WARM CLIMATES IN EARTH HISTORY

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Warm climate forcing mechanisms

PAUL J. VALDES

ABSTRACT

Warm climates present a particularly challenging test of our understanding of climate system processes. This paper reviews the possible mechanisms that can affect climate on long timescales. Increased radiatively active gases can act to warm climate but cannot by themselves simulate correctly the temperature gradient between equator and pole. Ocean and atmospheric heat transport, large bodies of water, a modified cryosphere, clouds, and surface vegetation are all thought to be of importance, but our limited ability to model some of these processes means that at present we cannot fully explain the mechanisms behind warm periods in earth history.

INTRODUCTION

Warm climates in the distant past provide one of the most challenging tests of our understanding of climate system processes and our ability to predict them. The differences between these climates and our present one are dramatic. It is hard to imagine a world in which, for instance, there were no ice caps at either pole. Why were these periods so warm? What processes could lead to a climate regime that was so radically different from our present one? What are the parallels with possible warmer worlds that are predicted for the future? These questions provide the motivation for paleoclimate studies of these periods, and elucidating them is of fundamental importance for our understanding of climate dynamics in warmer climate regimes.

The evidence for warm climates in the past is addressed in other chapters in this volume (e.g., Crowley and Zachos, Chapter 3; Bice *et al.*, Chapter 4; Thomas *et al.*, Chapter 5; MacLeod *et al.*, Chapter 8). Data from a wide variety of ocean and terrestrial sources all agree on some basic aspects. The Mesozoic and early Tertiary were undoubtedly warmer than the present, especially at mid- and high latitudes. Ice cover was either absent or very restricted. Low latitude temperatures seem to have been similar to the present, at least within the bounds of observational error (*c.* 2 °C; (Crowley and Zachos, Chapter 3), and thus latitudinal temperature gradients appear to have been very much reduced. This chapter discusses the possible forcing mechanisms that could produce such a warm, low gradient climate and examines the role of climate modeling in explaining past warm periods.

DEFINITION OF THE PROBLEM

The increasing amount of knowledge that we are gaining of past climates enables us to ask ever more subtle questions, but there are probably three main issues. Firstly, why was the mean global climate warmer for some periods in the past? Secondly, why was the temperature gradient between equator and pole reduced? Geological evidence indicates that the greatest warming was at high latitudes, and that temperatures in the tropics were similar to, or even slightly cooler than, those at present (Crowley and Zachos, Chapter 3, this volume). Early climate model simulations for the Cretaceous suggested that the amount of CO_2 (or other radiatively active gases) required to produce sufficient warmth at high latitudes would result in overheating in the tropics (Schneider *et al.*, 1985). Thus warm periods in the past cannot be explained solely by increased concentrations of CO_2 and other radiatively active gases. Furthermore, a reduced temperature gradient between equator and pole suggests relatively sluggish atmospheric zonal winds (although it should be noted that the geological record measures only the surface temperature gradient; a vertically averaged temperature gradient would be more useful but is geologically unattainable).

The third issue of importance, not least because of the contradictions between model results and data, is continental interior temperatures. The geological data suggest that these were more equable than today: for example winter temperatures are thought to have been above freezing in the Eocene and Cretaceous (e.g., Wing and Greenwood, 1993; Markwick, 1994; Herman and Spicer, 1997; Norris *et al.*, Chapter 6, this volume; Wing *et al.*, Chapter 7, this volume). However, most model simulations to date have produced too low winter temperatures (e.g., Barron and Washington, 1984; Barron *et al.*, 1993). This result arises from the simple fact that land has a relatively low effective heat capacity compared with sea ice or ocean (a ratio of 1:5:60), and thus cools rapidly in winter. The contradiction between the data and model results has led to claims that either the data are incorrect or the models and our understanding of climate are fundamentally flawed (Sloan and Barron, 1990; Wing, 1991).

A SIMPLE CLIMATE MODEL

Before discussing possible mechanisms it will be useful to develop a conceptual model of how our climate system works. To a first approximation, and for time periods in excess of decadal, the climate system must reach an equilibrium in which the global annual mean of incoming solar energy is balanced by the outgoing long-wave radiation (Crowley and North, 1991). If this were not true then the earth would warm or cool until a new equilibrium was reached. This ‘energy balance’ can be written as

$$S_0 \times (1 - \alpha_p) = \varepsilon_p \sigma T_s^4, \qquad (1.1)$$

where S_0 is the solar constant (currently about 1370 W m^{-2} but 4.5% less in the late Ordovician); α_p is the planetary albedo (proportion of solar radiation reflected back to space, currently approximately 0.30); ε_p is proportional to the infrared emissivity, and is currently approximately 0.61; σ is the Stefan–Boltzmann constant (5.67×10^{-8} W m^{-2} K^{-4}), and T_s is the surface temperature (in kelvin). The flux of energy released from the earth's core is small compared with the other terms and can probably be ignored (but see Roach, 1998).

For the earth at present, this corresponds to a global annual mean surface temperature of 15 °C, which is approximately correct. The model is so simple that it has to apply in all climate regimes, whether past, present, or future. Unfortunately, it is also so simple that it is not really a predictive model but rather a diagnostic one. There are many parameters that have to be specified but which can and do vary depending on the climate system. For instance, the planetary albedo depends on the surface conditions (especially ice cover, land extent, and vegetation type) and also on the amount, type, and microphysical properties of clouds. The infrared emissivity depends on atmospheric composition and cloud cover. In addition, these parameters strongly depend on latitude, so a refinement to the simple energy balance model is to include latitude and time dependence:

$$\begin{aligned} C \times (\text{rate of temperature change}) = S_0 \times (1 - \alpha) - \varepsilon\sigma T_s^4 \\ - \text{divergence of latitudinal heat transport,} \end{aligned} \tag{1.2}$$

where C is the heat capacity of the climate system, and C, S, α, ε, and T are all now functions of latitude. S depends on the orbital parameters, and α generally varies with temperature, including a marked change in albedo when the temperature decreases and ice forms (typical values are from 75% for ice-covered conditions to 25% for ice-free conditions). If the climate is in equilibrium, the rate of change of temperature will be zero and we return to pure energy balance. The most important new term, however, is related to the total transport of heat by the atmosphere and ocean. This transport is a necessary part of the climate system because the solar input is greater in the tropics than at high latitudes but the long-wave cooling is more uniform with latitude. Thus the tropics receive more energy than they emit, and the polar regions emit more energy than they absorb, and hence there must be transport of heat from the equator toward the poles.

Again, such a simple model has to apply to all periods, but there are still many parameters that need to be specified rather than predicted. Indeed, there are now more parameters since we have to know the latitudinal transport of heat (which is largely unknown for periods other than the present). For the present climate, the transport is divided approximately equally between the ocean and the atmosphere. For the atmosphere, observations suggest that three main processes accomplish the heat transport. Firstly, the zonal mean flow (the Hadley and Ferrel cell) and the zonal variations in the mean flow induced by orography and land/sea temperature contrasts transport heat toward the pole. Secondly, mid-latitude depressions transport a substantial amount of heat toward the pole. Indeed, it can be argued that the only reason we have these transient systems is because of their efficiency in transporting heat from the subtropics to polar regions. Thirdly, observations suggest that a substantial amount of heat transport is associated with movement of water vapor (Peixoto and Oort, 1992).

A common approach in these simple models is to make transport of heat proportional to the latitudinal temperature gradient: the transport is weaker if the equator-to-pole temperature gradient is weaker. This leads to a potential paradox when considering a climate that is warmer than the present at high latitudes (Bice *et al.*, Chapter 4, this volume; DeConto *et al.*, Chapter 9, this volume). Equation (1.2) would suggest that a warmer pole would require more heat being transported from tropical regions. However, the reduction in the temperature gradient implies a reduction in poleward heat transport.

In part, this paradox arises from a too simplistic treatment of the equation. Firstly, in a warmer climate there would be less ice and hence more solar radiation would be absorbed at higher latitudes (Rind and Chandler, 1991). This would be only partially offset by increased emission of long-wave radiation (and both processes could be influenced by changes in cloud cover). Thus there would be less need for transport of heat to the poles. Secondly, the processes that transport heat from equator to poles are more complicated than can be modeled based on temperature gradient alone. Changes in the land/sea contrast, orography, and ocean seaways could all be important, as could the vertical ocean temperature gradient. Thus the paradox can be addressed but the challenge is to determine if we can answer it and do so in a quantifiable way.

An additional point to note is that complex climate models (called general circulation models, GCMs) are just an extension of Equation (1.2) in an attempt to remove some of the arbitrariness of the parameters. Rather than specifying albedo, emissivity, and heat transport, they calculate them from ‘first principles’ by considering Newton’s laws of motion and the first and second laws of thermodynamics. However, the models cannot solve these equations exactly and the approximations (called parameterization schemes) effectively have parameters which are the equivalents of those in Equation (1.2) but are generally less understandable. The parameterizations are derived from present-day observations of the physical processes but may in reality depend heavily on the climate state. It is this which is the main cause of uncertainties in climate model predictions of the future and is why it is vital to test the models on past climate regimes. Further, this testing must be for climate regimes warmer than present so that the models can be properly evaluated in an appropriate parameter range.

CLIMATE PROCESSES

The earth’s climate is one of the most complex physical systems studied by science. It encompasses a huge range of different processes and time and spatial scales. It is potentially a highly non-linear system, and thus to separate ‘cause and effect’, or forcing from response, is often extremely difficult. However, for the purposes of this paper it is useful to separate processes that force climate and climate change from the response of the climate system. It is also useful to further subdivide forcing mechanisms into external processes, which are completely unaffected by the state of the earth's climate, and internal processes, which can be modified and feed-back with climate.

Wigley (1981) suggested the following processes:

- changes in the position of the solar system relative to the galactic center
- evolution of the sun and solar variability
- changes in the orbit of the earth
- continental drift and orography
- volcanic activity
- evolution of the atmosphere
- albedo feedbacks and the land surface
- ocean and atmospheric circulation.

The first four processes are external to the climate system, while the subsequent processes become more and more internal. Ocean and atmospheric circulation should probably be viewed as part of the response of the climate system, rather than a forcing process.

The effect of the first process is difficult to quantify. It may be of importance (McCrea, 1975), but there is insufficient knowledge, data, and understanding to assess this and thus it is not discussed further here. The other processes are discussed below.

Evolution of the sun and solar variability

Solar evolution models (e.g., Endal and Sofia, 1981) generally suggest that solar output has increased by about 5% over the last 500 Ma, and by no more than 1% over the last 100 Ma. We can use our simple energy balance model to estimate the importance of these changes. If we assume that both albedo and emissivity are unchanged from present-day values, then a 1% decrease in solar constant would imply a temperature decrease of approximately 1 °C. We will see later that if we include feedback processes that modify the albedo and emissivity this could double the result. However, the key point is that solar variability seems to be in the wrong direction to explain past warm periods, and we need to find other processes that will more than compensate for these solar changes (Gibbs *et al.*, Chapter 13, this volume).

It should also be noted that some solar physics models (e.g., Gough, 1977) have suggested that Kelvin–Helmholtz instabilities could cause short-term solar output variability of up to 5% (generally a reduction). Such changes would last for the order of 10 Ma. The corresponding temperature changes would be substantial and, in general, there is little evidence of short-term cold spells during the last few hundred million years.

Changes in the orbit of the earth

Orbital changes are thought to be the main cause of the climate variations over the last 2 Ma. These orbital changes are caused by interactions between the gravitational forces of the earth, moon, and sun, and other planets. The changes modify the seasonal distribution of incoming solar radiation, but have very little effect on the annual average radiation. Feedback processes in the climate system, especially in the cryosphere and the ocean, convert the seasonal changes into annual mean changes. For instance, at 115 000 years ago the orbit was such that there was less insolation during the northern hemisphere summer. This resulted in cooler summers, winter snow not completely melting, and a gradual build-up of snow and ice. This increased the planetary albedo and cooled climate, in terms of both the global and the annual average. This is a very strong positive feedback process and greatly amplifies the direct orbital forcing.

Orbital changes have certainly been occurring in the distant past, but it is not possible to calculate the precise orbital configuration beyond approximately 5 Ma ago (Berger *et al.*, 1989). As a result, any study of the effects of orbital changes during the Mesozoic and early Tertiary can be based only on taking both typical and extreme values for the eccentricity, obliquity, and precession of the earth's orbit (e.g., Crowley and Baum, 1995). Such studies have shown that orbital changes can strongly influence climate, especially at high latitudes and on a seasonal basis. Whether this gets converted into an annual mean change will depend largely on the state of the cryosphere: if there is no ice cover at either pole the impact of orbital changes is likely to be very subdued (compared with the present); if ice is present, then it becomes likely that the orbital variations will manifest themselves in large-amplitude oscillations of temperature at mid- and high latitudes.

It has also been suggested that the obliquity of the earth was substantially smaller during the Eocene (Wolfe, 1978), based on the presence of light-limited floras. However, if this were true it would act to increase the equator-to-pole temperature gradient (Barron, 1984), and thus would not help to explain past warm periods.

There are a number of further caveats to the above discussion. Firstly, only in some cases can the geological data resolve the relatively rapid orbital oscillations. In such cases the amplitude appears to be large, even if we are not sure of the response mechanism. There is a small possibility that there could be a preservational bias to one particular phase of the orbital variations. If this were the case, the data would give a misleading view of the warm periods. Although this idea cannot be completely discounted, it is probably unlikely because of the large range of proxy climate indicators. Further, Oglesby and Park (1989), Valdes *et al.* (1995), and others have shown that, at least for the Cretaceous and the Jurassic, even using extreme orbital parameters cannot reproduce mid-latitude continental warmth.

Secondly, there is a problem with evaluating the effects of orbital variations. Many GCMs have failed to simulate glacial inception, at 115 000 years ago (e.g., Rind *et al.*, 1989; Phillips and Held, 1994). There is currently only one atmospheric GCM that has successfully predicted snow to last over the summer season (Dong and Valdes, 1995). Other GCMs have simulated glacial inception when ocean processes (Sytkus *et al.*, 1994) or vegetation feedbacks have been included (Gallimore and Kutzbach, 1996). Thus there is some uncertainty about the ability of models to correctly simulate the climate change associated with orbital parameter changes.

Continental drift and orography

Changes in continental configuration and orographic relief have been widely discussed as possible mechanisms for climate change (e.g., Barron and Washington, 1984). They can operate in several ways. Firstly, the distribution of land can profoundly alter the climate regime and circulation. Large continents will generally have large seasonal variations of temperature because the thermal capacity of land is lower than that of ocean or ice. If the continents are at tropical latitudes, summers will be characterized by a monsoon-type circulation with heavy rainfall (Kutzbach and Gallimore, 1989). At higher latitudes, continents near the pole can also strongly influence snow accumulation (e.g., Crowley *et al.*, 1987; Gibbs *et al.*, Chapter 13, this volume). Barron and Washington (1984) found that paleogeography could explain a substantial fraction of the Cretaceous warmth. However, Barron *et al.* (1993) found that the inclusion of a seasonal cycle resulted in a major change in these conclusions. In these simulations, the paleogeography was found to be of less importance than CO_2 and sea-ice distribution.

This issue of continentality has often been raised as central to discussions of mid- and high latitude warmth in the middle Cretaceous and Eocene. The effects of lakes (Sloan, 1994) and interior seaways (Valdes *et al.*, 1996) have both been proposed as possible mechanisms of amelioration of low winter temperatures. Such processes might explain the warmth of some regions, but results from Eurasia (Herman and Spicer, 1997) show that high continental temperatures, especially in winter, are widespread and therefore more difficult to explain by the introduction of lakes or seaways.

Secondly, the distribution of continents can profoundly influence ocean circulation. The opening up of seaways has been argued to be important for the formation of the Antarctic ice cap (Kennett, 1977), and for late Cenozoic climate change when the Panama isthmus closed (Maier-Reimer *et al.*, 1990). Also, during periods with a Pangean supercontinent there was effectively only one ocean and thus only one western boundary current (compared with the two at present: the Kuroshio Current and the Gulf Stream). This might have resulted in considerably reduced transport of heat by the ocean (Valdes, 1993).

Orography can also play a major part in altering regional climate (Norris *et al.*, Chapter 6, this volume), although the direct effects on global mean temperatures are less pronounced (e.g., Barron, 1985). Obvious local responses include reduced surface temperature in the vicinity of the mountains and rain shadows on the downwind side but larger-scale effects of orography can be more important. Much recent attention has focused on the effects of the growth of Tibet on climate (see the review by Molnar *et al.*, 1993). Tibet acts as a heat source that intensifies the Asian monsoon and hence strongly influences the climate regime of a large part of the tropics. Further, the increased weathering due to Tibetan uplift has been suggested to be important for atmospheric CO_2 concentrations and it has been claimed that the growth of Tibet resulted in global cooling during the Cenozoic (Raymo *et al.*, 1988; Raymo and Ruddiman, 1992).

A further effect of orography is that the cold elevated surfaces may be where ice sheets start to form, first as small mountain glaciers and then spreading to form a larger area. The ice sheets have a high albedo, and hence reflect more solar radiation back to space, resulting in a positive feedback.

Volcanic activity

Volcanic eruptions can influence climate on a short timescale by the emission of sulfate aerosols which cool the climate system. If these aerosols remain only in the troposphere they are relatively rapidly (in a few weeks) removed by precipitation processes and do not have a strong effect on climate. However, if the eruption is sufficiently intense to emit the aerosols into the stratosphere the climatic effect of the eruption will last for considerably longer. From experience of modern eruptions (Robock and Free, 1995), the timescale is a decade and the climatic effect is a cooling of a few degrees Celsius.

Most geological indicators of climate have very crude temporal resolution, so that individual volcanic activity would be climatically recordable in the rock record only if there were periods of sustained activity well beyond any bounds based on the recent past. Bralower *et al.* (1997) have suggested that massive volcanic activity played a role in enhancing low latitude intermediate water production, leading to a methane clathrate disassociation (see below). Thus this indirect effect of volcanism can result in substantial warming.

Volcanic activity also emits CO_2 and this could have a more important effect on long-term climate. Long-term accumulation of CO_2 from outgassing from oceanic ridge systems as well as from intra-plate volcanism can result in substantial variations of atmospheric CO_2. Thus periods with above-average total areas of mid-ocean spreading are also likely to have enhanced atmospheric and oceanic CO_2 levels (see next section).

Evolution of the atmosphere

The most common explanation for the global warmth of distant past climates is elevated concentrations of atmospheric CO_2 (e.g., Barron *et al.*, 1993; Sloan and Rea, 1996). There is much evidence, from sources such as the isotopic carbon ratio in paleosols (e.g., Cerling, 1991) and geochemical modeling (e.g., Berner, 1994), that suggests that this was true for most of the Mesozoic and early Tertiary. Geochemical modeling can estimate CO_2 concentrations throughout the last 500 Ma, although the uncertainties become large further back in time. Even for the late Mesozoic, estimates are from two to eight times present-day, pre-industrial values. The radiative effect of CO_2 is proportional to the logarithm of the concentration, and thus some of the proposed large changes in CO_2 are less dramatic than they at first appear.

An increase in CO_2 concentration results in a reduction in the emissivity and hence warming. In the absence of other feedback processes, Equation (1.1) would suggest that for a doubling of CO_2 the global annual mean warming would be about 1.1 °C. An eight-fold increase would suggets a warming of about 3.3 °C. The geological data imply a 4–6 °C global warming during the middle Cretaceous, and thus an eightfold change appears to be insufficient. However, the starting assumption of no feedbacks is almost certainly incorrect.

Less is known about other radiatively important gases, including methane. Methane is known to vary naturally in glacial and interglacial times. However, there is no direct or indirect estimate of methane concentrations during warm periods in the past. There are also no model estimates for methane. In part this is because we do not have a full understanding of the sources and sinks, even for the present-day climate. Further, the residence timescale for methane can be of the order of 10–20 years and thus methane concentrations are closely linked to short-timescale processes.

Recently there has been increased interest in the effect of catastrophic releases of methane on both future climate change (Harvey and Huang, 1995) and past climates (e.g., Dickens *et al.*, 1997). It is suggested that large oceanic reservoirs of natural gas hydrates can occasionally release large amounts of methane, which is then oxidized to CO_2. These dramatic releases have been suggested to be important for short-lived intervals of extreme warmth, such as that seen at the Paleocene/Eocene boundary. It has also been proposed that increased levels of methane could lead to enhanced polar stratospheric clouds and substantial polar warming (Sloan *et al.*, 1992).

Oxygen itself is not an especially important radiatively active gas, but ozone is. The concentration of ozone is proportional to the concentration of oxygen. Thus if oxygen concentrations varied, stratospheric ozone concentrations would vary proportionately. This would result in large changes in temperature in the stratosphere but more modest changes near the surface. Hansen *et al.* (1997) showed that the climate response to changing ozone concentrations can vary from a 1–3 °C regional warming (if ozone is removed from above 10 mb) to a similar cooling if ozone is removed from between 70 and 250 mb.

Another important radiatively active gas is water vapor. Water vapor has a very short residence time in the atmosphere and is heavily influenced by the state of the climate. The amount of water vapor in the atmosphere strongly depends on temperature. Typical mixing concentrations for the surface in the tropics are 20 g kg^{-1} whereas near the pole they can be less than 1 g kg^{-1}. Thus water vapor is normally considered to be a feedback of the climate system rather than an explicit forcing. GCMs predict the concentration of water vapor based on estimates of evaporation, precipitation, and moisture transport. The simpler models incorporate the effect of water vapor through modifications to the emissivity in Equations (1.1) and (1.2).

Water vapor is a very strong absorber of long-wave radiation, and since its concentration increases with temperature it generally represents a positive feedback process. For example, if an increase in CO_2 caused some warming this would result in more water vapor in the atmosphere, which would enhance (or amplify) the initial CO_2-induced climate change. Estimates suggest that the water vapor feedback can almost double the effect on the climate system of changes in CO_2 (e.g., Cess, 1989), so that for a doubling of CO_2 the total warming is expected to be 1.8 °C (cf. 1.1 °C for no feedback).

However, this argument has been challenged by some researchers (e.g., Lindzen, 1990). Near the surface there is already so much water vapor that increasing it has no effect on the radiative properties of the atmosphere. Higher in the atmosphere, water vapor is sparser and therefore here it is more effective as a climate forcing process. Lindzen (1990) suggests that in a warmer atmosphere convection is likely to be more vigorous and this would result in a drying of the upper troposphere. The reduced water vapor concentrations would act as a cooling mechanism and thus water vapor feedback would be a damping (or negative feedback) process, with a doubling of CO_2 producing a less than 1 °C warming. Such arguments are at the heart of the debate on future climate change. From a paleoclimate perspective this presents a further challenge: almost all attempts at explaining warm climate periods invoke CO_2, at least as part of the forcing mechanism; if the sensitivity of the climate system to increases in CO_2 is indeed so small then a different mechanism will have to be found.

A further problem arises when considering CO_2-induced warming. In simple and complex models an increase in CO_2 leads to warming everywhere, even in the tropics. Barron and Washington (1985) found that for a CO_2 concentration four times that of the present day the global temperature increase was reasonable but the tropics were too warm. One possible explanation for this can be deduced from Equation (1.2). Increased transport of heat between equator and poles would cool the tropics and warm high latitudes. Sloan *et al.* (1995) proposed this for the Eocene. In simple models the total heat transport is specified (or modeled in a very simple way) whereas atmospheric GCMs predict the atmospheric component but specify the ocean transport. The resulting simulation cannot be viewed as a complete prediction because the results are strongly dependent on the choice of ocean transport. However, the results show that a combination of increased radiatively active gases plus changes to total heat transport could produce a world that was warmer and had a reduced equator-to-pole temperature gradient. However, uncertainty in the ocean heat transport element has resulted in many authors electing to specify sea surface temperature, which allows a fuller investigation of other parameter space.

Albedo feedbacks and the land surface

Equations (1.1) and (1.2) show that the energy balance of the earth is strongly influenced by planetary albedo. The earth's albedo is controlled by surface conditions and by cloud cover. At the earth's surface the most marked effects occur for ice and snow. Fresh, clean snow can have an albedo of more than 90%, older snow and ice will have somewhat lower albedos and sea ice may be nearer 55%, but all of these are much larger than other surface types. In addition, sea ice plays a crucial role in insulating the atmosphere from the ocean, and has a smaller heat capacity than open ocean. Barron *et al.* (1993) found that changes in sea-ice distribution were of fundamental importance in explaining Cretaceous warmth in GCM simulations.

Snow and ice act to cool the system by reflecting solar radiation back to space before it can warm the earth, but this should not be viewed as an external forcing factor. Like water vapor, snow and ice are strongly dependent on climate. If the earth warms, ice will retreat and the amount of solar radiation absorbed by the surface will increase. This will further warm climate and thus snow/ice albedo is a positive feedback mechanism. This could enhance the warming caused by a doubling of CO_2 from 1.8 °C to 2.3 °C.

The effect of ice albedo can dramatically modify the need for transport of heat from equator to poles. In warmer periods when ice caps were not present at either pole the albedo was reduced and more solar energy absorbed. Thus it is possible to have a warmer high latitude climate without necessarily increasing the transport of heat between equator and poles (Rind and Chandler, 1991).

There are again some caveats to the above arguments. In a recent set of GCM simulations (Randall *et al.*, 1994) it was found that a few models did not show snow albedo feedback as positive. In these models, when snow melted it was replaced by cloudy conditions so that the planetary albedo was either unaltered or even slightly increased. Even for those models that showed a positive feedback there was considerable variation in the magnitude of the response. This again shows that we lack a full understanding of the climate system.

These results emphasize that it is not just ice but also cloud cover that controls the high latitude albedo. Unfortunately the prediction of clouds is one of the most unreliable parts of any climate model yet clouds have profound effects on the energy balance of the earth. They reflect solar radiation and therefore act to cool climate. They also prevent long-wave radiation escaping to space and in this way act to warm climate. Observations of the present climate system suggest that currently the net effect of clouds is to cool the climate system (e.g., Ramanathan *et al.*, 1989). However, the key issue is to understand how that will change in different climate regimes. Further, at high latitudes in the present climate the insulating effect of clouds appears to dominate over the albedo effect and hence clouds act to warm the high latitudes. Some GCM simulations have suggested this to be important (Sloan *et al.*, 1992; Sellwood and Valdes, 1997) but these results should be treated with caution because our current ability to predict clouds is limited. Thus we cannot prove or disprove the idea that warm periods were significantly cloudier at mid- to high latitudes in winter.

Other albedo effects are related to vegetation. A tropical rainforest has an albedo of 13%, grassland an albedo of 20%, and desert an albedo as high as 40%. In addition, high latitude forests can modify the snow albedo feedback because snow falls through the branches of the trees and hence the albedo remains low, even when the snow is deep (Bonan *et al.*, 1992). Vegetation also alters the transfer of heat, momentum, and moisture between the surface and the atmosphere and thus affects atmospheric circulation. Further, Charney *et al.* (1977) have suggested that in semi-arid regions there are important vegetation/climate feedbacks. If vegetation cover decreases, the albedo increases and this results in a decrease in precipitation that would be liable to result in further vegetation decreases. Further, Clausen (1994) showed that in semi-arid Africa there was the possibility of multiple equilibria. If the region was initially vegetated, there continued to be enough precipitation to maintain the vegetation. However, if the region was initially unvegetated, then the precipitation remained weak and vegetation was not able to grow. Thus the distribution of vegetation has the potential to affect climate. On a global mean basis, GCM simulations suggest that vegetation warms climate by 1–2 °C (Dutton and Barron, 1996). However, regionally this effect can be much larger and vegetation may be an important component in explaining the problems of warm continental interiors (Dutton and Barron, 1996; Otto-Bliesner and Upchurch, 1997; DeConto *et al.*, Chapter 9, this volume). It should be noted that changing the albedo of high latitude forests is most important in springtime; during winter there is only limited sunlight.

Ocean and atmospheric circulation

Circulation changes in the atmosphere and ocean depend on all of the factors discussed above and these cannot really be considered as a forcing process. They are part of the response of the climate system to the other forcing mechanisms. The atmosphere is the fastest-changing component of the whole climate system and any attempt to understand climate change must consider the role of the atmosphere. In particular, a GCM dynamically simulates the transport of heat and moisture by the atmosphere, given suitable boundary conditions such as the solar constant, orbital variations, atmospheric composition, land/sea contrast, orography, and surface type. An atmospheric GCM also requires specification of the sea surface temperatures, or a simple ocean model may be used which requires specification of the heat transport in the ocean. The models are then run until some form of dynamic equilibrium is achieved. Typically this requires the model to simulate a decade or more. The resulting ‘prediction’ of climate can be compared with the geological record and successes or failures can be noted. The models are becoming increasingly sophisticated and fewer factors have to be specified as boundary conditions (see DeConto *et al.*, Chapter 2, this volume), but this means that there are many more parameterizations that may, in part, depend on present-day climate.

As discussed earlier, GCMs are not perfect and one of the most important areas where they appear to fail is with regard to continental interiors. There continues to be some debate about whether the models and the data are truly in disagreement because the data are not as far into the continents as would be desirable. However, it is useful to address the question of what processes could warm continents in winter. Our simple energy balance model acts as a useful guide. One way of warming the interior would be to increase the transport of heat into the interior. Heat is currently being transported from the relatively warm oceans on to the continents. The mid-latitude flow will always be from the west (unless the poles were warmer than the tropics) and a key question is where is the nearest water upstream to the continental interior? Sloan (1994) and Valdes *et al.* (1996) have shown that if a water mass is added near to the region in question, this will act to warm the climate. Thus this solution to cold continental interiors is not a ‘general’ solution but depends on finding evidence for large water masses upstream of the data points.

Another aspect of the problem is the role of orography (Norris *et al.*, Chapter 6, this volume). Continental sedimentary deposits tend to record the climate best at lower elevations, where the sediment will accumulate. Most GCMs have a relatively coarse resolution which has a tendency to smooth orography. Peaks are smaller but valleys are higher. The crudest method for correcting this error is to correct the temperature, assuming a uniform lapse rate of approximately 6 K km^{-1}

(e.g., Rees *et al.*, Chapter 10, this volume). Further sophistication can be achieved through the use of downscaling, or limited area models.

Other methods for increasing heat transport are less successful. Heat transport can be enhanced if the winds strengthen or if the nearest water masses are warmed. Climate model simulations (e.g., Schneider *et al.*, 1985; Valdes and Sellwood, 1992) have found that there is a canceling effect in that warmer high latitude sea surface temperatures reduce the windspeed so that the net effect of the warming is diminished. However, this is a zonally symmetric viewpoint and it is possible to imagine a situation whereby, for instance, the introduction of a mountain could lead to a local enhancement in the transport of heat from the south. This emphasizes the importance of correct reconstructions of orography (see Norris *et al.*, Chapter 6, this volume).

Finally, we need to consider the role of the oceans. Until recently all climate model simulations had used either specified sea surface temperature or a specified transport of heat in a simplistic ocean model. The resulting simulations could generally not be tested against ocean data since this was part of the input to the model. The ideal would be to run an ocean model as well as an atmospheric model, and this is beginning to happen (e.g., Barron and Peterson, 1990; Schmidt and Mysak, 1996; Bush and Philander, 1997; Bice *et al.*, Chapter 4, this volume). The work has two benefits over atmosphere-only models. Firstly, it is now possible to validate the models over the entire globe. The simulations are independent of ocean data and hence we have a much richer source of data for comparisons (MacLeod *et al.*, Chapter 8, this volume). This is especially important because coupled ocean–atmosphere models are now the main tools for predicting future climate change and these models must be validated against past climate data. Secondly, by predicting the ocean as well as the atmosphere, we can come much nearer to a complete answer to the question posed at the start of this paper: why were there warm periods in the past, with no polar ice caps?

However, there is one very important limitation on ocean models being driven by atmospheric models. Our current understanding of the present-day coupled system is very poor, and we cannot model it well. Many models for the present and future have to resort to arbitrary 'flux' corrections in order to get the present-day climate anywhere near correct. It has been suggested that this may be related to problems with the representation of clouds in the atmospheric component of these models (Gleckler *et al.*, 1995). Thus the use of coupled models for paleoclimate simulations should currently be viewed as an approximate method more suited for examining processes than for its truly predictive capability.

Further, it is worth pointing out that the problem of continental interiors can still be addressed without the need for ocean–atmosphere models. Sensitivity experiments using specified sea surface temperatures allow an examination of the parameter space to determine if there are any combinations of sea surface temperature that could produce warm winters. If such a distribution is found, then an ocean–atmosphere model will be required in order to examine if the sea surface temperatures are in any sense consistent with the rest of the climate system.

SUMMARY AND FINAL REMARKS

In order to answer the three questions posed at the start of this paper, we must seek physical processes which

1. decrease the global average planetary albedo and/or emissivity (e.g., changes in CO_2, changes in CH_4, large decreases in high clouds)
2. increase the poleward heat transport, and/or decrease the high latitude albedo (e.g., stronger ocean currents, or removal of polar ice caps)
3. cause changes in cloud cover, land surface type, decreased continentality and/or increased heat transport from ocean to land (e.g., boreal forests).

The first two factors suggest that there are possible processes that change the mean global temperature and the equator-to-pole temperature gradient. More problematic is the third issue relating to continental interior temperatures. Some success in the matching between data and models has been achieved only through the consideration of 'special cases', such as lakes or interior seaways, coupled to other feedback processes such as vegetation. This raises another issue. Normal scientific method should always include an estimate of the uncertainties and error bars due to the various processes. For our purpose we need to define both the geological proxy data errors and the model errors. Gates *et al.* (1996) showed that there is large intermodel variability in the prediction of present-day surface temperature for atmospheric GCMs. At high latitudes the differences can exceed 10 °C. Simulations with coupled ocean–atmosphere models will almost certainly have an even wider spread of results. Simulations of the past will have a similar range of scatter. Thus it could be said that the models and data agree to within the error bars. However, this interpretation of modeling results is controversial since a similar argument applied to future climate predictions would suggest that the predicted change in future climates in mid- and high latitudes does not exceed the modeling errors!

Finally it is worth considering a further complicating factor, that of chaos. Chaos theory has been applied to weather forecasting, and has sometimes been referred to as 'the butterfly effect.' The basic idea is that we will never be able to observe perfectly the state of the atmosphere (perhaps because a butterfly flaps its wings). The uncertainty will grow rapidly and result in an inability to deterministically forecast the weather beyond a timescale of a few weeks. The extent to which chaos theory applies to climate is still open to debate. There are some indications that climate is not as deterministically predictable as we would like to think. Several studies have shown that the climate system can exhibit multiple equilibria. Examples include the ice albedo feedback (as discussed above; see also Gibbs *et al.*, Chapter 13, this volume) and the thermohaline circulation in the North Atlantic (Manabe and Stouffer, 1988). For a given forcing, the climate can be in two (or more) different states and thus the present climate state depends both on the current forcing processes and on its history. For modeling work this would correspond to the simulation being sensitive to the initial conditions. Further, in the case of the thermohaline circulation, it has been suggested that if the climate system is close to a transition point then relatively small differences could result in major changes to climate (Rahmstorf, 1996).

If such considerations are truly important then we cannot ask the question 'Why were some periods in the past warm?' without asking an additional question about the state of the system beforehand. For instance, the real reason for the Cretaceous being ice free may not be high CO_2 levels or changes in ocean circulation but because the previous period(s) were also ice free. The key aspect to be explained would then be what causes the changes from one climate regime to another.

In practice, our knowledge of the climate system and especially particular climates is still sufficiently elementary that it is premature to pursue the issue of chaos too far. However, such richness of behavior of the climate system will insure that there will continue to be important challenges for scientists studying past and future climate change.

A corollary to the above discussion is that the climate system is non-linear and the climate dynamics of 'warm' periods are likely to be very different from those of 'cold' periods, such as the Last Glacial Maximum. This highlights the importance of studying warm periods. It is essential that we use our knowledge of these past periods in combination with the climate modeling tools that are being used to predict the future. These models must be thoroughly tested against past climates before we can have any real confidence in their ability to predict future climate.

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