Results are presented from the Community Climate System Model, version 4 (CCSM4), simulation of the Last Glacial Maximum (LGM) from phase 5 of the Coupled Model Intercomparison Project (CMIP5) at the standard 1° resolution, the same resolution as the majority of the CCSM4 CMIP5 long-term simulations for the historical and future projection scenarios. The forcings and boundary conditions for this simulation follow the protocols of the Paleoclimate Modeling Intercomparison Project, version 3 (PMIP3). Two additional CCSM4 CO₂ sensitivity simulations, in which the concentrations are abruptly changed at the start of the simulation to the low 185 ppm LGM concentrations (LGMCO₂) and to a quadrupling of the preindustrial concentration (4×CO₂), are also analyzed. For the full LGM simulation, the estimated equilibrium cooling of the global mean annual surface temperature is 5.5°C with an estimated radiative forcing of −6.2 W m⁻². The radiative forcing includes the effects of the reduced LGM greenhouse gases, ice sheets, continental distribution with sea level lowered by approximately 120 m from the present, and orbital parameters, but not changes to atmospheric aerosols or vegetation biogeography. The LGM simulation has an equilibrium climate sensitivity (ECS) of 3.1(±0.3)°C, comparable to the CCSM4 4×CO₂ result. The LGMCO₂ simulation shows a greater ECS of 4.2°C. Other responses found at the LGM in CCSM4 include a global precipitation rate decrease at a rate of −2% °C⁻¹, similar to climate change simulations in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4); a strengthening of the Atlantic meridional overturning circulation (AMOC) with a shoaling of North Atlantic Deep Water and a filling of the deep basin up to sill depth with Antarctic Bottom Water; and an enhanced seasonal cycle accompanied by reduced ENSO variability in the eastern Pacific Ocean’s SSTs.

1. Introduction

The Paleoclimate Modeling Intercomparison Project (version 3; PMIP3) and the Coupled Model Intercomparison Project (phase 5; CMIP5) proposed that modeling groups perform a simulation of the Last Glacial Maximum (LGM) with the same models and at the same resolutions as simulations being done to simulate the twentieth century and scenarios into the future. Traditionally, simulations of past climate have been done with lower-resolution versions by the modeling groups or different versions of the models [e.g., for the Community Climate System Model (CCSM), see Shin et al. (2003) and Otto-Bliesner et al. (2003, 2006); for PMIP models, see Table 1 and Otto-Bliesner et al. (2009)]. Using the same version for both past and future climate change experiments should enhance progress in addressing some of the outstanding scientific questions of the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4).

One rationale for simulating the LGM is to allow evaluation of the model response to ice-age boundary conditions relative to paleodata. New global proxy reconstructions and syntheses have been produced in the last five years to allow assessments of spatial patterns of LGM sea surface temperature cooling (Waelbroeck et al. 2009), continental-scale changes of temperature and moisture (Bartlein et al. 2011), and changes in fire regimes (Power et al. 2008). Assessments of the reliability of the individual proxy records and techniques used for these reconstructions have also been provided.

A second purpose of the CMIP5 LGM simulations is to attempt to provide empirical constraints on global
climate sensitivity. IPCC assessments of climate sensitivity to doubled CO$_2$ have remained fairly consistent since the first assessment with the range in the IPCC AR4 stated as 2°–4.5°C with a best estimate of about 3°C with greater than 66% likelihood (Solomon et al. 2007). There is still debate on whether warm versus cold climate states have the same climate sensitivity. The PMIP2 LGM simulations give estimates of the effective climate sensitivity of 2.6°–3.1°C, but vary among the models on whether there is greater or less sensitivity at LGM than with doubled CO$_2$ (Crucifix 2006; Otto-Bliesner et al. 2009). Attempts have also been made to constrain climate sensitivity using climate models with paleorecords (Schneider von Deimling et al. 2006; Hargreaves et al. 2007; Edwards et al. 2007; Schmittner et al. 2011).

In this paper, we analyze the response of CCSM4 to the forcings of the Last Glacial Maximum. We separate out the response to just the lowered atmospheric concentrations of CO$_2$ at LGM from the full LGM forcings and boundary conditions. In addition, we assess the similarities and differences between the glacial-simulated responses compared with the CMIP5 multicentury simulation in which the atmospheric CO$_2$ is abruptly increased to 4 times its preindustrial value.

2. Model description

The Community Climate System Model is a general circulation climate model with components for the atmosphere, ocean, land, and sea ice. These components exchange state information and fluxes through a coupler. The version used in this study, CCSM4, was released to the community in April 2010. The development of CCSM4 and a documentation of its 1850 Common Era (CE) preindustrial control simulation can be found in Gent et al. (2011). A large suite of additional simulations has been completed with this version as part of CMIP5, including Last Millennium (Landrum et al. 2013) and future scenario simulations (Meehl et al. 2012). The simulations described in this paper are performed with the nominal 1° version of CCSM4.

The atmospheric model is the Community Atmosphere Model (version 4; CAM4) which uses the Lin–Rood finite volume core with a uniform resolution of 1.25° in latitude by 0.9° in longitude (Neale et al. 2013). CAM4 uses 26 vertical layers. All simulations, except the LGM simulation, use the standard CAM4 second-order divergence damping operator. The LGM simulation adopted the more scale-selective fourth-order divergence damping for high-latitude stability by alleviating grid-scale noise associated with the steeply sided ice sheets over North America and Greenland (Lauritzen et al. 2012). Improvements in CAM4 over CAM3 include a much better representation of the spatial and temporal aspects of ENSO (Richter and Rasch 2008; Neale et al. 2008; Deser et al. 2012).

The land model is version 4 of the Community Land Model (CLM4; Lawrence et al. 2012) and adopts the same horizontal resolution as CAM4. Relative to previous versions, CLM4 includes an improved hydrology and a carbon–nitrogen biogeochemistry model that can impact the seasonal and interannual vegetation phenology. Importantly for the climate change simulations presented in this study, although the plant functional type distribution is fixed at preindustrial values (Lawrence and Chase 2007), the leaf area index and vegetation height are prognostic and can be affected by the changed climate (Thornton et al. 2007). Also, the land component improvements result in better latent heat flux into the atmosphere and river runoff into the ocean, and reduce biases in the simulation of surface temperature (Gent et al. 2011).

The CCSM4 ocean component is based on the Parallel Ocean Program, version 2, of the Los Alamos National Laboratory (Smith et al. 2010). We use the standard displaced ocean grid with poles in Greenland and Antarctica and 60 levels in the vertical (Danabasoglu et al. 2012a). The nominal 1° horizontal resolution is a uniform 1.1° in longitude and variable in latitude from 0.27° at the equator to 0.54° at 33° latitude. A new overflow parameterization improves the path of the Gulf Stream and the meridional overturning circulation in the North Atlantic Ocean. These changes also improve the sea
surface temperature and salinity biases in the North Atlantic relative to CCSM3 (Danabasoglu et al. 2012a). The greater vertical resolution relative to older versions of CCSM, particularly in the upper ocean, helps to reduce the SST errors in the main upwelling regions and improve the mean and annual cycle of SST along the equator in the eastern Pacific Ocean (Gent et al. 2011).

The sea ice model is based on the Community Ice Code, version 4 (Hunke and Lipscomb 2010; Holland et al. 2012), and uses the same horizontal grid as the ocean component. It includes an improved treatment of the surface albedo and shortwave radiative transfer in the ice and overlying snowpack (Briegleb and Light 2007). These improvements to the ice model have resulted in a better simulation of Arctic sea ice extent and thickness than in CCSM3 (Jahn et al. 2012). Antarctic sea ice is more extensive in CCSM4 than in observations; however, its simulated variability compares well with observations (Landrum et al. 2012).

3. Boundary conditions and simulation setup

The forcings and boundary conditions for the LGM simulation follow the protocols of PMIP3 and CMIP5 (https://wiki.lsce.ipsl.fr/pmip3/doku.php/pmip3:design:21k:final). Those that can be estimated from reconstructions are set to values appropriate for the LGM; those that are not provided in the PMIP3 protocols use the same values as in the CCSM4 CMIP5 1850 CE preindustrial (PI) control simulation (Gent et al. 2011).

Changes to the ocean model overflow parameterizations were made in the LGM simulation because of the increased land areas and ice shelves relative to the PI simulation. The PI simulation has two overflow regions in the Southern Hemisphere (SH) defined in the Ross and Weddell Seas, which have been removed for the LGM simulation because the LGM ice shelves extend over the specified source regions in the overflow parameterization. The two overflow regions specified in the NH, in the Denmark Strait and Iceland–Scotland region, are retained as specified in the PI simulation. However, to alleviate an advective instability near the location of the Denmark Strait overflow caused by a narrowing of
the Nordic Seas, coupled with the impact of strengthened winds and a more southward displacement of the sea ice edge, the number of levels from the bottom over which the tidal mixing parameterization is applied is increased from 2 to 6 in the LGM simulation.

The initial conditions for temperature and salinity are taken from restart files from the CCSM4 simulation with CO₂ concentrations lowered to the LGM value, 185 ppm, as discussed below, after 600 years of simulation were completed. As per PMIP3 protocols, an offset of 1 psu is added to the salinity globally to account for the increase in ice volume. All prognostic fields associated with velocity (pressure fields, baroclinic velocities) are zeroed out at startup. The LGM simulation was run for 600 years, and then a minor change was made to the North American eastern coastline in the mid-Atlantic region in order to improve the anomalously large vertical velocities found in single grid boxes near the coast. These large vertical velocities caused an unrealistic latitudinal jump in the maximum Atlantic meridional overturning circulation (AMOC). After restoring the coastline and bathymetry along the western boundary of the North Atlantic to the preindustrial condition, the simulation was continued for another 400 years. The northward ocean heat transport in the North Atlantic was not affected, even though the change reduced the maximum AMOC by \( \sim 10 \) Sv (1 Sv \( = 10^6 \) m³ s⁻¹).

In addition to the CCSM4 LGM simulation, we analyze two other CCSM4 experiments where the only radiative forcing is a change to the concentrations of CO₂ at the start of the run, LGMCO₂ and 4×CO₂. The atmospheric CO₂ level is lowered to 185 ppm in LGMCO₂, with an initial condition taken from the preindustrial control. In 4×CO₂, the concentration of atmospheric CO₂ is increased to 4 times the preindustrial control level. All other forcings and boundary conditions are unchanged from the PI control simulation, described in more detail in Gent et al. (2011). For quantifying the responses shown here, averages are computed over the last 30 years of each integration unless otherwise noted. The experiments are summarized in Table 1.

4. Comparison with proxy glacial reconstructions

To validate the LGM CCSM simulation, we compare the mean annual LGM surface cooling relative to the PI control with the latest available proxy reconstructions (Fig. 2). To compare with proxy reconstructions of SST, the annual mean surface temperature from the atmospheric model is replaced by the ocean freezing
temperature of sea ice, \(-1.8^\circ C\) in sea ice regions. Reconstructed LGM \(\rightarrow\) PI SST differences from the Multi-proxy Approach for the Reconstruction of the Glacial Ocean Surface (MARGO) reconstruction (Waelbroeck et al. 2009) and mean annual temperature (MAT) differences from the land-based reconstruction of Bartlein et al. (2011) are overlaid for comparison. Additional land-based proxy data from Schmittner et al. (2011), available at http://mgg.coas.oregonstate.edu/~andreas/data/schmittner11sci/ and not included in Bartlein et al. (2011), are also shown, including estimates from ice cores over Greenland and Antarctica.

The surface temperature anomaly relative to the PI control of the CCSM4 LGM simulation captures the broad spatial response suggested in the proxy reconstructions, especially when comparing the proxy data with model output sampled at the proxy data locations and plotted as a function of latitude as shown in Fig. 3. The model, however, fails to produce the glacial warming found at some of the proxy data sites (Figs. 2 and 3) such as over northern Alaska in the terrestrial proxy data and in the Nordic Seas in the proxy SST data. Over Alaska, the LGM surface temperature response is sensitive to uncertainties in the prescription of the Laurentide ice sheet, which can cause large changes in atmospheric circulation as found in previous CCSM3 LGM simulations (Otto-Bliesner et al. 2006). The notable warming in the proxy reconstructions in the Nordic Seas is found in or near regions covered by sea ice in the model and is associated with large uncertainties (Waelbroeck et al. 2009).

The LGM surface air temperature at the nearest model grid point to the Summit, Greenland, ice core site is colder than the PI control by \(8.6^\circ C\) and at Vostok station in East Antarctica by \(9.9^\circ C\). The simulated LGM cooling at the Vostok site agrees well with the estimate of \(\sim 9(\pm 2)^\circ C\) from Antarctic reconstructions (Masson-Delmotte et al. 2006); however, the simulated Summit, Greenland, cooling underestimates the \(19^\circ\)–23\(^\circ\)C cooling estimated by borehole reconstructions (Dahl-Jensen et al. 1998; Masson-Delmotte et al. 2006). The true elevations at Summit Greenland and Vostok are well resolved by the \(\sim 1^\circ\) CAM4 orography (Table 2), so no correction for elevation is done to the temperature here (Table 2). Over land, the CCSM4 LGM shows greater cooling at the southeastern border of the Scandinavian ice sheet in comparison with the proxy reconstruction.

Despite some large local model–proxy data discrepancies, estimates of regional mean LGM cooling show good agreement with regional means estimated from the proxy data (Table 3) although all regions show a weak cold bias in the simulated cooling. When computed on the proxy data grid, the global mean CCSM4 LGM land region cools by \(6.5^\circ C\) compared with a cooling of \(6.1(\pm 1.5)^\circ C\) estimated from the Bartlein et al. (2011) proxy data. The CCSM4 LGM terrestrial cooling estimated on the proxy data grid is less than the cooling of \(7.5^\circ C\) computed for all land areas on the full model grid. There is good agreement as well over the ocean regions where the Waelbroeck et al. (2009) proxy data indicates a global mean LGM SST cooling of \(2.1(\pm 1.8)^\circ C\), computed on the proxy data grid, compared with \(2.7^\circ C\) obtained for the CCSM4 LGM SST cooling computed on the same grid. The CCSM4 cooling on the proxy grid is slightly greater than the LGM SST cooling of \(2.4^\circ C\) computed for the full ocean grid.
The NH extratropical regional mean (30°–90°N) terrestrial cooling of 8.2°C, obtained for the model sampled at the proxy data locations, agrees well with the mean cooling of 7.5(±1.6)°C estimated from the proxy data over the same region. The regional NH extratropical cooling estimated using model output sampled at only the proxy locations, however, underpredicts the total LGM cooling of 12.5°C averaged over all land areas from 30° to 90°N. This is because the proxy locations miss large regions of strong surface cooling in the LGM simulation exceeding ~30°C, found over the upper elevations of the NH ice sheets (Fig. 2). The CCSM4-simulated glacial cooling on the full model grid over NH extratropical land regions is also colder than estimated by Bintanja et al. (2005), who reconstructed a cooling of 7.5(±1.5)°C by an inversion of a global sea level record in conjunction with an ice sheet/shelf/bedrock model. The simulated LGM SST cooling in the NH extratropics of 3.7°C, averaged over the proxy-sampled locations, agrees well with the LGM SST cooling of 3.2°C obtained for the MARGO proxy SST reconstruction.

Averaged over the entire tropical ocean domain (30°N–30°S), the simulated LGM SST cooling is 2.1°C (Table 3). Averaged over the proxy SST sites (from 30°S...
to 30°N), the simulated tropical LGM SST cooling of 2.2°C is greater but within the range of cooling of 1.5 (±1.2)°C estimated by the MARGO reconstruction (Waelbroeck et al. 2009). However, the model tropical LGM SST cooling is somewhat weaker than the cooling found by Ballantyne et al. (2005) of 2.7(±0.5)°C. The simulated tropical SST cooling pattern does not reproduce the basin-scale heterogeneity found in the reconstructions with greater cooling in the Atlantic basin and relatively little cooling, or even weak warming in the central Pacific (Waelbroeck et al. 2009). However, the CCSM4 LGM predicts greater cooling in the eastern equatorial basins, as suggested by the MARGO proxy data (Fig. 2) and discussed later (Fig. 14).

Computed using output at only the Bartlein et al. (2011) proxy data locations, the averaged simulated tropical (30°S–30°N) terrestrial LGM cooling of 3.5°C agrees with the tropical cooling of 3.1(±1.2)°C found for the Bartlein et al. (2011) pollen-based proxy temperatures. These area-weighted averages agree well with the LGM cooling of 2.9°C, computed over all model tropical land regions from 30°N to 30°S. The simulated LGM tropical terrestrial surface air temperature cooling of 2.9°C is just more than half of the 5.4(±0.3)°C cooling

Table 2. Summary of climate changes. Annual means are computed over the final 30 years of each integration. Niño-3.4 statistics are computed over last 200 years of monthly time series with the mean monthly cycle and trend removed. Model elevations are shown in parentheses.

<table>
<thead>
<tr>
<th></th>
<th>PI control</th>
<th>LGMCO2</th>
<th>LGM</th>
<th>4×CO2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Global precip (mm day⁻¹)</td>
<td>2.93</td>
<td>2.79</td>
<td>2.61</td>
<td>3.20</td>
</tr>
<tr>
<td>Tropical precip (mm day⁻¹)</td>
<td>4.29</td>
<td>4.18</td>
<td>3.93</td>
<td>4.55</td>
</tr>
<tr>
<td>Global surface T (°C)</td>
<td>14.01</td>
<td>11.55</td>
<td>9.04</td>
<td>19.47</td>
</tr>
<tr>
<td>Tropical land surface T (°C)</td>
<td>23.59</td>
<td>21.98</td>
<td>20.89</td>
<td>29.08</td>
</tr>
<tr>
<td>Greenland summit MAT (K)(True elev. 3207 m)</td>
<td>243.93 (3073)</td>
<td>237.87</td>
<td>235.39 (3037)</td>
<td>251.87</td>
</tr>
<tr>
<td>Vostok MAT (K)(True elev. 3500 m)</td>
<td>220.28 (3431)</td>
<td>217.49</td>
<td>210.31 (3680)</td>
<td>230.12</td>
</tr>
<tr>
<td>Global precipitable water (mm)</td>
<td>23.93</td>
<td>21.14</td>
<td>18.84</td>
<td>32.91</td>
</tr>
<tr>
<td>Tropical SST (°C)</td>
<td>26.94</td>
<td>25.72</td>
<td>24.78</td>
<td>30.40</td>
</tr>
<tr>
<td>Niño-3.4 σ (°C)</td>
<td>1.02</td>
<td>1.09</td>
<td>0.82</td>
<td></td>
</tr>
<tr>
<td>Transports (Sv)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bering Strait</td>
<td>0.96</td>
<td>0.99</td>
<td>Closed</td>
<td>1.14</td>
</tr>
<tr>
<td>Florida Strait</td>
<td>29.5</td>
<td>28.0</td>
<td>32.3</td>
<td>27.0</td>
</tr>
<tr>
<td>Drake Passage</td>
<td>175.5</td>
<td>214.7</td>
<td>231.7</td>
<td>184.6</td>
</tr>
<tr>
<td>AMOC (Sv)</td>
<td>25.7</td>
<td>23.5</td>
<td>35.8</td>
<td>21.7</td>
</tr>
<tr>
<td>NH max</td>
<td>18.3</td>
<td>16.5</td>
<td>20.5</td>
<td>15.3</td>
</tr>
<tr>
<td>Max@34’S</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sea ice area (10⁶ km²)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>NH</td>
<td>11.70</td>
<td>17.60</td>
<td>8.64</td>
<td>3.84</td>
</tr>
<tr>
<td>SH</td>
<td>16.98</td>
<td>21.91</td>
<td>27.88</td>
<td>5.81</td>
</tr>
</tbody>
</table>

Table 3. Area-weighted regional LGM – PI mean annual surface temperature difference proxy data model comparison (°C). Estimates of LGM cooling relative to the PI control mean annual air temperature difference over land, or SST over ocean for proxy reconstructions relative to the model. Because the model grid is much finer than the proxy data grids, the model output is first averaged over a representative box at the proxy data locations (CCSM@proxy), which is either 2° lat × 2° lon for the terrestrial proxy data grid, or 5° lat × 5° lon for the SST proxy. The 2-m reference height air temperature from the model is used over land to compare with the land proxy MAT. Over the open ocean region, the surface temperature is used to compare with the SST proxy data. To obtain an estimate of SST under sea ice, the minimum ocean freezing temperature of −1.8°C is used. For comparison with the averages estimated over the proxy locations, the simulated regional mean temperature differences computed on the full model grid over the land and ocean domains are shown. The uncertainties given are estimated as the root mean of the area-weighted squared data errors. Land differences on the CCSM grid are taken after the mean temperatures are computed using the appropriate land masks for the LGM and PI simulations. A mean is not provided over SH land region because of the paucity of terrestrial proxy data in the SH in Bartlein et al. (2011).

<table>
<thead>
<tr>
<th>Region</th>
<th>Land (°C)</th>
<th>Ocean (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Proxy</td>
<td>CCSM@proxy</td>
</tr>
<tr>
<td></td>
<td>Proxy</td>
<td>CCSM@proxy</td>
</tr>
<tr>
<td>Global</td>
<td>−6.1(±1.5)</td>
<td>−6.5</td>
</tr>
<tr>
<td>Tropics (30°N–30°S)</td>
<td>−3.1(±1.2)</td>
<td>−3.5</td>
</tr>
<tr>
<td>NH-extra (30°–90°N)</td>
<td>−7.5(±1.6)</td>
<td>−8.2</td>
</tr>
<tr>
<td>SH-extra (30°–90°S)</td>
<td>−4.0</td>
<td></td>
</tr>
</tbody>
</table>
estimated by the meta-analysis of pollen, snow-line, and noble gas proxies of Ballantyne et al. (2005), although some of the proxies used in Ballantyne et al. (2005), such as snow lines, are also influenced by changes in precipitation.

The model LGM cooling estimates yield a simulated tropical land MAT/SST cooling ratio of 1.4 using the full grid averages, or 1.6 using the means estimated for the proxy-sampled region, which underpredicts the ratio found by combining the Waelbroeck et al. (2009) and Bartlein et al. (2011) proxy reconstructions, or by using the Ballantyne et al. (2005) reconstruction, both of which yield a ratio of ~2, but with an O(1) relative error if the regional temperature errors are propagated. These simulated LGM cooling ratios agree well with the average land/sea warming ratio response to increased GHGs of ~1.5 from 40°N to 40°S reported in Sutton et al. (2007) for an IPCC AR4 multimodel ensemble.

The simulated LGM sea ice coverage agrees well with available proxy reconstructions. The SH winter-time sea ice area (not shown) increases in the LGM simulation over the preindustrial control by 56%, smaller than the 100% increase estimated by proxies (Gersonde et al. 2005). The equatorward expansion of the 15% sea ice concentration contour is in excellent agreement with the proxy evidence, showing the greatest expansion, to ~47°N, in the Atlantic and Indian Ocean sectors and more modest expansion to ~55°N in the Pacific sector, relative to 47°N and 57°N, respectively, as estimated by Gersonde et al. (2005). The spatial coverage of the annual mean LGM sea ice concentration (Fig. 10, described in greater detail below) agrees reasonably well in the western North Atlantic to the pattern of glacial sea ice distribution found by de Vernal and Hillaire-Marcel (2000), although the simulated concentrations are lower than estimated in the proxy reconstruction. The Norwegian Sea up to ~70°N is open in the LGM simulation in northern summer as suggested by the Waelbroeck et al. (2009) reconstruction.

5. Analysis of global climate sensitivity in the CCSM4

a. Climate feedback parameter

To compare the sensitivity of the climate system to different radiative forcings in the CCSM4, we start by estimating the climate feedback parameter $\lambda$ which reflects the strength of the net feedbacks to an imposed radiative forcing $\Delta F$ as given by the global energy balance,

$$\Delta Q = \Delta F - \lambda \Delta TS. \quad (1)$$

Adopting the terminology of Knutti and Hegerl (2008), $\Delta Q$ is the net heat uptake of the climate system equal to net downward heat flux at the top of the model atmosphere (TOA). The last term on the right-hand side of (1) is the radiative response of the climate system, assumed to be linearly proportional to the global and annual mean surface temperature change $\Delta TS$. The climate feedback parameter $\lambda$ (W m$^{-2}$ °C$^{-1}$) is equal to $\Delta F/\Delta TS$ at equilibrium, where $\Delta Q = 0$. In these simulations, $\Delta F$ arises from changes to greenhouse gas concentrations and changes in surface albedo and orography from the imposed glacial ice sheets and glacial land sea mask.

Because of limited computer resources, none of these coupled simulations was integrated to a complete statistical equilibrium, although the drifts in $\Delta TS$ and $\Delta Q$ are reasonably small toward the end the LGM and LGMCO$_2$ simulations (not shown). Thus, we estimate $\lambda$ using a method proposed by Gregory et al. (2004) to find $\Delta TS$ at equilibrium, by regressing $\Delta Q$ against $\Delta TS$ (Fig. 4) to obtain a “projected” equilibrium temperature change $\Delta TSe$ as the $\Delta TS$ intercept of the regression line found by least squares fit (i.e., where $\Delta Q = 0$). Gregory et al. (2004) warns that this method may underestimate the actual equilibrium temperature response if (1) is not linear. Danabasoglu and Gent (2009) show that this method, applied to integrations longer than ~150 years, can be used to get close to the equilibrium $\Delta TS$, a constraint that all these simulations satisfy.

Using this method on the annual mean time series for the available length of the integrations, we project an equilibrium temperature response of ~5.5°C in the LGM simulation, ~2.6°C for the LGMCO$_2$ simulation, and 6.2°C for the 4×CO$_2$ simulation. In comparison, by the end of the integration the LGM simulation has cooled globally by 5.0°C at the surface, ~90% of the projected equilibrium cooling (Table 2). The LGMCO$_2$ simulation has cooled by 2.5°C by the end of its integration, ~96% of the projected equilibrium surface cooling. The 4×CO$_2$ simulation shows a global mean surface warming of 5.5°C, ~88% of the projected equilibrium cooling. Following this analysis, the 4×CO$_2$ simulation was integrated for another 1000 years but is not yet available. However, the global mean surface temperature difference from the PI control at the end of the longer simulation is found to be ~6.2°C (G. Branstator and H. Teng 2012, personal communication), in good agreement with the projected equilibrium warming found here using only the first 269 years of the simulation.
An estimate for the radiative forcing $\Delta F$ in (1) is difficult to obtain from coupled model output alone and becomes a source of uncertainty here. To obtain an estimate of the CO$_2$ forcing, consistent with Bitz et al. (2012) and Kay et al. (2012), we use the logarithmic formula in Ramaswamy et al. (2001) scaled to obtain the stratospherically adjusted radiative forcing for a doubling of CO$_2$ in CAM4 $F_{2x}$ of 3.5 W m$^{-2}$ as found in Kay et al. (2012). For determining the LGM GHG forcing, we follow Crucifix (2006) and use the formula $F_{GHG} = 3.5(-2.85/3.7)$ W m$^{-2}$, where $-2.85$ and $3.7$ are the GHG forcings (W m$^{-2}$) for the LGM and a CO$_2$ doubling as obtained from the formula in Ramaswamy et al. (2001) and 3.5 W m$^{-2}$ is $F_{2x}$, defined above. This results in an LGM GHG forcing of $-2.7$ W m$^{-2}$.

Quantifying $\Delta F$ for the LGM simulation requires an estimate of the additional forcing because of the imposed ice sheets. The radiative forcing caused by the added ice sheet and exposed land areas from the lowered sea level is estimated directly using the approximate partial radiative perturbation (APRP) method of Taylor et al. (2007). The forcing caused by the ice sheet and exposed land areas is computed from the change in net TOA shortwave flux from changes in the surface albedo over the region where the NH ice sheets are added (i.e., excluding Greenland and Antarctica) and globally where the land mask is changed (Fig. 5) to account for the effect of replacing ocean with land. We do not account for the snow cover response over the ice sheet, so this forcing estimate includes the minor effect of surface albedo feedbacks in these regions. This method produces a global mean radiative forcing caused by changes in surface albedo of $-3.5$ W m$^{-2}$, with $-2.7$ W m$^{-2}$ contributed from the ice sheets and $-0.8$ W m$^{-2}$ from the non–ice sheet additional land area.

This estimate of $-3.5$ W m$^{-2}$ for the radiative forcing from the added ice sheets and land area is at the high end of the range of previous estimates found in the literature and computed using the APRP method for the PMIP2 models (Braconnot et al. 2012). The CCSM4 estimate is larger in comparison with the radiative forcing for the ice sheet and sea level for the CCSM3 LGM simulation estimated at $-2.7$ W m$^{-2}$ (Otto-Bliesner et al. 2009). Added to an estimate of LGM GHG forcing of $-2.7$ W m$^{-2}$, a total forcing of $-6.2$ W m$^{-2}$ is obtained. This estimate lies within the range discussed in the literature [see Yoshimori et al. (2009) for a discussion]. We adopt an uncertainty of $\pm10\%$ for the LGM $\Delta F$ to
account for the error in using the APRP method in comparison with the more accurate PRP method as shown in Taylor et al. (2007), the error in the estimate for the GHG forcing based on $F_{2\times}$, the inclusion of surface albedo feedbacks from snow over the NH added ice sheets in the forcing estimate, and the exclusion of the negligible radiative forcing from the change in orbital parameters.

Based on the method in Gregory et al. (2004), we estimate $\lambda$ in (1), denoted as $\lambda_e$, using the estimates of the radiative forcing $\Delta F$ and the estimated equilibrium temperature response $\Delta T_{Se}$, such that

$$\lambda_e = \frac{\Delta F}{\Delta T_{Se}}. \tag{2}$$

The $\lambda_e$ estimated for the simulation with full LGM forcing of 1.16 W m$^{-2}$°C$^{-1}$ compares well with the estimate obtained for the $4\times CO_2$ simulation of 1.13 W m$^{-2}$°C$^{-1}$ (Table 4) and also with the climate feedback parameter of 1.16 W m$^{-2}$°C$^{-1}$ obtained from a doubled CO$_2$ CCSM4 simulation configured with the slab ocean model (SOM) (Bitz et al. 2012). However, the smaller $\lambda_e$ estimated for the LGMCO$_2$ simulation of 0.84 W m$^{-2}$°C$^{-1}$ suggests the CCSM4 has a greater climate sensitivity to lowered CO$_2$ forcing alone.

**b. Effective climate sensitivity**

The climate feedback parameter $\lambda$ in (1), estimated at any given time, yields an “effective” climate feedback parameter $\lambda_{eff}$ as in Gregory and Mitchell (1997) and Senior and Mitchell (2000), where

$$\lambda_{eff} = \frac{\Delta F \Delta Q(t)}{\Delta T_S(t)}. \tag{3}$$

The effective climate feedback parameters found here using global and annual means of $\Delta Q$ and $\Delta T_S$ computed over the last 30 years of each integration relative to the PI control simulation, and $\Delta F$ found above (Table 4) compare well with $\lambda_e$ found earlier using the projected equilibrium temperature response.

The $\lambda_{eff}$ can be used to obtain an effective climate sensitivity [ECS (°C)], an estimate of the equilibrium temperature response to a doubling of CO$_2$ such that

$$\text{ECS} = \frac{F_{2\times}}{\lambda_{eff}}, \tag{4}$$

where $F_{2\times}$ is the radiative forcing from a doubling of CO$_2$. Taking $F_{2\times} = 3.5$ W m$^{-2}$ as found in Kay et al. (2012), we estimate an ECS from the $\lambda_{eff}$ from LGM and $4\times CO_2$ comparable with the equilibrium climate sensitivity of 3.2°C found in Bitz et al. (2012) for CCSM4 in a slab ocean configuration at the same atmospheric resolution, whereas an ECS of 4.1°C, estimated from the $\lambda_{eff}$ from LGMCO$_2$, is much larger (Table 4).

The higher ECS predicted for the LGMCO$_2$ suggests that either weaker negative feedbacks or stronger positive feedbacks arise from a lowering of atmospheric CO$_2$ alone compared with either increasing CO$_2$ as in

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**TABLE 4.** Summary of the estimated climate sensitivity parameters. The total LGM forcing is the sum of the shortwave contribution, estimated from the APRP method of Taylor et al. (2007) and the estimated LGM GHG forcing. The ECS is estimated from $\lambda_e$ (or $\lambda_{eff}$).

<table>
<thead>
<tr>
<th>Case</th>
<th>$\Delta F$ (W m$^{-2}$)</th>
<th>$\Delta T_{Se}$ (°C)</th>
<th>$\lambda_e = \Delta F/\Delta T_{Se}$ (W m$^{-2}$°C$^{-1}$)</th>
<th>$\lambda_{eff}$ (W m$^{-2}$°C$^{-1}$)</th>
<th>$F_{2\times}/\lambda_e$ (°C)</th>
<th>$F_{2\times}/\lambda_{eff}$ (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LGM</td>
<td>-6.23 (±0.6)</td>
<td>-5.51</td>
<td>1.13 (±0.1)</td>
<td>1.22</td>
<td>3.1 (±0.3°C)</td>
<td>2.9</td>
</tr>
<tr>
<td>LGMCO$_2$</td>
<td>-2.18</td>
<td>-2.60</td>
<td>0.84</td>
<td>0.85</td>
<td>4.2</td>
<td>4.1</td>
</tr>
<tr>
<td>$4\times CO_2$</td>
<td>7.0</td>
<td>6.22</td>
<td>1.13</td>
<td>1.09</td>
<td>3.1</td>
<td>3.2</td>
</tr>
</tbody>
</table>
4×CO₂, or from the full glacial forcing in the CCSM4. The addition of ice sheets, lowered sea level, reductions of additional radiatively important trace gas constituents such as methane and ozone, and a small but positive annual and global mean orbitally induced insolation change in the CCSM4 brings the ECS back in line with the higher CO₂ forced simulations.

c. Analysis of global feedbacks

To investigate the processes responsible for the differing response of the CCSM4 to different radiative forcings, as manifested by the different climate feedback parameters found, we diagnose the individual feedbacks that comprise the climate response in (1). Following the conventions of Gregory and Mitchell (1997), we decompose λ in (1) into individual feedback parameters. First, we decompose the net TOA radiative imbalance into longwave and shortwave components with ΔQ = ΔQ_LW - ΔQ_SW, where the net absorbed TOA shortwave flux ΔQ_SW is defined to be net positive down; the net outgoing longwave TOA flux ΔQ_LW is defined as net positive up; and we assume that all components of the radiative response are linearly proportional to ΔTS. Because the LGM simulation has a large shortwave radiative forcing, we partition the radiative forcing ΔF into the longwave and shortwave constituents, ΔF = ΔF_LW + ΔF_SW, and expand (1) in terms of λ as

\[ \lambda = (\Delta F_{LW} + \Delta Q_{LW})/\Delta TS - (\Delta Q_{SW} - \Delta F_{SW})/\Delta TS \] (5a)

\[ \lambda = \lambda_{LW} - \lambda_{SW}. \] (5b)

Using this convention, a positive shortwave feedback acts to reduce the magnitude of the total climate feedback parameter λ and hence tends to enhance the temperature change ΔTS. As shown in Table 5 and Fig. 6, the total outgoing longwave feedback λ_LW is of nearly equal magnitude for all simulations. The total shortwave feedback λ_SW, positive for all simulations, is largest for the LGM simulation and smallest for LGMCO₂, with a large relative spread of ~40%.

We decompose the longwave and shortwave feedbacks further to separate out the effects of clear-sky and cloud processes:

\[ \lambda = \lambda_{LWCLR} - \lambda_{WCF} - \lambda_{SWCLR} - \lambda_{SWCF}, \] (6)

with the convention that the outgoing longwave clear-sky feedback λ_LWCLR is reduced by positive feedbacks from longwave cloud forcing λ_WCF, shortwave clear-sky λ_SWCLR, and shortwave cloud λ_SWCF processes.

Table 5. Analysis of climate feedback parameters (W m⁻² °C⁻¹). Shortwave feedbacks marked with an asterisk * are estimated using the APRP technique (Taylor et al. 2007).

<table>
<thead>
<tr>
<th>Feedbacks (W m⁻² °C⁻¹)</th>
<th>LGM</th>
<th>LGMCO₂</th>
<th>4×CO₂</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total λ “effective”</td>
<td>1.22</td>
<td>0.85</td>
<td>1.09</td>
</tr>
<tr>
<td>λ_LW</td>
<td>1.87</td>
<td>1.83</td>
<td>1.89</td>
</tr>
<tr>
<td>λ_LWCLR</td>
<td>1.87</td>
<td>1.92</td>
<td>1.75</td>
</tr>
<tr>
<td>λ_LWCF</td>
<td>0.004</td>
<td>0.09</td>
<td>-0.13</td>
</tr>
<tr>
<td>λ_SW</td>
<td>0.64</td>
<td>0.98</td>
<td>0.79</td>
</tr>
<tr>
<td>λ_SWCLR</td>
<td>1.30</td>
<td>1.59</td>
<td>0.94</td>
</tr>
<tr>
<td>λ_SWCF</td>
<td>-0.66</td>
<td>-0.62</td>
<td>-0.14</td>
</tr>
<tr>
<td>λ_alb*</td>
<td>0.58</td>
<td>0.93</td>
<td>0.48</td>
</tr>
<tr>
<td>λ_cld*</td>
<td>-0.24</td>
<td>-0.41</td>
<td>-0.01</td>
</tr>
<tr>
<td>λ_clr*</td>
<td>0.16</td>
<td>0.12</td>
<td>0.10</td>
</tr>
</tbody>
</table>

These component feedbacks are computed as follows (Table 5):

\[ \lambda_{LWCLR} = \frac{\Delta F_{LW} + \Delta Q_{LWCLR}}{\Delta TS}, \] (7a)

\[ \lambda_{LWCF} = \frac{\Delta Q_{LWCLR} - \Delta Q_{LW}}{\Delta TS}, \] (7b)

\[ \lambda_{SWCLR} = \frac{\Delta Q_{SWCLR} - \Delta F_{SWCLR}}{\Delta TS}, \] (7c)

\[ \lambda_{SWCF} = \frac{\Delta Q_{SW} - \Delta Q_{SWCLR}}{\Delta TS}. \] (7d)

The clear-sky longwave feedback (7a) has a similar magnitude for all simulations, and the longwave feedback from clouds (7b) is a relatively minor component. Thus, the longwave component feedbacks do not explain the smaller λ and hence higher sensitivity found in LGMCO₂.

Instead, the larger positive shortwave feedback most likely explains the smaller total climate feedback parameter found in LGMCO₂. However, the shortwave feedback clear-sky component determined using (7c) cannot be directly related to changes in the surface albedo from the masking effect of clouds. Thus, to better identify the processes associated with the large positive shortwave feedback, we use the APRP method of Taylor et al. (2007) on the mean monthly climatological fields to isolate and directly estimate the changes in the global and annual mean TOA shortwave fluxes from changes in surface albedo, changes in cloud fraction, and changes in clear-sky scattering and absorption properties, and their associated feedbacks (λ_alb, λ_cld, and λ_clr, respectively; Table 5, Fig. 6).

Using the APRP method, we find that the surface albedo feedback λ_alb is largest for the LGMCO₂ simulation and smallest for the 4×CO₂ simulation and...
contributes to the bulk of the total shortwave feedback (Table 5). The largest changes in the net TOA shortwave flux from surface albedo changes are associated with expansion in the LGM and LGMCO2 simulations (or contraction in the 4×CO2 simulation) of the highly reflective snow/sea ice cover in high latitudes (Fig. 7). Interestingly, in the Arctic the reduction of net TOA shortwave flux from surface albedo changes in the LGM simulation is similar to that found in the LGMCO2 simulation. A maximum upper threshold for both surface albedo and sea ice fraction is nearly reached in response to just the lowering of CO2 to the glacial level of the LGMCO2 simulation.

The global mean shortwave feedback from cloud changes λ_cld is relatively large for both of the glacial cases, though smaller than λ_alb, and represents a compensating negative feedback. The magnitude of λ_cld is negligible for the 4×CO2 simulation, as also found in the doubled CO2 CAM4 analysis of Kay et al. (2012). In fact, the components of the feedback parameter estimated here for the 4×CO2 simulation agree with those estimated for the CO2 doubling experiment using the CAM4 coupled to the slab ocean model by Kay et al. (2012). The separate shortwave feedback components found using the APRP method do not sum to the total shortwave feedback because of simplifications in the radiative transfer used in the method and nonlinear contributions not explicitly estimated (Taylor et al. 2007). Although the residual is larger in the LGMCO2 case relative to the other cases, the enhancement of the residual is smaller than for the surface albedo feedback.

6. Spatial response of surface temperature and sea ice

The spatial distribution of the surface temperature response is highly amplified in the polar regions in all simulations relative to the tropics (Fig. 8). As discussed earlier, the largest cooling in the LGM simulation is found over the NH continental ice sheets. In the LGMCO2 and 4×CO2 simulations, the greatest surface temperature response is found in the high latitudes associated with the surface albedo changes particularly related to the changes in sea ice coverage.

In the SH, sea ice area expands in the annual mean by 64% and 29% in the LGM and LGMCO2 simulations, respectively, and decreases by 66% in the 4×CO2 simulation (Table 2, Fig. 9). The NH sea ice area increases in the annual mean by 50% in the LGMCO2 simulation, and decreases by 67% in the 4×CO2 simulation (Table 2, Fig. 10). The relative increase in NH sea ice area in the LGM simulation is not easily evaluated because the Arctic Ocean area is reduced from the imposed ice sheets and expansion of land. Sea ice expands significantly in the Pacific subpolar region and in the Barents Sea and extends farther into the North Atlantic from...
the Labrador Sea in the LGM simulation (Fig. 10) than in the PI simulation. Reduced sea ice coverage is found in the Labrador Sea in the LGM simulation relative to the LGMCO2 simulation owing to increased downsloping winds caused by the nearby presence of the large ice sheet, and increased northward ocean heat transport in the Atlantic Ocean related to the increase in the strength of the AMOC, discussed below.

Polar amplification is evaluated by normalizing the zonally averaged annual mean temperature change by the global mean temperature change for each simulation (Fig. 11). The polar amplification shifts equatorward.
in the colder climates with the LGM and LGMCO2, showing greater polar amplification in the mid- to high-latitude NH regions than the $4\times CO_2$ simulation. The LGMCO2 simulation exhibits the greatest amplification north of 55°N, related to the greater equatorward expansion of NH sea ice in the LGMCO2 simulation (Fig. 10). The LGM simulation exhibits the largest response in the NH midlatitudes because of the presence of the imposed Laurentide ice sheet. In the SH, the polar amplification is greatest in the LGM simulation, especially over Antarctica which is elevated, and weakest in the LGMCO2 simulation. In the glacial simulations, the tropics cool at a similar rate of $\sim 0.5 ^\circ C$ of tropical cooling per 1°C of global cooling, whereas in the $4\times CO_2$ simulation the rate at which the tropics warm increases to $0.7 ^\circ C$ °C$^{-1}$ (of global warming).

7. Sensitivity of other climate features

a. Precipitation

Globally, the atmosphere is moister for the warmer climate of $4\times CO_2$ with precipitable water increasing by about 6% °C$^{-1}$ (of global temperature warming), and drier in the glacial simulation with a decrease of about 5% °C$^{-1}$ (of global temperature cooling; Table 2). The global precipitation rate decreases in both the full glacial forcing and the lowered CO2 case and increases in the higher CO2 case, at a rate of $\sim 2% ^\circ C$ °C$^{-1}$ (of global temperature change), consistent with results for the twentieth-century and future climate change experiments in the IPCC AR4 (Held and Soden 2006). The global mean precipitation rate decrease in the full glacial case is twice the magnitude of the lowered CO2 alone.

Regionally, the patterns of precipitation change are more complex (Fig. 12). With colder (warmer) temperatures in the glacial (higher CO2) simulations, precipitation generally decreases (increases) in the tropics and midlatitude storm tracks. Although the AMOC strengthens in the full LGM simulation and weakens in both CO2-only forced simulations (see discussion below), no significant shifts in the location of the tropical Atlantic ITCZ occurs (Fig. 12). The warming and/or cooling of these regions is relatively uniform in our simulations, consistent with Otto-Bliesner and Brady (2010), who suggest from an analysis of freshwater forcing simulations that the gradients between the northern subtropical and southern subtropical SSTs are important for the ITCZ shifts.

b. Surface wind stress over the oceans

The largest change in surface wind stress over the ocean (shown where sea ice coverage is less than 50%) is found in the full glacial case in the NH subpolar Pacific and Atlantic where the westerlies strengthen because of the presence of the large Laurentide ice sheet (Fig. 13).
The NH zonal mean westerlies shift equatorward and strengthen in the LGM case compared with the PI control. A weak strengthening of the NH westerlies is noted in the LGMCO2 simulation and weakening in the 4xCO2 simulation.

In the tropics, the NH easterlies strengthen (weaken) for the glacial (increased CO2) simulations with the greatest response shown to full glacial forcing in the Atlantic basin in the LGM simulation. In the SH, the southeasterly trade winds strengthen in the LGM and to a much lesser extent in the LGMCO2 simulation in all three ocean basins. In the 4xCO2 simulation, the southeasterly trades weaken in the Atlantic and Indian Oceans and strengthen in the Pacific over South America to the west of the cold tongue as discussed above.

The SH westerlies shift equatorward (poleward) in the glacial (higher CO2) cases, but the response in the LGM simulation is accompanied by a significant increase in strength. This strengthening is particularly pronounced just equatorward of the sea ice edge as denoted by the 50% contour (Fig. 9). In general, the direction of the latitudinal shifts in the westerlies in response to CO2 is consistent with Toggweiler et al. (2006); however, the size of the effect is relatively minor in these simulations.

**FIG. 9.** Spatial maps of mean annual sea ice concentration (%) for the Southern Hemisphere for (a) LGM, (b) LGMCO2, (c) PI, and (d) 4xCO2 simulations.
Tropical sea surface temperature is strongly affected by radiative forcing as shown by the changes to the mean and variance of SST (Fig. 14). The mean tropical SST in the Pacific basin cools (warms) in response to glacial (increased CO2) forcing (Figs. 14a–c), with the response more pronounced in the Pacific cold tongue region than the response in the warm pool. This has a tendency to strengthen (weaken) the zonal SST gradient across the Pacific for glacial (increased CO2) forcing. However, the increased cooling in the eastern Pacific in the LGM simulation is still not as pronounced as suggested by some proxy evidence (Ballantyne et al. 2005; Waelbroeck et al. 2009). In addition, the basin-to-basin heterogeneity noted in the proxies is not found in the CCSM4 glacial simulations. Interestingly, in both the LGM and the 4xCO2 simulations, the amplitude of the equatorial SST variability is reduced, although the reduction is greatest in the 4xCO2 case compared with the LGM case. Thus, changes in ENSO amplitude do not appear to be only related to the response of the zonal SST gradient in the Pacific. In the LGM simulation, the amplitude of the seasonal cycle of SST is increased (not shown). Both the magnitude and sign of the ENSO response to glacial forcing remain a large uncertainty for,
as discussed in Zheng et al. (2008), neither the different paleo ENSO proxies nor the group of PMIP2 LGM model simulations is in agreement.

de. Atlantic meridional overturning circulation and ocean transports

It is suggested that the strength of the Atlantic meridional overturning circulation may respond significantly to both glacial forcings and in response to future scenarios (Solomon et al. 2007). Paleoclimate reconstructions suggest that not only was the glacial AMOC different from today in key aspects such as vertical extent, but it may have declined significantly at times in the past 20 ka in response to large meltwater events associated with de-glacial warming and sea level rise (McManus et al. 2004; Lynch-Stieglitz et al. 2007). Large disagreement in the simulated response of AMOC to glacial forcing was found by the PMIP2 even with a similar AMOC for modern conditions (Otto-Bliesner et al. 2007). A slowdown in AMOC implies a reduction in northward ocean heat transport, a tendency that cools the NH and warms the SH, redistributing heat through the bipolar see-saw and thus not affecting the global mean. However, as the mean position of the sea ice edge (Schmittner et al. 2002;
Bitz et al. (2005) is maintained by ocean heat transport, reductions in AMOC may affect the positive sea ice albedo feedback effect.

Figure 15 compares the mean Eulerian component of the meridional overturning streamfunction in the North Atlantic, contours of which represent streamlines for zonally integrated ocean transport. The AMOC from the PI control simulation as presented in Danabasoglu et al. (2012b) is included for comparison. The location of the maximum AMOC, at a depth of ~1000 m and a latitude of ~35°N, does not show sensitivity to forcing. At this latitude, the AMOC shows a tight recirculation cell that varies in magnitude depending on the case. This tight recirculation cell is a result of strong coastal upwelling in only a few grid boxes adjacent to the boundary south of 35°N as discussed in an earlier section. The full LGM simulation appears to be particularly sensitive to this problem, as shown by the largest difference between the maximum overturning and the transport at the outflow latitude, 34°S (Table 2). This is largely caused by the southward displacement and strengthening of the glacial westerlies over the North Atlantic.

The strength of both the maximum overturning found at 35°N and the outflow transport at 34°S weakens in response to either an increase or decrease of CO2. Greater reductions are found in the 4×CO2 simulation (Table 2), although the reduction lessens after the first 250 years of the 4×CO2 simulation. The AMOC maximum in the 4×CO2 simulation undergoes a large-amplitude transient adjustment early in the simulation, with a sharp reduction to ~15 Sv in the first 50 years followed by a rapid recovery to ~22 Sv after ~150 years and remains stable thereafter with a small trend. In a millennial-length extension of the 4×CO2 simulation (not available for this analysis), the AMOC maximum slowly increases over the next ~500 years to reach an equilibrium of ~23 Sv (H. Teng 2012, personal communication). In contrast, the strength of AMOC increases in the full LGM simulation both at the outflow latitude and the maximum.

In both simulations forced with CO2 only, there is a shoaling of the zero contour separating the clockwise circulation of North Atlantic Deep Water (NADW), produced from the formation and sinking in high northern latitudes, from the counterclockwise circulation of simulated Antarctic Bottom Water (AABW). The zero contour shoals from ~3200-m depth in the PI control to ~3000-m depth in the LGMC02 case, and to ~2600-m depth in the 4×CO2 case. Interestingly, the AABW cell penetrates farther north in all simulations compared with the PI control. The northward extent of the AABW cell penetrates about 20° farther poleward in the

FIG. 13. Spatial maps of the mean annual surface wind stress anomalies (N m⁻²) over the ocean (where sea ice fraction is less than 50%) computed and labeled as in Fig. 8.
LGM simulation than the PI simulation, whereas AABW fills the deep basin up to the depth of the Nordic sills in both the $4 \times CO_2$ and full LGM simulations.

The barotropic transport through the Drake Passage shows a strong sensitivity to glacial forcing with an increase of ~32% in the LGM simulation and of ~22% in the LGMCO$_2$, (c) $4 \times CO_2$, and (d) PI simulations compared with the PI control (Table 2). Interestingly, the Drake Passage transport also increases in the $4 \times CO_2$ simulation though weakly by ~5%. The barotropic transport through the Florida Straits (FS) is not particularly sensitive to either glacial forcing or increased CO$_2$ with less than 10% change in magnitude. The weak increase of ~10% in the LGM simulation contradicts Lynch-Steiglitz et al. (1999), who found evidence for a weaker transport through the FS at LGM, using proxy data. The FS transport in the $4 \times CO_2$ simulation undergoes an ~5-Sv transient oscillation over the first 250 years of the simulation and thus the decrease is not significant. Similarly, the Bering Strait transport shows only a weak sensitivity to changes in CO$_2$ with an ~5% weakening in the LGMCO$_2$ simulation, and an increase of ~9% in the $4 \times CO_2$ simulation.

8. Comparison with previous glacial modeling with the CCSM3

There are considerable model differences between the CCSM3 and the updated CCSM4, associated with significant improvements made in the component model physics. In addition, the CCSM4 atmospheric component has greater horizontal resolution, at ~1° versus ~2.85° in the T42 spectral truncation of the CCSM3 LGM simulation, which, in addition to being of lower horizontal resolution, tends to smooth orography, effectively lowering the elevations of the imposed ice.
sheets. These updates and improvements have led to significant coupled climate differences and improvements alleviating some regional biases for the preindustrial climate simulation discussed in Neale et al. (2013). We present some of the differences between the full LGM CCSM4 simulation with PMIP3 forcing, and the CCSM3 LGM simulation using PMIP2 forcing reported in Otto-Bliesner et al. (2006). The previous glacial modeling study using the CCSM3, completed as part of PMIP2, used the ICE-5G ice sheets (Peltier 2004), which is the most significant difference from the forcing used in the CCSM4 LGM simulation discussed here. A comparison of the CCSM3 LGM with the CCSM1 LGM simulation of Shin et al. (2003) is found in Otto-Bliesner et al. (2006) and will not be repeated here.

Compared with the CCSM3 LGM simulation, both the tropical and global mean cooling is greater in the CCSM4 by ~0.5°C. The spatial pattern of cooling also shows some interesting regional differences. In particular, the CCSM3 LGM simulation exhibited a warming relative to the PI over Beringia and the Gulf of Alaska, which was found to be related to the higher elevation of ICE-5G North American ice sheet compared with the ICE-4G ice sheet used in PMIP1 (Otto-Bliesner et al. 2006). This subpolar warming is not evident in the CCSM4 LGM response; however, a region of weaker cooling is found in the northern Gulf of Alaska in a similar location. The zonal mean surface cooling found in the CCSM4 LGM is about ~2°C colder poleward of 60°N, about the same over the tropics, and warmer by ~4°C poleward of 40°S than the CCSM3 LGM simulation.

FIG. 15. Eulerian mean meridional overturning circulation in the Atlantic Ocean basin for the (a) LGM, (b) LGMCO2, (c) PI control, and (d) 4xCO2 simulations. Positive (negative) contours reflect clockwise (counterclockwise) circulation. Contour interval is 3 Sv.
The zonal heterogeneity of the tropical CCSM4 LGM SST response with greater cooling in the eastern basins than the western warm pool region is a notable improvement in better agreement with the proxy evidence, over the CCSM3 LGM response, which showed a relatively uniform cooling along the equator similar to the other PMIP2 models (Otto-Bliesner et al. 2009). Both LGM simulations show a reduction in the standard deviation of the monthly SST variability in the Niño-3.4 region with an ~20% reduction in the LGM CCSM4 and a slightly greater reduction of ~29% in the CCSM3, compared with their respective PI simulations. Both LGM simulations show a strengthening of the mean annual cycle of SST in the eastern Pacific as well.

The wind stress response to glacial forcing over the North Atlantic is qualitatively similar in the CCSM4 to the CCSM3 as shown by comparison with Fig. 7 of Otto-Bliesner et al. (2006). The NH westerlies shift southward in the subpolar North Atlantic; however, they strengthen in the CCSM4 simulation. The North Atlantic easterlies strengthen to a greater degree in the CCSM4, with little change noted in the CCSM3 glacial simulation. In the Pacific basin, the NH westerlies also shift southward in both the CCSM4 and CCSM3 LGM simulations compared with the respective PI simulations; however, they strengthen in the CCSM3 simulation but not in the CCSM4. Similar decreases in sea ice area are found in both the CCSM4 and CCSM3 LGM simulations because of the decrease in the Arctic Ocean area from the expansion of the imposed ice sheets; however, sea ice expands to a greater degree in the subpolar North Atlantic in the CCSM3 LGM than the CCSM4.

Relative to the respective PI simulations, the mid-latitude SH westerlies strengthen at the latitude of the Drake Passage in both the CCSM3 and CCSM4 LGM simulations with a similar magnitude of response, although the SH westerlies in the CCSM3 PI simulation are notably too strong. The SH sea ice expansion in CCSM4 with an increase of ~64%, shows less sensitivity than in the CCSM3, which increased by 142%. This may explain the weaker high-latitude SH cooling found in the CCSM4.

The response of the important ocean transports to LGM forcing in the CCSM4 shows some similarities and differences compared with the response in the CCSM3. In the CCSM3, both the maximum strength of the AMOC and the export of NADW at 34°S are reduced compared with the PI simulation, whereas the AMOC strengthens in the CCSM4 LGM simulation. Both the CCSM3 and CCSM4 LGM simulations show a shoaling of the NADW cell, with the AABW cell taking up more of the deep North Atlantic; however, the depth of maximum transport is insensitive to LGM forcing in the CCSM4. Both the CCSM4 and CCSM3 LGM simulations show increased barotropic transports through the Drake Passage and the Florida Straits, although the increases of 64% and 22%, respectively, in the CCSM3 LGM are twice as large as the increases of 32% and 10% found in the CCSM4 LGM simulation relative to the PI.

9. Summary and outstanding issues

We have analyzed the CCSM4 Last Glacial Maximum simulation at the standard 1° resolution, the same resolution as the majority of the CCSM4 CMIP5 long-term simulations for the historical and future projection scenarios. The forcings and boundary conditions for the LGM simulation follow the protocols of PMIP3 and CMIP5, including changes in the atmospheric concentrations of the principal long-lived greenhouse gases and the orbital parameters, and the effects of the height, extent, and volume of the large continental ice sheets. Aerosols and vegetation are prescribed at their pre-industrial distributions, although the seasonal and interannual vegetation phenology, including the total leaf and stem area indices and canopy heights, respond to LGM climate forcing. Two additional CCSM4 CO2 sensitivity simulations, in which the concentrations are instantaneously changed and fixed at the start of the simulation to the lowered glacial value of 185 ppm in LGMCO2 and to a fourfold increase of the preindustrial concentration in 4×CO2, are also analyzed.

To summarize the results:

1) The simulated LGM surface cooling and the expansion of sea ice compares well with the latest proxy climate reconstructions. Regional means computed over the same locations as the proxy data are found to be in excellent agreement, although the simulated LGM shows consistently greater mean cooling.

2) The equilibrium global mean annual surface temperature is projected to cool by 5.5°C in the full LGM simulation and by 2.6°C in the LGMCO2 simulation, suggesting that just less than half of the global LGM temperature response is from the lowered CO2. The projected equilibrium warming in the 4×CO2 simulation is estimated at 6.2°C, consistent with an equilibrium warming of 3.2°C found in the 2×CO2 CAM4 experiment coupled to a slab ocean model (Bitz et al. 2012).

3) The corresponding TOA radiative forcing, estimated using the APRP method of Taylor et al. (2007), yields an estimate of −6.2(±0.6) W m\(^{-2}\) for the full LGM simulation with a forcing associated with surface
albedo changes caused by the LGM ice sheet and additional land areas accounting for \(-3.5 \text{ W m}^{-2}\) and GHG forcing accounting for \(-2.7 \text{ W m}^{-2}\). This simulation does not include forcing caused by changes in aerosols. Predicted changes in the vegetation phenology impact surface albedo changes and contribute to the surface albedo feedback response.

4) The full LGM and \(4\times\text{CO}_2\) simulations show similar global climate sensitivities with an effective climate sensitivity (ECS) comparable to the equilibrium climate sensitivity of the CCSM4 slab ocean model result of 3.2°C (Bitz et al. 2012). The LGMCO2 simulation shows a greater ECS of \(\sim 4.2\)°C from a greater expansion of NH sea ice allowing a larger positive sea ice/snow albedo feedback response. This agrees with the finding of an increased positive feedback associated with surface albedo changes for LGMCO2, relative to the other simulations, as diagnosed using the APRP method of Taylor et al. (2007).

5) The LGMCO2 simulation is more sensitive at high northern latitudes, with a significant expansion of sea ice and a reduced AMOC. The extensive NH ice sheets in the full LGM simulation produce downsloping winds that keep the North Atlantic warmer with weaker sea ice expansion than in LGMCO2 and a more vigorous AMOC.

6) The strength of the AMOC is reduced in the LGMCO2 and \(4\times\text{CO}_2\) simulations and increased in the full LGM simulation. A weakening of AMOC, accompanied by decreased northward heat transport into the high latitudes, may enhance the expansion of sea ice and lead to greater positive ice/snow albedo feedback in the LGMCO2 simulation. This response may help explain some of the differences in climate sensitivities found in these simulations. Both CO2-forced-only simulations and the full LGM simulation show a shoaling of the clockwise cell associated with the formation of North Atlantic Deep Water, with the deep counterclockwise circulating cell associated with Antarctic Bottom Water filling the deep basin up to sill depth in the \(4\times\text{CO}_2\) and LGM simulations.

7) Globally, the atmosphere is moister in the warmer climate of \(4\times\text{CO}_2\) and drier in the glacial simulations. Globally, the precipitation rate decreases in both the LGM and the LGMCO2 case and increases in the \(4\times\text{CO}_2\) case, at a rate of \(-2\% \text{ °C}^{-1}\) (of temperature change), similar to climate change simulations in the IPCC AR4 (Held and Soden 2006).

8) The directions of the latitudinal shifts in the westerlies are consistent with the response to CO2 discussed by Toggweiler et al. (2006); however, the size of the effect is relatively minor in these simulations.

9) The eastern Pacific cold tongue SST shows a greater response to TOA radiative forcing than the western Pacific warm pool SST, with greater cooling in both glacial simulations and greater warming in the increased CO2 simulation than the PI. ENSO variability is reduced in the full LGM and \(4\times\text{CO}_2\) simulations but is not significantly different than PI in the LGMCO2 simulation when computed over the last 200 simulated years. Shorter segments do show significant variations. The LGM simulation shows a significant strengthening of the mean annual cycle of SST in the eastern Pacific.

There are still a number of outstanding uncertainties in our understanding of the forcings and boundary conditions for the LGM. Our CCSM4 LGM simulation prescribed the vegetation and aerosols at their preindustrial distributions to be consistent with the CCSM4 long-term control, twentieth-century, and future projection simulations, which also prescribe these forcings. Although the vegetation biogeography did not change in our LGM simulation, the phenology response to the LGM climate is included. Prediction of aerosols and vegetation will be included in Community Earth System Model (CESM) LGM simulations and are expected to have important regional effects. The PMIP3/CMIP5 LGM ice sheet reconstruction is a blended product obtained by averaging three different ice sheet reconstructions in light of comments by experts from the geologic community. All three reconstructions are constrained by observational evidence of ice margin locations but vary in the shape of the LGM ice sheets depending on their method. Because the radiative forcing caused by the ice sheets is of the same order of magnitude as all other glacial forcings summed together, assessing the sensitivity to the uncertainties in the ice sheet reconstructions should be included in future LGM simulations.

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APPENDIX

The Extension of the LGM Simulation

To examine whether the results presented here are not affected by surface climate trends, the LGM simulation was extended an additional 500 years. Over the extension period, the global annual mean surface temperature cools an additional 0.64°C and surface temperature over land cools an additional 0.56°C globally. The tropics cool by ~0.3°C and the NH and SH regions (averaged for the 20°–90° latitude range) cool by 0.5° and 1.1°C, respectively. In the last 100 years of the extension period, the surface trends nearly vanish in the NH with a cooling of less than ~0.02°C averaged over the region accompanied by no additional expansion in NH sea ice. However, surface temperature cools by ~0.2°C over the SH region over the last 100 years of the extension, accompanied by a 2% expansion of sea ice area. The trends in surface cooling minimally impact polar amplification (Fig. 11) at the end of the extension, which shows no change in the tropics, a small reduction in the high-latitude NH to ~2.6 from ~2.8, and a small increase over SH sea ice to ~2 from ~1.8. The other significant changes in Table 2 are an increase in the ocean barotropic transport through the Drake Passage to 253 Sv in response to the additional SH cooling, a weakening of the AMOC maximum at 34°S by ~2 Sv to be more comparable to the PI case, and a weakening in the AMOC maximum north of 28°N and below 500 m by just less than 4 Sv, which is still greater than the PI case.

In Fig. 4, the slope of the regression line continues to flatten with the addition of the 500-yr extension, resulting in an increase of the projected equilibrium temperature change, ΔTse in Table 4, to 6.1°C. This modestly increases the estimate of ECS in Table 4 to 3.4°C, still comparable to ECS estimated for the 4×CO2 simulation. The TOA radiative fluxes respond to the surface climate adjustments; however, when the global feedback parameters are reanalyzed using 30-yr means at the end of the extension period, there is little change to the results. The λeff in Table 4, reduces to 1.09 W m⁻² °C⁻¹, resulting in better agreement with the 4×CO2 case and yielding an effective climate sensitivity in excellent agreement with Bitz et al. (2012). In Table 5, the longwave feedback parameters change insignificantly by less than 0.02 W m⁻² °C⁻¹. The total shortwave feedback λsw increases to 0.77 W m⁻² °C⁻¹, resulting in better agreement with the 4×CO2 case, mostly because of an increase in λswclr of 0.09 W m⁻² °C⁻¹. For the shortwave feedback components estimated using the APRP method, λ_alb is increased modestly by 0.07 W m⁻² °C⁻¹, λ_cld is decreased in magnitude by 0.04 W m⁻² °C⁻¹, and λ_clr is not changed.

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