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The Climate System in the Recent Geological Past

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SUMMARY

To model climate in the distant geological past, we assume that the physical processes involved are the same as those built into models of present-day climate, but that certain of the boundary conditions were different. As regards the Mesozoic, the major changes in boundary condition would be to continental positions, mountain elevations, and sea level; these change only on timescales of the order 10^6 years. However, during the past million years, with today's boundary conditions, climate has undergone enormous and rapid fluctuations. The examination of these more recent changes gives us very considerable insight into the myriad interactions whereby oceanic and atmospheric chemical composition, the marine and continental biosphere, the surface albedo as well as temperate-latitude ice volume and low-latitude aridity have all fluctuated in response to rather subtle changes in the latitudinal and seasonal redistribution of solar energy. At present we can only construct a general circulation model of the climate of today, or of the last ice age, by specifying many of the components of the global climate system that ought to be treated as unknown variables. The models constructed by Gallée *et al.* (1991, 1992, 1993) point to the characteristics required to simulate a glacial–interglacial cycle. Despite this, we are a long way from actually understanding how climate actually changes on this timescale. In particular, it is difficult to understand how the climate system was able to build ice sheets as quickly as they appear to have formed (Rind *et al.* 1989).

1. INTRODUCTION

Twenty thousand years ago, the northern continents were covered by massive ice sheets of the order 3 km thick; as a result sea level was some 130 m below its present level with about 3% of the present ocean mass stored as ice. A similar situation prevailed at 140 ka before present (BP) yet by 125 ka BP Hippopotamus roamed the Yorkshire Dales. In northern Africa, the Sahara–Sahel boundary moved northwards by 10° of latitude between 18 ka BP and 8 ka BP (Petit-Maire 1991); the study of the history of African lakes documents the same contrast (Street & Grove 1976). From Alaska to the Cascade Range, the Mexican volcanoes and south to the Chilean Andes the mountain snowline was almost 1 km lower at 18 ka BP than today; a similar figure can be documented in many other areas (summarized in Broecker & Denton 1989). In central Antarctica, the parallel temperature change was about 9°C (Jouzel *et al.* 1987). These changes have all taken place without significant variation in those boundary conditions that can be specified by a geologist working in the Mesozoic. It is common practise to 'test' a simple model, by asking it to predict the present climate, before using it to model the climate of the distant past. If we view the climate as being in steady state with respect to the boundary conditions, why do we not demand that it predicts the glacial climate that appears to be more probable under present boundary conditions, before accepting that it is adequate?

2. RESPONSE TO ORBITAL FORCING

During the past twenty years a great deal of effort has been devoted to describing the manner in which the earth climate system has changed on timescales of the order 10^{-4} to 10^{-5} cycles per year, and to examining the record in terms of response to the so-called Milankovitch forcing. Hays *et al.* (1976) examined records from a pair of deep-sea sediment cores in the subantarctic Indian Ocean. Together, the cores provided coverage of the last 400 ka. One of the records that they obtained, $\delta^{18}O$ in calcite foraminiferal tests, reflects variations in the volume of continental scale ice sheets in the northern hemisphere (Shackleton & Opdyke 1973). The other records were derived from analysis of the radiolarian assemblages in the sediment and provided a description of changing sea surface temperature and water mass structure (Imbrie & Kipp 1971; Hays *et al.* 1976). Several of the findings of Hays *et al.* (1976) have been central to subsequent investigations. First, different components of the climate system (atmosphere, hydrosphere, cryosphere) react with different time constants to external forcing. Second, the responses to changes in the seasonal and latitudinal distribution of solar insolation are complex, providing the opportunity to probe the working of the climate system that would not be present if it were the case that we simply observed parallel changes in temperature across the Earth (Imbrie *et al.* 1989; Imbrie *et al.* 1992). Third, there is a strong coupling between Northern and Southern hemispheres that is

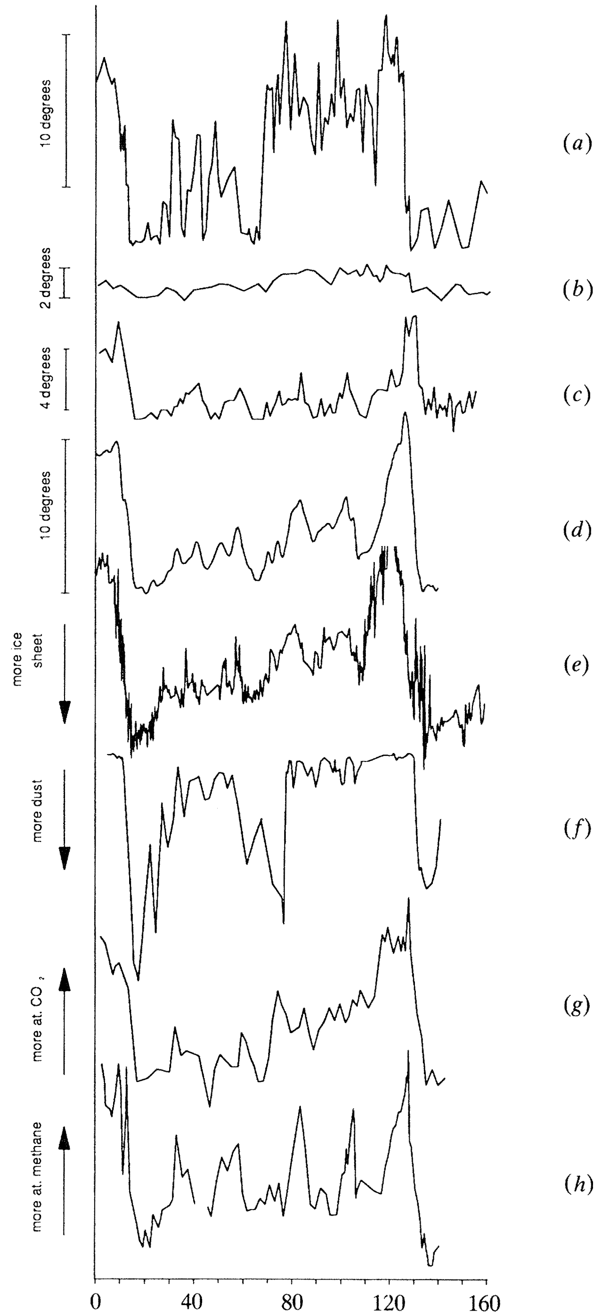

Figure 1. Records covering the past 160 ka. (*a*) Sea surface temperature 50°N (Ruddiman & McIntyre 1984); (*b*) sea surface temperature 2°S (Le 1992); (*c*) sea surface temperature 44°S (Hays *et al.* 1976); (*d*) air temperature 78°S (Jouzel *et al.* 1987); (*e*) $\delta^{18}O$ core V19-30 (Shackleton & Pisias 1985); (*f*) dust in Vostok ice (de Angelis *et al.* 1987); (*g*) atmospheric CO_2 (Barnola *et al.* 1987); (*h*) atmospheric methane (Chappellaz *et al.* 1990). (*d*), (*f*), (*g*) and (*h*) are replotted on the timescale of Shackleton *et al.* (1992).

still not well understood (Broecker & Denton 1989).

Imbrie *et al.* (1992) have examined a large number of records from various components of the climate system and have shown that in each record variance is concentrated in the frequency bands of changing eccentricity of the earth orbit; obliquity of the earth rotational axis; and climatic precession. Figure 1 shows short sections (160 ka) of a few records that exemplify the data examined by Imbrie *et al.* (1992).

The records shown in figure 1 have, in common, that all appear to show extreme 'glacial' values at about 20 ka BP. However, one of the most important aspects of the study of glacial cycles emerges from the study of phase relationships. In each of the three frequency bands mentioned, a characteristic time constant, detected from a phase lag of that component with respect ot the forcing, can be determined. The best example of such a time constant is, of course, the ice sheets themselves (Weertman 1964). A continental scale ice sheet has a long time constant, both because annual accumulation on the surface is very small in relation to overall thickness, and because the response of the underlying lithosphere to changing load is slow due to the very high viscosity.

Figure 1 shows two atmospheric records: the concentration of carbon dioxide (Barnola *et al.* 1987) and of methane (Chappellaz *et al.* 1990). This phase relationships of these records are in dispute because the measurements are made in ice cored from central Antarctica rather than in marine sediments. There are uncertainties both in the age of the ice, and in the age of the air that ultimately becomes trapped in the ice as the space between the snow crystals eventually forms isolated bubbles (Sowers *et al.* 1991; Petit *et al.* 1990; Shackleton *et al.* 1992). Shackleton & Pisias (1985) working with a marine proxy for atmospheric carbon dioxide, provided evidence that the phase atmospheric CO_2 is intermediate between that of the insolation forcing, and that of the major ice sheets. We do not understand the origin of these CO_2 variations, but it is certain that the ocean is involved through circulation changes and/or nutrient redistribution (Broecker 1982; Boyle 1988; Shackleton & Pisias 1985; Mix *et al.* 1991). It is not at present known whether the concentration of atmospheric CO_2 would have varied over such a large range (almost a factor of two) in the absence of cyclic continental glaciation.

On a longer timescale, atmospheric CO_2 may have varied over an even larger range. Many claims for significantly higher levels in the distant geological past are purely hypothetical but Berner *et al.* (1983) attempted to model the long-term history of atmospheric CO_2, obtaining values a factor of ten higher than today for the Mesozoic. Bender (1984) laid out the basis for determining past CO_2 levels, but concluded that the exercise could not be completed, since he could not put constraints on the dissolved CO_2 in the ocean. However, Shackleton (1985) showed that a constraint is available through carbon isotope data, arguing that at least for the past 10^8 years atmospheric CO_2 has not exceeded about twice its present level. This approach was expanded with elegant nomograms by Berger & Spitzy (1988); these workers suggested as an upper limit, a level 2.5 times greater than today for the Eocene and possibly 10 times greater for the Cretaceous (lack of $\delta^{13}C$ data limits our ability to put a tighter constraint on the Cretaceous level). More recently an independent approach has emerged through the control that pCO_2 exerts on $\delta^{13}C$ fractionation during photosynthesis (Rau *et al.* 1989; Jasper & Hayes 1990). It is likely that the combination of the two approaches will at least provide a low-

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resolution record of atmospheric CO_2 that is more tightly constrained than what is available at present.

The origin of the natural CH_4 variations should probably be attributed to changing input from low latitude wetlands (Chappellaz *et al.* 1990). The character of the record is consistent with a direct forcing by low-latitude insolation forcing in the same manner that the monsoon system appears to respond, particularly if expressed on a timescale derived from deep-sea sediment records. The methane record does not suggest a coupling to ice sheet dimensions. Thus, atmospheric methane concentration must be a potential feedback on all geological timescales. In the context of this volume, the important point is that both atmospheric CO_2 and CH_4 have varied in response to insolation forcing over a range of the order of a factor of two. Thus, their concentrations are not functions only of the slowly evolving boundary conditions. Chappellaz *et al.* (1990) argue that, taking into account chemical feedbacks, the combined effect of the observed natural variations in the concentrations of CO_2 and CH_4 in the atmosphere is equivalent to a 2.3°C forcing on global climate.

The important conclusion is that during the past million years, with essentially 'fixed' boundary conditions, climate states characterised by continent-scale ice sheets, by abundant lakes across the present Sahara, by Hippopotamus basking in the Yorkshire Dales, alternated on timescales of only thousands or tens of thousand years. Although enormous progress has been made in understanding and modelling these climate states and transitions between them, it is extremely unlikely that they would ever have been predicted by a climate model. Overall feedback factors approach a factor of ten, but even now some features such as generating the snowfall on North America needed for building ice sheets on a timescale that is out of dispute, have eluded modellers.

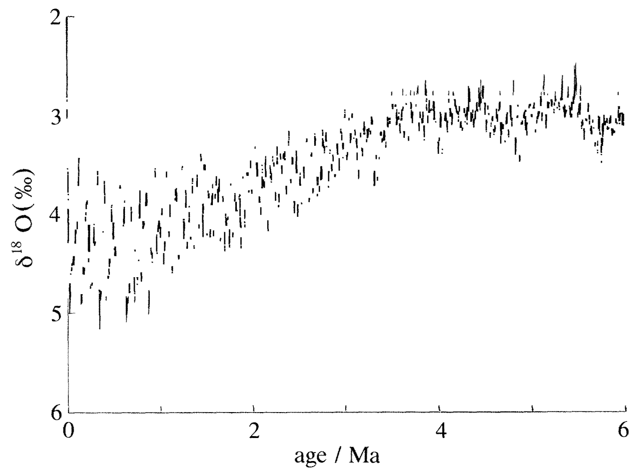

Figure 2. Composite oxygen isotope record for the past six million years. Analyses are benthonic foraminifera from core V19-30 (0–0.34 Ma; Shackleton & Pisias 1985), ODP Site 677 (0.34–1.7 Ma; Shackleton *et al.* 1990), and ODP 846 (1.7–6.0 Ma; Shackleton *et al.* 1993), plotted at 0.003 Ma intervals.

3. LONG-TERM TRENDS IN THE CLIMATE SYSTEM

Several workers in recent years have attempted to model the evolution of glacial cycles through the last few million years. These studies have been motivated by the increasing detailed records that have been developed of climate variability. Figure 2 shows a recent and very detailed record for the past six million years (Shackleton *et al.* 1993). It is visually obvious that the record is not statistically stationary. The character of the insolation forcing has not changed to a significant extent over the past few million years, but the response to that forcing has changed enormously. The coldest extremes have all been experienced during the past million years. Less is known of the extreme warm episodes. We draw attention in particular to evidence from Antarctica showing that at some time about three million years ago there was a significant amount of vegetation on that continent (see, for example, Webb & Harwood 1991). In northern Greenland at 82°N there is extensive evidence for the former presence of a boreal forest in deposits with an estimated age of about 2 Ma (Funder *et al.* 1984). Although this age estimate could be too young, the deposits are certainly not older than Pliocene and it is possible that exceptionally warm conditions prevailed at the same time in Antarctica and northern Greenland. There is considerable evidence for a high sea-level stand in the mid Pliocene suggestive of a time of reduced glaciation in Antarctica. Cronin (1988) discusses the evidence for a considerably warmer North Atlantic at the time of this marine high-stand.

One factor that has been discussed extensively is mountain uplift. On the one hand, modelling studies indicate that the Tibetan Plateau is indeed a significant factor in determining the climate that the present solar input maintains (Ruddiman & Kutzbach 1989; 1991). On the other hand, Raymo *et al.* (1988) and Raymo & Ruddiman (1992) have discussed the major impacts that mountain building must have had on ocean chemistry. Unfortunately, there is no agreement as to the history of the Tibetan Plateau, so that we cannot accurately test the hypothesis that the non-stationarity evident in figure 2 derives from mountain uplift. Another factor that has been invoked is the closure of the marine connection across Central America, which probably occurred about three million years ago. However, determining the geological history of this type of connection requires very accurately dated stratigraphic records for both marine and terrestrial sections either side of the connection. For the more remote geological past the detailed history of such gateways cannot be well constrained. Again, modelling shows that this particular connection has a significant effect on ocean circulation (Maier-Reimer *et al.* 1990), but we are still some way from having any convincing match between predicted climatic effects of this closure and their verification in the geological record.

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4. CONCLUSIONS

During the past million years, climate has typically undergone enormous changes within intervals of a few thousand years, under the influence of very subtle changes in the seasonal and latitudinal distribution of the incoming solar energy. We cannot even model the atmospheric circulation in states other than today's; even less can we model the important interactions with the ocean, the cryosphere, the biosphere and the lithosphere that have played a key part in this relatively recent climatic variability. Models of climate of the distant geological past may lead us to seek new geological data, but we should be very careful not to imagine that their outputs actually describe the operation of the climate system in the past.

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