C, Hauglustaine, D.A., and François, L.M., 1992. The faint young sun climate paradox: A simulation with an interactive seasonal climate-sea ice model. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **97**, 133–150.
Gibbs, M.T., Barron, E.J., and Kump, L.R., 1997. An atmospheric pCO_2 threshold for glaciation in the Late Ordovician. *Geology*, **25**, 447–450.
Gibbs, M.T., Bice, K.L., Barron, E.J., and Kump, L.R., 2000. Glaciation in the early Paleozoic "greenhouse": The roles of paleogeography and atmospheric CO_2. In Huber, B.T. et al., (eds), *Warm Climates in Earth History.* Cambridge: Cambridge University Press, pp. 386–422.
Gibbs, M.T., Rees, P.M., Kutzbach, J.E., Ziegler, A.M., Behling, P.J., and Rowley, D.B., 2002. Simulations of Permian climate and comparisons with climate-sensitive sediments. *J. Geol.*, **110**, 33–55.
Hay, W.W., and Wold, C.N., 1998. The role of mountains and plateaus in a Triassic climate model. In Crowley, T.J., and Burke, K. (eds.), *Tectonic Boundary Conditions for Climate Reconstructions.* New York: Oxford University Press, pp. 116–143.
Haywood, A.M., Sellwood, B.W., and Valdes, P.J., 2000a. Regional warming: Pliocene (3 Ma) paleoclimate of Europe and Mediterranean. *Geology*, **28**, 1063–1066.
Haywood, A.M., Valdes, P.J., and Sellwood, B.W., 2000b. Global scale palaeoclimate reconstruction of the middle Pliocene climate using the UKMO GCM: Initial results. *Global Planet. Change*, **25**, 239–256.
Held, I.M., and Suarez, M.J., 1974. Simple albedo feedback models of the icecaps. *Icarus*, **26**, 613–628.
Henderson-Sellers, B., and Henderson-Sellers, A., 1989. Modelling the ocean climate for the early Archaean. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **75**, 195–221.
Herrmann, A.D., Patzkowsky, M.E., and Pollard, D., 2003. Obliquity forcing with 8–12 times preindustrial levels of atmospheric pCO_2 during the Late Ordovician glaciation. *Geology*, **31**, 485–488.
Hoffman, P.F., Kaufman, A.J., Halverson, G.P., and Schrag, D.P., 1998. A Neoproterozoic snowball earth. *Science*, **281**, 1342–1346.
Hotinski, R.M., and Toggweiller, J.R., 2003. Impact of a Tethyan circumglobal passage on ocean heat transport and "equable" climate. *Paleoceanog.*, doi:10.1029/2001PA000730.
Huber, M., and Sloan, L.C., 2000. Climatic response to tropical sea surface temperature changes on a "greenhouse" Earth. *Paleoceanography*, **15**, 443–450.
Huber, M., and Sloan, L.C., 2001. Heat transport, deep waters, and thermal gradients: Coupled simulation of an Eocene greenhouse climate. *Geophys. Res. Lett.*, **28**, 3481–3484.
Hunt, B.G., 1979. The effects of past variations of the Earth's rotation rate on climate. *Nature*, **281**, 188–191.
Hyde, W.T., Crowley, T.J., Baum, S.K., and Peltier, W.R., 2000. Neoproterozoic 'snowball Earth' simulations with a coupled climate/ice-sheet model. *Nature*, **405**, 425–429.

Hyde, W.T., Crowley, T.J., Tarasov, L., and Peltier, W.R., 1999. The Pangean ice age: Studies with a coupled climate-ice sheet model. *Clim. Dyn.*, **15**, 619–629.

Jenkins, G.S., 2000. Global climate model high-obliquity solutions to the ancient climate puzzles of the Faint-Young Sun Paradox and low-latitude Proterozoic glaciation. *J. Geophys. Res.*, **105**, 7357–7370.

Jenkins, G.S., Marshall, H.G., and Kuhn, W.R., 1993. Precambrian climate: The effects of land area and Earth's rotation rate. *J. Geophys. Res.*, **98**, 8785–8791.

Jenkins, G.S., and Smith, S.R., 1999. GCM simulations of Snowball Earth conditions during the late Proterozoic. *Geophys. Res. Lett.*, **26**, 2263–2266.

Kasting, J.F., Pollack, J.B., and Crisp, D., 1984. Effects of high CO_2 levels on surface temperatures and atmospheric oxidation state of the early earth. *J. Atmos. Chem.*, **1**, 403–428.

Kennett, J.P., 1977. Cenozoic evolution of Antarctic glaciation, the circum-Antarctic oceans and their impact on global paleoceanography. *J. Geophys. Res.*, **82**, 3843–3859.

Kiehl, J.T., and Dickinson, R.E., 1987. A study of the radiative effects of enhanced atmospheric CO_2 and CH_4 on early Earth surface temperatures. *J. Geophys. Res.*, **92**, 2991–2998.

Kirschvink, J.L., 1992. Late Proterozoic low-latitude global glaciation: The snowball Earth. In Schoff, J.W., and Klein, C. (eds.), *The Proterozoic biosphere.* New York: Cambridge University Press, pp. 51–52.

Kuhn, W.R., and Kasting, J.F., 1983. Effects of increased CO_2 concentrations on surface temperatures of the early Earth. *Nature*, **301**, 53–55.

Kuhn, W.R., Walker, J.C.G., and Marshall, H.G., 1989. The effect on Earth's surface temperature from variations in rotation rate, continent formation, solar luminosity, and carbon dioxide. *J. Geophys. Res.*, **94**, 11129–11136.

Kump, L.R., and Slingerland, R.L., 1999. Circulation and stratification of the early Turonian Western Interior Seaway: Sensitivity to a variety of forcings. In Barrera, E., and Johnson, C.C. (eds.), *Evolution of the Cretaceous Ocean-Climate System.* Boulder, CO: Geological Society of America Special Paper 332, 181–190.

Kutzbach, J.E., 1994. Idealized Pangean climates: Sensitivity to orbital change, In Klein G.D. (ed.) *Pangea: Paleoclimate, Tectonics, and Sedimentation During Accretion, Zenith, and Breakup of a Supercontinent.* Boulder, CO: Geological Society of America Special Paper 288, 41–55.

Kutzbach, J.E., and Gallimore, R.G., 1989. Pangaean climates: Megamonsoons of the megacontinent. *J. Geophys. Res.*, **94**, 3341–3357.

Kutzbach, J.E., and Ziegler, A.M., 1993. Simulation of Permian climate and biomes with an atmosphere-ocean model: Comparisons with observations. *Phil. Trans. R. Soc. Lond.*, **341**, 327–340.

Kutzbach, J.E., Guetter, P.J., Ruddiman, W.F., and Prell, W.L., 1989. Sensitivity of climate to late Cenozoic uplift in Southern Asia and the American West: Numerical experiments. *J. Geophys. Res.*, **94**, 18393–18407.

Kutzbach, J.E., Prell, W.L., and Ruddiman, W.F., 1993. Sensitivity of Eurasian climate to surface uplift of the Tibetan Plateau. *J. Geology*, **101**, 177–190.

Longdoz, B., and Francois, L.M., 1997. The faint young sun climatic paradox: Influence of the continental configuration and of the seasonal cycle on the climatic stability. *Global Planet. Change*, **14**, 97–112.

Maier-Reimer, E., Mikolajewicz, U., and Crowley, T., 1990. Ocean general circulation model sensitivity experiment with an open Central American Isthmus. *Paleoceanography*, **5**, 349–366.

Manabe, S., and Broccoli, A.J., 1990. Mountains and arid climates of middle latitudes. *Science*, **247**, 192–195.

Manabe, S., and Bryan, K. Jr., 1985. CO_2-induced change in a coupled ocean-atmosphere model and its implications. *J. Geophys. Res.*, **90**, 11689–11707.

McGuffie, K., and Henderson-Sellers 1997. *A Climate Modeling Primer.* New York: Wiley, 253 pp.

Mikolajewicz, U., Maier-Reimer, E., Crowley, T.J., and Kim, KY., 1993. Effect of Drake Passage and Panamanian gateways on the circulation of an ocean model. *Paleoceanography*, **8**, 409–426.

Moore, G.T., Hayashida, D.N., Ross, C.A., and Jacobson, S.R., 1992. Paleoclimate of the Kimmeridgian/Tithonian (Late Jurassic) world. I. Results using a general circulation model. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **93**, 113–150.

O'Connell, S., Chandler, M.A., and Reudy, R., 1996. Implications for the creation of warm saline deep water: Late Paleocene reconstructions and global climate model simulations. *Geol. Soc. Am. Bull.*, **108**, 270–284.

Oglesby, R.J., 1989. A GCM study of Antarctic glaciation. *Clim. Dyn.*, **3**, 135–156.

Oglesby, R.J., and Ogg, J.G., 1999. The effect of large fluctuations in obliquity on climates of the late Proterozoic. *Paleoclimates*, **2**, 293–316.

Otto-Bliesner, B.L., 1993. Tropical mountains and coal formation: a climate model study of the Westphalian (306 MA). *Geophys. Res. Lett.*, **20**, 1947–1950.

Otto-Bliesner, B.L., 1998. Effects of tropical mountain elevations on the climate of the Late Carboniferous: Climate model simulations. In Crowley, T.J., and Burke, K. (eds.), *Tectonic Boundary Conditions for Climate Reconstructions.* New York: Oxford University Press, pp. 100–115.

Otto-Bliesner, B.L., and Upchurch, G.R. Jr., 1997. Vegetation-induced warming of high-latitude regions during the Late Cretaceous period. *Nature*, **385**, 804–807.

Owen, T., Cess, R.D., and Ramanathan, V., 1979. Enhanced CO_2 greenhouse to compensate for reduced solar luminosity on early Earth. *Nature*, **277**, 640–642.

Parrish, J.T., Ziegler, A.M., and Scotese, C.R., 1982. Rainfall patterns and the distribution of coals and evaporates in the Mesozoic and Cenozoic. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **40**, 67–101.

Pavlov, A.A., Kasting, J.F., Brown, L.L., Rages, K.A., and Freedman, R., 2000. Greenhouse warming by CH_4 in the atmosphere of early Earth. *J. Geophys. Res.*, **105**, 11981–11990.

Pollard, D., and Schulz, M., 1994. A model for the potential locations of Triassic evaporate basins driven by paleoclimatic GCM simulations. *Global. Planet. Change*, **9**, 233–249.

Poulsen, C.J., Seidov, D., Barron, E.J., and Peterson, W.H., 1998. The impact of paleogeographic evolution on the surface oceanic circulation and the marine environment within the mid-Cretaceous Tethys. *Paleoceanography*, **13**, 546–559.

Poulsen, C.J., Barron, E.J., Johnson, C.C., and Fawcett, P., 1999. Links between major climatic factors and regional oceanic circulation in the mid-Cretaceous. In Barrera, E., and Johnson, C.C. (eds.), *Evolution of the Cretaceous Ocean-Climate System.* Boulder, CO: Geological Society of America Special Paper 332, 73–89.

Poulsen, C.J., Barron, E.J., Arthur, M.A., and Peterson, W.H., 2001a. Response of the mid-Cretaceous global oceanic circulation to tectonic and CO_2 forcings. *Paleoceanography*, **16**,576–592.

Poulsen, C.J., Pierrehumbert, R.T., and Jacob, R.L., 2001b. Impact of ocean dynamics on the simulation of the Neoproterozoic "snowball Earth." *Geophys. Res. Lett.*, **28**, 1575–1578.

Poulsen, C.J., Jacob, R.L., Pierrehumbert, R.T., and Huynh, T.T., 2002. Testing paleogeographic controls on a Neoproterozoic snowball Earth. *Geophys. Res. Lett.*, doi: 10.1029/2001GL014352.

Poulsen, C.J., Gendaszek, A.S., and Jacob, R.L., 2003. Did the rifting of the Atlantic Ocean cause the Cretaceous thermal maximum? *Geology*, **31**, 115–118.

Poussart, P.F., Weaver, A.J., and Barnes, C.R., 1999. Late Ordovician glaciation under high atmospheric CO_2: A coupled model analysis. *Paleoceanography*, **14**, 542–558.

Prell, W.L., and Kutzbach, J.E., 1992. Sensitivity of the Indian monsoon to forcing parameters and implications for its evolution. *Nature*, **360**, 647–652.

PSUCLIM 1999a. Storm activity in ancient climates 1. Sensitivity of severe storms to climate forcing factors on geologic timescales. *J. Geophys. Res.*, **104**, 27277–27293.

PSUCLIM 1999b. Storm activity in ancient climates 2. An analysis using climate simulations and sedimentary structures. *J. Geophys. Res.*, **104**, 27295–27320.

Rees, P.M., Gibbs, M.T., Ziegler, A.M., Kutzbach, J.E., and Behling, P.J., 1999. Permian climates: Evaluating model predictions using global paleobotanical data. *Geology*, **27**, 891–894.

Rees, P.M., Ziegler, A.M., Gibbs, M.T., Kutzbach, J.E., Behling, P.J., and Rowley, D.B., 2002. Permian phytogeographic patterns and climate data/model comparisons. *J. Geol.*, **110**, 1–31.

Rind, D., and Chandler, M., 1991. Increased ocean heat transports and warmer climates. *J. Geophys. Res.*, **96**, 7437–7461.

Ruddiman, R.F., and Kutzbach, J.E., 1989. Forcing of late Cenozoic northern hemisphere climate by plateau uplift in Southern Asia and the American West. *J. Geophys. Res.*, **94**, 18409–18427.

Ruddiman, W.F., Prell, W.L., and Raymo, M.E., 1989. Late Cenozoic uplift in Southern Asia and the American West: Rationale for general circulation modeling experiments. *J. Geophys. Res.*, **94**, 18379–18391.
Sagan, C., and Mullen, G., 1972. Earth and Mars: Evolution of atmospheres and surface temperatures. *Science*, **177**, 52–56.
Schneider, S.H., Thompson, S.L., and Barron, E.J., 1985. Mid-Cretaceous continental surface temperatures: Are high CO_2 concentrations needed to simulated above-freezing winter conditions? In Sundquist, E.T., and Broecker, W.S. (eds.), *The carbon cycle and atmospheric CO_2: Natural variations, Archean to present.* Washington, DC: American Geophysical Union Geophysical Monograph 32, 546–553.
Seidov, D.G., 1986. Numerical modeling of the ocean circulation and paleocirculation. In Hsu, K.J. (ed.), *Mesozoic and Cenozoic Oceans.* Washington, DC: Geodynamics Series, American Geophysical Union, 15, 11–26.
Sellers, W.D., 1969. A global model based on the energy balance of the Earth-Atmosphere system. *J. Appl. Meteorol.*, **8**, 392–300.
Sellwood, B., and Valdes, P.J., 1997. Geological evaluation of climate General Circulation Models and model implications for Mesozoic cloud cover. *Terra Nova*, **2**, 75–78.
Sewall, J.O, Sloan, L.C., Huber, M., and Wing, S., 2000. Climate sensitivity to changes in land surface characteristics. *Global. Planet. Change*, **26**, 445–465.
Shellito, C., Sloan, L., and Huber, M., 2003. Climate model sensitivity to atmospheric CO_2 levels in the Early-Middle Paleogene. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **193**, 113–123.
Sloan, L.C., and Barron, E.J., 1990. “Equable” climates during Earth history? *Geology*, **18**, 489–492.
Sloan, L.C., and Barron, E.J., 1992. A comparison of Eocene climate model results to quantified paleoclimatic interpretations. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **93**, 183–202.
Sloan, L. Cirbus 1994. Equable climates during the early Eocene: Significance of regional paleogeography for North American climate. *Geology*, **22**, 881–884.
Sloan, L.C., and Morrill, C., 1998. Orbital forcing and Eocene continental temperatures. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **144**, 21–35.
Sloan, L.C., and Pollard, D., 1998. Polar stratospheric clouds: A high latitude warming mechanism in an ancient greenhouse world. *Geophys. Res. Lett.*, **25**, 3517–3520.
Sloan, L.C., and Rea, D.K., 1995. Atmospheric carbon dioxide and early Eocene climate: A general circulation modeling sensitivity study. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **119**, 275–292.
Sloan, L.C., Crowley, T.J., and Pollard, D., 1996. Modeling of middle Pliocene climate with the NCAR GENESIS general circulation model. *Mar. Micropaleontol.*, **27**, 51–61.
Sloan, L.C., Huber, M., Crowley, T.J., Sewall, J.O., and Baum, S., 2001. Effect of sea surface temperature configuration on model simulations of “equable” climate in the Early Eocene. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **167**, 321–335.
Thompson, S.L., and Barron, E.J., 1981. Comparison of Cretaceous and present Earth albedos: Implications for the causes of paleoclimates. *J. Geol.*, **89**, 143–167.
Toggweiler, J.R., and Bjornsson, H., 2000. Drake Passage and palaeoclimate. *J. Quat. Sci.*, **15**, 319–328.
Toggweiler, J.R., and Samuels, B., 1995. Effect of Drake Passage on the global thermohaline circulation. *Deep Sea Res. I*, **42**, 477–500.
Trenberth, K.E., 1992. *Climate system modeling.* New York: Cambridge University Press, 788 pp.
Upchurch, G.R. Jr., Otto-Bliesner, B.L., and Scotese, C.R., 1999. Terrestrial vegetation and its effects on climate during the latest Cretaceous, In Barrera, E., and Johnson, C.C. (eds.), *Evolution of the Cretaceous Ocean-Climate System.* Boulder, CO: Geological Society of America Special Paper 332, 407–426.
Valdes, P.J., 1994. Atmospheric general circulation models of the Jurassic. In Allen, J.R.L., Hoskins, B.J., Sellwood, B.W., Spicer, R.A., and Valdes, P.J. (eds.), *Palaeoclimates and Their Modeling: With Special Reference to the Mesozoic Era.* London: Chapman and Hall, pp. 109–118.
Valdes, P.J., and Sellwood, B.W., 1992. A palaeoclimate model of the Kimmeridgian. *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, **95**, 47–72.
Walker, J.C.G., and Zahnle, K.J., 1986. Lunar nodal tides and distance to the Moon during the Precambrian. *Nature*, **320**, 600–602.
Washington, W.M., and Parkinson, C.L., 1986. *An introduction to three-dimensional climate modeling.* New York: University Science Books, 422 pp.
Williams, G.E., 1975. Late Precambrian glacial climate and the Earth’s obliquity. *Geol. Mag.*, **112**, 441–465.
Williams, G.E., 1986. Precambrian permafrost horizons as indicators of palaeoclimate. *Precambrian Res.*, **32**, 233–242.
Wilson, K.M., Pollard, D., Hay, W.W., Thompson, S.L., and Wold, C.N., 1994. General circulation model simulations of Triassic climates: Preliminary results. In Klein, G.D. (ed.), *Pangea: Paleoclimate, Tectonics, and Sedimentation During Accretion, Zenith, and Breakup of a Supercontinent.* Boulder, CO: Geological Society of America Special Paper 288, 91–116.
Yapp, C.H., and Poths, H., 1992. Ancient atmospheric CO_2 inferred from natural goethites, *Nature*, **355**, 342–344.
Ziegler, A.M., Hansen, K.S., Johnson, M.E., Kelly, M.A., Scotese, C.R., and Van der Voo, R., 1977. Silurian continental distributions, paleogeography, climatology, and biogeography. *Tectonophysics*, **40**, 13–51.

Cross-references

Carbon Isotope Variations over Geologic Time
Cenozoic Climate Change
Climate Change, Causes
Climate Forcing
Cretaceous Warm Climates
Faint Young Sun Paradox
Early Paleozoic Climates (Cambrian-Devonian)
Glaciations, Pre-Quaternary
Heat Transport, Oceanic and Atmospheric
Late Paleozoic Paleoclimates (Carboniferous-Permian)
Mesozoic Climates
Mountain Uplift and Climate Change
Neogene Climates
Obliquity
Paleocene-Eocene Thermal Maximum
Paleocean Modeling
Paleogene Climates
Plate Tectonics and Climate Change
Snowball Earth Hypothesis

http://www.springer.com/978-1-4020-4551-6

Encyclopedia of Paleoclimatology and Ancient Environments
Gornitz, V. (Ed.)
2009, XXVIII, 1049 p. 585 illus., 38 illus. in color., Hardcover
ISBN: 978-1-4020-4551-6