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The impact of dynamic sea-ice on the climatology and climate sensitivity of a GCM: a study of past, present, and future climates

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Abstract We assess two parametrisations of sea-ice in a coupled atmosphere–mixed layer ocean–sea-ice model. One parametrisation represents the thermodynamic properties of sea-ice formation alone (THERM), while the other also includes advection of the ice (DYN). The inclusion of some sea-ice dynamics improves the model's simulation of the present day sea-ice cover when compared to observations. Two climate change scenarios are used to investigate the effect of these different parametrisations on the model's climate sensitivity. The scenarios are the equilibrium response to a doubling of atmospheric CO_2 and the response to imposed glacial boundary conditions. DYN produces a smaller temperature response to a doubling of CO_2 than THERM. The temperature response of THERM is more similar to DYN in the glacial case than in the $2 \times CO_2$ case which implies that the climate sensitivity of THERM and DYN varies with the nature of the forcing. The different responses can largely be explained by the different distribution of Southern Hemisphere sea-ice cover in the control simulations, with the inclusion of ice dynamics playing an important part in producing the differences. This emphasises the importance of realistically simulating the reference climatic state when attempting to simulate a climate change to a prescribed forcing. The simulated glacial sea-ice cover is consistent with the limited palaeodata in both THERM and DYN, but DYN simulates a more realistic present day sea-ice cover. We conclude that the inclusion of simple ice dynamics in our model increases our confidence in the simulation of the anomaly climate.

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1 Introduction

It has long been recognised that changes in sea ice due to increasing greenhouse gas concentrations may have a large impact on the size of the equilibrium global warming through changes in surface albedo (e.g. Spelman and Manabe 1984, Ingram et al. 1989, Meehl and Washington 1990). Coupled ocean–atmosphere general circulation models (GCMs) have become an important tool for investigating the sensitivity of the climate system to increases in greenhouse gas concentrations. Many of these coupled GCMs have only included a representation of the simple thermodynamics of sea ice, and have ignored the important process of sea-ice advection (e.g. Houghton et al. 1990, 1992).

A number of parametrisations of sea-ice advection for use in GCMs have been developed (e.g. Flato and Hibler 1990, 1992). Pollard and Thompson (1994) performed a study using two versions of the GENESIS GCM with and without sea-ice dynamics. They found that the simulated present day sea-ice was far more realistic if sea-ice dynamics were included. They also found that the climate sensitivity to a doubling of CO_2 was reduced when they included the sea-ice dynamics. Rind et al. (1997) have shown that the sensitivity of the GISS mixed layer ocean model to a doubling of CO_2 depends on the areal coverage of sea ice in the control in the Southern Hemisphere, and on the thickness of sea ice in the control in the Northern Hemisphere. These two studies (Pollard and Thompson 1994; Rind et al. 1997) show that the response of a model depends on the reference state and highlight the potential importance of including sea ice advection in GCMs in order to realistically simulate a CO_2-induced global warming.

Recent experiments with coupled-ocean atmosphere GCMs at the Hadley Centre at the Met Office (Johns et al. 1997; Gordon et al. 2000) have incorporated a simplified representation of sea-ice dynamics in which ice is advected by the surface currents in the ocean model. This work will assess the effect on present-day climate (Sect. 3) and climate sensitivity (Sect. 4) of including this scheme for sea-ice dynamics using a version of the Hadley Centre atmospheric model coupled to a simple mixed layer ocean model (Sect. 2). The climate sensitivity is investigated in two climate change scenarios, one with atmospheric CO_2 concentrations doubled from present-day levels, and the other with boundary conditions representative of the last glacial maximum. A qualitative comparison is made between the model simulations of glacial sea-ice cover and some recent reconstructions of glacial sea-ice cover for the Southern Ocean and the North Atlantic (Sect. 5).

2 Model description

A version of the Met Office Hadley Centre atmospheric climate model is coupled to a simple mixed layer ocean model at a horizontal resolution of 2.5° in latitude by 3.75° in longitude. The 19 level atmospheric model (HadAM2b) is similar to that described by Johns et al. (1997), but includes changes to the model physics as described by Hewitt and Mitchell (1997). The mixed layer ocean model represents the effects of the thermodynamics of a 50-m deep well-mixed layer of water. This model of the mixed layer does not attempt to simulate ocean currents or the deep ocean. The surface temperature of the ocean in ice-free waters for simulations of the present day climate is maintained close to climatological values by the use of a seasonally varying additive heat flux which is diagnosed in a calibration experiment where sea surface temperatures (SSTs) are restored instantaneously back to climatological values. This heat flux accounts for the transfer of heat in the ocean due to ocean dynamics as well as model errors. The ocean temperature under ice is maintained at a value that gives an ocean to ice heat flux sufficient to produce a stable climatology of ice thickness and fractional cover (which represents an areal concentration) without the need for a specified correction to the ice thickness. Again this is done via a heat flux correction to the ocean temperature.

Experiments were performed for two model formulations that differed only in their treatment of sea ice. One included a zero-layer thermodynamic only representation of sea ice (henceforth referred to as THERM) and the second also included a simple representation of sea-ice dynamics (henceforth referred to as DYN). THERM is the formulation that was used in the Hadley Centre's first transient climate change experiment (Murphy 1995; Murphy and Mitchell 1995). The DYN sea-ice model has been used in subsequent Hadley Centre coupled models, HadCM2 and HadCM3 (Johns et al. 1997; Gordon et al. 2000, respectively). A detailed description of the sea ice model parametrisation (dynamics and thermodynamics) is given by Cattle and Crossley (1995). Unlike coupled ocean–atmosphere models, the sea ice in the mixed layer ocean model does not feedback on the hydrological cycle. The values of the heat flux were calculated in a five year calibration experiment for each of DYN and THERM.

2.1 Sea-ice thermodynamics

The thermodynamics of the model is based on the zero-layer model of Semtner (1976). A parametrisation of ice concentration is included, based on that of Hibler (1979). The ice concentration is not allowed to exceed 0.995 in the Arctic and 0.980 in the Antarctic since completely unbroken ice cover on the scale of the model grid boxes is rarely observed in reality, even in pack ice. Ice thickness can be increased by the formation of "white ice" (Ledley 1985) where the weight of snow forces the ice-snow interface below the water line. All rainfall is assumed to reach the ocean through leads, while snowfall is allowed to accumulate on the ice surface and contributes to the overall budget of the mixed layer ocean through assumed melt over the leads fraction.

Surface fluxes over the ice and leads fractions of each grid box, and surface temperatures, are calculated separately within the atmosphere component of the model, assuming a linear temperature profile in the ice/snow layer. The oceanic heat flux into the base of the ice is related to the temperature difference between the ocean top level and the base of the ice (assumed to be at freezing point of -1.8 °C), with a coupling coefficient of 20 Wm^{-2} K^{-1}.

Over sea-ice the surface albedo varies linearly from 0.8 for a surface temperature at or below 263.15 K to 0.5 for a surface temperature greater than 273.15 K. The albedo of bare ice and snow-covered ice are not discriminated.

2.2 Sea-ice dynamics

A simple parametrisation of sea-ice dynamics, based on Bryan (1969), is included in the second series of experiments (DYN). Ice thickness, concentration and snow depth are advected using the top layer ocean current, which is prescribed from values diagnosed in the HadCM2 coupled ocean–atmosphere model (Johns et al. 1997), and an upstream advection scheme. Ice rheology is only crudely represented by preventing convergence of ice once the ice thickness reaches 4 m. As in Murphy (1995), the penetrative part of shortwave radiation into leads is allowed to warm the ocean as it would at open ocean grid boxes, but all other surface heat fluxes into leads are split, so that a fraction of them proportional to the ice concentration is used to form or melt ice by the ice model. The remainder is absorbed in the ocean mixed layer.

2.3 Experimental design

Each of the models (THERM and DYN) has been used to simulate three climates: a control experiment to simulate the present-day climate (henceforth referred to as CON), an experiment identical to CON but with atmospheric CO_2 concentrations instantaneously doubled (henceforth referred to as CO2), and an experiment to represent the climate at the last glacial maximum 21 000 years before present (henceforth referred to as LGM). The latter climatic period necessitates several changes to the experimental setup of CON, atmospheric CO_2 is reduced from the control value of 323 ppmv to 230 ppmv to represent the fractional reduction in CO_2 levels from mid-Holocene to glacial levels; the model's land distribution and topography are modified to take into account the lower sea level and large continental ice sheets that existed at the last glacial maximum; and the Earth's orbital parameters are modified to be appropriate for the LGM (see Berger 1978 for a description). A more detailed description of these changes can be found in Hewitt and Mitchell (1997). The prescribed ocean currents used in the dynamic sea-ice model and the additive heat flux used in the mixed layer ocean model (which accounts for ocean heat transports) do not change in the anomaly simulations from the values used in the control simulations.

The length of each simulation is shown in Table 1. There is an adjustment, or "spin up" period that is designed to allow the atmosphere–mixed layer ocean–sea-ice coupled climate system to reach quasi-equilibrium. A longer spin up phase was needed for the LGM experiments. Results presented in the following section are averages over the "analysis period" after the "spin up", as listed in Table 1. The global average annual mean surface temperature of the control runs of DYN and THERM are stable over the analysis period (not shown).

3 Present-day climatology

In the Southern Hemisphere the area averaged sea-ice concentration is stable throughout both the THERM and DYN runs (not shown), and is much larger in THERM than in DYN (Table 2). However, the total area of grid boxes containing ice (the "ice extent") is about 30% smaller in THERM. The Southern Hemisphere sea-ice thickness in DYN has an annual average of about 0.4 m, thinner than in THERM (Fig. 1a and Table 2). In THERM there is a steady increase of Southern Hemisphere sea-ice thickness, with the annual average increasing from 0.8 m to 1.8 m over the 20 year run (Fig. 1a). This increase occurs as a result of thermodynamic growth over a small area around the Antarctic coast (an area less than 4% of the total Antarctic sea ice extent). In reality, ice formed here is continuously advected equatorwards preventing it from building up. This process is represented in DYN which does not produce a continual thickening of the sea ice. The ice extends considerably farther north in DYN, as is expected due to strong meridional currents moving ice away from the coast, reducing the compactness and thickness of ice close to the coast and increasing the extent of ice equatorwards, albeit very thin and broken. In Antarctic summer (Fig. 2e, f), more ice survives in THERM since the ice in the interior has become extremely thick and compact during the winter and hence is less prone to melting. The sea-ice distributions simulated by DYN are generally much more realistic than THERM when compared to observations (e.g. the SMMR satellite data in Gloersen et al. 1992) although DYN removes too much ice during the summer in the Weddell Sea. The SMMR satellite data has estimated that the total area of the Southern Hemisphere ocean surface that has at least 15% ice concentration (Gloersen and Campbell 1988) varies between a summer-time minimum of about 2 million km^2 and a winter-time maximum of about 15 million km^2 (Table 3). The DYN simulation produces a credible, although slightly too high, seasonal range for total sea-ice area (Table 3, calculated as the total area where the sea-ice concentration exceeds 15% in line with the estimates from satellite data). THERM however generally has too much of the ocean covered (Table 3, for example the minimum

Table 1 Length, in years, of the spin up period and the analysis period of the CON, CO2, and LGM experiments for both THERM and DYN

Model	Experiment	Length of spin up	Length of analysis
THERM	CON	10	10
	CO2	10	10
	LGM	18	20
DYN	CON	10	20
	CO2	10	20
	LGM	16	20

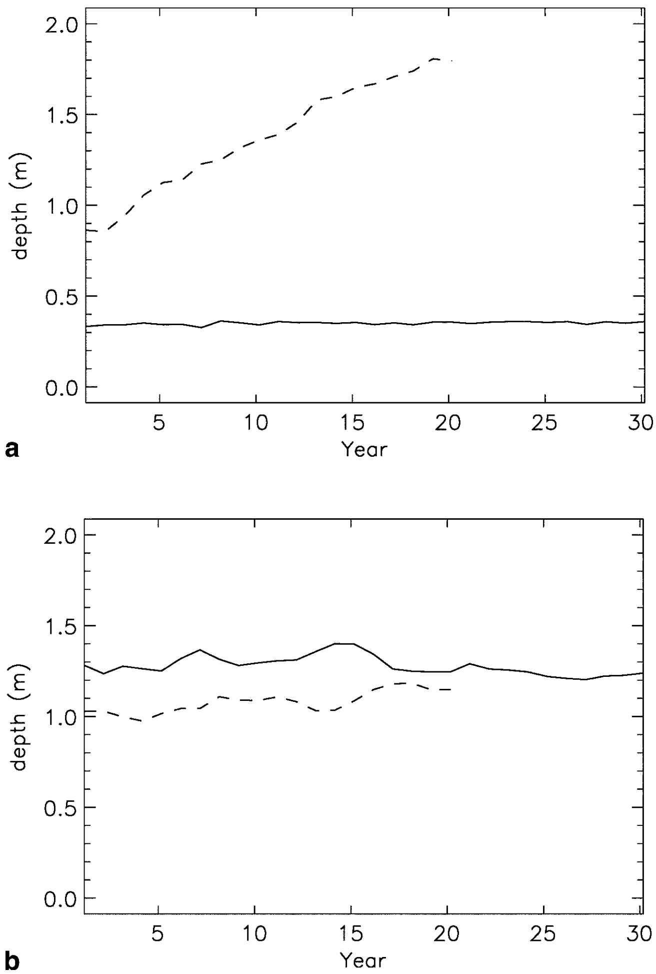

Fig. 1a, b Annual mean sea-ice thickness, in m, averaged over ice points (with an ice concentration > 0.0001) as a function of year from the control experiment. *Solid line* DYN control, *dashed line* THERM control. **a** Southern Hemisphere, **b** Northern Hemisphere

Table 2 Annual mean sea-ice thickness, in m, average areal concentration, and ice extent, in million km^2, for CON, CO2, and LGM averaged over all Arctic and Antarctic sea-ice points with an ice concentration >0.0001

	CON		CO2		LGM	
	DYN	THERM	DYN	THERM	DYN	THERM
Antarctic thickness	0.4 (0.2)	1.3 (0.7)	0.3	0.5	0.2	2.1
Antarctic concentration	0.3 (0.2)	0.5 (0.5)	0.3	0.5	0.3	0.5
Antarctic extent	32.1 (28.2)	24.8 (20.9)	27.9	15.4	36.6	30.5
Arctic thickness	1.2 (1.3)	1.0 (1.0)	0.8	0.7	3.3	4.9
Arctic concentration	0.6 (0.6)	0.6 (0.6)	0.5	0.6	0.4	0.4
Arctic extent	19.9 (10.4)	19.2 (10.2)	16.1	15.2	23.4	22.3

Figures in brackets are the values for CON if the LGM land-sea mask is used. Note that the average sea-ice concentration in the LGM experiments is relatively low compared to the control because much of the low latitude expanded sea ice is well broken with low fractional coverage area is about three times as large as the satellite estimate), largely due to the relatively high sea-ice concentrations compared to DYN.

In the Northern Hemisphere the ice extents and average ice concentrations are similar in the two experiments (Table 2). The average ice thickness in THERM is about 20% lower than in DYN, and the thicknesses are fairly stable in both runs (Fig. 1b). Areas of divergent currents will tend to thin the ice in DYN making it less compact and thinner than in THERM, most notably in the eastern Arctic Ocean (Fig. 3). In regions where the currents are convergent, such as adjacent to the northern coasts of Canada, Greenland, and Siberia, the opposite is true. The geographical constraint on ice extent in the Arctic basin means that over much of the basin the sea-ice is thicker and more compact in DYN.

DYN produces a more realistic distribution of sea ice than THERM with the thickest, and most compact, sea ice off the Canadian Arctic archipelago and Northern Greenland in good agreement with observations (see e.g. the submarine derived data of Bourke and McLaren 1992). When the ice is at its maximum cover, i.e. winter-time, the ice concentration is maintained close to 100% over much of the Arctic Ocean in both experiments (not shown), and the models are in good agreement with satellite-derived concentrations (e.g. Gloersen et al. 1992). The differences between the annual mean sea-ice distributions shown in Fig. 3 are largely due to differences that occur at the time of minimum ice cover, i.e. summer-time (Fig. 3e, f). DYN maintains thicker and more compact sea ice in the Beaufort Sea and off the Canadian Arctic archipelago and Northern Greenland, again in good agreement with the observations. This is largely due to the convergence of sea-ice advected from neighbouring regions. In THERM the ice thickness reduces fairly monotonically southwards from polar regions producing thinner less compact sea ice in the regions where the currents in DYN are convergent, and thicker more compact sea ice in regions where those currents diverge. Both models produce reasonable winter-time and summer-time total Arctic sea-ice area (Table 3) when compared to the satellite data, although the seasonal range is larger than the data suggests.

Table 3 Minimum and maximum ice area, in million km^2, from the DYN and THERM control simulations and satellite estimates from Gloersen and Campbell (1988) averaged over all Arctic and Antarctic sea-ice points with an ice concentration >0.15, to be consistent with the satellite-derived observational datasets. The minimum and maximum ice area from the control simulations are calculated from monthly mean values

	Antarctic		Arctic	
	Minimum	Maximum	Minimum	Maximum
DYN	1.7	18.5	5.1	15.7
THERM	6.0	18.8	5.1	16.1
Satellite estimates	2	15	6	13

4 Climate sensitivity

4.1 Surface temperature response

The seasonal cycles of the surface temperature responses of THERM and DYN to both a doubling of CO_2 and glacial boundary conditions have a similar zonal mean pattern (Fig. 4), with a large response at high latitudes, especially in winter, and a smaller response in the tropics and over the Arctic in summer.

The high latitude amplification of the temperature response is seen in most model simulations of CO_2 induced climate change (e.g. Houghton et al. 1990; Kattenberg et al. 1996), as well as model simulations of the response to glacial boundary conditions (e.g. Broccoli and Manabe 1987). The amplification is a result of the well-documented positive sea-ice albedo feedback (e.g. Ingram et al. 1989). The response is enhanced in the winter half of the year due largely to the higher static stability of the atmosphere which produces a relatively shallow surface layer to warm or cool. However, summer-time changes to sea ice and oceanic heat storage affect the subsequent winter-time growth of sea ice and enhance the winter response. For example, in the case of doubling CO_2, the warmer summer-time oceanic mixed layer delays the formation of sea ice in autumn and winter and reduces the amount of ice that eventually forms. This results in an enhanced warming during autumn and winter due to the much larger heat fluxes from the thinly ice-covered or uncovered warm ocean into the cold atmosphere. During the Arctic summer the surface temperature over sea-ice points is at melting point in the control and so the change in heating at the surface does not change the surface temperature of the sea ice, but is instead absorbed by the mixed layer or used to change the sea-ice extent and thickness.

The global average surface temperature response to a doubling of CO_2 is larger in THERM than in DYN, arising mainly from differences in the Southern Hemisphere (Table 4). However, Table 4 shows that the global average surface temperature response to LGM boundary conditions is more similar in THERM and DYN, as are the average surface temperature response for each hemisphere, and for land and ocean points only.

We shall now concentrate on the response over the ocean since the only difference between the THERM and DYN models is the treatment of sea-ice advection. The zonally averaged surface temperature response over the ocean of THERM compared to DYN (Fig. 5) for both a doubling of CO_2 and LGM boundary conditions raises two interesting questions that we shall discuss in more detail. Firstly, why does THERM exhibit a higher sensitivity to a doubling of CO_2 than DYN? Secondly, why is the Southern Hemisphere response of THERM more similar to DYN in the LGM experiments than in the CO2 experiments. In particular why does DYN show a larger sensitivity than THERM close to Antarctica? These questions imply that local feedbacks and the dis-

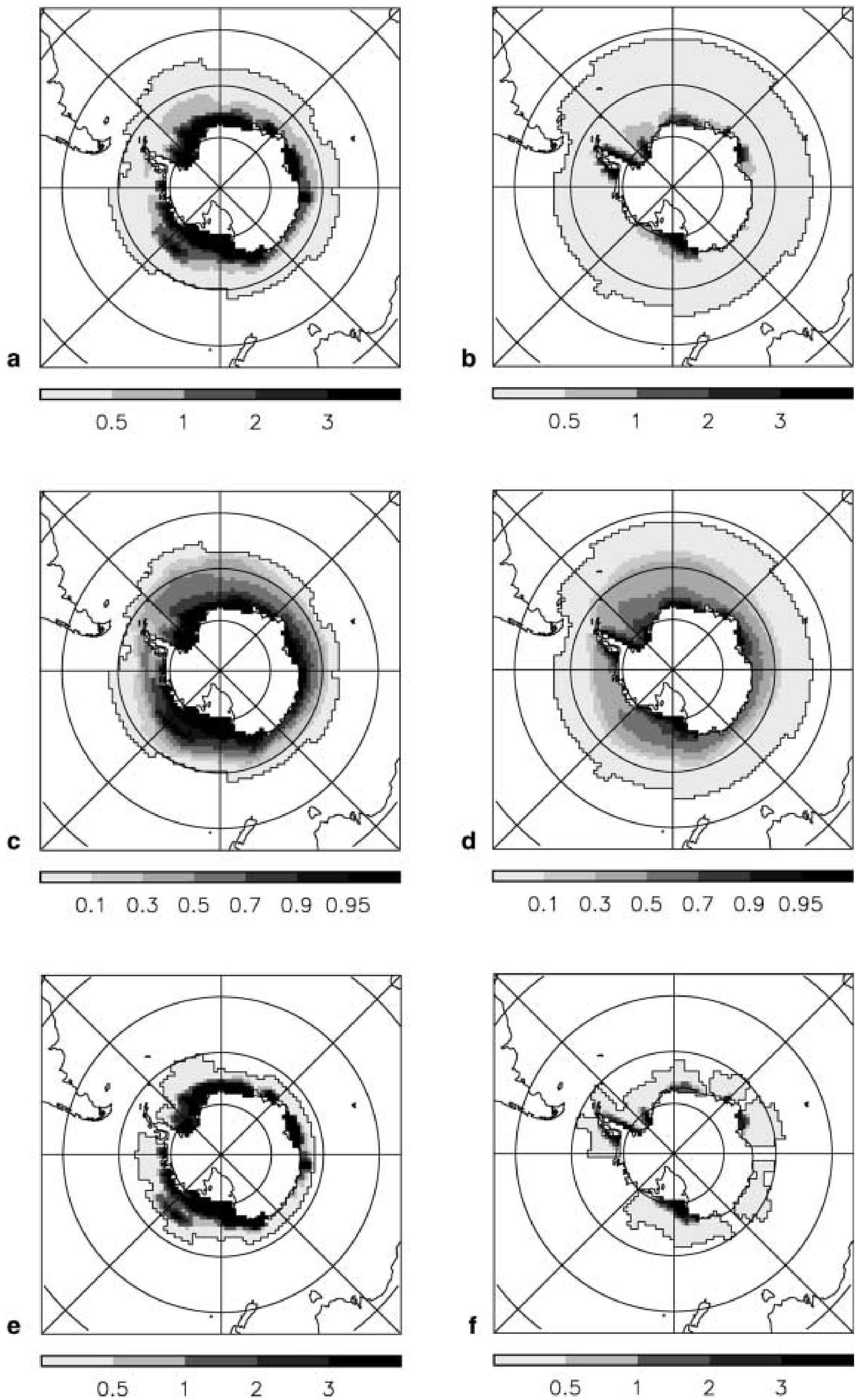

Fig. 2a–f Southern Hemisphere (40°–90°S) time mean sea-ice thickness, in m, and concentration, as % coverage of a grid box, from the control experiments. Latitude lines marked every 15°. **a** THERM annual mean thickness. **b** DYN annual mean thickness. **c** THERM annual mean concentration. **d** DYN annual mean concentration. **e** THERM March thickness. **f** DYN March thickness

tribution of the forcing may be important in determining the climate sensitivity, and that the global average response can not necessarily be deduced from the global average forcing (see Hansen et al. 1997).

To help answer these questions we shall concentrate on the response of the Southern Hemisphere sea ice. The Southern Hemisphere sea-ice changes tend to dominate the global surface temperature response largely because the Southern Hemisphere sea ice has a stronger albedo feedback effect being at lower latitudes than ice in the Arctic. Sea-ice concentrations strongly influence surface heat fluxes, which affect climate. Climate and surface heat fluxes strongly impact sea-ice thickness by determining thermodynamic growth and ice dynamics.

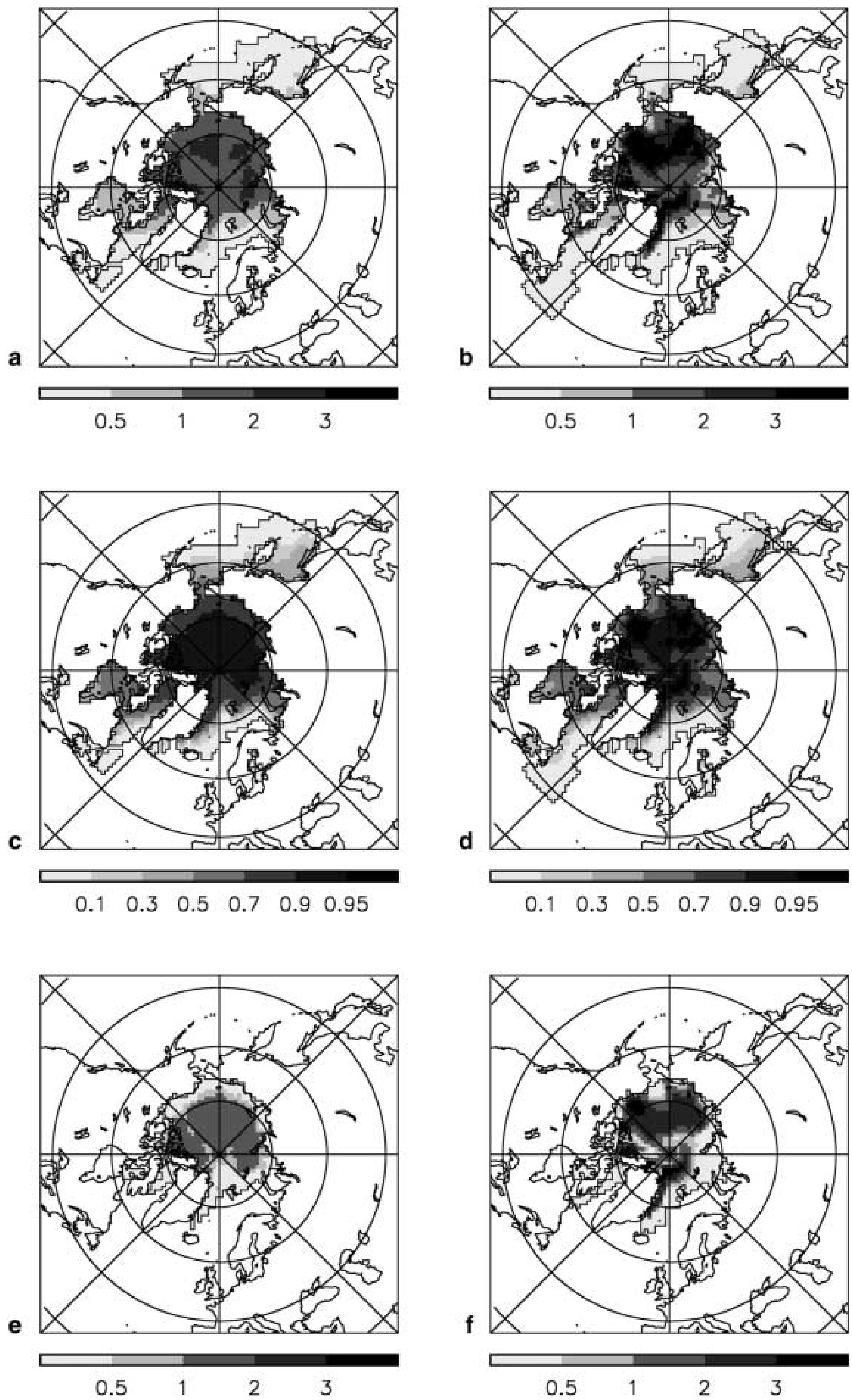

Fig. 3a–f Northern Hemisphere (40°–90°N) time mean sea-ice thickness, in m, and concentration, as % coverage of a grid box, from the control experiments. Latitude lines marked every 15°. **a** THERM annual mean thickness. **b** DYN annual mean thickness. **c** THERM annual mean concentration. **d** DYN annual mean concentration. **e** THERM September thickness. **f** DYN September thickness

Sea-ice thickness (particularly thinning ice) can feed back to changes in sea-ice concentration. If the concentration increases then the heat flux from the ocean to the atmosphere decreases, and the surface albedo increases, both leading to a cooling (Fig. 6a, b). The thickness will also increase as the areal concentration increases (Fig. 6c, d), but once the concentration has reached its maximum permitted value any further increases in sea-ice thickness will produce relatively small changes to the surface heat fluxes. These factors are important year round, with surface albedo changes being particularly effective in the summer months when the

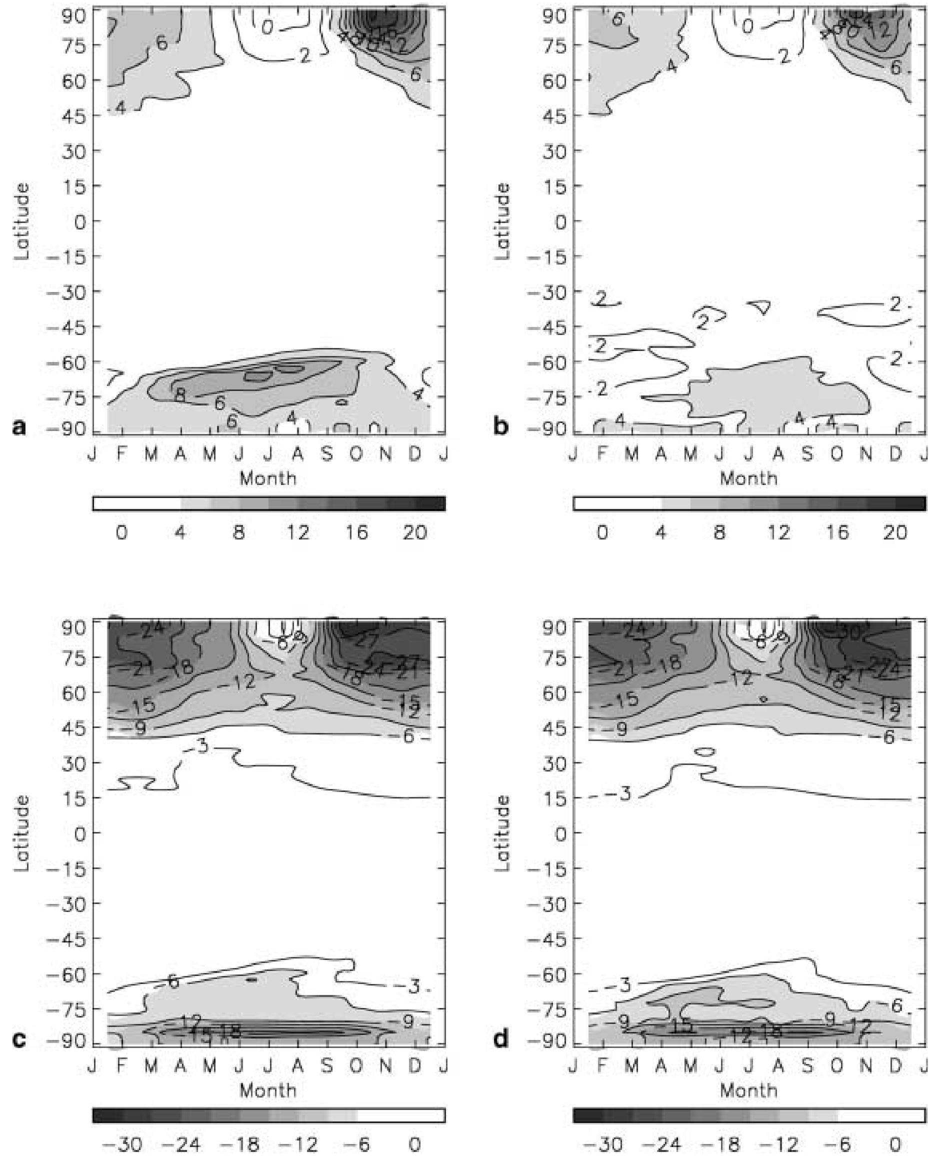

Fig. 4a–d Time-latitude plot of change in zonally averaged surface air temperature as a function of month. The *top two panels* are for $2 \times CO_2$ – control, with contours every 2 K, *shaded* above 4 K; the *bottom two panels* are for LGM – control, contours every 3 K, *shaded* below −6 K. **a** THERM. **b** DYN. **c** THERM. **d** DYN

Table 4 Area averaged annual mean surface temperature response for 2 × CO2 minus present (CO2–CON), and LGM minus present (LGM–CON), in K

		Global	Northern Hemisphere	Southern Hemisphere	Land	Ocean
CO2–CON	DYN	2.8	3.1	2.4	3.9	2.3
	THERM	3.3	3.4	3.2	4.3	2.9
LGM–CON	DYN	−4.5	−6.7	−2.3	−8.5	−1.6
	THERM	−4.4	−6.6	−2.2	−8.3	−1.5

insolation is high, and the upward heat fluxes from the ocean to the atmosphere are strongest in the winter months when the atmosphere is coldest.

4.2 Sea-ice response

First, we consider why the response to a doubling of CO_2 is larger in THERM than DYN.

The surface warming is generally largest in the regions where the control run sea-ice is relatively thick and compact (compare e.g. Fig. 7a with Fig. 2a) since these are the regions where the reduction in sea-ice thickness and compactness is generally largest. Similarly, the surface warming is smallest in the regions where the control run sea ice is relatively thin and broken. This mechanism, which shall be referred to as mechanism 1, explains most of the sea-ice changes seen in the CO2 experiments. The surface warming is then larger in THERM than DYN (Fig. 8a) in the regions where THERM has thicker, more compact sea-ice than DYN in the control (Fig. 2a–d) and a larger reduction in sea-ice than DYN (Fig. 8b). However, equatorward of about 55°S, mostly beyond the ice edge of THERM, there is ice in the DYN control, but the concentration in the control is less than 10% and the changes are small. Hence the effect on the temperature of the sea-ice feedback is very weak and the larger changes polewards in THERM dominate the difference in response. Overall, on doubling CO_2, the net result is a stronger sea-ice albedo feedback in THERM than in DYN, with a much larger reduction in the total Antarctic sea-ice extent (Table 2) in THERM.

Second, we consider the sea-ice response to the LGM boundary conditions, and describe why the Southern Hemisphere response is similar in the two LGM experiments while the response to a doubling of CO_2 is so different (Fig. 5 and Table 4).

In the LGM simulations, mechanism 1 operates as in the CO2 experiments, but in reverse, so we would expect the model with less sea ice in the control to cool the most. This is true, for example, around the Ross Sea where the sea ice in the control run of DYN is markedly thinner than in THERM (Fig. 2) and consequently, the glacial increases in sea-ice thickness and compactness are larger in DYN than THERM and DYN cools more than THERM (Fig. 8c, d).

However, mechanism 1 clearly does not explain the relative changes of THERM compared to DYN seen in the Weddell Sea and the South Atlantic sector of the Southern Ocean (Fig. 8c, d), where the response of THERM is larger than DYN even though DYN has thinner sea-ice in the control than THERM. A different mechanism becomes dominant here.

The thermodynamics of the sea-ice model will grow more sea ice due to the surface cooling produced by the glacial boundary conditions in both THERM and DYN. In regions of strong divergent currents, for example in

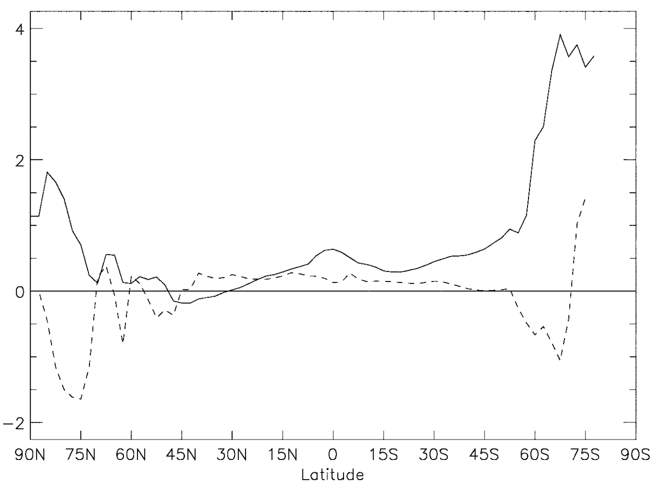

Fig. 5 Zonal average of annual mean difference of difference, as THERM – DYN anomaly – control, for surface temperature, in K, over ocean and sea ice. *Solid line* is $2 \times CO_2$ – control and *dashed line* is LGM – control

the Weddell Sea, the advective tendencies of the dynamic sea-ice model in DYN will act to reduce the sea-ice thickness and concentration, opposing the changes due to the thermodynamics. The result is that the increases in ice concentration are larger in THERM than DYN, even though the sea ice is slightly thicker in the control in THERM, and THERM cools more. In the regions where the currents are convergent, advection tends to reinforce the growth from thermodynamics, and so the increases in ice concentration can be larger in DYN than THERM. In both cases, this effect of the dynamics, which will be referred to as mechanism 2, opposes mechanism 1 (see Appendix for a mathematical illustration). In a few places, the model's crude representation of ice rheology means that once the sea-ice becomes more than 4 m thick there is no further increase of ice thickness due to dynamics (as described in Sect. 2.2), and so mechanism 2 is limited in some regions of convergent currents, for example along the Antarctic Peninsula. However, once the sea ice has become as thick as 4 m any further changes to sea-ice thickness do not produce a large change to the surface temperature (Fig. 6), which reduces the impact of mechanism 2 further in these localised regions.

Mechanism 2 is important in the LGM experiments since more ice is available for advection due to the thermodynamic growth of sea ice, while in the CO2 experiments less sea ice is available for advection due to the thermodynamic reduction in sea ice. Mechanism 1 tends to dominate in regions where the sea ice in the control of one model is relatively thick compared to the other model, and mechanism 2 can be important in regions where the advective tendencies are large. The net result is that both models produce a comparable increase in the total Antarctic sea-ice extent (Table 2).

These two mechanisms also operate in the Northern Hemisphere, but the patterns of changes are more complicated than in the Southern Hemisphere because most of the sea-ice cover of the Northern Hemisphere is in the relatively land-locked Arctic Ocean basin. The land barriers produce a complicated geographical distribution of sea-ice thickness and concentration (Fig. 3). The pattern of differences between control sea-ice thickness and concentration of THERM and DYN is also more complicated than in the Southern Hemisphere, and so, for brevity, will not be discussed here. As discussed earlier, the Southern Hemisphere dominates the difference in global climate sensitivity (Table 4).

In summary, the reason that the Southern Hemisphere response of THERM is more similar to DYN in the LGM experiments than in the CO2 experiments is because mechanism 2 tends to oppose mechanism 1 in the LGM simulations. Closer to the Antarctic coast, where the sea ice is thick in the control of THERM, mechanism 1 dominates over mechanism 2 and here DYN is more sensitive than THERM. For the CO2 experiments mechanism 2 reinforces mechanism 1, and the mainly divergent currents around Antarctica mean that THERM is more sensitive than DYN.

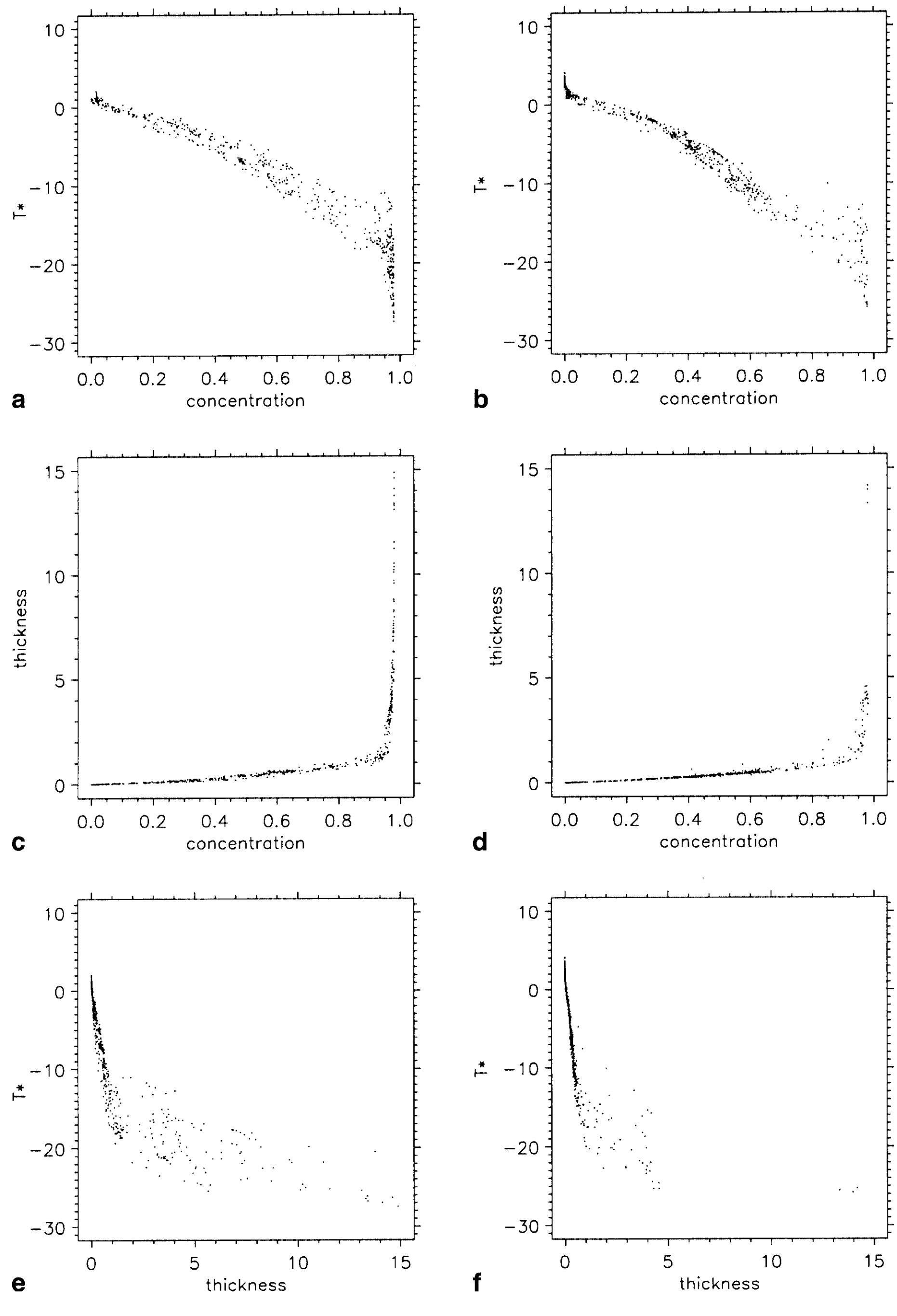

Fig. 6a–f Scatter plots showing relationships between annual mean surface temperature (T*), in °C, sea-ice concentration, and sea-ice thickness, in m, for every Southern Hemisphere sea-ice point. **a** Temperature against concentration for THERM. **b** Temperature against concentration for DYN. **c** Thickness against concentration for THERM. **d** Thickness against concentration for DYN. **e** Temperature against thickness for THERM. **f** Temperature against thickness for DYN

5 Comparison of LGM sea-ice extents to palaeoclimatic data reconstructions

The most commonly used reconstruction of sea-ice extents, based on SSTs, for the last glacial maximum is that of the Climate: Long-Range Investigation, Mapping and Prediction project (CLIMAP Project Members 1981), which used planktonic microfossils as a proxy indicator for SST. However, there has been much speculation as to the robustness of the CLIMAP dataset, particularly for regional studies of SST changes, but its continued widespread use is largely due to it being the only *global* SST reconstruction for the LGM. There are a few problems with the CLIMAP database specific to sea-ice regions that are relevant here. First, the planktonic foraminifera used do not necessarily provide a proxy record of the ocean surface temperature since they are mobile in the water column and often inhabit an intermediate layer of the upper waters of the ocean. Second, planktonic foraminifera are not very sensitive to cold environments such as those that occur in high-latitude ocean basins where the temperature is below 5 °C. Finally, the CLIMAP dataset was established prior to

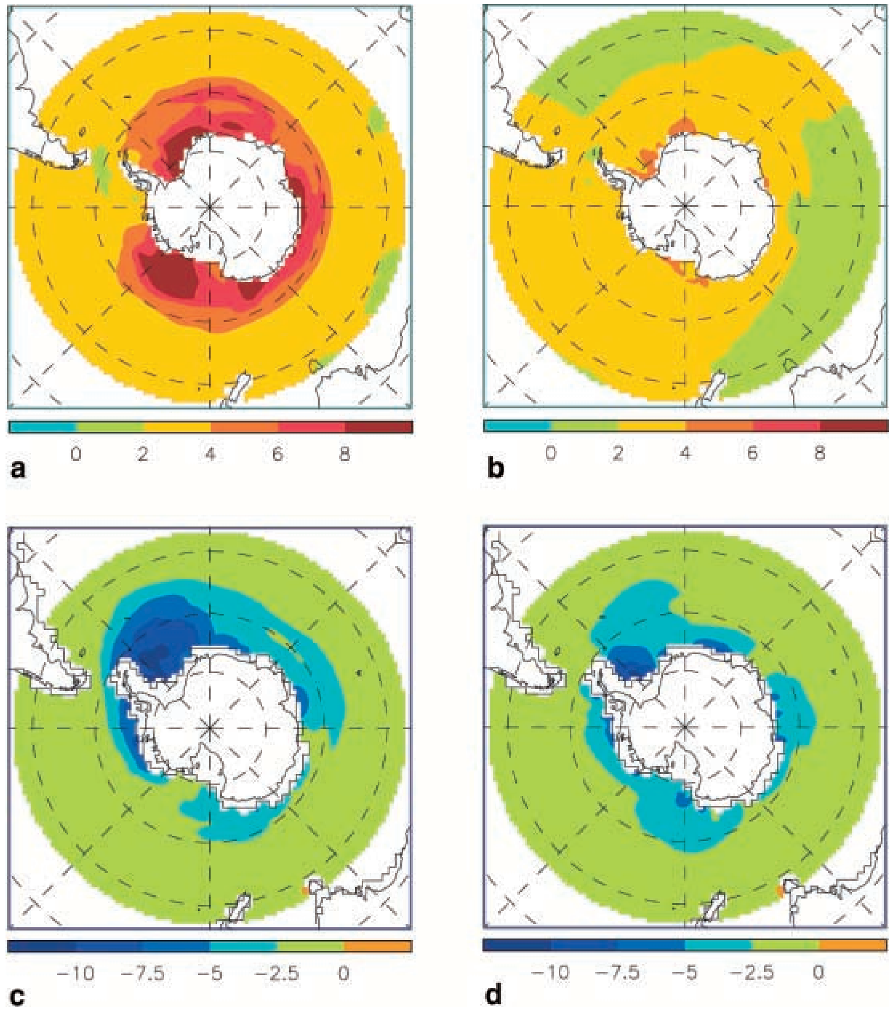

Fig. 7a–d Southern Hemisphere annual average change in surface temperature, in K. **a** $2 \times CO_2$ – control for THERM. **b** $2 \times CO_2$ – control for DYN. **c** LGM – control for THERM. **d** LGM – control for DYN

the definition of a high resolution stratigraphical scheme in the North Atlantic, and probably includes "Heinrich" layers which actually correspond to events colder than the LGM time slice centred around 21 000 years before present (21 kBP). The CLIMAP dataset includes records covering a timespan of 18 k ± 2 K ^{14}C BP years (where 18 k ^{14}C years is equivalent to the 21 k sidereal years referred to in this study).

We use two relatively recent reconstructions of glacial sea-ice cover, using different proxy indicators, in the crude qualitative, but illustrative, model–data comparison below. One for the Southern Ocean uses a modern analogue technique (MAT, Crosta et al. 1998) applied to diatoms, and the other for the North Atlantic uses dinoflagellate cysts (GEOTOP, de Vernal et al. 1994). Both methods provide quantitative estimates of sea-ice presence in terms of months cover per year during the LGM. This diagnostic is intended to be used to compare to, or constrain, GCMs. Permanent sea-ice cover in the reconstructions is considered to be comparable to an annual mean ice concentration of greater then 0.9 from the GCM (Fig. 9), 9–12 months cover per year from the reconstruction is comparable to an annual mean GCM concentration of 0.5–0.9, and less than 3 months cover per year in the reconstructions is comparable to an annual mean GCM concentration of less than 0.1 (Anne de Vernal personal communication).

The MAT reconstruction of Crosta et al. (1998) produces a broadly similar winter sea-ice limit to that given by CLIMAP, with the maximum ice extent at about 55°S in the Pacific Sector of the Southern Ocean and at about 50°S in the Indian Ocean sector. North of the Weddell Sea the MAT reconstruction produces a greater extension than CLIMAP, out to about 45°S. Both THERM and DYN reproduce some of the seasonal features of the MAT LGM reconstruction for the Southern Ocean, with a marked expansion of the sea-ice cover compared to present day (Table 2 and Fig. 9a, b). The equatorward component of the currents in DYN means that it produces a far greater winter sea-ice extent than THERM. This greater sea-ice extent, albeit very thin, is in better agreement with the MAT reconstruction. However, we feel that it is not possible at this stage to make firm conclusions as to which model, if either, is producing a more realistic seasonal simulation of the LGM sea ice in the Southern Ocean for the following

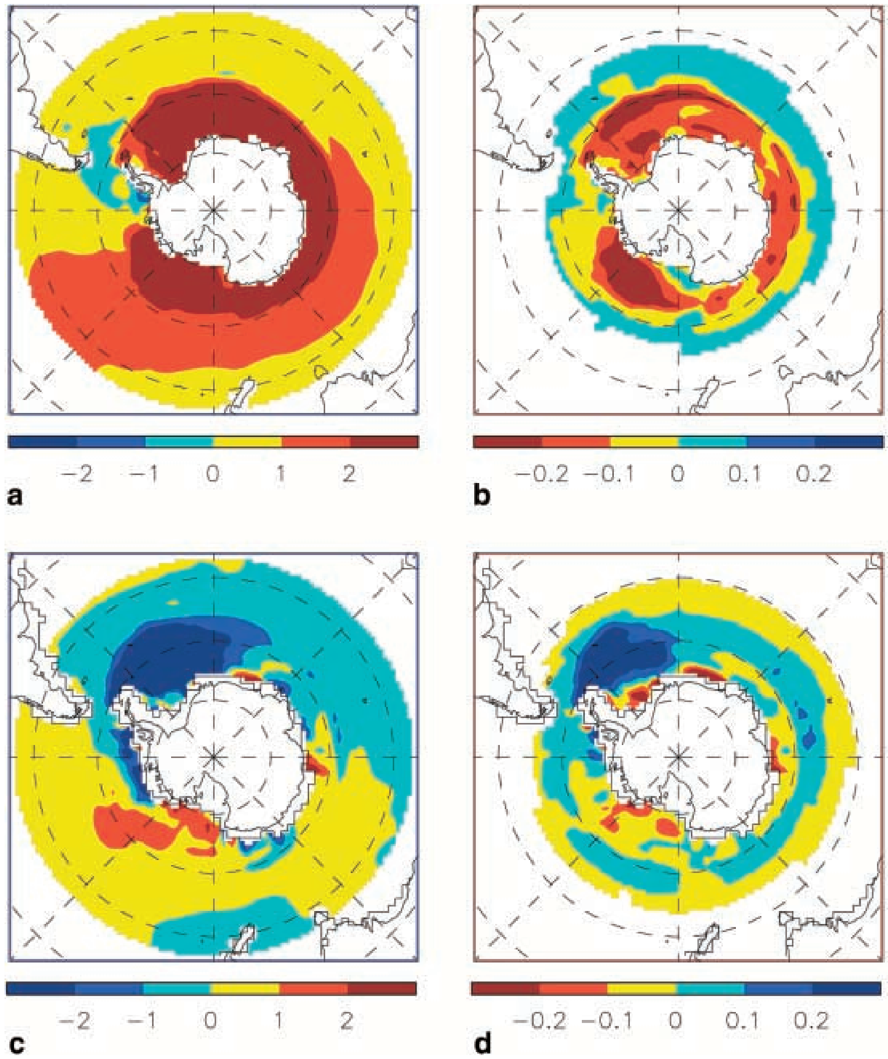

Fig. 8a–d Southern Hemisphere annual average difference for THERM – DYN. **a** 2 × CO_2 – control surface temperature, in K. **b** 2 × CO_2 – control ice concentration. **c** LGM – control surface temperature, in K. **d** LGM – control ice concentration

three reasons. Firstly, the Southern Ocean is a data sparse region for reconstructions of LGM sea ice. Secondly as stated, there is the possibility that the data includes events colder than the LGM which would affect the sea-ice extents. Thirdly, Crosta et al. (1998) conclude that additional work is needed to resolve the LGM summer sea-ice extent.

The GEOTOP reconstruction avoided data from the Heinrich layers and concentrated on the northwest North Atlantic. The reconstruction provides evidence for perennial sea-ice cover throughout Baffin Bay, extending along the coasts of Greenland and eastern Canada. Records from the Labrador Sea indicate ice cover for much of the year along the continental margin, but seasonally ice-free conditions offshore. While the CLIMAP reconstruction is consistent with these sea-ice extents there are differences over much of the rest of the North Atlantic. CLIMAP indicates winter sea ice extending fairly zonally all the way across the North Atlantic from New England to France with the summer sea-ice limit retreating to Iceland and Northern Ireland. The GEOTOP reconstruction however has a much more ice-free northeastern North Atlantic with the sector near Europe largely free from glacial sea ice and more of a southwest–northeast sea-ice limit in the northwestern North Atlantic.

The DYN and THERM simulations produce similar seasonal coverage over the North Atlantic (Fig. 9c, d) and they agree very well with the GEOTOP data in terms of the limits of perennial sea ice, seasonal sea ice, as well as a completely ice free sector in the northeast Atlantic off the coast of Europe. The main difference between the two simulations is the greater extent of maximum sea-ice cover in DYN (marked by the 0 to 0.1 contour range), particularly the greater southward extent down to 37.5°N in the western Atlantic. Unfortunately, there is at present no data in this region to compare the model simulations to, but an analysis of several more cores is underway. It should be borne in mind that the reason DYN extends the sea ice further south is because the currents applied to the model have a southward component in that region. It is possible that

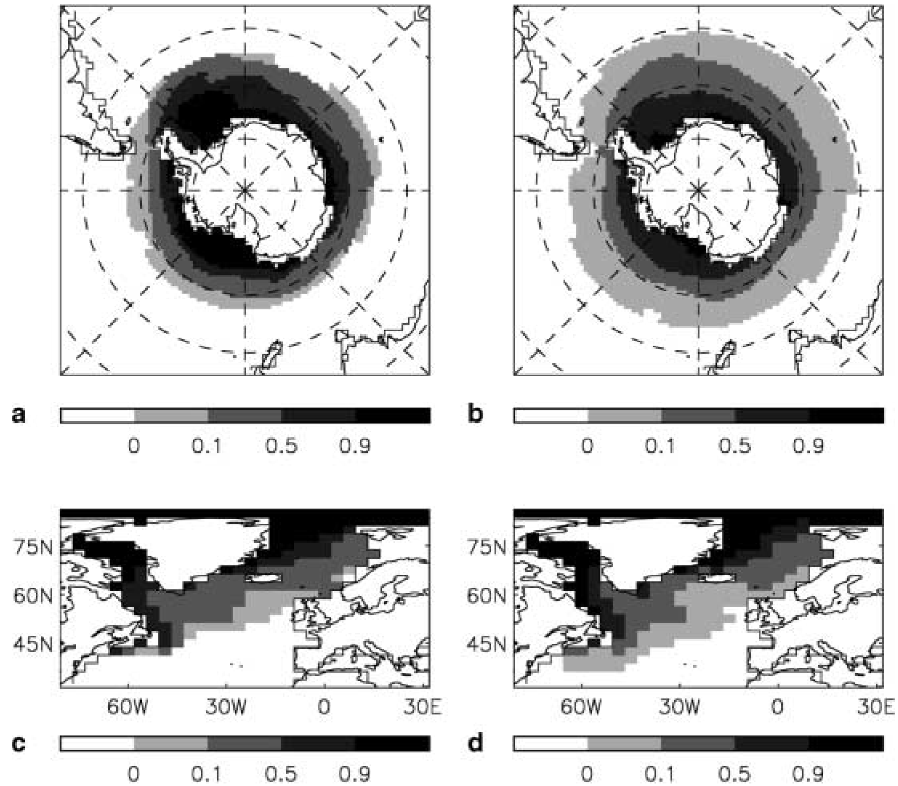

Fig. 9a–d Annual mean LGM sea-ice concentration. **a** THERM Southern Hemisphere. **b** DYN Southern Hemisphere. **c** THERM Northern Hemisphere. **d** DYN Northern Hemisphere

the real glacial currents in the North Atlantic were very different to the currents that have been determined from the HadCM2 coupled model simulation for the present day.

6 Concluding remarks

We have assessed the effect of including a more complete description of sea-ice physics in our coupled atmosphere–mixed layer ocean–sea-ice model by conducting sensitivity experiments using two parametrisations of sea ice. The inclusion of a simple "ocean-drift" parametrisation for sea-ice dynamics improves the model's simulation of sea-ice thickness and compactness, especially in the Antarctic region, when compared to observational data for the present day. This parametrisation, which uses the ocean surface current to advect the sea ice gives comparable results to those found when a more detailed cavitating fluids parametrisation (Flato and Hibler 1990, 1992) has been included in a different mixed layer ocean model (the GENESIS model, Pollard and Thompson 1994).

The sensitivity of the model's surface climate to a doubling of CO_2 is reduced by 15% on the inclusion of sea-ice dynamics, although much larger differences occur over sea-ice areas in autumn and winter. The larger response in THERM is mostly attributable to the simulated present-day sea-ice cover. The response of the model is qualitatively similar to that of Pollard and Thompson (1994) who found a 9% reduction in the climate sensitivity to a doubling of CO_2 on including sea-ice dynamics. Pollard and Thompson (1994) found strikingly similar local changes in Antarctica, and they also note the dependence of the response on the simulated present-day sea-ice cover, particularly for the Southern Hemisphere.

Rind et al. (1997) have used the GISS mixed layer ocean model to study the influence that different present-day sea-ice distributions have on the response to a doubling of CO_2. They found that the Southern Hemisphere sea-ice changes contributed more to global warming than the Northern Hemisphere, primarily due to the albedo feedback occurring at lower latitudes, consistent with this study. However, they found that the Southern Hemisphere changes were greatest where the sea-ice coverage was largest, which seems to contradict the findings of our study and Pollard and Thompson's (1994) work. In fact, the mechanisms that control our sea-ice response are broadly consistent with Rind et al.'s (1997) results, but the balance of the competing mechanisms is different in our simulations, again indicating the importance of the control sea-ice thickness, concentration and extent.

When glacial boundary conditions are applied to the THERM and DYN models the response shows some qualitative similarities, allowing for the reverse in sign, to the $2 \times CO_2$ response, such as the high-latitude amplification to the temperature response. However, the climate responses of the two models are more similar to each other in the glacial case than in the doubled CO_2 case. At the LGM both THERM and DYN produce a global average cooling of similar magnitude. It is the response of the Southern Hemisphere sea ice in the two models that largely determines the different global average temperature changes, partly because the Southern Hemisphere sea ice has a stronger albedo feedback being at lower latitudes than the Arctic sea ice, and partly because the mechanisms that produce changes in sea ice combine differently for the CO_2 experiment and the LGM experiment.

Quantitative comparisons of model results with reconstructions of glacial sea ice extent in the Southern Ocean and North Atlantic are problematical, but a crude comparison suggests that the model's glacial sea ice extents are consistent with the palaeodata. More palaeoclimatic data is needed to resolve the boundaries of the North Atlantic seasonal glacial sea ice in order to validate the model's LGM sea ice simulation. However, since the two models differ most in their simulations of Southern Hemisphere sea ice, it is this data sparse region that could provide useful palaeoclimatic data, and model–data comparisons, for assessing whether DYN is capable of producing a more realistic sea-ice response than THERM.

We have shown that for our coupled atmosphere–mixed layer ocean–sea-ice model the different nature of the anomaly responses can be attributed largely to the sea-ice distributions in the control simulations. The present-day sea ice simulated by DYN is more realistic than THERM based on present-day observational data, and the glacial responses of both DYN and THERM are consistent with the palaeodata. We therefore conclude that by including a simple parametrisation for sea-ice advection, and therefore by including a more complete description of sea-ice physics, we can have more confidence in the simulation of anomaly climates. This study reinforces previous findings that since the climate sensitivity of the GCM to perturbations is dependent on the control simulation, it is essential to simulate the present-day climate as realistically as possible, so that any climate change scenario is applied to a realistic reference climatic state.

We have also shown that the global mean climate sensitivity can vary for different forcings and for different models. Therefore, even if we could reconstruct a global average temperature and a global average radiative forcing for a past climate state, such as the LGM, from proxy data or from a climate model, we would not necessarily be able to determine the globally averaged temperature change to a different forcing. Such an approach, if possible, would have been useful to "narrow the range of uncertainty" of future anthropogenic-induced globally averaged temperature changes, but we can still investigate the mechanisms that produce the different climate sensitivities. Further work is needed to quantify the potential size of feedbacks, for example due to changes to sea ice, cloud, vegetation, and ocean–atmosphere interactions, and also to assess the more realistic sea-ice models as they become incorporated into coupled ocean–atmosphere models.

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Appendix 1: effect of dynamic tendencies on sea-ice response

We illustrate, in a simplified manner, the effect of changes in sea-ice thickness on sea-ice dynamics. We will assume, for simplicity, that no other feedbacks operate.

The thickness of sea-ice, H_{DCON}, at a point in the control of DYN will be given by

$$H_{DCON} = H_{TCON} - \nabla \cdot (\mathbf{v}H_{TCON}) \quad (1)$$

$$= H_{TCON} - [(\nabla H_{TCON}).\mathbf{v} + H_{TCON}(\nabla.\mathbf{v})] \quad (2)$$

where $\mathbf{v}$ is the surface current velocity, and H_{TCON} is the thickness of sea ice in the control of THERM.

The second term on the right hand side of Eq. 1 represents the advective tendencies of the sea-ice model, and involves the horizontal gradients of sea-ice thickness and the divergence of ocean currents (Eq. 2). The divergence of the currents is dominant over much of the ocean surrounding Antarctica (not shown). In regions of divergent currents ($\nabla \cdot \mathbf{v} > 0$) the advective term is generally negative and the sea-ice is generally thinner in DYN than THERM, i.e. $H_{DCON} < H_{TCON}$, and vice versa in regions of convergent currents.

Now consider a climate change that changes the ice thickness by a fraction a in THERM, where a is positive for a warmer climate, such as in the $2 \times CO_2$ experiments, and a is negative for a cooler climate, such as in the LGM experiments. The thickness of the sea ice in the anomaly experiments of THERM (H_{TANOM}) and DYN (H_{DANOM}) is then given by

$$H_{TANOM} = (1 - a)H_{TCON}$$

$$H_{DANOM} = H_{TANOM} - \nabla \cdot (\mathbf{v}H_{TANOM})$$

$$= (1 - a)(H_{TCON} - \nabla \cdot (\mathbf{v}H_{TCON}))$$

The change in sea-ice thickness between the anomaly experiments and the controls in THERM (ΔH_T) and DYN (ΔH_D) is then

$$\Delta H_T = -aH_{TCON}$$

$$\Delta H_D = -a(H_{TCON} - \nabla \cdot (\mathbf{v}H_{TCON}))$$

and the relative response of THERM compared to DYN is given by

$$\Delta H_T - \Delta H_D = -a\nabla \cdot (\mathbf{v}H_{TCON})$$

For the $2 \times CO_2$ experiments a is defined to be positive, and ΔH_T and ΔH_D are *negative*. In regions of divergent currents, $\nabla \cdot (\mathbf{v}H_{TCON})$ is generally positive, and so THERM is more responsive than DYN. Conversely, in regions of convergent currents DYN is more responsive than THERM.

Since $H_{TCON} > H_{DCON}$ in regions of divergent currents, and $H_{TCON} < H_{DCON}$ in regions of convergent currents, the effect, in a change of climate, of the combined dynamic and thermodynamic tendencies in DYN compared to the thermodynamic-only tendencies in THERM is to produce a larger change in the model that has the most ice in the control, and this *reinforces* mechanism 1.

For the LGM experiments a is negative, and ΔH_T and ΔH_D are positive. In regions of divergent currents, $\nabla \cdot (\mathbf{v}H)$ is generally positive, and therefore THERM is more responsive than DYN, and vice versa.

As in the $2 \times CO_2$ experiments, $H_{TCON} > H_{DCON}$ in regions of divergent currents, and $H_{TCON} < H_{DCON}$ in regions of convergent currents. However, in the LGM climate change, mechanism 1 suggests that the model with less ice in the control will respond the most. Therefore, the effect of the dynamic and thermodynamic tendencies in DYN compared to the thermodynamic-only tendencies in THERM is to *oppose* mechanism 1. This effect is referred to as mechanism 2.