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High-latitude climate sensitivity to ice-sheet forcing over the last 120 kyr

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ABSTRACT

Interpretation of ice-core records is currently limited by paucity of modelling at adequate temporal and spatial resolutions. Several key questions relate to mechanisms of polar amplification and inter-hemispheric coupling on glacial/interglacial timescales. Here, we present the first results from a large set of global ocean–atmosphere climate model 'snap-shot' simulations covering the last 120 000 years using the Hadley Centre climate model (HadCM3) at up to 1 kyr temporal resolution. Two sets of simulations were performed in order to examine the roles of orbit and greenhouse gases versus ice-sheet forcing of orbital-scale climate change. A series of idealised Heinrich events were also simulated, but no changes to aerosols or vegetation were prescribed. This paper focuses on high latitudes and inter-hemispheric linkages. The simulations reproduce polar temperature trends well compared to ice-core reconstructions, although the magnitude is underestimated. Polar amplification varies with obliquity, but this variability is dampened by including variations in land ice coverage, while the overall amplification factor increases. The relatively constant amplification of Antarctic temperatures (with ice-sheet forcing included) suggests it is possible to use Antarctic temperature reconstructions to estimate global changes (which are roughly half the magnitude). Atlantic Ocean overturning circulation varies considerably only with the introduction of Northern Hemisphere ice sheets, but only weakens in the North Atlantic in the deep glacial, when ocean–sea-ice feedbacks result in the movement of the region of deep convection to lower latitudes and with the introduction of freshwater to the surface North Atlantic in order to simulate Heinrich events.

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1. Introduction

Polar regions have in recent decades experienced temperature increases that are several times larger than the global average (Trenberth et al., 2007). Future climate change simulations demonstrate a range of polar amplification of global warming between 1.5 and 4.5 (Holland and Bitz, 2003). This amplification is a result of several factors including ice-albedo feedback, greater fractional changes in heating rather than evaporation at high latitudes, as well as impacts on ocean-to-atmosphere heat transfer from the changing sea-ice cover (ACIA, 2004). The Arctic and Antarctic Peninsula are consequently sensitive indicators of global change, and particularly vulnerable to such changes.

Past variations in polar temperature can be reconstructed from the water stable-isotope composition of ice in deep ice cores from the Arctic and Antarctic, scaled by borehole temperatures and gas fractionation to account for changes in seasonality (e.g. Masson-Delmotte et al. 2006a; Jouzel et al., 2007; Kawamura et al., 2007). Deep ice cores are unique in providing high temporal resolution data, which span several glacial cycles in the case of the EPICA (European Project for Ice Coring in Antarctica) and Vostok cores (Petit et al., 1999; EPICA community members, 2004). Recent temperature reconstructions from ice-core proxy data suggest that during the Last Glacial Maximum (LGM) temperatures at the EPICA dome C core site were 9 °C cooler than present day (Jouzel et al., 2007) and Greenland was 19–22 °C colder than present day (Masson-Delmotte et al., 2006a). During the mid-Holocene (6 kyr BP; MH) temperatures estimated from stable isotopes were up to 0.8 and 0.5–0.9 °C warmer than present over east Antarctica and central Greenland respectively (Masson et al., 2000, Masson-Delmotte et al., 2005). At the start of the last interglacial (130 ± 1 kyr BP; LIG) polar temperatures were much warmer than at present, 3–5 °C in central Greenland and roughly 4 °C at the EPICA site.

The question of how these orbital-scale high latitude temperature changes amplify global change, and what teleconnections and feedbacks are operating, can be investigated using climate models. The

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Paleoclimate Model Intercomparison Project (PMIP1 and PMIP2; Braconnot et al., 2007a,b) simulations have been used to quantify polar amplification at the LGM, MH and pre-industrial (PI). For example, fully coupled ocean–atmosphere model LGM simulations from the second phase of the project (PMIP2) suggest an average polar amplification factor for Greenland of 2.7 (Masson-Delmotte et al., 2006b), due to changes in topography, albedo feedbacks and atmosphere/ocean circulation changes. Antarctic amplification from the same LGM model simulations was 2.1 (with no correction for elevation changes). A feature of all PMIP2 models is the underestimation of Greenland temperature change at the LGM, while some models do manage to capture the magnitude of Antarctic cooling. Central Greenland simulated LGM to PI temperature change was 6.2–14.5 °C, which is roughly half that obtained from stable isotopes; possibly a result of neglecting aerosol loading and vegetation changes in the model, which probably have greater impact over Greenland than Antarctica (Masson-Delmotte et al., 2006a).

Both Greenland and Antarctica temperature anomalies from the PMIP2 range of models were linearly related to the modelled global change. However, only a few scenarios have been extensively compared: the LGM and MH, as well as $2 \times CO_2$ and $4 \times CO_2$ scenarios (Masson-Delmotte et al., 2006b). The polar amplification factor was smaller for $2 \times CO_2$ and $4 \times CO_2$ scenarios than for the LGM. It has been suggested that the Antarctic in particular can be used to estimate global temperature changes over glacial–interglacial (G-IG) timescales, partly because it will be less sensitive to local changes in land surface cover, and the polar amplification makes it a sensitive indicator (e.g. Genthon et al., 1987; Masson-Delmotte et al., 2006b; Hargreaves et al., 2007). One interpretation of the PMIP2 results is that Antarctic temperature changes in ice cores may be used to represent global changes if divided by roughly a factor of two (Masson-Delmotte et al., 2006b). Antarctic climate change may also be useful to constrain global climate sensitivity (Hargreaves et al., 2007). Following PMIP2 and other studies, it would be useful to investigate the variability of polar amplification throughout the last glacial cycle.

The majority of modelling studies with general circulation models (GCMs) have focussed on key time periods in the last glacial cycle, such as the LGM and MH. Previous attempts at modelling the evolution of climate through the last glacial–interglacial cycle have used energy balance models (EBM) or Earth system models of intermediate complexity (EMIC) (e.g. Short and Mengel, 1986; Gallee et al., 1992; Tarasov and Peltier, 1999; Berger et al., 1998), or have used GCMs in which the boundary condition forcing is accelerated by up to a factor of 100 to cope with the large computational effort, although the internal timescales and physics of the model are unchanged (i.e. Jackson and Broccoli, 2003; Kutzbach et al., 2008; Lorenz and Lohmann, 2004). In most GCM-based simulations, the studies have so far only focussed on orbital forcing, using either a simple slab ocean model (Jackson and Broccoli, 2003) or a low resolution GCM (Kutzbach et al., 2008). Such models have demonstrated, for example, the importance of both orbital forcing and decline in atmospheric CO_2 for Northern Hemisphere ice volume (e.g. Berger et al., 1998), and the importance of sea-ice feedbacks in mid to high latitude response to orbital forcing to enable glaciation (Jackson and Broccoli, 2003). One recent study (Liu et al., 2009) used a GCM to model the evolution of the deglaciation from 21 kyr to 14 kyr, which highlighted the importance of changes in the Atlantic overturning and CO_2 to achieve the abrupt warming at the Bølling–Allerød transition.

In this study we present the first results from a large number of simulations with a fully coupled GCM that cover the last glacial cycle. We performed snapshot simulations at up to 1 kyr intervals over the time period 120 kyr BP to present. Two experiments were performed: in one experiment the orbital configuration and atmospheric greenhouse gas concentrations were varied, and in the other experiment variations in ice sheets were prescribed in addition. Primarily, climate changes on orbital timescales are examined, as the simulations are not fully transient. The only millennial-scale variation to be investigated is the impact of Heinrich events, which are modelled in a series of idealised freshwater hosing experiments. Here, in particular, we use these sensitivity experiments to address the question of the variability of polar amplification and mechanisms producing asymmetry in climate change in the Northern and Southern Hemispheres over the last 120 kyr BP.

2. Methods

2.1. General model description

We performed multiple "snapshot" simulations with the Hadley Centre climate model, HadCM3 (Gordon et al., 2000; Pope et al., 2000). HadCM3 is a state-of-the-art GCM that was heavily used in both the third and fourth assessment reports of the Intergovernmental Panel on Climate Change (IPCC, 2001, 2007). The GCM consists of a linked atmospheric model, ocean model and sea ice model. The resolution of the atmospheric model is 2.5° in latitude by 3.75° in longitude by 19 unequally spaced levels in the vertical. The spatial resolution over the ocean in HadCM3 is 1.25° by 1.25° by 20 unequally spaced layers in the ocean extending to a depth of 5200 m. The model contains a typical range of parameterisations in the atmosphere and ocean, including a detailed radiation scheme that can represent the effects of minor trace gases (Edwards and Slingo, 1996). The land surface scheme includes the representation of the freezing and melting of soil moisture, and terrestrial evaporation includes the dependence of stomatal resistance on temperature, vapour pressure and CO_2 concentration (Cox et al., 1999). In this version of the model, interactive vegetation is not included. The ocean model uses the Gent–McWilliams mixing scheme (Gent and McWilliams, 1990). The sea ice model uses a simple thermodynamic scheme and contains parameterisations of ice drift and leads (Cattle and Crossley, 1995). We extensively benchmarked our version of the model to ensure that it statistically gave the same control climate as previously published.

2.2. Boundary conditions

A complete Earth system model would be able to attempt to simulate the last glacial–interglacial cycle using orbital forcing only. However, the HadCM3 GCM does not include interactive ice, carbon cycle, or methane. Hence we must force the model with prescribed changes in orbit, greenhouse gases and ice-sheet evolution. The former two are relatively well constrained. Orbital parameters are taken from Berger and Loutre (1991) and were straightforward to include within the GCM. Similarly atmospheric concentrations of CO_2 were taken from Vostok (Petit et al., 1999; Loulergue et al., 2008) and CH_4, and N_2O were taken from EPICA (Spahni et al., 2005). All ice-core data were on the same EDC3 timescale (Fig. 1; Parrenin et al., 2007).

Ice-sheet extent and elevation, and associated changes in sea level and isostatic adjustment, are more poorly constrained, particularly for the period before the Last Glacial Maximum (21 kyr BP). For GCM modelling, we need to know the evolution of all of the major ice sheets and there are relatively few studies which have attempted to reconstruct together all of the major ice sheets, namely the North American, Greenland, Fennoscandian, and Antarctic ice sheets. We chose to develop our ice-sheet reconstructions using the ICE5G model of Peltier (2004). This was used by the PMIP2 project as a boundary condition for the LGM simulations. The dataset also includes a detailed evolution of the ice thickness, extent, and

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