The Arctic cryosphere in the Mid-Pliocene and the future

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The Mid-Pliocene (ca 3 Myr ago) was a relatively warm period, with increased atmospheric CO₂ relative to pre-industrial. It has therefore been highlighted as a possible palaeo-analogue for the future. However, changed vegetation patterns, orography and smaller ice sheets also influenced the Mid-Pliocene climate. Here, using a general circulation model and ice-sheet model, we determine the relative contribution of vegetation and soils, orography and ice, and CO₂ to the Mid-Pliocene Arctic climate and cryosphere. Compared with pre-industrial, we find that increased Mid-Pliocene CO₂ contributes 35 per cent, lower orography and ice-sheet feedbacks contribute 42 per cent, and vegetation changes contribute 23 per cent of Arctic temperature change. The simulated Mid-Pliocene Greenland ice sheet is substantially smaller than that of modern, mostly due to the higher CO₂. However, our simulations of future climate change indicate that the same increase in CO₂ is not sufficient to melt the modern ice sheet substantially. We conclude that, although the Mid-Pliocene resembles the future in some respects, care must be taken when interpreting it as an exact analogue due to vegetation and ice-sheet feedbacks. These act to intensify Mid-Pliocene Arctic climate change, and act on a longer time scale than the century scale usually addressed in future climate prediction.

Keywords: Pliocene; ice sheets; cryosphere; Greenland; palaeoclimate

1. Introduction

A considerable number of modelling studies have focused on the climate of the Mid-Pliocene (3.29–2.97 Myr ago) (e.g. Chandler et al. 1994; Sloan et al. 1996; Haywood & Valdes 2004; Haywood et al. 2005). It was a period of relative global
warmth (e.g. Haywood & Valdes 2004), and was followed by global cooling, which marked the onset of intense glacial–interglacial cycles that characterize the last 2 Myr (Lisiecki & Raymo 2005; Lawrence et al. 2006). However, the causes of the warm climate of the Mid-Pliocene are not currently well understood. Records derived from stomatal densities in fossilized leaves (Kurschner et al. 1996) and measurements of marine δ13C (Raymo et al. 1996) indicate elevated CO₂ levels of approximately 380–400 ppmv, with excursions as high as 425 ppmv, and this undoubtedly contributed to the warmer climate. Additionally, sea-level data indicate a reduced extent of the Greenland and Antarctic ice sheets relative to modern (Dowsett & Cronin 1990), which would also have further increased the local and global temperatures due to reduced albedo. Reconstructions of vegetation from pollen indicate a poleward shift of boreal forest in the Mid-Pliocene relative to modern, in particular in Siberia and northern Canada (Salzmann et al. 2008). This was most likely a response to increased temperatures, but the associated decreased albedo would also have contributed to maintaining a warmer Mid-Pliocene climate. Additionally, Late Miocene and Pliocene tectonic uplift may have been responsible for cooling into the Quaternary (Ruddiman & Kutzbach 1989), in particular when the height of the Rocky Mountains started to extend above the equilibrium line altitude, leading to cooling due to more extensive snow cover. Most likely, a combination of all of the above contributed to enhanced Mid-Pliocene warmth.

The Intergovernmental Panel on Climate Change (IPCC) have highlighted the Mid-Pliocene as a potential analogue for future climate change (Jansen et al. 2007): ‘The Mid-Pliocene (about 3.3–3.0 Ma) is the most recent time in Earth’s history when mean global temperatures were substantially warmer for a sustained period..., providing an accessible example of a world that is similar in many respects to what models estimate could be the Earth of the late 21st century.’ One of the greatest concerns about future climate change is its effects at high latitudes, in particular the possible implications for ice-sheet stability. The rise in sea level following the melting of the ice sheets would clearly have disastrous implications for populations situated close to sea level, and a significant impact on the global economy. Sea surface temperature (SST) reconstructions derived from faunal analysis techniques (e.g. Dowsett et al. 1999; Dowsett 2007) indicate that the greatest temperature increases at the Mid-Pliocene relative to modern were at higher latitudes.

In this paper, we use a modelling strategy to investigate the Mid-Pliocene and future climate changes, with a focus on the Atlantic sector of Northern Hemisphere high latitudes and the Arctic cryosphere. For the Mid-Pliocene, we investigate the relative role of CO₂, vegetation, ice sheet and tectonics in determining the characteristics of Northern Hemisphere climate and cryosphere, in particular the Greenland ice sheet. We do this by carrying out an ensemble of general circulation model (GCM) simulations in which we move from the climate of the pre-industrial to that of the Mid-Pliocene, in a series of incremental steps, and use the resulting climatologies to force an offline ice-sheet model. This leads to an understanding of the important mechanisms that have determined the past Arctic climate change. For the future, we investigate the response of the Arctic cryosphere to elevated anthropogenic atmospheric CO₂ levels, by carrying out an additional ensemble of GCM simulations in which we elevate CO₂ to various levels under otherwise pre-industrial conditions. This provides a guide to the
nature of climate impacts, given certain future CO$_2$ stabilization scenarios. Finally, by comparing the Mid-Pliocene and future climates, we ask whether the Mid-Pliocene represents a good analogue for future high-latitude Arctic climate and ice-sheet change. This indicates to what extent Mid-Pliocene palaeodata can be used as an indicator of possible future climate change.

2. Model descriptions

In this study we make use of two models. Firstly, the global GCM, HadCM3 (Gordon et al. 2000): a fully coupled atmosphere–ocean GCM, which does not require the use of flux correction to maintain a realistic present-day climate (Gregory & Mitchell 1997). HadCM3 has been used in the IPCC third and fourth assessment reports (Houghton et al. 2001; Solomon et al. 2007), and performs well in a number of tests relative to other global GCMs (Covey et al. 2003; Solomon et al. 2007). The horizontal resolution of the atmospheric model is 2.5° in latitude by 3.75° in longitude, with 19 vertical layers. The atmospheric model has a time step of 30 min and includes many parametrizations representing subgrid-scale effects, such as convection (Gregory & Rowntree 1990) and boundary-layer mixing (Smith 1993). The land surface scheme includes the representation of the freezing and melting of soil moisture. The representation of evaporation includes the dependence of stomatal resistance on temperature, vapour pressure and CO$_2$ concentration (Cox et al. 1999). The spatial resolution in the ocean in HadCM3 is 1.25°×1.25°, with 20 vertical layers. The model uses the Gent & McWilliams (1990) mixing scheme, and there is no explicit horizontal tracer diffusion. The horizontal resolution allows the use of a smaller coefficient of horizontal momentum viscosity, leading to an improved simulation of ocean velocities compared with earlier versions of the model. The sea-ice model uses a simple thermodynamic scheme and contains parametrizations of sea-ice drift and leads (Cattle & Crossley 1995). For one of the GCM simulations, where a long run is needed in order to reach equilibrium, we use a version of the model with a lower resolution in the ocean—2.5° in latitude by 3.75° in longitude (this version of the model is known as HadCM3L).

Secondly, we use a three-dimensional thermomechanical ice-sheet model, GLIMMER v. 1.0.4., configured over the Greenland region. The core of the model is based on the ice-sheet model described by Payne (1999). The ice dynamics is represented with the widely used shallow-ice approximation, and a full three-dimensional thermodynamic model is used to determine the ice flow law parameter. The model is formulated on a Cartesian $x$–$y$ grid, and takes as input the surface mass balance and air temperature at each time step. In the present work, the ice dynamics time step is 1 year. To simulate the surface mass balance, the present work uses the positive degree day (PDD) approach described by Reeh (1991). The basis of the PDD method is the assumption that the melt that takes place at the surface of the ice sheet is proportional to the time-integrated temperature above freezing point, known as the PDD. The method described by Reeh (1991) and that implemented here is somewhat more sophisticated, in that two PDD factors are used, one each for snow and ice, to take account of the different albedos and densities of these materials. The use of the PDD mass-balance models is well established in coupled atmosphere–ice-sheet palaeoclimate modelling studies (e.g. DeConto & Pollard 2003; Lunt et al. 2008a, b).
GLIMMER includes a representation of the isostatic response of the lithosphere, which is assumed to behave elastically, based on the model of Lambeck & Nakiboglu (1980). The forcing data from HadCM3 are transformed onto the ice model grid using bilinear interpolation, which ensures that precipitation is conserved in the atmosphere–ice-sheet coupling. In the case of the surface air temperature field, a vertical lapse-rate correction is used to take account of the difference between the high-resolution topography seen within GLIMMER, and that represented with HadCM3. The use of a lapse-rate correction to better represent the local temperature is established in the previous work (e.g. Pollard & Thompson 1997).

The experimental design is different from that described in a previous study, investigating the role of Panama Seaway closure on Northern Hemisphere glaciation (Lunt et al. 2008a). In that study, the absolute temperatures and precipitation from the GCM were used to drive the ice-sheet model. In part to account for the cold bias in the HadCM3-simulated winter temperature over Greenland (Murphy et al. 2002), the PDD factors were tuned in that study to give the best agreement with the present-day Greenland ice sheet. Here, we use anomaly coupling to force the ice-sheet model, and use more realistic values for the PDD factors (0.008 for ice and 0.003 m d$^{-1}$ K$^{-1}$ for snow), consistent with the previous simulations of the future evolution of the Greenland ice sheet (e.g. Ridley et al. 2005). For the baseline climate to which the anomalies are applied, we use a standard temperature climatology derived from observations (Ohmura 1987), and precipitation from ERA-40 reanalysis (which agrees reasonably well with the satellite and ice-core data over Greenland (Hanna et al. 2001)). The GLIMMER ice-sheet model uses a single value for the lapse-rate correction. We set this to a value of 7.0 K km$^{-1}$, which is intermediate between the summer and annual mean values of 7.992 and 6.277 as defined in the EISMINT3 standard (Huybrechts 1998). For consistency with the previous work (e.g. Ridley et al. 2005), we use the Greenland bedrock topography of Letreguilly et al. (1991), as opposed to that of Bamber et al. (2001) in our previous study. We use initial conditions for the ice-sheet model simulations, which are as consistent as possible with the prescribed ice sheets in the corresponding GCM simulation. The initial condition for the ice-sheet simulations derived from GCM simulations with a Pliocene ice sheet ($Plio$ and $Plio_{modCO_2}$) is a state with no ice and a relaxed bedrock topography. The initial condition for the ice-sheet simulations derived from GCM simulations with a present-day ice sheet ($Mod_{400}$ ppmv, $Mod_{560}$ ppmv and $Mod_{1120}$ ppmv) is the present-day observed configuration of the ice and bedrock (Letreguilly et al. 1991). The length of the ice-sheet simulations is 50 000 years, as opposed to 20 000 in the previous Lunt et al. (2008a) study; this is to allow equilibrium to be reached in the Pliocene simulations. Additionally, we use a more recent version of the GLIMMER ice-sheet model (v. 1.0.4 as opposed to 0.5.6), which includes, among other changes, a bug fix related to the surface mass balance scheme.

3. Experimental design

In order to assess the relative importance of different processes that have determined the evolution of the Arctic cryosphere since the Pliocene, we first carry out an ensemble of GCM simulations. In this ensemble, we move from the
climate of the pre-industrial to that of the Mid-Pliocene in a series of incremental steps. The two extreme members of this ensemble are a standard pre-industrial simulation and a standard Mid-Pliocene simulation (Haywood & Valdes 2004; Lunt et al. 2008a, b). The differences between these two simulations are the atmospheric CO2 concentration, and the surface boundary conditions: two-dimensional fields that are prescribed and stay fixed during the simulations. There are in fact 22 such surface boundary conditions that differ between the two simulations; examples include snow-free vegetation albedo, orography and heat capacity of the soil. It is clearly not possible to vary each one of these independently (more than four million simulations), or even sequentially (23 simulations). Instead, we group them together into classes, together representing similar processes. We choose two such classes: (i) vegetation and soils, and (ii) orography and ice. Together with the atmospheric CO2 concentration, this makes three classes; moving sequentially between these classes gives us a total of four simulations.

In order to assess the impacts of future elevated CO2 concentrations, and the relevance of the Mid-Pliocene as an analogue for future climate change, we carry out three additional GCM simulations with pre-industrial boundary conditions, but atmospheric CO2 set at 400, 560 and 1120 ppmv, respectively.

The boundary conditions associated with the total of seven simulations are summarized in table 1.

Table 1. Summary of GCM simulations. (M is for pre-industrial (modern) boundary conditions, P is for Mid-Pliocene boundary conditions and F is for future boundary conditions (F1 is 560 ppmv CO2 and F2 is 1120 ppmv CO2.).)

<table>
<thead>
<tr>
<th>simulation name</th>
<th>CO2</th>
<th>orography and ice</th>
<th>vegetation and soils</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mod</td>
<td>M</td>
<td>M</td>
<td>M</td>
</tr>
<tr>
<td>Modplioveg</td>
<td>M</td>
<td>M</td>
<td>P</td>
</tr>
<tr>
<td>Plio_modCO2</td>
<td>M</td>
<td>P</td>
<td>P</td>
</tr>
<tr>
<td>Plio</td>
<td>P</td>
<td>P</td>
<td>P</td>
</tr>
<tr>
<td>Mod400 ppmv</td>
<td>P</td>
<td>M</td>
<td>M</td>
</tr>
<tr>
<td>Mod560 ppmv</td>
<td>F1</td>
<td>M</td>
<td>M</td>
</tr>
<tr>
<td>Mod1120 ppmv</td>
<td>F2</td>
<td>M</td>
<td>M</td>
</tr>
</tbody>
</table>

Mid-Pliocene and future Arctic cryosphere

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The difference between the prescribed pre-industrial and Mid-Pliocene vegetation and soils is shown in figure 1, in terms of the snow-free albedo in the two cases. For the pre-industrial, this is derived from the Wilson & Henderson-Sellers (1985) dataset of land coverage; for the Mid-Pliocene, this is derived from the PRISM2 reconstruction (Thompson & Fleming 1996; Dowsett et al. 1999). The largest differences are in the Arctic high latitudes, where there are bare soil/tundra/glaciers (relatively high albedo) in the pre-industrial, and boreal forest (lower albedo) in the Mid-Pliocene. This is particularly noticeable on the margins of Greenland, the Canadian Archipelago and southern Alaska. Outside of Arctic high-latitude regions, the biggest change is in the Sahara, where Mid-Pliocene vegetation encroaches into regions that are desert in the pre-industrial.

The difference between the pre-industrial and Mid-Pliocene orography and ice is shown in figure 2, in terms of the surface height and ice-sheet extent. Again, the Mid-Pliocene dataset is from the PRISM2 reconstruction. The most significant changes in terms of surface height in the Northern Hemisphere are
the Rockies, which are higher by approximately 1000 m in the pre-industrial compared with the Mid-Pliocene, and over Greenland itself, where the increase in height is due to the larger pre-industrial ice sheet, as well as some uplifted topography around the margins of Greenland, consistent with the data indicating Late Cenozoic uplift around the North Atlantic (Japsen & Chalmers 2000).

4. Results

Figure 3 shows the time evolution of the seven GCM simulations described above, in terms of global annual mean surface air temperature. The pre-industrial control has been integrated for approximately 1000 years, and the Mid-Pliocene control for over 1050 years. The other simulations are between 200 and 900 years long. It can be seen that the simulations with pre-industrial boundary conditions and high CO$_2$ ($\text{Mod}_{400 \text{ ppmv}}$, $\text{Mod}_{560 \text{ ppmv}}$ and $\text{Mod}_{1120 \text{ ppmv}}$) are approaching equilibrium, but there is still a residual trend in temperature relative to the control state. These simulations should therefore be regarded as representing a minimum likely equilibrium future change, given their CO$_2$ forcing. In order to obtain a better degree of equilibrium in the $\text{Mod}_{1120 \text{ ppmv}}$ case, we used a version of the GCM with a lower ocean resolution (HadCM3L). We also repeated the $\text{Mod}_{560 \text{ ppmv}}$ simulations with the lower-resolution model, and found only small differences in the mean Arctic climate (in table 2, a change of 2.4°C compared with 2.8°C), and similar ice-sheet volumes when the climatologies were used to force the ice-sheet model (to within 4%). We therefore consider the HadCM3 and HadCM3L simulations to be comparable. There is an apparent slight upward trend in the Mid-Pliocene control simulation ($\text{Plio}$). This is due to the fact that the pre-industrial and the Mid-Pliocene simulations have been carried out in phases over a
number of years of ‘real-world’ time, on a number of different architectures and
with a number of different compilers. Changes between machines and compilers
introduce small variations in climate sensitivity. The important point here is that
the last 200 years (at least) of all the HadCM3 simulations described in this paper
were carried out on the same machine and with the same compiler. Therefore, the
mean climatologies of all the simulations (calculated over the 30-year period
indicated with thick black lines in figure 3) are fully intercomparable.

Figure 2. Orography and ice boundary conditions. (a,b) Northern Hemisphere orography in (a) the
pre-industrial and (b) the Mid-Pliocene. (c,d) Ice sheet extent in (c) the pre-industrial and
(d) the Mid-Pliocene.

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We focus our discussion here on the climatic changes that occur in the North Atlantic and Greenland, in terms of temperature, precipitation, sea ice and ice-sheet volume. A subsequent paper will discuss the ensemble of Mid-Pliocene simulations from a global perspective. Table 2 summarizes the results of the simulations, which are discussed in detail below.

Table 2. Changes in Atlantic sector Arctic (55° N: 85° N, 85° W: 5° W) summer (June–August) temperature (°C), annual mean precipitation (mm d⁻¹) and ice volume (as a percentage of the modern control ice volume) between pairs of simulations representing the three processes investigated: vegetation and soils, orography and ice, and CO₂.

<table>
<thead>
<tr>
<th>process and soils (Mod to Plio)</th>
<th>Mod_{plioveg} - Mod</th>
<th>temp_{JJA} (°C)</th>
<th>precip_{ANN} (mm d⁻¹)</th>
<th>ice volume (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>vegetation and soils</td>
<td>+1.2</td>
<td>+0.09</td>
<td>-3.8</td>
<td></td>
</tr>
<tr>
<td>orography and ice (Mod to Plio)</td>
<td>Plio_{modCO₂} - Mod_{plioveg}</td>
<td>+2.2</td>
<td>+0.18</td>
<td>-2.2</td>
</tr>
<tr>
<td>CO₂ (280 to 400 ppmv)</td>
<td>Plio - Plio_{modCO₂}</td>
<td>+1.8</td>
<td>+0.24</td>
<td>-78</td>
</tr>
<tr>
<td>CO₂ (280 to 560 ppmv)</td>
<td>Mod_{400 ppmv} - Mod</td>
<td>+1.1</td>
<td>+0.11</td>
<td>-1.6</td>
</tr>
<tr>
<td>CO₂ (280 to 1120 ppmv)</td>
<td>Mod_{1200 ppmv} - Mod</td>
<td>+5.0</td>
<td>+0.59</td>
<td>-88</td>
</tr>
</tbody>
</table>

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(a) Vegetation and soils

Figure 4a shows the temperature difference between the Mod_{plioveg} and Mod simulations in June–July–August (JJA). This shows the response of the system to a change from the pre-industrial to the Mid-Pliocene vegetation and soil properties. The focus on the boreal summer season is because this is when the majority of ablation occurs on the Greenland ice sheet. The most significant change is a warming in the northeast of Greenland and in Ellesmere Island. This is a direct result of the change from tundra and bare soil in the pre-industrial to boreal forest in the Mid-Pliocene. This decreases the albedo (figure 1), in particular when the surface is covered in deep snow (from 0.8 to 0.3), leading to greater absorption of solar radiation and surface warming. There is less Arctic sea ice in Mod_{plioveg} than in the Mod simulation, particularly on the coastal regions of the Northern Hemisphere Arctic continents. However, there is still polar sea ice present in summer in the Mod_{plioveg} simulation. The sea-ice changes act to amplify the temperature response to the vegetation change via the albedo feedback. Over the region shown in figure 4, the mean temperature change following the change in vegetation and soils is an increase of 1.2°C. Figure 4d shows the annual mean precipitation anomaly following the change in vegetation and soils. Much of the Northern Hemisphere becomes wetter, and this is true also over much of the Atlantic sector and Greenland itself. In fact, there is a strong correlation between temperature and precipitation change in all the anomalies presented in this paper, resulting from increased evaporation in warmer climates. The largest increase in precipitation over Greenland is in the northeast, where the temperature anomaly is also the greatest. There is a
decrease in precipitation over the southeastern margins of Greenland, which extends into the North Atlantic.

Figure 5a shows the pre-industrial ice sheet, as predicted by GLIMMER, given the observations of modern temperature and precipitation. This is the ice sheet associated with the Mod climate. Figure 5b shows the ice sheet associated with the Mod\textsubscript{plioveg} climate, which is obtained by forcing GLIMMER with observations, plus an anomaly calculated from the difference between the Mod\textsubscript{plioveg} and Mod climates. The loss of mass compared with the Mod case is equivalent to 4 per cent of the total Mod ice volume, and occurs in the north of the island in the regions of the greatest temperature anomaly. In the centre of the ice sheet, where the temperature change is relatively small and ablation is reduced due to the low temperatures, the height of the ice sheet is almost unchanged from the pre-industrial state.

\begin{enumerate}
  \item \textit{Orography and ice}

  Figure 4b shows the temperature difference between the Plio\textsubscript{modCO2} and Mod\textsubscript{plioveg} simulations, in JJA. This shows the response of the system to a change from pre-industrial orography and ice sheets to Mid-Pliocene orography and ice sheets. The largest change in temperature in the region considered occurs over the Greenland ice sheet itself. The temperature increase is caused by three factors: the decrease in altitude due to the lower ice-sheet elevation; the decrease in albedo due to the less extensive ice sheet; and a decrease in temperature due to the upwind orography change in the Rockies. Comparison with previous studies (e.g. Lunt \textit{et al.} 2004, 2008\textsuperscript{b}) indicates that the response to the local orography and albedo changes is likely to dominate, and to be of similar magnitude in JJA. Over the region shown in figure 4, the mean temperature change following the change in orography and ice is an increase of 2.2\textdegree C. The associated precipitation change (figure 4e) shows increased precipitation over much of the warmer ice-sheet surface. However, there is a decrease in precipitation over the southern tip of Greenland, which is clearly not directly related to temperature. Instead, it is related to the fact that, in Mod\textsubscript{plioveg}, the presence of the high-altitude ice sheet in the poleward margins of the North Atlantic storm track leads to enhanced precipitation, as the moist air masses rise. In Plio\textsubscript{modCO2} simulation, the ice-sheet surface is lower, and so there is less orographically induced precipitation. When the Plio\textsubscript{modCO2} climate is used to force the ice-sheet model, the resulting Greenland ice-sheet geometry is as shown in figure 5c. Compared with the Mod\textsubscript{plioveg} ice sheet in figure 5b, there is a further loss of mass, equivalent to 2 per cent of the total Mod ice volume. The main loss of mass again occurs in the northern regions of Greenland. Interestingly, there is an increase in ice-sheet elevation around the summit of the ice sheet, probably related to the different configuration of the ice sheet in the GCM. The increase in precipitation could also contribute to this increased height.

  \item \textit{CO\textsubscript{2}}

  Figure 4c shows the temperature difference between the Plio and Plio\textsubscript{modCO2} simulations, in JJA. This shows the response of the system to a change from the pre-industrial to the Mid-Pliocene atmospheric CO\textsubscript{2} concentrations (280 to
Over the region shown in figure 4, the mean temperature change following the change in CO₂ is 1.8°C. This greenhouse warming extends over most of the globe, and is the greatest at Antarctic high latitudes due to sea ice and albedo feedbacks. There is a slight cooling in the Barents Sea, due to a decrease in meridional heat transport by the Atlantic Ocean, associated with an increase in sea ice. All these changes are consistent with the findings from many previous studies, which have investigated the response of the climate system to various greenhouse gas emissions. Figure 4 shows the surface climate anomalies associated with a change from pre-industrial to Mid-Pliocene. The simulated surface temperature change for the Modplioveg_−_Mod, Plio_−_Mod, and Plio_−_Mod CO₂ scenarios are displayed in panels (a), (b), and (c), respectively. Panels (d), (e), and (f) depict the simulated precipitation change for the Modplioveg_−_Mod, Plio_−_Mod, and Plio_−_Mod CO₂ scenarios, respectively.
increases in CO₂ (e.g. Solomon et al. 2007). The high CO₂ leads to enhanced sea-ice melting at the North Pole in summer; however, the pole still does not become ice free at any stage in the year in Plio simulation. Also in agreement with previous work, the increase in CO₂ leads to an increase in precipitation in the region considered (figure 4f). However, in terms of ice-sheet response, the temperature change dominates the precipitation change, and is sufficient to melt the majority of the Greenland ice sheet (figure 5d compared with figure 5c). This additional mass loss is equivalent to approximately 80 per cent of the total Mod ice volume. The ice sheet associated with the Plio climate is constrained to the regions of relatively high altitude in the east, where ablation is limited. The large difference between the Plio and Plio modCO₂ ice sheets indicates that a decline in CO₂ controlled the enhancement of Greenland glaciation in the Late Pliocene. This is in agreement with our recent study (Lunt et al. 2008b), which compared the importance of CO₂ in controlling the development of the Pliocene Greenland ice sheet with three other mechanisms: Panama Seaway closure; tectonic uplift; and termination of a permanent El Niño. The Mod, Plio and Plio modCO₂ GCM simulations presented in this paper are all continuations of the equivalent simulations in the Lunt et al. (2008b) paper.

It is also instructive at this point to make a comparison between the GCM and ice-sheet Plio simulations with some observations of the climate of the Mid-Pliocene, in order to evaluate the model performance. On a global scale, the model does a reasonable job relative to the estimates of Mid-Pliocene SST from faunal analysis of foraminifera (Haywood & Valdes 2004). The agreement improves further when the alkenone data from the tropics are also included (Haywood et al. 2005). Hill et al. (2007) provided a summary of reconstructions of Mid-Pliocene continental Arctic temperatures. Comparison with these data (not shown) indicates that we have reasonable agreement with the observations, in particular in the Canadian Archipelago where both the model and the data indicate temperatures of the order of 10–15°C greater than modern. However, work by Cronin et al. (1993) pointed to an ice-free Arctic in Mid-Pliocene summer, whereas we have permanent sea ice at the North Pole in Plio simulation. Additionally, comparison with SST estimates from foraminiferal
assemblages (Dowsett et al. 1999; Dowsett 2007) points to a cold bias in the GCM of the order of 2–4°C in the Arctic, and approximately 6°C in the Barents Sea. However, uncertainties do exist with all of the data reconstructions, especially in terms of spatial scale (the data represent a point location whereas the model represents an average over a large area). Nonetheless, we do obtain good agreement with the data in terms of the extent of the Mid-Pliocene Greenland ice sheet. Hill et al. (2007) present data from several sites in Greenland that indicate ice-free conditions at these locations. In all of these regions (Dye 3, GRIP, North GRIP, Ile de France and Kap Kobenhavn), we also have ice-free conditions. In summary, given the apparent cold Arctic bias of the GCM, it appears that our simulated absolute Mid-Pliocene temperatures should be interpreted as representing a minimum warming relative to modern.

(d) Future (CO₂ = 400, 560 and 1120 ppmv)

Figure 6a shows the temperature difference between the Mod₄₀₀ ppmv and Mod simulations, in JJA. Similar to §4c, this shows the response of the system to a change from pre-industrial to Mid-Pliocene CO₂ (280 to 400 ppmv), but in this case, the CO₂ change is applied under otherwise pre-industrial conditions, as opposed to Mid-Pliocene conditions. Interestingly, the change in temperature is not so large in this case, 1.1°C compared with 1.8°C (figure 6a compared with figure 4c). The reason for this is probably that the more extensive Greenland ice sheet of the pre-industrial relative to Mid-Pliocene means that the snow–albedo feedback is less strong, as the underlying surface has high albedo even if there is no snow cover. It is also interesting to compare the absolute Mod₄₀₀ ppmv Arctic climate with that of the Mid-Pliocene (Plio simulation). Even ignoring the Greenland region itself, where the differences in surface temperature are partly due to the different elevation of the ice-sheet surface, the absolute summer temperature in Mod₄₀₀ ppmv simulation is 4.1°C, which is cooler than the 6.1°C in Plio simulation. This shows that in the short term, before vegetation and ice-sheet feedbacks have amplified the climate change, the Arctic climate of the future is likely to be cooler than that of the Mid-Pliocene, even if they have the same atmospheric CO₂ concentration. As expected, there is a decrease in the summer and autumn sea ice in the Arctic in Mod₄₀₀ ppmv simulation compared with pre-industrial. There is an increase in precipitation in the North Atlantic sector, and in particular enhanced precipitation in the southern and eastern margins of the Greenland ice sheet (figure 6d). Further increasing CO₂ to 560 ppmv (figure 6b,e) and 1120 ppmv (figure 6c,f) leads to further warming and further increased precipitation. The spatial pattern of change is very similar in the three cases. In terms of Arctic sea ice, there is still summer sea ice with CO₂ at 560 ppmv, but an increase up to 1120 ppmv leads to ice-free conditions in summer. Again, ignoring Greenland itself, the absolute summer temperature in the region shown in figure 6 is 5.0°C in Mod₅₆₀ ppmv simulation, and 8.1°C in Mod₈₁₂₀ ppmv simulation (again, for comparison with 6.1°C in the Mid-Pliocene). Therefore, it is not until atmospheric CO₂ stabilizes above 560 ppmv that the short-term future Arctic climate begins to resemble that of the Mid-Pliocene.

Despite the temperature increase, the equilibrium Greenland ice sheet under 400 ppmv is not reduced a great deal relative to modern (less than 2% of the total Mod ice volume, figure 7a). This is very different from the case of the
Mid-Pliocene, where a change in CO$_2$ from 280 to 400 ppmv led to a significant decrease in ice volume of 78 per cent. It shows that the present-day ice sheet is more stable than the Mid-Pliocene ice sheet to an equivalent CO$_2$ forcing, due to vegetation and ice/climate feedbacks. Also note that, whereas the ice-sheet response to a change in CO$_2$ in the Mid-Pliocene was simulated with an ice-free initial condition, in the future case this was simulated with a glaciated initial condition. The additional warming associated with the doubling of CO$_2$ from pre-industrial values results in increased melting of the Greenland ice sheet (equivalent to 7% of the total $Mod$ ice volume, figure 7b). The further warming associated with the quadrupling of CO$_2$ from pre-industrial values results in a collapsed Greenland ice sheet (total ice volume loss equivalent to approx. 90% of that of $Mod$, figure 7c). This implies that, under otherwise pre-industrial conditions, there is a critical value for CO$_2$, somewhere between 560 and 1120 ppmv, past which the Greenland ice sheet will become unstable. However, because we have not fully coupled the ice sheet and climate models, there is some uncertainty in these values (see §5).

In contrast to the Mid-Pliocene, where some degree of confidence was gained in our results by comparing with palaeodata, here we compare our future climate simulations with previous modelling work. Ridley et al. (2005) carried out a future climate/ice-sheet simulation under the same quadrupled CO$_2$ as our $Mod_{1120}$ ppmv simulation. They used the same GCM, but used a different ice-sheet model, and asynchronously coupled the GCM and ice-sheet models, using 1 year of GCM output to drive the ice-sheet model for 10 years. The resulting Greenland ice sheet, after more than 3000 years of simulation, collapsed to 7 per cent of the original volume. This is for comparison with 12 per cent in our $Mod_{1120}$ ppmv ice-sheet simulation. The extra ice in our simulation is situated in the south of the island (figure 7c), a region that is ice free in the simulation of Ridley et al. (2005). Driesschaert et al. (2007) simulated the response of the Greenland ice sheet to a variety of CO$_2$ stabilization scenarios, including ones similar to our $Mod_{120}$ ppmv (their IS4CO2), our $Mod_{560}$ ppmv (their IS2CO2) and our $Mod_{400}$ ppmv (their SP350 or SP450) simulations. A direct comparison with their results is not possible due to the fact that the equilibrium states of their simulations are not presented; rather, they give the ice volume after 1000 years of integration. However, they do note that their IS4CO2 scenario results in a complete melting of Greenland after 3000 years, and it appears from their results that in the IS2CO2 case there is less than 10 per cent melting of the ice sheet, and in the SP350 and SP450 cases, closer to 2 per cent. All of these results are consistent with our own findings. Mikolajewicz et al. (2007) carried out simulations with two times CO$_2$ and four times CO$_2$, and integrated their fully coupled climate/ice-sheet model for 600 years. Again, direct comparison is not possible, but at the end of their simulations, the two times CO$_2$ ice sheet was reduced by approximately 10 per cent, and the four times CO$_2$ ice sheet by approximately 35 per cent (and still decreasing rapidly). In the two times CO$_2$ case, the effect of doubling CO$_2$ appears to have more of an effect on the Greenland ice sheet than in our simulations, and could lead to substantial melting in the longer term. Therefore, the threshold that we obtain for substantial Greenland melt, being somewhere between 560 and 1120 ppmv, should be regarded with some caution.
In this section, we discuss several aspects of this work, mostly related to the experimental design. Firstly, we have not coupled together the GCM and ice-sheet models. In reality, there are a number of feedbacks between ice and climate, albedo and altitude being the most important. Account is taken of the temperature–altitude feedback, because the temperature predicted by the GCM is lapse-rate corrected by the ice-sheet model, allowing, for example, the surface temperature of the ice sheet to decrease as the ice sheet grows and increases in altitude. However, no account is taken of the albedo feedback, and in the model a growing ice sheet does not influence the surface energy budget through albedo changes. GCMs are currently too computationally expensive to allow full synchronous coupling with an ice-sheet model, but asynchronous coupling could be implemented in future work, especially if a lower-resolution version of the GCM were used.

Ice-sheet models themselves are currently undergoing a period of development, with particular attention being paid to improved representation of ice dynamics (e.g. Pattyn 2003), the accurate representation of ice streams and their coupling with ice shelves (e.g. Pattyn et al. 2006; Schoof 2006, 2007), improved modelling of surface mass balance (e.g. Bougamont et al. 2007) and treatments of basal sliding that take potential positive feedbacks into account (e.g. Parizek & Alley 2004; Price et al. 2008). As such, future development of the ice-sheet model to improve the representation of these processes may lead to different behaviour under warm climate conditions.

Since the publication of the PRISM2 dataset, more recent data have indicated some inadequacies with the Mid-Pliocene boundary conditions applied in the GCM. In particular, it appears that the height of the Mid-Pliocene US Rockies is
probably too low, although the Mid-Pliocene Canadian Rockies are well represented (G. Foster 2008, unpublished data). Also, Salzmann et al. (2008) have produced a more recent representation of global Mid-Pliocene vegetation, which has some significant differences from that presented in PRISM2, including at high Northern Hemisphere latitudes. Future work should look at the effect of including these new boundary conditions in the GCM. Additionally, sensitivity studies to the level of prescribed atmospheric CO₂ should be performed, given the uncertainties in reconstructing pre-Quaternary trace gas concentrations (e.g. Kurschner et al. 1996; Raymo et al. 1996; Pearson & Palmer 2000).

Simulations from the UK Met Office in the recent IPCC fourth assessment report (Solomon et al. 2007) were carried out with not only HadCM3 but also HadGEM (Johns et al. 2006). HadGEM runs at a higher resolution than HadCM3, and includes several improvements to the model physics. It is possible that a higher-resolution GCM would improve the simulation of Mid-Pliocene climate, in particular the cold bias at high latitudes, by increasing the poleward heat transport, thereby leading to enhanced polar warmth. The higher resolution would also lead to a better representation of orographic effects, and allow a better matching of resolution with the ice-sheet model. However, HadGEM is much more computationally expensive than HadCM3, and a balance is needed between resolutions and, given computational constraints, the tunability of the control climate and attainability of equilibrium.

6. Conclusions

We have simulated the climate of the pre-industrial and the Mid-Pliocene, and carried out an ensemble of simulations that attribute the warmer Arctic climate and reduced ice-sheet volume of the Mid-Pliocene to a variety of mechanisms. We have shown that the warmer Mid-Pliocene Arctic climate can be attributed to increased
CO₂ (35%), lower orography and ice-sheet feedbacks (42%), and vegetation changes (23%). The Mid-Pliocene Greenland ice sheet is much reduced relative to modern, mostly due to the high levels of CO₂; reducing CO₂ to pre-industrial values under otherwise Pliocene conditions leads to a Greenland ice sheet that is similar in volume to that of modern. In terms of future climate change, our results indicate that a CO₂ stabilization at 400 ppmv would maintain some 98 per cent of the current ice-sheet volume in the short term, and stabilization at 560 ppmv would maintain some 93 per cent of the current ice-sheet volume, whereas an increase to 1120 ppmv would result in a large reduction to only 12 per cent of the current volume. However, the exact threshold value of CO₂ should be regarded with caution, due to the lack of interactive coupling of the GCM and ice-sheet model, and possible missing physical processes in the ice-sheet model.

In terms of the suitability of the Mid-Pliocene as an analogue for climate change at the end of this century, we have shown that, in terms of the Arctic cryosphere, the Mid-Pliocene climate (with an atmospheric CO₂ concentration of 400 ppmv) is warmer than a future climate simulation with stabilization at the same level of CO₂. In our case, it is not until CO₂ is raised above 560 ppmv that the future climate starts resembling that of the simulated Mid-Pliocene, and the Greenland ice sheet collapses and is confined to the eastern mountains. The reason for this is mostly a result of differences in time scale between the century-scale future and the more equilibrated Mid-Pliocene. This finding is also supported by biome modelling presented in Salzmann et al. (2009). The prescribed Mid-Pliocene vegetation is in equilibrium with the warm conditions, and intensifies warming in the Arctic. This process is not accounted for in the future simulations, which is realistic for the century-scale time scales we are considering here. Similarly, the reduced size of the Greenland ice sheet in the Mid-Pliocene GCM simulations essentially represents ice-albedo feedback acting over thousands of years, whereas the future climate simulations do not represent this process, again, probably realistic for the century-scale time scale considered here. However, in the longer-term future, it is likely that a stabilization of CO₂ at relatively modest values (approx. 400 ppmv) will lead to climate resembling that of the Pliocene more closely, as long time-scale vegetation–climate–ice-sheet interactions amplify the CO₂-only forcing. This possibility should be investigated with fully coupled climate–ice-sheet–vegetation models, as millennial-scale simulations become practicable as a result of increasing computational speed.

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