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Modelling of Palaeoclimates: Examples from the Recent Past

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Modelling of palaeoclimates: examples from the recent past

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SUMMARY

Three-dimensional general circulation models have been used in equilibrium studies of past climates by fixing the slowly changing parts of the climate system (orbital parameters, ice sheets, atmospheric composition). Results are presented from studies of the mid-Holocene, the last glacial maximum and the initiation of the last ice age.

1. INTRODUCTION

Over the last two and a half decades, advances in computing power have made possible the development of detailed three-dimensional global models of climate (general circulation models, or GCMS). These models are based on physical principles (see for example, Schlesinger 1988; Trenberth 1992), have been validated extensively against present climate, and have shown some skill in reproducing the main features of the general circulation of the atmosphere (Houghton *et al*. 1990, 1992). Averaged over areas of 106 km^2, typical seasonal mean errors are a few degrees Celsius in surface temperature, and several tens of per cent in precipitation (Houghton *et al.* 1990). The validation of ocean GCMS is less complete, due partly to there being less observational data available, and the inherently smaller scale of synoptic disturbances in the oceans (10–100 km as opposed to 1000 km in the atmosphere).

The main application of such models has been in attempts to understand and predict the climatic effects of increases in greenhouse gases (for example, Mitchell 1989; Houghton *et al.* 1990, 1992). Although there is broad agreement between different models on the nature of changes at the largest scales, there is poor agreement on regional scales, and the response of the models is often dependent on the parametrization of processes not resolved on the models' grid.

General circulation models have been applied extensively to past climates (see, for example, reviews by Hecht 1985; Saltzman 1990). At present, the length of GCM experiments is limited to decades by the availability of computer resources, so we cannot yet attempt to simulate the evolution of climate over geological timescales. Instead, it is generally assumed that on short timescales (hundreds of years) climate is in approximate equilibrium which is determined uniquely by the state of the slower changing components of the system, such as the major ice sheets, atmospheric composition and the Earth's orbital parameters. These components, or 'boundary conditions', are prescribed and the response of the faster changing components of climate, such as cloud, sea-ice and ground wetness are simulated by the model. This approach gives 'snapshots' of specific times in the past. It is assumed that the final equilibrium is independent of the initial conditions. Considerable progress has been made in producing synoptic data for specific periods from the last glacial maximum to present (see for example, Climate Long Range Investigation Mapping and Prediction project members 1976, 1981 (henceforth referred to as (CLIMAP 1976, 1981)); Co-operative Holocene Mapping Project (COHMAP) members 1988). Note, however, that there are problems of interpretation when comparing model grid values, which are representative of areas of 105 km, with observational data from specific sites, even with instrumental data for our present climate (Reed 1986).

If the changes in boundary conditions are well established, the model can be used to determine the accompanying changes in atmospheric circulation, frequency of storms and other parameters not easily derived from field data. For periods for which the changes in boundary conditions are not well established, one can investigate the effect of changing different boundary conditions singly or in different combinations. By comparing the simulated and reconstructed climates, one may gain some insight into the causes of the change in climate, and the accompanying changes in circulation. Model simulations may help to highlight where competing theories of change diverge. This would allow one to optimize the choice of site to obtain data to discriminate between different hypotheses.

In principle, one can use the 'snapshot' approach to determine whether or not the simulated climate is in long term equilibrium with the prescribed boundary conditions. If one can determine the rate of change of the slowly changing boundary conditions from equilibrium experiments in which the boundary conditions are prescribed, it may be possible to parametrize those rates of change as a function of the boundary

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conditions themselves. One could then model the evolution of the slowly varying components of climate in a much simplified model.

2. MODELS

In atmospheric GCMs, prognostic equations governing the conservation of momentum, heat, mass and water substance are solved numerically over the globe at a set of discrete levels in the vertical. Current atmospheric models use a horizontal grid which is typically 300 km or 30 spectral waves for climate applications, and 10 to 20 vertical levels (see, for example, table 1). The largest source of uncertainty is associated with the parametrization of processes not resolved on the numerical grid (for example, cloud, radiation, cumulus convection, surface evaporation and surface drag). For some processes (for example, radiative transfer) parametrizations are based on the underlying mathematical theory. However, for many processes, no mathematical basis has been established, so parametrizations are developed from observational studies, laboratory experiments, numerical experiments on a fine numerical mesh or sensitivity experiments, as appropriate. Ocean GCMs are constructed in a similar way to atmospheric models, though the parametrization of physical processes is generally simpler.

When atmospheric and oceanic models are coupled together, errors in both components lead to the accumulation of substantial errors in the simulation of sea surface temperature and sea-ice extents. In order to produce a realistic simulation of present climate, some investigators have prescribed fluxes of heat and water at the ocean surface (Manabe & Stouffer 1988; Sausen *et al.* 1988), it being argued that climate change can be regarded as a perturbation about a basic state. These prescribed fluxes keep the same value in anomaly experiments. Unfortunately, the corrective fluxes required are large in certain regions (notably in areas of strong sea temperature gradients).

3. SIMULATIONS OF THE MID-HOLOCENE

The Holocene (12000–0 years before present) is of particular interest for a variety of reasons. First, good data is plentiful, largely as a result of the COHMAP project (COHMAP members 1988). Second, the known changes in boundary conditions were simple (see, for example, Kutzbach & Street-Perrott 1985). Changes in the Earth's orbital parameters (increased obliquity, and a change in perihelion from northern winter to northern summer) led to increases of up to 8% in Northern Hemisphere summer insolation, and decreases in winter, giving an enhancement of the seasonal cycle. At 9 ka BP, there was still a substantial Laurentide ice-sheet, but by 6 ka BP, this had virtually disappeared. Most simulations have been for 9 ka BP, when the changes in insolation were largest, or 6 ka BP, when the ice sheets were similar to present, and there is more data. Results for 6 ka BP are similar to those at 9 ka BP, but less pronounced (Kutzbach & Guetter 1986). Here we summarize results for both periods.

The increased insolation in the boreal summer produces a general warming, especially in the Northern Hemisphere. The continental surface warms relative to that of the ocean (figure 1*a*) because because the ocean temperatures are prescribed (or in models with an interactive ocean – see table 1 – because of the larger oceanic thermal inertia). The enhanced warming of the atmospheric column over land leads to a lowering of surface pressure (figure 1*b*), and stronger low level convergence from the oceans near the surface. This increases the precipitation associated with the Northern Hemisphere monsoons over India, the Sahel and Venezuela (figure 1*c*). The ascent over the northern continents is accompanied by subsidence over the neighbouring oceans, a drying of the marine atmosphere and a reduction in cloud cover. In short, the enhancement of the seasonal cycle of radiation leads to more vigorous Northern Hemisphere monsoons.

The increased land sea contrast produces stronger monsoon flow over the northern Indian Ocean in summer. Luther *et al.* (1990) and Bigg *et al.* (1991) used surface winds from atmospheric model simulations to drive mesoscale ocean models of the Indian Ocean. Both found increased and more extensive upwelling in the Arabian sea in summer, consistent with palaeoclimatic indices (for example, Prell & Kutzbach 1987).

In northern winter, insolation is reduced, cooling the continents, increasing surface pressure in the tropics and reducing precipitation over tropical land (see, for example, Kutzbach & Guetter 1986; Mitchell *et al.* 1988). The ground wetness (figure 1*d*) increases in the northern subtropics and reduces in mid-latitudes (Gallimore & Kutzbach 1989; Mitchell 1990). This is reasonably consistent with estimates of lake levels for the period (Street-Perrott & Roberts 1983).

In simulations with a simple ocean, the sea temperature changes lag the land changes by two months or so. The extra solar heat in summer is stored in the ocean and released in late autumn and early winter, delaying the onset of winter cooling over the northern continents. The changes in the Earth's orbital parameters over the last 125 ka have altered the seasonal distribution of insolation, but not the total amount of radiation reaching the top of the atmosphere over the year. Thus some nonlinearity of response is required to produce an annual mean warming (or cooling). The reduction in cloud over the northern oceans in summer enhances the warming of the oceans. The reductions in cloud (and sea-ice) increase the storage of heat in the mixed layer in summer, and further delay the winter cooling over the continents. Indeed, Mitchell *et al.* (1988) found that the northern extratropics were warmer throughout winter, despite reduced insolation in that season, and there was a global annual mean warming of 0.5 K.

The simulated climate at 9 ka BP is sensitive to changes in other boundary conditions. Mitchell *et al.* (1988) found that including an idealized Laurentide ice-sheet at 9 ka BP led to cooling, not warming over the northern extratropics in winter, and reduced the

Table 1. *GCM simulations of the Holocene*

(Resolution is either degrees (latitude by longitude) or number of spectral waves, R or T for rhomboidal or triangular truncated respectively. L indicates the number of levels. ML = mixed layer. SST = sea surface temperature. ka = 1000 years before present).

study	period	seasonal cycle	atmosphere horizontal and vertical resolution	ocean, or SSTs	land ice	ice-free land albedo	length of experiment
Kutzbach & Otto Bliesner (1982)	9 ka	yes	R10, L5	present	present & 9 ka	present	14 months
Kutzbach & Guetter (1986) (also Prell & Kutzbach 1987)	3, 6 ka 9 ka 12 ka[a]	Jan, Jly	R15, L9	present 12 ka	present 9 ka 12 ka	present 12 ka	450 days (Jan) 150 or 450 (July)
Rind *et al.* (1986)	11 ka	yes	8×10, L9	present & 18 ka	present & 18 ka	present	4 years
Kutzbach & Gallimore (1988)	9 ka	yes	R10, L5	ML	present	present	9 years[b]
Mitchell *et al.* (1988) (see also Mitchell 1990)	9 ka	yes	5×7.5, L11	ML	present & 9 ka	present	13 years[b]
Street-Perrott *et al.* (1990)	9 ka	yes	5×7.5, L11	ML	present	9 ka	6 years[b]
Liao (1992)	6 ka	yes	5×7.5, L11	ML	present	present	10 years[b]
Mitchell & Hewitt (this paper)	6, 9 ka	yes	2.5×3.75, L19	present	present	present	5 years
Hall (personal communication)	6 ka	yes	T31, L19	present	present	present	in progress

[a] CO_2 levels were slightly reduced.
[b] Does not include time to spin-up equilibrium.

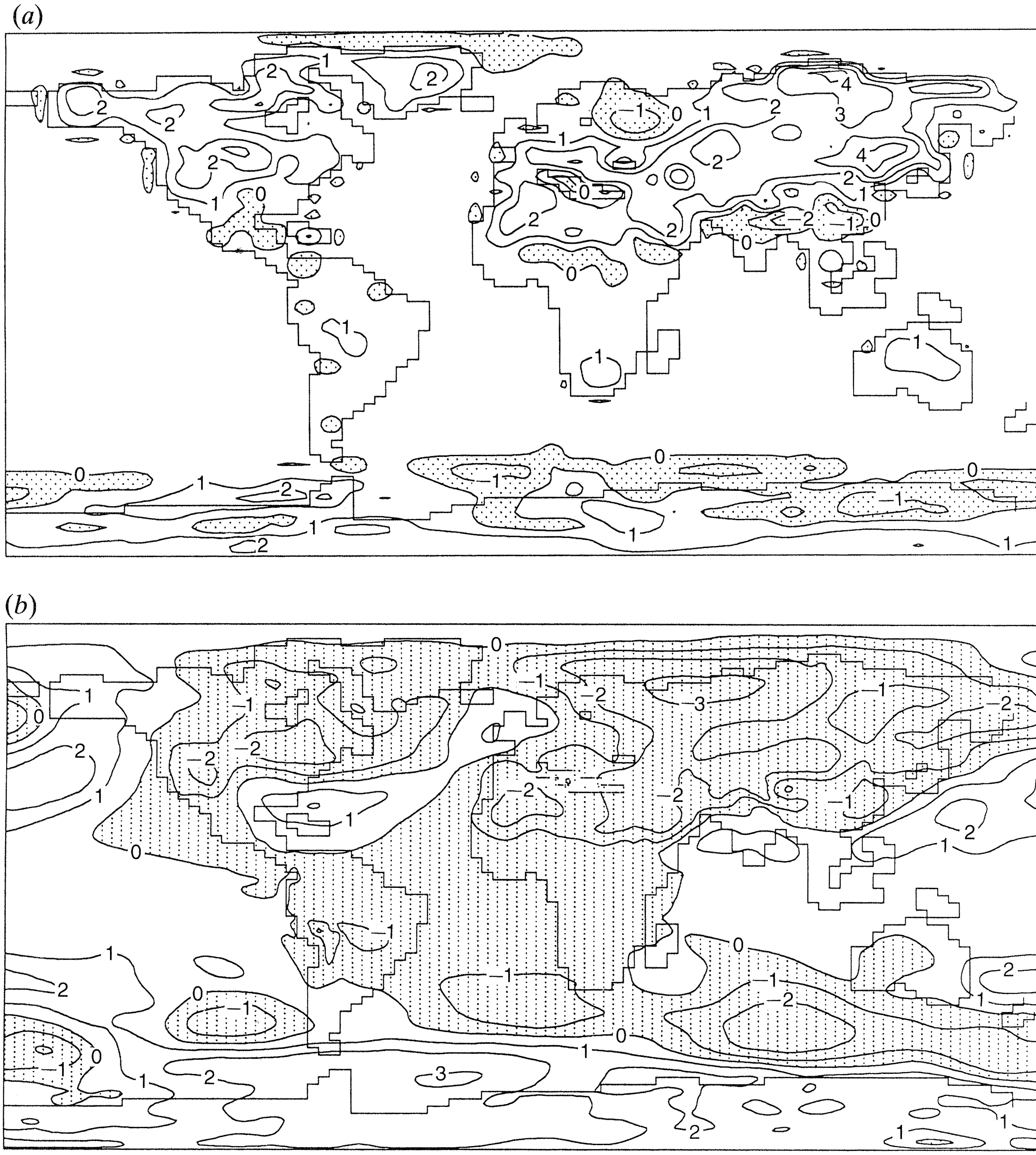

Figure 1. Changes (6 ka BP-present) during June, July and August simulated using a preliminary version (Cullen 1993) of the Meteorological Office unified forecast/climate model (see table 1). (*a*) Surface temperature . Contours every 1°C; areas of decrease stippled. (*b*) Surface pressure. Contours every 1 mb; areas of decrease are stippled. (*c*) Precipitation. Contours at 0, ±1, 2 and then every 5 mm per day; areas of decrease are stippled. (*d*) Soil moisture. Contours at 0, ±1, 2 and then every 5 cm; areas of decrease are stippled.

warming over land in July by 20%. Changes in vegetation reconstructed from palaeodata would give a lower surface albedo over much of the northern subtropics (Petit-Maire *et al.* 1988). Street-Perrott *et al.* (1990) found that reducing surface albedo over north Africa and Arabia enhanced the intensification of monsoon precipitation at 9 ka BP.

4. SIMULATIONS OF THE LAST GLACIAL MAXIMUM

The climatic changes during the last glacial maximum (about 18 ka BP) were much larger than at 6 or 9 ka BP. However, the cause of the changes is not well established. The Earth's orbital parameters, which

(*c*)

(*d*)

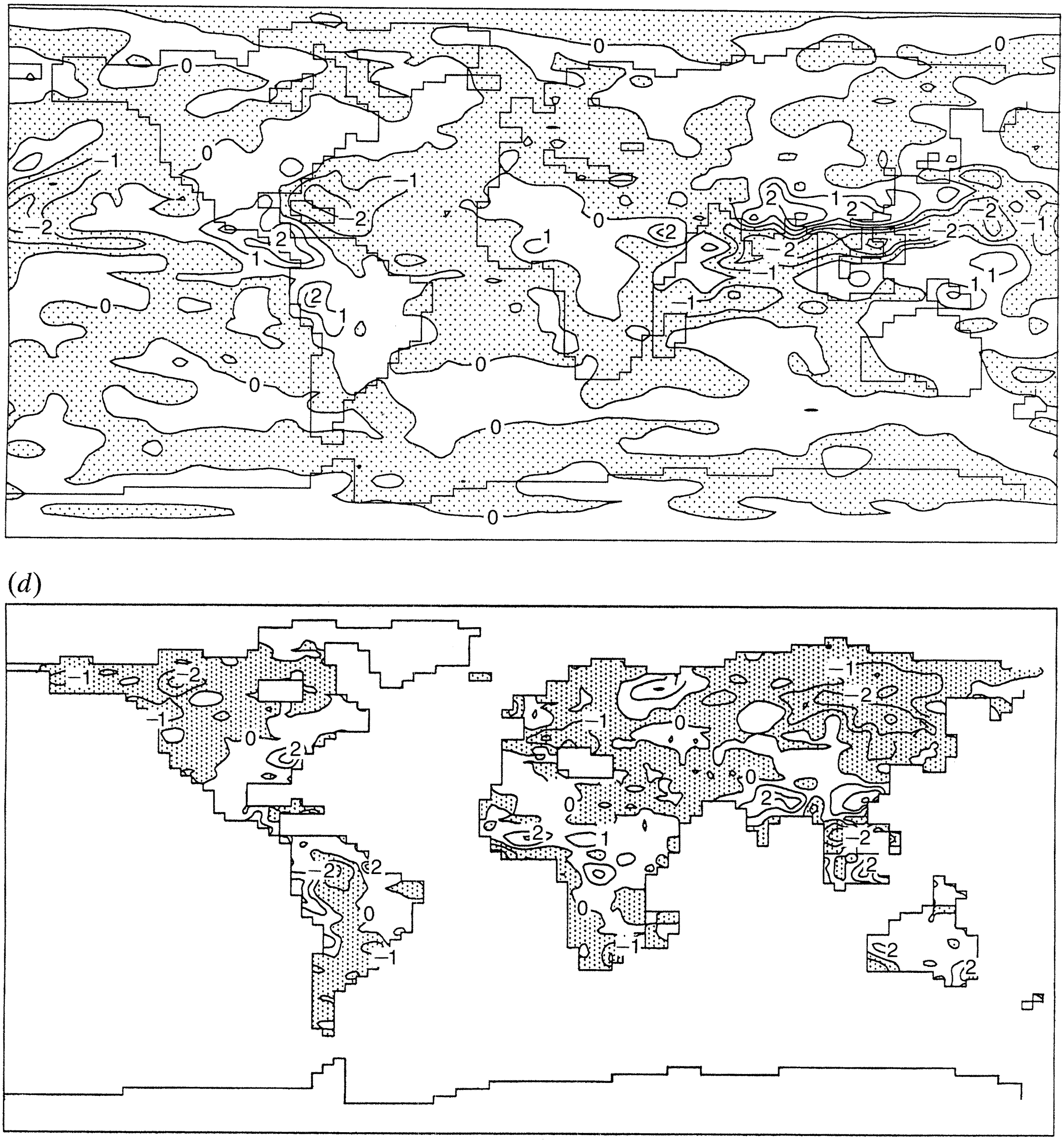

Figure 1 (*continued*)

appear to explain most of the differences in the mid-Holocene, were fairly similar to present (Berger 1978). The largest differences in the distribution of radiation occurred at the equinoxes, and changes in solsticial seasons in high latitudes were small. There were large ice sheets over North America and northwest Europe (Denton & Hughes 1981), sea temperatures were generally cooler, especially in the North Atlantic ocean, sea-ice was more extensive and snow-free land albedos were higher in much of the tropics and subtropics (CLIMAP 1976, 1981). CO_2 levels were less than two thirds their present value (Barnola *et al.* 1987).

Most experiments to date have taken into account the ice-sheets over the northern continents, with the changes in sea surface temperatures and sea-ice cover either prescribed or calculated in the model (see table 2). These changes lead to a simulated global mean cooling which, in general, is greater over land than the neighbouring ocean. This produces higher surface pressure over much of the northern extratropical continents, and reduced precipitation (for example, Manabe & Broccoli 1985*a*; Kutzbach & Guetter 1986; Rind 1987). The ice sheets displace the jet stream and the associated storm tracks southwards, particularly over North America, producing a marked reduction

Table 2. *GCM simulations of the last glacial maximum*

(Resolution is either degrees (latitude by longitude) or number of spectral waves, R or T for rhomboidal or triangular truncation respectively. L indicates the number of levels. ML = mixed layer. ka = 1000 years before present. SST = sea surface temperature. CL76(81) = CLIMAP 1976 (1981).)

study	seasonal cycle (period covered)	atmospheric horizontal & vertical resolution	ocean or SSTs[a]	CO_2	land ice	snow-free land albedo	orbital parms.	length of study
Williams *et al.* (1974)	Jan, Jly	5 × 5, L6	18 ka	0 ka	18 ka	18 ka	0 ka	80 days
Gates (1976)	yes (Jly)	4 × 5, L2	CL76	18 ka	18 ka	18 ka?	0 ka	92 days
Manabe & Hahn (1977)	yes (Jly)	2.5 × 2.5, L11	CL76	0 ka	0 ka	0 ka	0 ka	113 days
					18 ka	18 ka	0 ka	133 days
Hansen *et al.* (1984)	yes	8 × 10, L9	CL81	18 ka	18 ka	18 ka	18 ka	6 years
Rind & Peteet (1985)	yes	8 × 10, L9	CL81	0 ka	18 ka	same as	18 ka	6 years
			CL81 less 2°C	0 ka	18 ka	control	18 ka	
Rind (1987)	yes	8 × 10, L9	CL81	0 ka	0 ka	0 ka	18 ka	4 years
			CL81	0 ka	18 ka	0 ka	18 ka	4 years
			CL81	0 ka	(10 m)	0 ka	18 ka	5 years
					18 ka (full)			
Kutzbach & Guetter (1986)	Jan	R15, L9	CL81	18 ka	18 ka	18 ka	18 ka	450 days
(also Prell & Kutzbach 1987)	Jly			0 ka	15 ka	15 ka?	15 ka	450 days (Jan) & 150 (Jly)
Broccoli & Manabe (1987)	yes	R15, L9	50 m ML	0 ka	18 ka	0 ka	0 ka	8 years
(also Broccoli & Manabe 1985*a*,*b*)				0 ka	18 ka	18 ka	0 ka	6 years
				18 ka	18 ka	18 ka	0 ka	8 years
Lautenschager & Hertereich (1990)	yes	T21, L16	CL81	18 ka	18 ka	18 ka	18 ka	6 years
Valdes (personal communication)	Jan, Jly	T42, L19	CL81	18 ka	18 ka	18 ka	18 ka	450 days in progress
	yes		CL81	18 ka	18 ka	18 ka	18 ka	
Manabe (1991)	annual mean	R15, L9	ocean GCM	18 ka	18 ka	18 ka	0 ka	8 years[b]

[a] CLIMAP SSTs were reconstructed for February and August; where a full annual cycle has been used, it has been obtained by interpolation.
[b] 1250 years in the ocean, 34000 years in the deep ocean.

Table 3. *Investigations of initiation of ice sheets*

(ka = 1000 years before present; mod = 5 times the change from 114 ka and 116 ka; CL81 = CLIMAP (1981).)

study	sea temperatures or ocean model	CO_2 (p.p.m.)	land ice	orbital parameters	length of experiment
Royer *et al.* (1983)	present	present	present	125 ka, 115 ka	1 year
Rind *et al.* (1989)	present	present	present	116 ka, mod	all 4 years
	mixed layer	present	present	116 ka, mod	
	mixed layer	present, −70 p.p.m.	10 m	mod	
	CL81, CL81 minus 2 K	−70 p.p.m.	10 m	mod	
Mitchell & Hewitt (this paper)	present	present	present	125 ka	5 years
				115 ka	
Oglesby (1990)	mixed layer	100 p.p.m., 200 p.p.m.	present	present	15–20 years
	present	330 p.p.m.	1 m snow	present	2 years

in precipitation immediately south of the ice sheets in most models. Westerly flow is also diverted around the north of the Laurentide ice sheet and then south between North America and Greenland. This cold southward outflow would have helped to maintain the eastern flank of the ice sheet, and cooled the northern Atlantic ocean. Even so, the simulated energy balance indicates that the Laurentide and Eurasian ice sheets would not be in equilibrium, but would melt, particularly at lower elevations (Manabe & Broccoli 1985*a*; Rind 1987).

Broccoli and Manabe (1987) have estimated the separate effects of the changes in land-ice, CO_2 and vegetation (through surface albedo) using an atmospheric model coupled to an oceanic mixed-layer model. The effects of the changes in vegetation are small. The increase in land-ice produces a large cooling in the Northern Hemisphere only. Reducing CO_2 has the largest effect, producing a cooling in both hemispheres, but particularly in the Southern Hemisphere where there is a strong sea-ice–temperature feedback.

The cooling of the low latitude ocean simulated by Broccoli & Manabe (1987) is larger than that reconstructed from palaeoclimatic data (CLIMAP 1981), particularly in the subtropics. For example, near 25°N, the zonal mean cooling is almost 3°C whereas the CLIMAP data gives a value of about 0.5°C. Manabe & Broccoli (1985*b*) found that changing from prescribed to model-generated cloud cover did not remove the discrepancy between simulated and CLIMAP sea surface temperatures.

The cooling deduced from data on mountains at these latitudes is also much larger than one would expect from nearby sea temperature changes (Rind & Peteet 1985). They find that a simulation with CLIMAP (1981) sea temperatures reduced everywhere by 2°C gives a better fit to the observed temperature reconstruction from tropical mountains, and a radiation balance at the top of the atmosphere which is closer to equilibrium than using the CLIMAP (1981) values.

Lautenschlager *et al.* (1992) have attempted to simulate the circulation of the ocean at the last glacial maximum, using CLIMAP (1981) sea surface temperatures, and wind stress and water fluxes from the atmospheric simulation of Lautenschlager and Herterich (1990, see table 2). The simulated changes are in broad agreement with palaeodata in the North Atlantic, but apparently overestimate the intensity of overturning in the northwest Pacific. The authors speculate that this may be due in part to shortcomings in the CLIMAP (1981) sea-ice extents.

Manabe & Stouffer (1988) produced two stable equilibria in a coupled ocean-atmosphere GCM using present day boundary conditions (insolation, land ice, CO_2 and snow free land albedos). One equilibrium was similar to present, with deep water formation in a relatively warm, salty North Atlantic. In the other equilibrium, the northern North Atlantic became fresher and colder, and meridional overturning almost ceased. The authors note that the second circulation bore some similarity to that believed to have occurred during the Younger Dryas. Manabe (1991) has also reported on the results from a coupled model simulation with ice-age boundary conditions (as in Broccoli & Manabe 1987).

5. INITIATION OF ICE SHEETS

The causes of glacial–interglacial cycles have yet to be established. One possibility is that changes in the Earth's orbital parameters may have led to an extension of perennial snowfields over northern North America and northwestern Europe, probably by reduced summer insolation. This may have been sufficient to build the major ice sheets, especially if augmented by other processes.

Changes in summer insolation in high northern latitudes around the beginning of the last glaciation reached a maximum around 125 ka BP, a minimum near 115 ka BP and a further maximum near 105 ka BP. Royer *et al.* (1983), comparing a simulation for 115 ka BP with one for 125 ka BP, found that temperatures fell by several degrees and annual mean soil moisture increased in the regions of the north American and European ice sheets. However Rind *et al.* (1989; see also table 3) and Mitchell & Hewitt (see table 3) were unable to produce permanent snowfields in these areas through changes in orbital parameters alone. Rind *et al.* (1989) produce stable permanent snow-ice fields in these regions only by including a 10 m ice sheet, reducing CO_2 concentrations, and using full ice-age sea-ice extents and sea temperatures reductions. The reductions in temperature, CO_2 and surface albedo are believed to have accompanied or even lagged the growth of the ice sheet: how could they then have contributed to the initiation of the ice sheet? Oglesby (1990) found that by reducing CO_2 to 200 p.p.m. or by prescribing an initial snow depth of 1 m over the areas of interest, he was able to obtain permanent snow cover over part of the region occupied by the major ice sheets. Note that an increase of 1 mm per day (water equivalent) in snow accumulation, or a reduction of 4 W m^{-2} heat available for melting, are equivalent to the growth of a 1 km ice-sheet in 1000 years; it is unlikely that current models can attain that accuracy on a regional scale in high latitudes.

6. CONCLUDING REMARKS

One of the main problems arising from the reconstruction of the last glacial maximum is the apparent discrepancy between estimates of sea temperature change in low latitudes, and the estimate from terrestrial data on nearby mountains. Current climate models suggest that the tropical sea temperatures should have been substantially lower than suggested by CLIMAP (1981). If not, models may be too sensitive in the tropics, and estimates of global warming due to increases in greenhouse gases may be exaggerated.

There are many areas for improvement in current models. There is a large range of sensitivity (1.9–5.4°C to doubling CO_2) which is mainly due to problems of representing cloud and associated processes (Cess *et al.* 1989; Senior & Mitchell 1993), and there is little confidence in the simulation of regional detail. Most studies to date have not taken into account changes in the ocean circulation and, as noted above, coupled ocean–atmosphere models require corrective adjustments to prevent the simulated climate from drifting unrealistically. By comparing results from different models, it may be possible to distinguish those aspects of the simulated response which are robust, and possibly more dependable, from those which are dependent on the model formulation. A step in this direction is the Palaeoclimate Modelling Intercomparison Project (PMIP) which will initially concentrate on simulations for 6 ka BP and the last glacial maximum.

I am grateful to Chris Hewett for carrying out the experiments for 6, 9, 115 and 125 ka BP.

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