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Phil. Trans. R. Soc. Lond. A 1999 **357**, 1873-1890
doi: 10.1098/rsta.1999.0405

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Modelling the climate response to orbital forcing

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Modelling the climatic response to orbital forcing is a challenging problem. There are currently two approaches. One uses simplified climate models that can be run for long periods and helps answer questions regarding the relative amplitudes of the climatic response to precession, obliquity, and eccentricity. The second approach uses more sophisticated climate models to examine the regional variations of climates associated with particular orbital forcing. Neither approach completely bridges the gap between astronomical forcing and the environmental response. In this paper, we review both methods, and their application to Quaternary and pre-Quaternary Milankovitch variations. The simplified models suggest that the temporal characteristic of variability will be different for an ice-free Earth but that there are mechanisms for the amplification of the 100 ka eccentricity variations. We also show that the more sophisticated models are relatively robust in predicting low latitude climate response, but that the predictions for higher latitude response are more variable. This can be explained in terms of whether the climate system goes through a threshold for the formation of snow and ice sheets. If a climate model has a warm bias, this threshold will be incorrectly represented.

Keywords: climate modelling; Jurassic climates; palaeoclimate; orbital forcing; Milankovitch

1. Introduction

The geological record shows evidence of climate variability on astronomical (Milankovitch) time-scales throughout a large part of the past 250 million years and beyond. At first sight, the existence of such variability is not surprising. The characteristic variability of the Earth's orbit (eccentricity, obliquity, and precession) is a product of the gravitational interaction between the Earth, Moon, Sun and other planets and such processes will have occurred throughout most of geological time. However, the changes in the Earth's orbit are not directly recorded in sedimentary rocks. The orbital variations modify the distribution of incoming solar insolation, which then influences climate. It is these variations in climate (and associated changes in sea level and other aspects of the environment) which are preserved in the rock record. If the characteristics of these interactions have changed, then this would have important consequences for the nature of the link between orbital variations and the geological record. Thus we need to have a full understanding of the links between orbital variations, climate and the sedimentary record before we can have complete confidence with astronomical calibration of the geological time-scale during the distant past.

Understanding the link between orbital forcing, climate change, and the sedimentary record is also important, because for many Mesozoic and Early Tertiary

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sequences, there is a clear signature of the 100 ka eccentricity variation, as well as the *ca.* 20 ka precession signal. Climate variability on the eccentricity time-scale (100 ka and 400 ka) has been a dominant signal (especially at high latitudes) for the last 700 ka and is thought to be related to land/lithosphere/ice feedbacks (see Crowley & North 1991). Prior to 700 ka, the dominant signal was nearer to the obliquity period (41 ka) and Saltzman & Sutera (1987), and Saltzman & Verbitsky (1993) have proposed that this transition was also related to changes in ice cover. Thus it appears that warmer periods in the past, such as the Mesozoic era in which permanent ice was either small or non-existent, are either unlikely to have large climate variability at the eccentricity periods, or any such variability would have a very different characteristic to that of the recent past. Moreover, the characteristics of climate variability depend on the region being considered. For the recent past, low latitudes are likely to be more affected by the precession signal whereas high latitudes are more influenced by obliquity. The 100 ka period eccentricity signal is seen at most latitudes. Thus if regional or global stratigraphic correlations are being attempted, we need to have a clear understanding of the geographical structure of the environmental variability recorded by the sedimentary record, and how this may have changed in past time periods.

In this paper we review our current understanding and ability to physically model climate change associated with orbital variations. Modelling such variability presents a particularly challenging set of problems. Over the past decade, there has been much progress with climate modelling but most emphasis has been focused towards understanding and predicting possible anthropogenic climate change during the next few centuries. State-of-the-art climate models used for such research now include a detailed representation of the atmosphere and ocean. Submodels also exist for simulating the terrestrial and marine biosphere, and the cryosphere. Progress is also being made in modelling of the carbon cycle. A model including all these processes would be an extremely powerful tool for past climate research but it would still be unable to answer key issues related to orbital forcing. This is because these sophisticated models could not be integrated for a glacial/interglacial cycle (due to computing limitations). Instead, the models can only be run in 'snapshot' mode. In this approach, processes that have intrinsically long time-scales (such as ice sheets) are prescribed rather than predicted.

An alternative approach is required if we wish to address issues related to the temporal variability of climate. For this purpose, some form of simplified climate model is required. Typically the most significant simplification is to reduce the dimensionality of the problem by, for example, considering only latitude and height (e.g. Gallée *et al*. 1991, 1992) or longitude and latitude (e.g. Deblonde & Peltier 1991, 1993). This results in a number of extra parametrizations of processes not explicitly represented in the model, but has the advantage that the model can be run to simulate climate over a number of glacial/interglacial cycles.

In the following sections we will discuss the relative strengths and weaknesses of the two approaches and the implications for understanding the links between orbital forcing and climate variability.

2. Insolation forcing

Before discussing climate models it is important to remember the basic characteristics of the insolation changes caused by orbital variations. The main forcing of all forms of climate models is the incoming solar radiation at the top of the atmosphere. On a global mean, annual mean basis, orbital variations have an almost negligible impact. It is the seasonality of radiation that is affected. More specifically, obliquity variations (period of 41 ka) control the magnitude of the seasonality, and precession variations (periods of 23 and 19 ka) influence the total heating in the season. Eccentricity variations (periods of 95, 123 and 410 ka) make an extremely small contribution to the global mean, annual mean, incoming solar radiation. However, eccentricity also modulates the precession signal. At times of low eccentricity (i.e. near circular orbits) the precessional effect will be diminished in magnitude.

Berger & Pestiaux (1984) and Berger *et al*. (1993) summarized the resulting time variations of incoming radiation as a function of latitude and time of the year. At low latitudes the precession signal dominates (at 23 and 19 ka), whereas at high latitudes, the obliquity signal is more important. At no latitude is the eccentricity signal substantial. Thus the large climate variability at the 100 ka period which is seen in Late Quaternary records requires special attention. Hays *et al*. (1976) and Imbrie *et al*. (1992, 1993) show that a simple linear model cannot explain this long time-period response. Thus the climate system and sedimentary record must be responding in a nonlinear sense to the imposed orbital forcing.

3. Long time-scale climate models

Almost all climate models are designed to examine some types of time variation. This includes the most sophisticated general circulation models. There is no explicit assumption of 'steady state'. However, the physical processes which are included in a particular model implicitly determine the appropriate time-scale. In general the processes included in any particular climate model can be split into three different time (and spatial) scales. There will always be a group of processes that are explicitly represented in the model. These are the resolved scales. There will also be a group of processes that vary on time (and spatial) scales much shorter than anything explicitly included in the model. The effects of such processes have to be parametrized (i.e. estimated). Finally, there will be a group of processes whose intrinsic time-scale is long compared to the resolved scale, and therefore do not change appreciably during integration. These processes act as 'boundary' conditions and have to be prescribed. For instance, a weather forecast model (for a prediction of a few days ahead) explicitly resolves the circulation of the atmosphere, but parametrizes small-scale turbulence. It also does not require a detailed model of the ocean because, on the time-scale of a few days, the sea surface temperatures do not change appreciably. Thus it is appropriate to prescribe sea surface temperature from observations.

Similarly, a climate model used to predict the next 100 years does not require a detailed submodel of the evolution of ice sheet elevation because any change in ice sheet height (on a 100 year time-scale) will generally have a small impact on the atmospheric circulation. They are potentially very important for sea level changes but the feedback onto climate is small. Thus ice sheet and sea level changes are a diagnostic aspect of future climate change simulations (on a 100 year time-scale).

Clearly, for even longer time-scales, such an approach is invalid. A climate model suitable for investigating orbital time-scale variability ideally requires predictive capabilities for all long time-scale processes. The very simplest models developed for examining the generic behaviour of the orbital forcing of climate variability avoid all

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detail and concentrate on an idealized representation of the key variables and climate processes. For instance, Maasch & Saltzman (1990) and Saltzman & Maasch (1991) represent the whole climate system in terms of three predictive variables and two diagnostic variables. They are able to reproduce a large-amplitude climate variation on a 100 ka time-scale, as well as the characteristic 'sawtooth' shape of the climate change. However, they suggest that this is near to a natural time-scale of the climate system that effectively resonates with the orbital forcing. This result suggests that the climate variations may not be entirely deterministically predictable (i.e. there is some chaotic aspect to the climate variability). If this is true, then orbital dating may have some problems.

A disadvantage of these highly simplified models is that their very simplicity may result in unphysical behaviour. This is because the models may typically include only one or two feedback processes, whereas the real climate system has many potential feedbacks and interactions. In addition, the high level of parametrizations of feedback processes may only be valid for small changes of climate. Thus the highly simplified models may be unable to simulate key aspects of the climate system.

To further investigate this, more complex models have been developed (e.g. Deblonde & Peltier 1993; Gallée *et al*. 1991; Berger *et al*. 1998). In these models, much greater detail is included and thus the models can be tested more specifically and used for understanding orbitally forced climate variability. The central physics of this type of model is the assumption of energy balance and thus these types of models are often referred to as energy balance models (EBMs). A good example of this approach is due to Gallée *et al*. (1991, 1992). This model includes representations of the atmosphere, cryosphere, and oceans, but simplifies longitudinal variations by considering sectors for each continental and ocean basin. This assumption means that the transport of heat and moisture by mid-latitude depressions is not explicitly represented because they are too small scale (both spatially and temporally). Instead, they have to be parametrized. In addition, it only includes the Northern Hemisphere. The model is able to reproduce the variations of ice volume over the past 200 ka, given an input of orbital variations and CO_2 (Berger *et al*. 1998), although the amplitude of the 100 ka period oscillations was still smaller than observed. Model sensitivity studies showed that CO_2 variations acted as a positive feedback, amplifying the direct orbital response. If atmospheric CO_2 concentrations are held constant at pre-industrial concentrations (i.e. 290 ppmv), then the amplitude of the predicted ice volume variations is too small, especially on eccentricity time-scales. A better simulation is achieved if atmospheric CO_2 concentrations are either held constant at relatively low values (210 ppmv) or if the observed variation of CO_2 is prescribed. For pre-Quaternary problems, this suggests that there may be a fundamental problem in modelling the detailed spectrum of climate variability because of lack of knowledge of the variations in CO_2. Ideally, atmospheric (and oceanic) CO_2 concentrations should be a predicted model variable but currently most carbon cycle models have difficulty in fully reproducing the variation of CO_2 over the past 150 ka.

This model has also been used to examine the onset and amplification of Northern Hemisphere glaciation at 3.5–5 Ma (Li *et al*. 1998; Maslin *et al*. 1998). The model results suggest that CO_2 variations (possibly caused by long-term tectonic changes) and changes in orbital parameters triggered the glaciation. The variation of CO_2 with time was important in determining the strength of the initial glaciation. Saltzman &

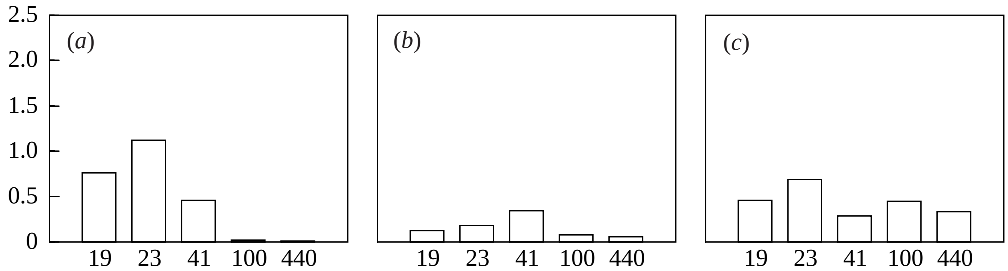

Figure 1. Total power (in kelvins) in the Milankovitch frequency bands predicted from a simple energy balance model (described in the text) for a point located over central Africa at 5° N. The FFT power spectra are for (*b*) the annual mean temperature, (*a*) the annual minimum temperature, and (*c*) the annual maximum temperature. The model was forced with orbital variations for the last 5 Ma and the last 4 Ma were analysed.

Verbitsky (1993) also suggested that the decline of atmospheric CO_2 played a major role in determining the characteristic variability of Late Cenozoic climate change.

All of these model results suggest that the characteristics of orbital forced climate variability are strongly controlled by feedbacks linked to the ice sheets. To examine this further, we have performed a number of simulations using a simple EBM, coupled to a dynamic ice-sheet model. The EBM component builds on the two-dimensional EBM described in Deblonde & Peltier (1993). The most significant alterations being a re-calibration and extension of the radiation parametrization based on modern observational (satellite) data, changes to the transport of heat to account for the nature of the Hadley circulation, and a prognostic (as opposed to a diagnostic) formulation for snow cover over land. This EBM had been coupled to an ice-sheet model based on the two-dimensional ice-flow model of Mahaffy (1976), described in Marsiat (1994), with the viscosity of the ice increased to help maintain flow at the much lower spatial resolutions used here. Lithospheric response is calculated using a locally damped bedrock model with a relaxation of 3000 years.

The model has been run for present day continental positions but with ice and snow processes disabled. This allows us to consider the nature of climate variability for warm climate regimes. Atmospheric CO_2 concentration was held constant at a pre-industrial value of 280 ppm and the model was integrated over 5 Ma forced only by the last 5 Ma of orbital insolation variations. In this simulation without ice and snow, the seasonal or annual mean temperature has little power in the 100 ka eccentricity period, either globally or locally. If, however, a nonlinear temperature diagnostic is analysed, such as the annual maximum temperature, significant power is observed at the eccentricity period. This is particularly apparent at low latitudes, where this diagnostic can be more nonlinear due to the possibility of a change in the time of year of the maximum temperature (figure 1). At high latitudes, there is a similar amplification in the variation on the eccentricity time-scales, but the relative magnitudes of the variation at shorter (precession and obliquity) time-scales are larger (figure 2).

Short *et al*. (1991) found a similar result in their EBM and explained it in terms of the equatorial land areas experiencing twice yearly peaks in solar forcing. This results in a maximum temperature occurring normally near the autumnal equinox but occasionally near the vernal equinox (depending on the precise orbital parameters). This 'rectification' of the precession signal results in power arising at eccentricity periods.

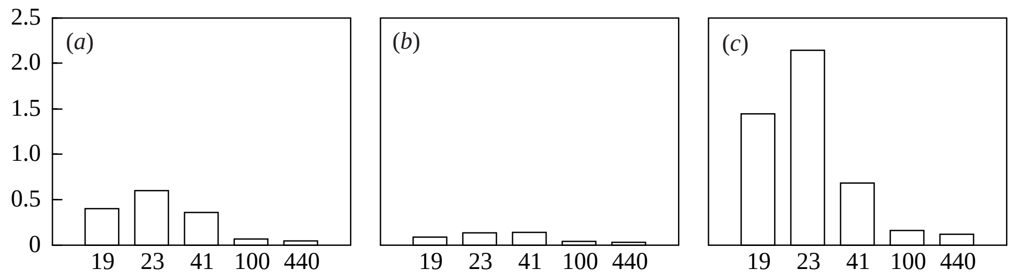

Figure 2. As in figure 1, but for a point at 50° N in central Asia.

Furthermore, Crowley *et al.* (1992) showed that this process was amplified at times of supercontinental land configuration, resulting in significant power at eccentricity periods. They also argue that this maximum temperature is a surrogate measure of monsoon strength and thus the hydrological cycle would also exhibit significant power at 100 ka periods. It is important to note that a key aspect here is that the diagnostic being considered is nonlinear, and the model is still approximately acting in a linear sense (Berger & Loutre 1997).

The hypothesis that the monsoon strength is coupled to the maximum temperature can only be fully tested using a more sophisticated model that incorporates a hydrological cycle. Although hydrology can be included in simplified models, it can only be included in a highly parametrized way. Probably the best tools for investigating this further are atmospheric general circulation models (GCMs).

4. Snapshot-style climate model simulations

As discussed in the introduction, the most sophisticated climate models include a detailed representation of the general circulation of the atmosphere and ocean. The models are based on the physical processes thought to govern the climate system but a number of approximations have to be made in order to solve the equations. Nonetheless the models are still extremely complicated and are expensive to run. Except for a few modern examples, most GCMs are run to simulate just a few decades. Therefore they cannot be used to examine the temporal variability of climate on orbital time-scales. Instead they have value because they can be used to examine the regional changes of climate for particular 'snapshots' of climate. In such simulations, the orbital parameters, ice sheets, and CO_2 concentrations are specified as boundary conditions and the models 'predict' the consequent changes in climate. It is particularly instructive to discuss the typical results of these snapshot integrations for two Quaternary times and also the pre-Quaternary.

(*a*) *Mid-Holocene* (6 ka)

By far the majority of GCM work has focused on the Late Quaternary climate. In particular, the Mid-Holocene was chosen by the Palaeoclimate Model Intercomparison Project (PMIP) to examine the climate sensitivity of GCMs to a well-defined past change of boundary conditions (Joussaume & Taylor 1995). More than 15 different atmospheric GCMs were used to perform exactly the same simulations. The models differ due to different approximations of the basic governing equations. Orbital

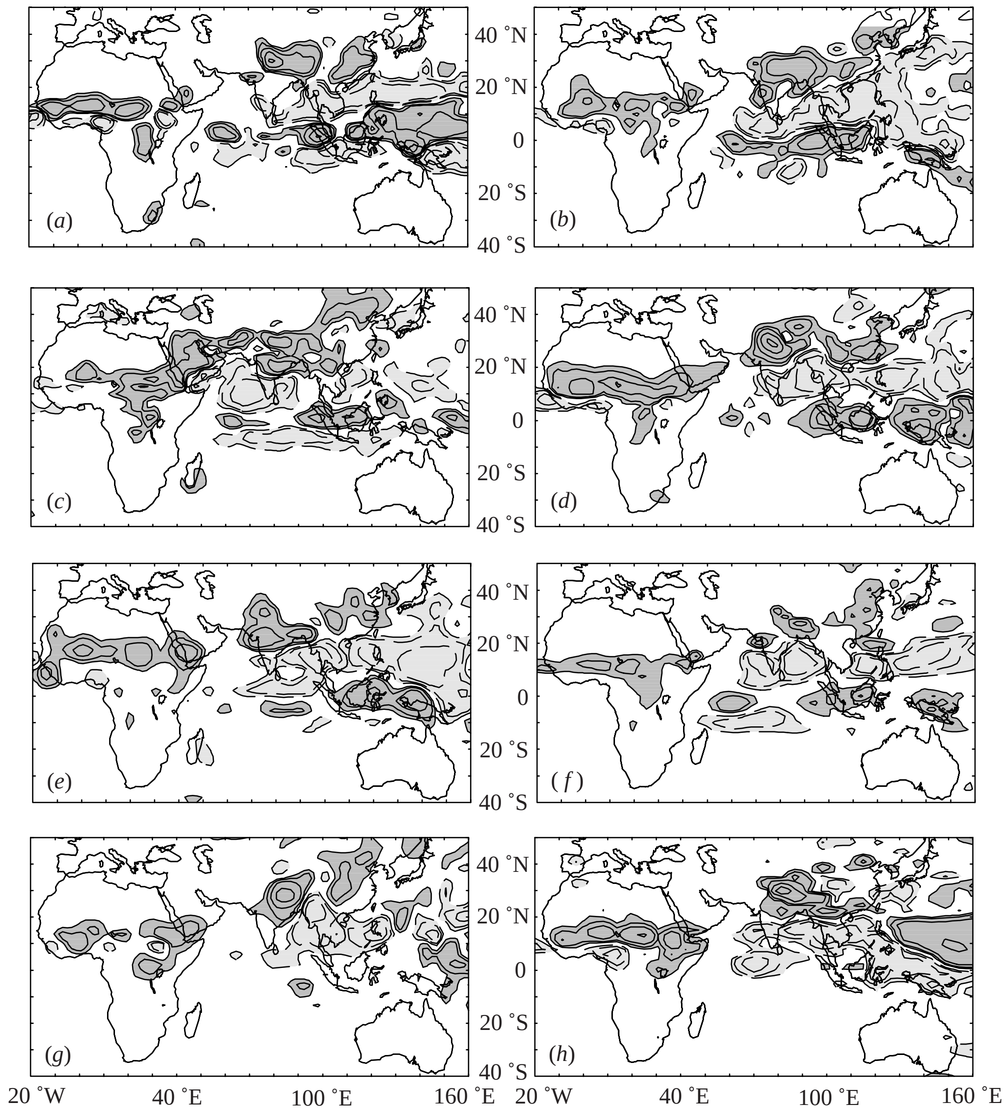

Figure 3. Changes in the June to September mean precipitation (Mid-Holocene to present day) between a number of GCM simulations. Units are mm d^{-1} and the contour intervals are at −10, −5, −2, −1, −0.5 0.5, 1, 2, 5, 10. Dark shading indicates wetter conditions during the Mid-Holocene, and light shading indicates drier conditions during the Mid-Holocene. (*a*) UGAMP; (*b*) ECHAM3; (*c*) CCM3; (*d*) GFDL; (*e*) GEN2; (*f*) UKMO; (*g*) CCC2.0; (*h*) LMCE/LMD5.

parameters and CO_2 levels were specified for 6 ka BP, but all other boundary conditions were kept at present day values. It was therefore not a complete simulation of the Mid-Holocene, but it is instructive to see what aspects of the orbitally forced climate change are relatively robust, and what aspects of the 'predicted' changes depend upon the details of the approximations used in the GCMs.

The changes in orbital forcing correspond to more incoming solar radiation in

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Northern Hemisphere summers. The resulting change in the precipitation for the June to September average (JJAS) is shown in figure 3 for eight different models. In the African and South-East Asian monsoon regions, all models show an increase in the strength of the monsoon and in the African sector; all models also show a northward shift. This is associated with an overall warming of the two continents in JJAS, as is to be expected given the orbital forcing. The results suggest that an enhancement of the monsoons at 6 ka BP is a relatively robust result. It is a very direct response to the orbital forcing and is consistent with lake level and other geological data. Hewitt & Mitchell (1998) and Kutzbach & Liu (1997) have shown that including ocean feedbacks results in a further amplification of the response but does not intrinsically change it. Similarly, Brostrom *et al*. (1998) have shown that land surface changes over North Africa can also act to further amplify the response but the general characteristics remain the same.

(*b*) *Glacial inception* (115 ka)

An interesting alternative period to study is 115 ka BP. This is often called the glacial inception period and is the time when ice sheets first grew after the previous interglacial. The orbital parameters for this period correspond to less incoming solar radiation in summer, almost the precise opposite of the Mid-Holocene. The reduced summer insolation results in cooler summer temperatures and hence a reduction in monsoon strength. Dong *et al*. (1996) and deNoblet *et al*. (1996*a*, *b*) compare the Mid-Holocene and glacial inception simulations in their GCMs and show that the monsoon responds in an almost linear manner. The change in Northern Hemisphere summer solar insolation during the glacial inception is almost equal and opposite to that during the Mid-Holocene and the change in the strength of the monsoon is also equal and opposite.

Crowley *et al*. (1992) suggest that the glacial inception is a time when the tropical temperatures should be at a maximum near the vernal equinox, rather than near the autumnal equinox. However, examination of the GCM results suggests some caution (figure 4). Only over the Amazon region is there a coherent region where the month of maximum temperature has changed to near the vernal equinox. Examination of the seasonal precipitation (not shown) suggests that the changes at 115 ka are relatively small, whereas during the Mid-Holocene the changes are much bigger.

Over the African and Asian monsoon regions the response if different. The GCM predicts no change in the timing of maximum temperature and thus these regions are not responding in the simple way predicted by the EBM. The lack of a shift seems to be related to changes in the hydrological cycle and associated changes in cloud cover. This has a strong impact on the net incoming surface solar radiation. Such a process is not explicitly modelled in simple EBMs. Moreover, Dong *et al*. (1996) show that the monsoon onset in Africa and Asia is barely changed. Thus it seems that large-scale processes, and not just the local radiative forcing are controlling these regions.

At higher latitudes, the glacial inception period has proved a particularly interesting period to study. The simplified, long time-scale models appear to successfully simulate the process where reduced summer insolation results in cooler summers and hence winter snow does not completely melt. This then results in a gradual build-up of the ice sheets. However, Royer *et al*. (1983) and Rind *et al*. (1989) showed that

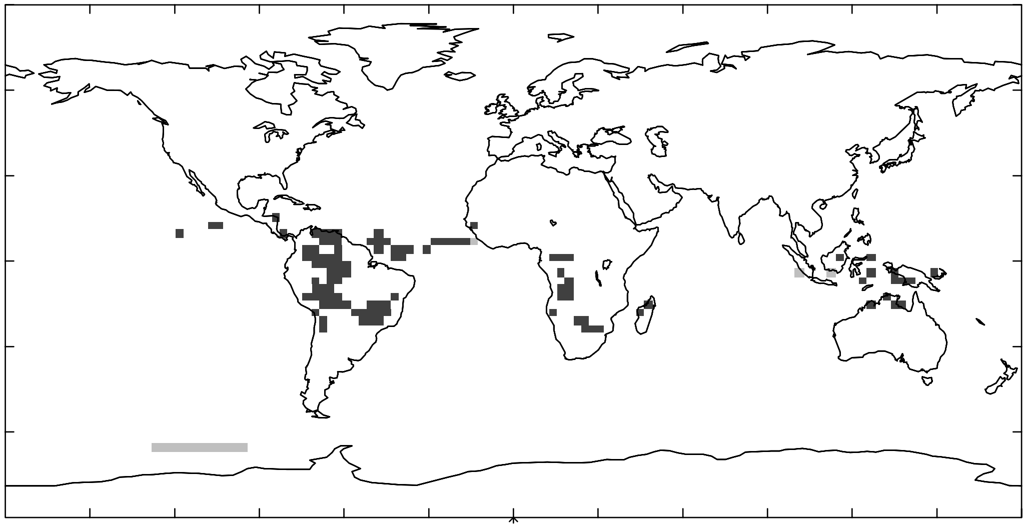

Figure 4. Dark shaded areas show the regions in which the maximum monthly mean temperature shifts from near the autumnal equinox in the present day model simulations to near the spring equinox in a simulation for 115 ka BP. Lighter shading shows regions in which the maximum monthly mean temperature shifts from near the spring equinox in the present day model simulations to near the autumn equinox in a simulation for 115 ka BP. All results are from the UGAMP GCM.

this was not the case in their GCMs. They were unable to get any winter snow to last throughout the summer. Further work showed that this was a relatively common result. However, Skytus *et al*. (1994) and Dong & Valdes (1995) did succeed in simulating some regions of snow accumulation. Since then, Gallimore & Kutzbach (1996) and deNoblet *et al*. (1996*b*) have also been able to simulate glacial inception, but only by invoking an additional feedback in which taiga is replaced by tundra. This results in an increase in the deep-snow-covered albedo and hence an enhancement of the cooling.

The key question is why did some models manage to reproduce glacial inception successfully with no additional feedback mechanism, whereas other models either cannot do it at all, or require additional feedbacks? It is almost impossible to answer this question definitively since all of the GCMs have many differences between each other. Moreover, in some publications, the climate and/or climate change were not included, so there is insufficient detail to fully evaluate the response mechanisms. However, an obvious aspect is that the survival of snow throughout the summer is a threshold-type process. If the summer temperatures rise sufficiently above freezing point, then snow melt will become rapid and the snow will not survive the summer, regardless of its initial thickness.

In most climate change experiments, the traditional approach is to show the differences between the present day and the perturbed climate (this is common for virtually all Quaternary simulations and all future climate change scenarios). The implicit assumption is that any small error in the present day control climate is unimportant. However, for a threshold mechanism, this is not true. Since the response of the system is highly nonlinear in the vicinity of the threshold, it is possible for a small error in either the control climate, or the model sensitivity to lead to gross errors

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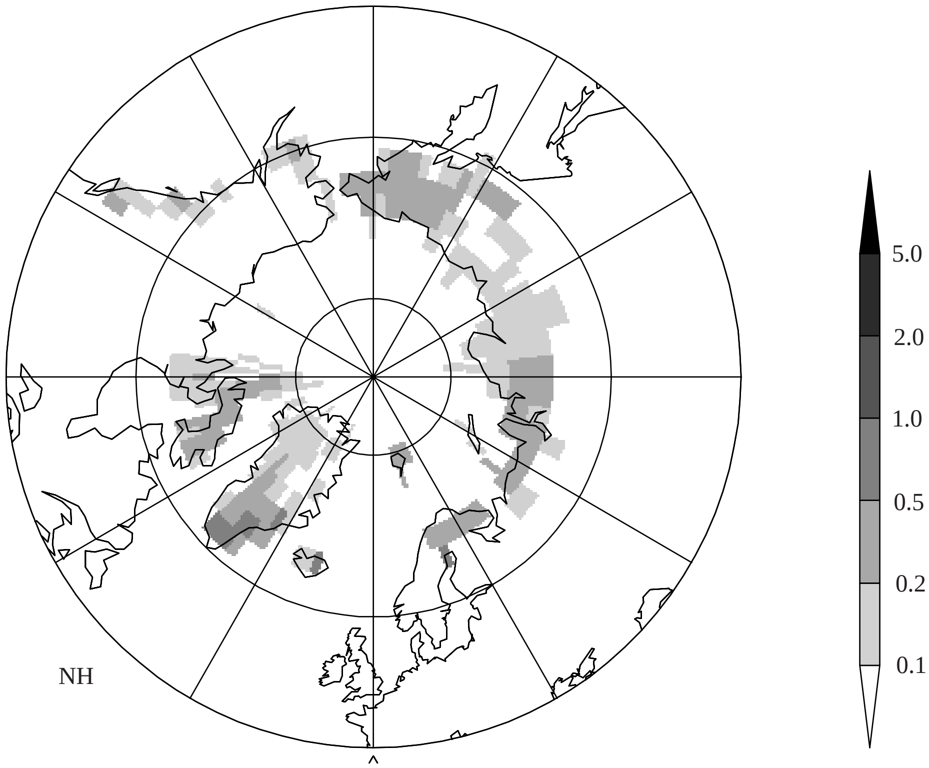

Figure 5. The surface snow/ice mass balance as simulated by the diagnostic model of Glover (1998) when forced with output from the UGAMP GCM simulation of 115 ka BP. The units are metres of water equivalent. Note that the contour interval is nonlinear.

in model output. In the case of glacial inception, the key parameter appears to be summer temperature. If the model has a warm bias in its control simulation, it is likely that the threshold temperature for glacial inception will not be achieved, even if the model is able to reproduce realistic changes in summer between the present day and 115 ka BP. If either of these components is incorrect, the model will fail to show the correct transition between snow-free summers in the present day, and snow-covered summers at 115 ka BP, unless of course, the errors fortuitously cancel.

To further investigate this, we have developed a diagnostic model of the snow mass balance (Glover 1999). The key inputs to the models are the temperatures, radiation budget, and precipitation from the atmospheric model (but also windspeed and surface humidity). The output is the net surface snow mass balance.

Dong & Valdes (1995) use the UK Universities Global Atmosphere Modelling Programme (UGAMP) GCM. If we use the UGAMP GCM results from Dong & Valdes (1995) for the present day and glacial inception simulations, we are basically able to reproduce their results (figure 5). For the present day, there is no net snow accumulation except over Greenland. At glacial inception, there are several large regions at high latitudes where snow is accumulating.

We can now repeat the offline diagnostic calculation of snow surface mass balance but using the present day model precipitation. This is shown in figure 6. In some regions, the amount of accumulation is altered (for instance, there are slightly larger

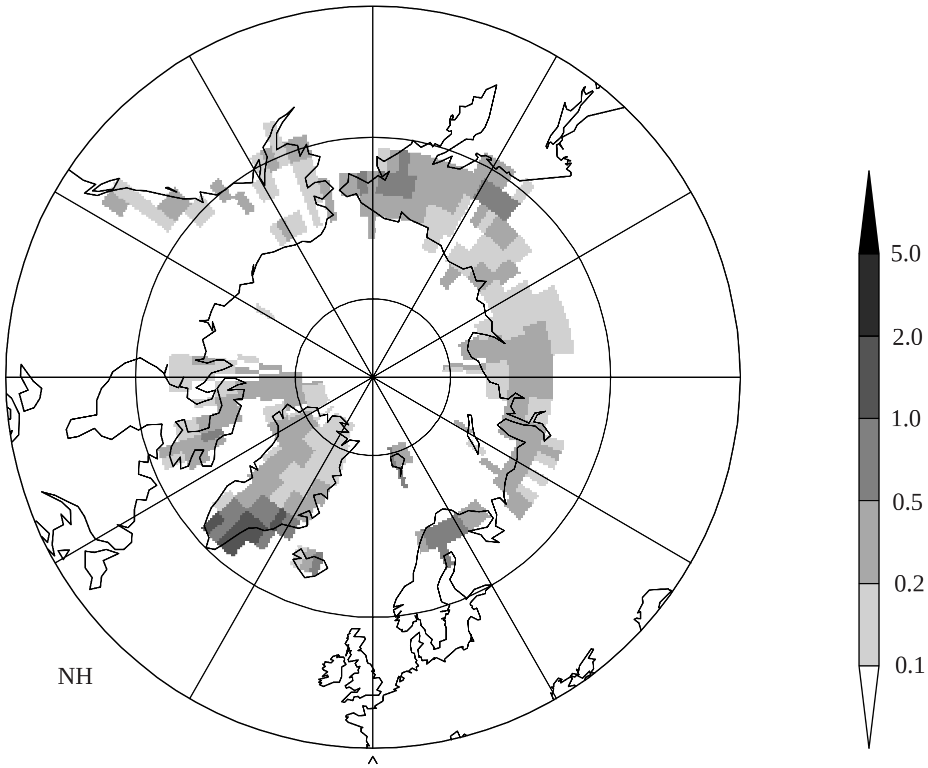

Figure 6. As figure 5, but replacing the 115 ka BP GCM precipitation with the present day model precipitation.

areas in Alaska and Siberia) but the regions which are sensitive to glacial inception are unchanged. Moreover, we get very similar results if we replace the present day model precipitation with observed precipitation. Thus models errors in simulating the present day precipitation, and the change in precipitation between present day and glacial inception do not seem to be important. This reinforces the idea that it is the changes in ablation (which depends on summer temperatures) that are the key.

It is possible to repeat the procedure using the observed present day temperature, but the model derived change of temperatures between the present day and glacial inception. The results are shown in figure 7. They show some reduction both in total accumulation and in areal coverage, but still indicate glacial inception. This suggests that while errors in the control (present day) climate of the UGAMP GCM model explain some of the accumulation during the glacial inception, much of it would remain if these errors were corrected (i.e. any small cold bias in the summer temperature in the control climate is not critical). In fact, the model does have a cold bias in winter but the results from the diagnostic model suggest that this is of lesser importance.

The conclusion is that Dong & Valdes (1995) manage to reproduce glacial inception because of a relatively large change in summer temperatures. The diagnostic model suggests that a cooling of at least 12 °C is required to reproduce the observed accumulation rates. This is a large change compared to Royer *et al*. (1983) and Rind *et al*. (1989). It is likely that part of this relatively large change was accom-

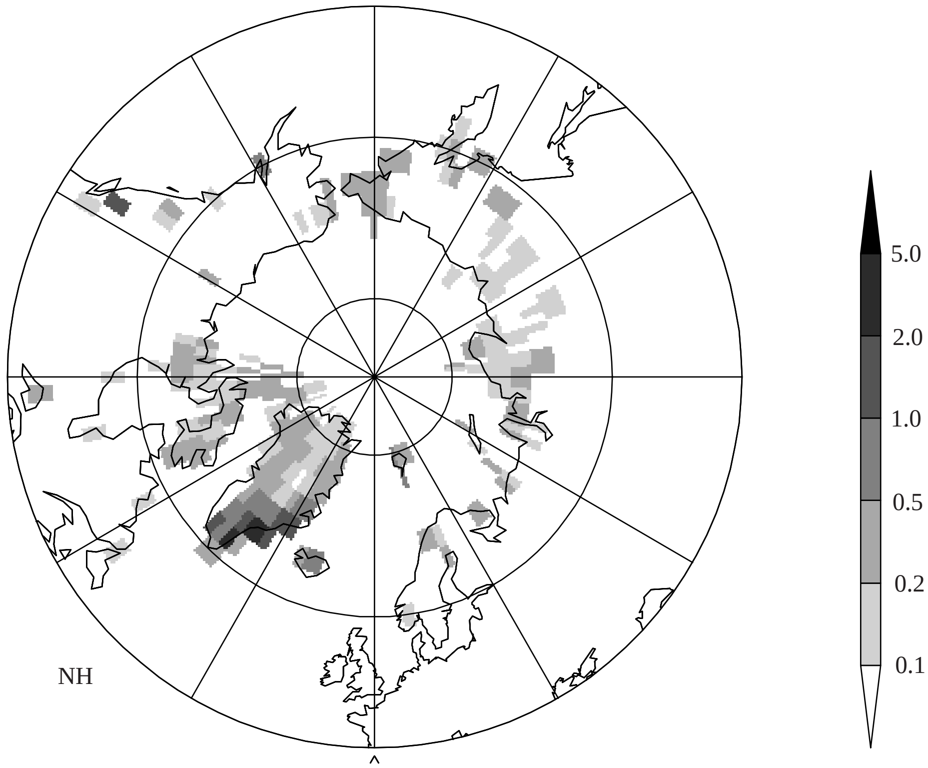

Figure 7. As figure 5, but using the observed present day temperatures and the GCM simulated change in temperature between 115 ka BP and the present day.

plished because the parametrization of snow albedo in the Dong & Valdes model only depends on the snow depth and not on the underlying vegetation type. Moreover, the deep snow albedo was relatively high. This therefore means that the snow albedo feedback was large. Effectively, the model implicitly included some of the aspects of the taiga–tundra transition.

(*c*) *Pre-Quaternary simulations*

The effect of orbital forcing on more distant past climates has also been investigated. Oglesby & Park (1989) and Park & Oglesby (1990, 1991) investigated the Mid-Cretaceous. Valdes *et al*. (1994) and Sellwood *et al*. (1999) investigated the Late Jurassic. Both sets of simulations showed that the tropical monsoon circulations were sensitive to changes in the orbital parameters and that these would effect run-off within these regions. This is the dominant signal in these modelling studies and the results from PMIP suggest that this is likely to be a fairly robust and believable aspect of the model results.

These results suggest that, in the tropics, the most likely climate signals related to orbital variations are via the hydrological cycle. However, the model results do not directly indicate the likely temporal variability. To investigate this, we have re-evaluated the Valdes *et al*. (1994) simulations to examine if the timing of the seasonal maximum temperature changes. A 'control' climate used present day orbital

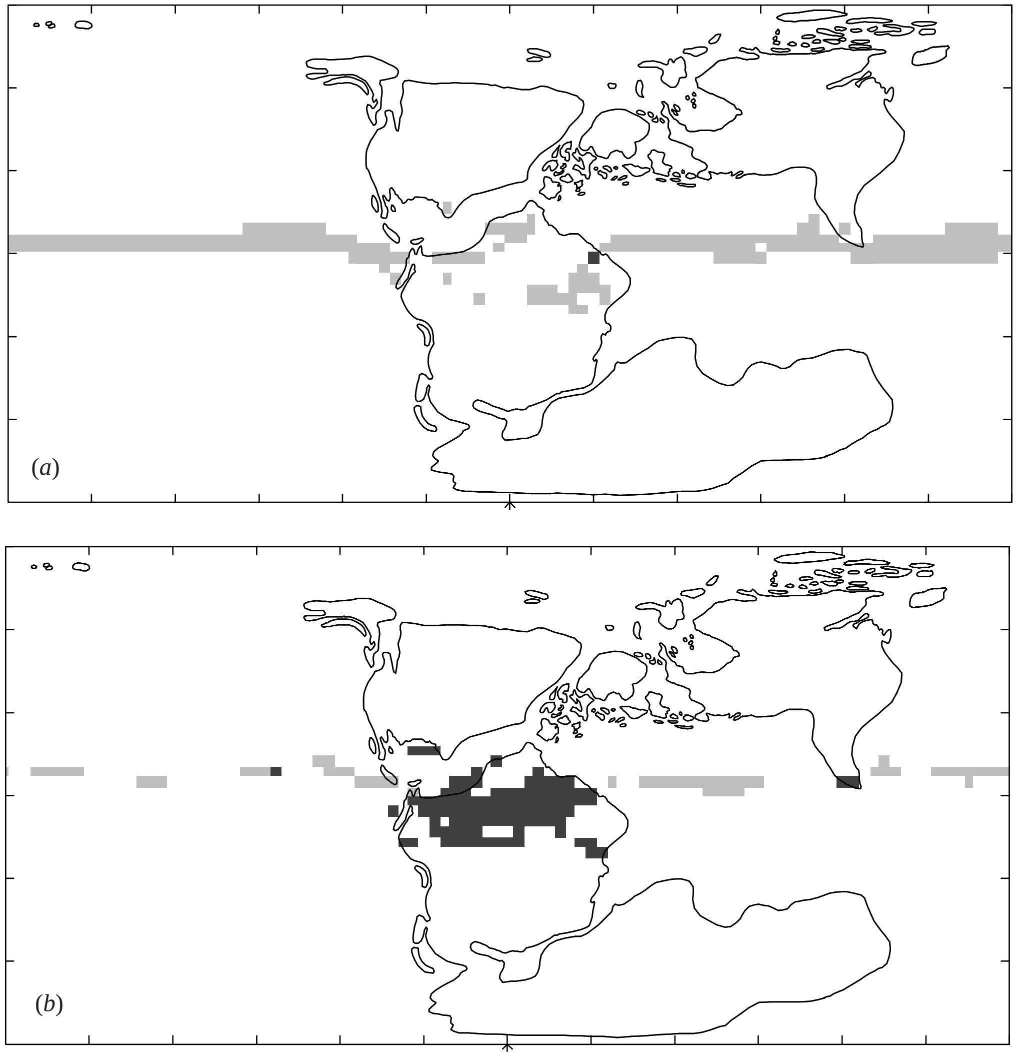

Figure 8. As figure 4 but for the Late Jurassic and using the changes between a simulation using present day orbital parameters and those using orbital parameters equivalent to (*a*) 9 ka BP, and (*b*) 115 ka BP.

parameters, and two additional simulations were completed using orbital parameters equivalent to 9 ka BP and 115 ka BP, chosen as two extremes from the last glacial/interglacial period. With orbital forcing equivalent to 115 ka BP, there are large regions of the tropics where the time of the year when temperatures reach their maximum is near the vernal equinox (figure 8). The area that is affected by this change is larger than that for the present day (figure 4). This effectively provides further confirmation of Crowley *et al.*'s (1992) suggestion that supercontinents would favour this mechanism for the amplification of the 100 ka eccentricity signal in maximum temperatures.

At mid-latitudes, the modelling predicts important changes in both temperature and the hydrological cycle. Valdes *et al.* (1994) found that some of the biggest changes occurred during winter. The changes in incoming solar radiation resulted in changes in surface temperature during winter, especially over land. Depending on the orientation of the coastline, this can increase the equator to pole temperature gradient which modifies the mid-latitude storm tracks that are the principal source of moisture for mid-latitudes during winter. In the Late Jurassic, the main region affected by this process is the Eurasian continent.

At higher latitudes, Crowley *et al*. (1993) and Valdes *et al*. (1994) found that orbital variations could influence the growth of ice sheets during the Carboniferous and Late Jurassic. The result for the Late Jurassic is particularly intriguing (and controversial) since this Period is normally assumed to be ice-free. However, the region covered by year-round snow is relatively small and centred over the present day Antarctic ice sheet and no data are available to confirm or contradict the result. Such results may help explain metre-scale sea-level variability during such time periods. However, any results about high latitude change must be treated with some caution. The discussion about the glacial inception period showed that high latitude orbitally forced climate change was difficult to model, because of the need to correctly simulate snow and ice.

5. Conclusions

This paper has attempted to review the current state of our understanding and modelling the climate response to orbital forcing. Such work is probably the most challenging problem in climate modelling because of the huge range of time-scales involved, and the many different processes that may play a significant role. The main conclusions are the following.

(1) For the Late Quaternary, ice feedbacks have dramatically altered the nature of the climate variability especially at high latitudes.

(2) There are processes that could result in substantial 100 ka climate variability in ice-free worlds, but they are unlikely to be important for palaeogeographies similar to the present (i.e. most of the Tertiary).

(3) The tropical response to orbital forcing is relatively linear and the precession signal is important. The most important aspect of change is the tropical hydrological cycle, especially within the monsoon regions.

(4) At high latitudes, our ability to model ice feedbacks is strongly constrained by the accuracy of the present day simulations (similar comments would also apply to any threshold mechanism, such as the thermohaline circulation).

In terms of cyclostratigraphy, it is clear the tropical response, especially in monsoon regions, will probably be a relatively stable feature of much of the rock record. There is a relatively direct cause-and-effect process whereby changes in orbital parameters result in change in the monsoon strength. This will get preserved in the sediment record through a number of different processes. Near the centre of the monsoon region, changes in humidity and run-off will alter lake levels. Changes in run-off will also alter $CaCO_3$ ratios in pelagic and hemipelagic environments (see Weedon 1993). Nearer to the edges of this region, small shifts in the monsoon would lead to changes between arid and humid climates and hence evaporite formation. Changes in pressure gradients related to the monsoon could also alter coastal upwelling and hence lead to changes in the fraction of organic carbon in sediments.

Mid-latitudes could be influenced by shifts or changes in the storm tracks. In the Quaternary, there are major changes in the storm tracks at the last glacial maximum (Hall *et al*. 1996) but much smaller changes at the Mid-Holocene and which were mainly concentrated in the Northern Hemisphere spring season (Hall & Valdes 1997). The changes were more dramatic at the last glacial maximum because of the strong temperature gradients associated with the greatly expanded sea ice edge. In the Jurassic, there were also changes in storm tracks but similar in magnitude to the Mid-Holocene example (Valdes *et al*. 1994).

In high latitudes the situation is more complex. Long time-scale models successfully simulate the waxing and waning of the great ice sheets, but this is sensitive to the imposed variations of CO_2. In addition, our diagnostic studies based on GCM output suggest that glacial inception is a threshold behaviour and thus very sensitive to the 'control' climate. The problem of simulating glacial inception is not the only example of a threshold-type process. Another obvious candidate is the thermohaline circulation. Computer climate modelling of such processes will always be a severe test of the models (a comment that is equally applicable to the future climate change debate).

One further vital step in our understanding is currently very weak, namely the link between climate and sediments. Any process which acts as some form of rectifier to the precession signal (such as significant run-off only occurring at one phase of the precession forcing) would result in a large eccentricity signal due to the modulation of precession by eccentricity. This includes both climatological and sedimentological processes. A simple example of this is river run-off from a large basin, such as the Amazon. Rainfall during December is mainly concentrated to the south of the basin whereas in June it maximizes to the north of the basin. Mid-Holocene-type orbits favour June–August precipitation and thus the northern part of the basin, whereas glacial-inception-type orbits would favour December–February precipitation and hence the southern part of the basin. In both cases, river run-off is enhanced and hence Amazon discharge would be a 'rectifier' of the precession signal and could be expected to have power on eccentricity time-scales.

Thus sediments may not be directly preserving the climate variability and it is important to bridge the gap between climate and sediments. Some limited initial progress towards linking climate and sediments has started to produce quantitative predictive models of sediments from climate (e.g. Price *et al*. 1995, 1997; Beerling *et al*. 1999). However, such quantitative models have not yet been applied to the modelling of orbital variations. This is clearly an area that requires much further work.

From a climate modelling perspective, most studies have so far focused on the atmospheric models coupled to at most one other component of the climate system. There are now a number of other studies with coupled ocean–atmosphere models and in the next few years there will be a tremendous expansion in models that include several additional components. Such Earth system models will at last allow us to fully tackle the link between orbital forcing, climate variability, and the sediment record and hence give us an understanding of the processes associated with cyclostratigraphy and astronomical dating.

PMIP project members contributed the data for the PMIP results. We thank Nick Shackleton and Marie-France Loutre for comments that greatly improved this paper. R.W.G. is funded by a NERC grant (GR3/10277).

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