A Comparison of Climate Feedback Strength between CO₂ Doubling and LGM Experiments

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ABSTRACT

Studies of the climate in the past potentially provide a constraint on the uncertainty of climate sensitivity, but previous studies warn against a simple scaling to the future. Climate sensitivity is determined by a number of feedback processes, and they may vary according to climate states and forcings. In this study, the similarities and differences in feedbacks for CO₂ doubling, a Last Glacial Maximum (LGM), and LGM greenhouse gas (GHG) forcing experiments are investigated using an atmospheric general circulation model coupled to a slab ocean model. After computing the radiative forcing, the individual feedback strengths of water vapor, lapse-rate, albedo, and cloud feedbacks are evaluated explicitly. For this particular model, the difference in the climate sensitivity between the experiments is attributed to the shortwave cloud feedback, in which there is a tendency for it to become weaker or even negative in cooling experiments. No significant difference is found in the water vapor feedback between warming and cooling experiments by GHGs. The weaker positive water vapor feedback in the LGM experiment resulting from a relatively weaker tropical forcing is compensated for by the stronger positive lapse-rate feedback resulting from a relatively stronger extratropical forcing. A hypothesis is proposed that explains the asymmetric cloud response between the warming and cooling experiments associated with a displacement of the region of mixed-phase clouds. The difference in the total feedback strength between the experiments is, however, relatively small compared to the current intermodel spread, and does not necessarily preclude the use of LGM climate as a future constraint.

1. Introduction

Climate sensitivity is a useful metric that represents the gross size of the climate system response to radiative perturbations, and it also affects the speed of the response (Hansen et al. 1984). Despite the tremendous efforts made to estimate the climate sensitivity over the past decades, there is still a considerable range of uncertainty (Edwards et al. 2007). The reduction of this uncertainty by using modern instrumental records is limited by the poorly known indirect effect of aerosols and the rate of ocean heat uptake, as well as a small signal-to-noise ratio resulting from unforced natural variability (e.g., Knutti et al. 2002; Forest et al. 2002). An extension to earlier historical periods also is limited by small forcings with relatively large uncertainties (Rind et al. 2004). In addition, a consensus has not been reached regarding reconstructed global or hemispheric mean temperature changes (Mann et al. 2008). As a complementary approach, the Last Glacial Maximum (LGM) has received attention.

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The LGM refers to the period approximately 21,000 yr ago when the expansion of continental ice sheets was most extensive during the last glacial cycle. The LGM greenhouse gas (GHG) forcing with respect to the preindustrial level is roughly comparable to the magnitude of concern for the near future. An attempt has been made to estimate the climate sensitivity based on reconstructed temperature changes (Genthon et al. 1987; Lorius et al. 1990; Hoffert and Covey 1992; Lea 2004). While useful, this approach is subject to the uncertainty of the dominant forcing for regions of interest, and the locality of temperature records. As an alternate approach, models with known climate sensitivities have been tested against reconstructions (e.g., Manabe and Broccoli 1985; Hewitt and Mitchell 1997; Pinot et al. 1999; Broccoli 2000; Masson-Delmotte et al. 2006; Braconnot et al. 2007). Recently, a perturbed physics ensemble approach was adopted in which subjectively selected influential model parameters were varied and constrained by both present-day observations and LGM reconstructions, producing the probability distribution function for the “accepted” climate sensitivity (Annan et al. 2005; Schneider von Deimling et al. 2006).

Whereas the above models do not assume that the climate sensitivity is identical between the LGM and the future, a fundamental question remains as to whether processes that determine the climate sensitivity operate in much the same way. Hewitt and Mitchell (1997) and Broccoli (2000) found that the climate sensitivities of their LGM simulations were 17% and 23% higher than those of their CO2 doubling (2 × CO2) experiments, respectively. Broccoli (2000) noted, however, that the values in his experiments converge if LGM ice points are excluded or only tropical points are considered. Recent studies by Crucifix (2006) for multimodel simulations and Hargreaves et al. (2007) for perturbed physics ensemble simulations show that the climate sensitivity differs between LGM and 2 × CO2 experiments, and warn against a simple scaling to the future.

To understand how climate sensitivity is determined, it is essential to quantify radiative forcing, feedback strength, and temperature response in a consistent manner. Nevertheless, simplified formulas are often used to calculate the LGM radiative forcing (Solomon et al. 2007, Fig. 6.5). In addition, there currently is substantial uncertainty in the LGM radiative forcing estimate, ranging from around −8.6 to −4.2 W m⁻² (Hansen et al. 1993; Hewitt and Mitchell 1997; Taylor et al. 2000; Broccoli 2000; Solomon et al. 2007). Given the intermodel spread in the computed radiative forcing even for 2 × CO2 (Solomon et al. 2007, Table 10.2), it is desirable to compute the LGM radiative forcing with the same model used for the simulations, as conducted here.

The conceptually simplest definition of climate feedback is the processes that result from surface temperature changes, and that result in net radiation changes at the top of the atmosphere (TOA) and consequent surface temperature changes. For 2 × CO2 and Nakicenovic and Swart’s (2000) A1B scenario simulations, the feedback strength has been systematically investigated and compared between the models (e.g., Colman 2003; Soden and Held 2006). On the other hand, there have been few studies that investigate the feedback strength in LGM simulations to a similar extent. Hansen et al. (1984) estimated the joint water vapor, lapse-rate, and cloud feedback strength by perturbing the prescribed LGM sea surface temperature (SST), but individual contributions were not quantified. Hewitt and Mitchell (1997) estimated sea ice albedo and cloud feedback strengths separately. Ramstein et al. (1998) compared the cloud feedback between 2 × CO2 and LGM experiments, but the LGM feedback loop is not closed because of the use of prescribed SST. Crucifix (2006) investigated 2 × CO2 and LGM feedback strengths for four coupled models, but the longwave feedback strength was decomposed into the cloud forcing (Cess and Potter 1988; Tsushima et al. 2005), which likely also contains the effect of noncloud feedback (Soden et al. 2004) and a residual. The simplified method proposed by Taylor et al. (2007) is only applicable for shortwave feedback, and the method proposed by Yokohata et al. (2005b,a, 2007) does not permit the evaluation of, for example, water vapor feedback alone. In the current study, we apply the partial radiative perturbation method, which is known to be the method closest to the formal definition of the climate feedback (Bony et al. 2006), and we explicitly evaluate the strength of water vapor, lapse-rate, albedo, and cloud feedbacks for both the 2 × CO2 and LGM experiments.

An understanding of how feedback processes operate under both LGM and idealized future forcings would highlight important processes and provide an insight into whether the future climate sensitivity could be constrained by past climate changes, and which geographical locations may provide useful information for the constraint. The corresponding coupled version of the model investigated in this study has been used for various Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report simulations (e.g., Nozawa et al. 2005; Kimoto 2005), and is also one of five coupled models that were run for the LGM in that report (Yanase and Abe-Ouchi 2007) as well as a midHolocene simulation (Ogaito and Abe-Ouchi 2007). Therefore, it would be valuable to present fundamental characteristics of the model, and to link past and future climate changes.
The outline of the rest of this paper is as follows: A brief description of the model is given in the next section, and the experimental design is described in section 3. Methods that are used to calculate the radiative forcing and the feedback strength are explained in section 4. In section 5, results are presented, followed by a summary and discussion in section 6.

2. Model description

The model used in this study is an atmospheric general circulation models (GCM) coupled to a mixed layer ocean model. The model also includes land surface and sea ice components, and was developed jointly by the Center for Climate System Research (CCSR), National Institute for Environmental Studies (NIES), and Frontier Research Center for Global Change (FRCGC) in Japan. The atmosphere and land components are identical to those in the coupled model with an ocean GCM, Model for Interdisciplinary Research on Climate 3.2, medium-resolution version [MIROC3.2(medres); K-1 Model Developers 2004].

The atmospheric component is based on primitive equations with a triangular spectral truncation at wave-number 42, approximately equivalent to a resolution of 2.81°. There are 20 vertical sigma levels, with the model top at $\sigma = 0.0001$, or about 30 km. Full-radiation calculations are carried out every 3 h with a diurnal cycle, which employ the two-stream discrete ordinate method and the k-distribution method with maximum random cloud overlapping. Aerosol species included are sulfate, organic carbon, black carbon, sea salt, and soil dust. The indirect effect of aerosols, a modification of cloud properties such as albedo and lifetime, also is included. Aerosol concentrations are calculated online with prescribed emissions for anthropogenic aerosols. The prognostic tracer includes total water, which is diagnostically decomposed into water vapor, cloud liquid water, and cloud ice water. Moist convection is parameterized based on the formulation of Arakawa and Schubert (1974), with further modification as described in K-1 Model Developers (2004).

The ocean component is represented by a 50-m-deep, motionless, mixed layer. To account for ocean heat transport, which is not represented by the model, heat flux anomalies are added to the ocean. This so-called $q$ flux is determined so that the model produces realistic annual cycles of observed modern SST and sea ice distribution. The identical $q$ flux is used for all experiments except for the modern ocean area replaced by ice sheets in the LGM. The use of the mixed layer ocean model with the same $q$ flux enables us to focus on the atmospheric part of “fast” feedback processes (Hansen et al. 2008), and is consistent with the method used to estimate the climate sensitivity in the IPCC reports. It provides first-order approximation for the equilibrium climate sensitivity and feedbacks of atmosphere–ocean GCMs (Yokohata et al. 2008). The thermodynamic sea ice model is employed, which does not include ice dynamics and the representation of leads, and sea ice is transported neither by wind nor ocean currents. The land surface component represents energy and water exchange with the atmosphere. It has one canopy layer, five soil layers with a total depth of 2 m, and a maximum of three snow layers.

3. Experimental design

Four equilibrium experiments are carried out. The model is spun up at least 30 yr for each experiment, and 30-yr stable simulations after spinning up are used for the analysis. However, only the last 10 yr are used for the climate feedback analysis because of the relatively high computational cost. The first experiment (CTRL) is conducted with preindustrial boundary conditions, and serves as a reference control experiment in our suite of experiments. In the second experiment ($2\times\text{CO}_2$), the CO$_2$ mixing ratio of the atmosphere is doubled from the CTRL experiment. The third experiment (LGM) simulates the climate at the LGM. The LGM boundary conditions include Peltier’s (2004) ice sheets, denoted as ICE-5G, GHGs, and the earth’s orbital geometry (Berger 1978), but not sea level and vegetation changes (Table 1). Additionally, a fourth experiment (LGMGHG) was carried out in which only GHGs of the CTRL experiment were varied to LGM levels. A comparison between the $2\times\text{CO}_2$ and LGMGHG experiments provides some insight into the symmetry of the model response to warming and cooling experiments. Although there may be a slight difference in the response of the model to CO$_2$ and non-CO$_2$ GHGs (e.g., Hansen et al. 2005; Ponater et al. 2006), it is expected to...

<table>
<thead>
<tr>
<th>Boundary conditions</th>
<th>CTRL</th>
<th>$2\times\text{CO}_2$</th>
<th>LGM</th>
<th>LGMGHG</th>
</tr>
</thead>
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<tr>
<td>CO$_2$ (ppmv)</td>
<td>285.431</td>
<td>570.862</td>
<td>185.000</td>
<td>185.000</td>
</tr>
<tr>
<td>CH$_4$ (ppbv)</td>
<td>863.303</td>
<td>863.303</td>
<td>350.000</td>
<td>350.000</td>
</tr>
<tr>
<td>N$_2$O (ppbv)</td>
<td>279.266</td>
<td>279.266</td>
<td>200.000</td>
<td>200.000</td>
</tr>
<tr>
<td>Perihelion (°)</td>
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<td>102.04</td>
<td>114.42</td>
<td>102.04</td>
</tr>
<tr>
<td>Obliquity (°)</td>
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<td>23.450</td>
<td>22.949</td>
<td>23.450</td>
</tr>
<tr>
<td>Eccentricity</td>
<td>0.01672</td>
<td>0.01672</td>
<td>0.018994</td>
<td>0.01672</td>
</tr>
<tr>
<td>Ice sheet</td>
<td>Modern</td>
<td>Modern</td>
<td>LGM</td>
<td>Modern</td>
</tr>
</tbody>
</table>
be too small to be of concern for our purpose. A comparison between the LGM and LGMGHG experiments, on the other hand, isolates the effect of LGM ice sheets and minor orbital influence (Hewitt and Mitchell 1997).

4. Analysis method

a. Radiative forcing

At least three definitions of radiative forcing have been proposed, depending on to what extent adjustment processes are included; these are the instantaneous, adjusted stratosphere, the adjusted troposphere, and stratosphere forcings (Shine et al. 2003; Hansen et al. 2005; Solomon et al. 2007). In the current study, we focus on the stratospheric-adjusted forcing because it is consistent with the feedback analysis method used, but we also present instantaneous forcing for a reference. The instantaneous forcing ($F_i$) is evaluated by net radiative imbalance at the tropopause after the radiative forcing constituents are altered, but before the climate system responds to the perturbation. On the other hand, the stratospheric-adjusted forcing ($F_a$) is evaluated at arbitrary levels between the tropopause and the TOA after the stratospheric temperature is radiatively adjusted to the applied perturbation. Because the adjustment time for the stratosphere is much faster than for the surface–troposphere system, $F_a$ is expected to represent a more effective forcing than $F_i$ (Hansen et al. 1997).

For the calculation of $F_a$, we employ the method described in Stuber et al. (2001), which is based on the seasonally evolving fixed dynamical assumption. In this method, the stratospheric temperature is updated every time step until the equilibrium is reached with the radiative heating rate calculated from the perturbed radiation and the dynamical heating rate calculated from the control experiment. In order for the stratospheric temperature to reach equilibrium, the domain of the stratosphere must be fixed during the course of integration, and thus annual mean tropopause levels, predetermined by the control experiment, are used with a threshold lapse rate of 0.1 K km$^{-1}$. Note that the use of the tropopause definition by the World Meteorological Organization (WMO 1957) alters the forcing by only about 1% for 2% CO$_2$ for both $F_i$ and $F_a$. While $F_i$ is estimated by a 1-yr integration, $F_a$ is estimated from the second year of a 2-yr integration, using the radiative transfer part of the GCM program.

As discussed in Hewitt and Mitchell (1997) and Broccoli (2000), a precise estimate of ice sheet forcing is usually not easy to make because of the presence of snow in the reference climate. In this study, we estimate the LGM ice sheet forcing by the radiative imbalance resulting from the difference in surface albedo between the LGM and CTRL experiments only over the LGM ice sheets (excluding present-day ice sheets). This means that the resulting LGM ice sheet forcing includes minor effects of snow albedo feedback in these areas. The effect of surface albedo changes in the rest of the regions is evaluated as surface albedo feedback, as described in the next subsection.

b. Climate feedback strength

As with radiative forcing, the definition of feedback processes depends on to what extent adjustment processes are included. Because the stratospheric adjustment does not result from surface temperature changes, it is not included in feedback. Because of the method we employ, however, we do include other fast tropospheric adjustments not resulting from surface temperature changes, such as the cloud response to atmospheric CO$_2$ concentration discussed in Gregory and Webb (2008).

The simplest model in describing the earth’s energy budget can be expressed by

$$\Delta H(t) = F_a + \lambda \Delta T_s(t),$$

where $\Delta H$ and $\Delta T_s$ are rate of change in heat content in the climate system (or essentially ocean heat uptake) and surface temperature changes, respectively (however, see Stephens 2005). Here, $\lambda$ is called the climate feedback parameter, and it must be negative for the system to be stable. In the statistical steady state, the left-hand side of (1) is zero, and thus the feedback parameter becomes a negative inverse of the climate sensitivity, defined as $\lambda = \Delta T_s/F_a = -\Lambda^{-1}$.

If nonlinear interactions between contributions of individual feedback processes to anomalous net radiative fluxes are negligible, that is, $\Delta R = \Sigma \Delta R_i$, the feedback parameter is written as a sum of individual feedback parameters, as

$$\Lambda = \Lambda_p + \Lambda_q + \Lambda_\Gamma + \Lambda_A + \Lambda_C + \Lambda_M,$$

where $P, q, \Gamma, A, C$, and $M$ denote Planck response, water vapor, lapse-rate, surface albedo (and emissivity), cloud, and aerosol feedbacks, respectively. Here, the $i$th feedback strength is defined by $\Lambda_i = \Delta R_i/\Delta T_s$. In the estimate of $\Delta R_i$, we use the so-called partial radiative perturbation (PRP) method (Wetherald and Manabe 1988). To eliminate a bias resulting from the choice of the reference climate state and the so-called decorrelation error (Colman and McAvaney 1997), a two-sided analysis is conducted as follows:
Forcing constituent $F_i$ and $F_a$ represent instantaneous and stratospheric-adjusted radiative forcing (W m$^{-2}$), respectively. The values corresponding to the $2 \times CO_2$, LGMGHG, and LGM experiments are highlighted in bold font.

<table>
<thead>
<tr>
<th>Forcing constituent</th>
<th>$F_i$</th>
<th>$F_a$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$2 \times CO_2$</td>
<td>3.78</td>
<td>3.26</td>
</tr>
<tr>
<td>LGM CO$_2$</td>
<td>-3.24</td>
<td>-1.97</td>
</tr>
<tr>
<td>LGM CH$_4$</td>
<td>-0.18</td>
<td>-0.18</td>
</tr>
<tr>
<td>LGM N$_2$O</td>
<td>-0.49</td>
<td>-0.50</td>
</tr>
<tr>
<td>LGM GHGs</td>
<td>-3.03</td>
<td>-2.67</td>
</tr>
<tr>
<td>LGM orbit</td>
<td>0.06</td>
<td>0.06</td>
</tr>
<tr>
<td>LGM ice sheets</td>
<td>-2.37</td>
<td>-2.35</td>
</tr>
<tr>
<td>LGM all</td>
<td>-5.33</td>
<td>-4.95</td>
</tr>
</tbody>
</table>

$\Delta R_i^{(0)} = R[x^{(0)}, \mu_i^{(0)}, \mu_{j\neq i}^{(0)}] - R[x^{(0)}, \mu_i^{(0)}, \mu_{j\neq i}^{(0)}]$, (3a)

$\Delta R_i^{(1)} = R[x^{(1)}, \mu_i^{(0)}, \mu_{j\neq i}^{(0)}] - R[x^{(1)}, \mu_i^{(1)}, \mu_{j\neq i}^{(0)}]$, and (3b)

$\Delta R_i = [\Delta R_i^{(0)} - \Delta R_i^{(1)}]/2$. (3c)

where $R$ is the net radiative flux at the TOA. Here superscripts (0) and (1) denote the control and perturbation experiments, respectively, $x$ represents a forcing constituent such as CO$_2$, and $\mu_i$ represents meteorological fields that are associated with the $i$th feedback process. The actual computation is performed using the radiative transfer part of the GCM program. In addition, total feedback is calculated by exchanging all of the meteorological fields simultaneously. The Planck response is defined here as a radiative flux anomaly caused by a hypothetical, vertically homogeneous temperature change for each atmospheric column. It is computed by exchanging air temperatures that vary by exactly the same amount as the corresponding surface temperature changes. In computing the lapse-rate feedback, total temperature feedback $\Lambda_T$ is first calculated by exchanging both the air and surface temperatures, and then the Planck response is subtracted,

$$\Lambda_T = [(\Lambda_T - \Lambda_P) \Delta T_s - \Delta R_{adj}]/\Delta T_s$$, (4)

where $\Delta R_{adj} = F_a - F_i^{TOA}$ is the change in net radiation at the TOA caused by the adjustment of stratospheric temperature to the forcing. The relevant meteorological fields are prestored every 3 h for 10 yr. Because the concentration and mode radius of the aerosols are allowed to vary, aerosol feedback strength also is computed. However, we find that its magnitude is negligible in all experiments, and thus it is not discussed further (see Takemura et al. 2008, with vegetation changes). Also, the effect of changes in the surface emissivity is negligible.

Fig. 1. Zonal and annual mean stratospheric-adjusted radiative forcing for $2 \times CO_2$, LGMGHG, and LGM experiments.

5. Results

a. Radiative forcing

Global and annual mean radiative forcing for individual forcing constituents as well as joint forcings corresponding to the three perturbation experiments are listed in Table 2. The difference between $F_i$ and $F_a$ is minor, except for CO$_2$. Note that the adjusted radiative forcing of $2 \times CO_2$ is relatively small compared to other models (Solomon et al. 2007, Table 10.2). The inclusion of the stratospheric adjustments results in less than 10% of the difference in the total LGM forcing. As stated in section 3, the effects of sea level and vegetation changes are not included, and it is not our purpose here to provide the best estimate of the LGM radiative forcing, but to link the forcing, feedback, and response within a single model. For both $F_i$ and $F_a$, the total LGM forcing is nearly equal to the sum of individual forcings. The LGM ice sheets contribute almost half of the total LGM forcing, and the rest is contributed by GHGs. The LGM orbital forcing is very small in the global and annual average, consistent with the insolation change at the TOA.

For an understanding of feedbacks, the spatial distribution of the forcing also is important. Zonal and annual mean stratospheric-adjusted forcing is shown in Fig. 1. The $2 \times CO_2$ and LGMGHG experiments exhibit an almost mirror image with reversed signs. The maximum magnitude of these forcing occurs in the subtropics, where the outgoing longwave radiation (OLR) is climatologically large. The LGM ice sheet has the largest local forcing in magnitude, with peaks about 60°N and 80°S. The spatial pattern of the LGM ice sheet forcing is shown in Fig. 2. The largest negative forcings

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locally are $-106.73$ and $-84.53$ W m$^{-2}$ over the Laurentide and Fennoscandian ice sheets, respectively. The LGM orbital forcing has meridional and seasonal variations in which there are more positive and negative forcings in low and summer high latitudes, respectively, resulting from a smaller obliquity. Both the contrast in the planetary albedo between high and low latitudes and the contrast in the meridional distribution of insolation result in a net positive, yet small, global and annual mean radiative forcing. The $2 \times \text{CO}_2$ and LGM experiments exhibit a large contrast in the meridional structure of the forcing in which there is fractionally more extratropical forcing in the LGM experiment.

b. Overview of the feedback strength

The LGM simulation is colder than the CTRL experiment by $2.7 \text{ K}$ for the tropical SST and $9.5 \text{ K}$ for the Antarctica. These values are consistent with the reconstructions of $2.7 \pm 0.5 \text{ K}$ (Ballantyne et al. 2005) and $9 \pm 2 \text{ K}$ (Masson-Delmotte et al. 2006), respectively. The simulated coolings of $15.7 \text{ K}$ over Greenland and $4.2 \text{ K}$ for the tropical land surface air temperature (SAT) are both slightly underestimated from the reconstructions of $19–22 \text{ K}$ (Masson-Delmotte et al. 2006) and $5.4 \pm 0.3$ (Ballantyne et al. 2005), respectively. Whereas the reconstructed LGM winter sea ice extension to about $55^\circ \text{ N}$ in the North Atlantic (de Vernal et al. 2006, and references therein) is reasonably well captured by the model, the reconstructed areal expansion of the Southern Hemisphere winter (September) sea ice by 100% (Gersonde et al. 2005) is underestimated by the model, which shows only a 49% increase.

Figure 3 shows global and annual mean feedback strengths for the $2 \times \text{CO}_2$, LGMGHG, and LGM experiments with respect to the CTRL experiment. In the feedback analysis, meteorological fields are exchanged between two experiments on the sigma levels. This may lead to some overestimation or underestimation of the “actual” feedback strength over the LGM ice sheets, where the surface pressure changes substantially due to the elevated surface. To assess the impact on the global mean feedback strength, we recalculated it excluding those ice grid points. The result turns out to be very
similar to what is shown in Fig. 3. As presented in a previous study (Colman and McAvaney 1997), it is verified that the strength of the total feedback is virtually identical between when individual feedback strengths are evaluated separately, and when they are evaluated simultaneously. Therefore, it is possible to decompose and attribute the total feedback strength to individual feedback contributions. Note that this approximate linearity of the feedback strength is also confirmed in all grid points. It does not, however, imply that individual feedback processes operate independently (Hansen et al. 1984).

As is well known for the $2 \times CO_2$ experiments (e.g., Colman 2003), water vapor feedback exhibits by far the strongest positive feedback in all three perturbation experiments. Because the amount of water vapor varies rapidly along with temperature under small changes in relative humidity according to the Clausius–Clapeyron relation, and the radiative effect of water vapor is approximately proportional to the logarithm of its concentration (Held and Soden 2000; Solomon et al. 2007, p. 632), the strength of the water vapor feedback is similar between the $2 \times CO_2$ and LGMGHG experiments (see also Colman et al. 1997). In the LGM experiment, on the other hand, water vapor feedback is about 78% of those experiments, with the reduction occurring mostly in the longwave component (Table 3). The lapse-rate feedback in the $2 \times CO_2$ and LGMGHG experiments is negative, as is usually the case, because the temperature change in the tropical upper troposphere is greater than at the surface, which amplifies the change in radiative damping (Colman 2003). Interestingly, the LGM experiment shows that it is actually positive. The combined water vapor and lapse-rate feedback shows a similar strength for all three perturbation experiments. This is consistent with the understanding that the vertical distribution of moisture and temperature anomalies is tightly coupled (Hansen et al. 1984; Zhang et al. 1994; Colman 2001), as discussed later in detail, and with the results that global mean water vapor and lapse-rate feedbacks are negatively correlated among the models (Colman 2003; Soden and Held 2006).

Somewhat surprisingly, the strength of albedo feedback is also nearly the same for the three perturbation experiments. On the other hand, the strength of the cloud feedback varies significantly between the LGM and the other two experiments. Nevertheless, cloud feedback is positive in all experiments. The strength of the cloud feedback in the LGM experiment is only about 32% that of the $2 \times CO_2$ experiment. The weaker cloud feedback in the LGM experiments is due to the shortwave component, which is negative (Table 3). It appears that the shortwave cloud feedback tends to be weaker or even negative with stronger coolings. Note that the strength of the shortwave and longwave feedback is different between the $2 \times CO_2$ and LGMGHG experiments, although the sum is nearly the same.

The efficacy of the LGMGHG and LGM forcing, defined by the ratio of climate sensitivity to that of CO$_2$ forcing (Hansen et al. 2005), becomes 0.94 and 0.86, respectively (Table 4). If we include the global mean q-flux difference between the LGM and $2 \times CO_2$ experiments, arising from the different land–sea mask, in the LGM radiative forcing the LGM efficacy becomes 0.79. Because the effect of the LGM ice sheet elevation, not included in the radiative forcing, causes global mean cooling of about 0.32 K, assuming a lapse rate of 5 K km$^{-1}$ (Abe-Ouchi et al. 2007), the “effective”

<table>
<thead>
<tr>
<th>Expt</th>
<th>$\Lambda_{q}^{SW}$</th>
<th>$\Lambda_{q}^{LW}$</th>
<th>$\Lambda_{C}^{SW}$</th>
<th>$\Lambda_{C}^{LW}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$2 \times CO_2$</td>
<td>0.35</td>
<td>1.45</td>
<td>0.59</td>
<td>0.20</td>
</tr>
<tr>
<td>LGMGHG</td>
<td>0.35</td>
<td>1.44</td>
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<td>0.40</td>
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<tr>
<td>LGM</td>
<td>0.33</td>
<td>1.08</td>
<td>−0.11</td>
<td>0.35</td>
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</table>

Table 3. Decomposition of water vapor and cloud feedbacks into shortwave and longwave components (W m$^{-2}$ K$^{-1}$); $\Lambda_{q}$ and $\Lambda_{C}$ denote water vapor and cloud feedbacks, respectively, and “SW” and “LW” denote shortwave and longwave components, respectively.
The efficacy of the LGM forcing may be considered to be as low as 0.74. This means that the LGM forcing is about 14%–26% less efficient in changing the SAT compared to the CO2 forcing with a given radiative forcing. The total feedback parameter in the LGM experiment is smaller than the $2 \times CO_2$ experiment by $0.13 \pm 0.28$ K (W m$^{-2}$)$^{-1}$. We interpret this difference to be explained by the difference in the “ALL” feedback strength of $-0.39$ K (W m$^{-2}$)$^{-1}$, which is partially counteracted by the difference in the Planck response of 0.14 K (W m$^{-2}$)$^{-1}$.

Figure 4a shows the zonal mean SAT changes normalized by global mean values for the three perturbation experiments with respect to the CTRL experiment, whereas Figs. 4b–d show the zonal mean feedback strength for each experiment. In all three experiments, SAT responses exhibit the well-known polar amplification. The pattern of the polar amplification is also seen in the feedback strength denoted by “ALL” in the $2 \times CO_2$ and LGM experiments, but it is stronger in low latitudes than in Northern Hemisphere high latitudes in the LGMGHG experiment. The same amount of radiative perturbation, regardless of whether it is being introduced by forcing or feedback processes, results in the larger response in higher latitudes. To balance the radiative imbalance at the TOA, the surface temperature is required to change more with the lower emission temperature (Joshi et al. 2003), and with stronger stratification (Manabe and Wetherald 1975; Yoshimori and Broccoli 2008). In addition, the total feedback strength does not necessarily coincide with the SAT changes, because the latter also is affected by the horizontal heat transport (Cai 2005). It is worthy of note that the combined water vapor and lapse-rate feedbacks overwhelm the albedo feedback in the Northern Hemisphere high latitudes, where polar amplifications are seen in all
cases. This is particularly prominent in the LGM experiment, and the lapse-rate feedback is the strongest feedback over Antarctica in all experiments. These features are often understated and are not widely recognized, and they emphasize the importance of feedback analysis in understanding polar amplification.

c. Water vapor feedback

Figure 5a shows the zonal and annual mean strength of the water vapor feedback. The maximum in the feedback strength is seen in low latitudes in all three experiments, reflecting the longwave component. In the LGM experiment, there is a secondary maximum in feedback strength around 50°N, where large local radiative forcing is imposed by the ice sheets. The shortwave component, on the other hand, is strongest in summer polar regions where the insolation is strong over a highly reflective surface (not shown).

The water vapor feedback in the LGM experiment is weaker than in the other two experiments in most latitudes except for the Northern Hemisphere midlatitudes. Because the meridional distributions of the forcing are different between the experiments, the meridional structure of the temperature response also is different. To investigate how the global mean water vapor feedback strength is affected by such an inhomogeneity, the zonal mean strength of the water vapor feedback is normalized by zonal mean SAT changes in each latitude, rather than by global mean changes (Fig. 5b). The result shows that water vapor feedback strength is, overall, similar between the experiments, and is tightly coupled with the local temperature response.

The greenhouse effect caused by water vapor is sensitive to change in its concentration in the upper troposphere (Solomon et al. 2007, p. 632). With a given temperature change under a fixed relative humidity assumption, a larger fractional change in the water vapor concentration, which determines the effect of longwave absorption, occurs in a colder region (Held and Soden 2000). Together with a climatologically large OLR, the enhanced upper-tropospheric temperature change constrained by the moist-adiabatic lapse rate therefore strengthens the water vapor feedback in low latitudes. On the other hand, the temperature change is confined in the shallow near-surface layer in the strongly stratified high latitudes, which results in a weak water vapor feedback. As a result, the forcing that produces more extratropical than tropical temperature changes leads to a weaker global mean water vapor feedback than the forcing that produces more tropical than extratropical temperature changes.

The spatial patterns of the water vapor feedback strength, normalized by global mean SAT changes, are shown in Fig. 6. The maximum feedback strength of 4.49 W m⁻² K⁻¹ is seen over the Laurentide ice sheet, and a weaker, but still relatively large, feedback strength of 2.55 W m⁻² K⁻¹ occurs over the Fennoscandian ice sheet. By comparing the LGMGHG and LGM experiments, it is clear that a local, strong water vapor feedback is invoked by the ice sheet forcing.

d. Lapse-rate feedback

Figure 7 shows the zonal and annual mean strength of the lapse-rate feedback. A common feature in the three perturbation experiments is that the lapse-rate feedback is negative in low latitudes and positive in high latitudes. The vertical temperature profile approximately follows
the moist adiabat associated with condensational heating in low latitudes, that is, an enhanced temperature response in the upper troposphere. As the enhanced temperature change at higher altitudes amplifies the radiative damping from the surface, it results in negative feedback. The opposite change in the vertical temperature profile occurs in high latitudes where the response is confined to the shallow near-surface layer under strong stratification, resulting in a positive lapse-rate feedback. Therefore, the global lapse-rate feedback results from competition between positive and negative feedbacks in high and low latitudes, respectively. Because a substantial fraction of the forcing exists in high latitudes in the LGM experiment, the net feedback becomes positive.

In low latitudes, LGM exhibits the smallest lapse-rate feedback in magnitude, but the difference from the other two experiments is not large enough to compensate for the difference in the water vapor feedback. It is in the Northern Hemisphere mid- to high latitudes where the rest of the difference is canceled out, so that the global mean combined water vapor and lapse-rate feedback become nearly equal among the three experiments.

Changes in the thermal structure of the atmosphere are shown in Figs. 8a,c,e. A large temperature response is seen near the surface both in polar regions and in the tropical upper troposphere. As mentioned earlier, radiative, though not dynamical, temperature changes in the stratosphere are not included in the lapse-rate feedback. The relation between the atmospheric

Fig. 6. Annual mean water vapor feedback strength with respect to the control experiment: (a) $2\times$ CO$_2$, (b) LGMGHG, and (c) LGM experiments.

Fig. 7. Zonal and annual mean lapse-rate feedback strength with respect to the control experiment.
FIG. 8. Changes in the annual mean thermal structure of the atmosphere with respect to the control experiment. (left) Difference between the two experiments, and (right) the corresponding surface temperature changes that are further subtracted in each grid point and normalized by the global mean surface air temperature changes: (a),(b) $2 \times$ CO$_2$, (c),(d) LGMGHG, and (e),(f) LGM experiments. Negative values are shaded.
thermal structure changes and lapse-rate feedback strength is more readily understood when the air temperature changes are displayed after subtracting temperature changes at the corresponding surface and dividing by the global mean SAT change (Figs. 8b,d,f). The latitudes of solid and dashed contours in the troposphere approximately represent latitudes of negative and positive lapse-rate feedbacks, respectively. The latitudes of positive lapse-rate feedback extend farther equatorward in the Southern Hemisphere in the LGMGHG compared to the $2 \times CO_2$ experiments, but this is counteracted by the weaker positive feedback in the Northern Hemisphere high latitudes. The LGM experiment exhibits the equatorward extension of positive lapse-rate feedback regions compared to the $2 \times CO_2$ experiment, and the effect of ice sheet forcing is clearly recognizable by a comparison between the LGM and LGMGHG experiments. It is also noticeable that the vertical temperature gradient change in low latitude is smaller than in the other two experiments.

The spatial patterns of the lapse-rate feedback strength are shown in Fig. 9. The most striking feature is that a large, positive feedback is seen over the LGM ice sheets: the maxima in feedback strength of 10.38 and 9.33 W m$^{-2}$ K$^{-1}$ are seen over the Fennoscandian and Laurentide ice sheets, respectively. The strong positive lapse feedback and relatively weak water vapor feedback are consistent with the strong near-surface cooling caused by the ice sheets. Because both water vapor and lapse-rate feedbacks are positive over the LGM ice sheets, these feedback processes together amplify the temperature response there. There is a fairly distinct difference in the strength of the lapse-rate feedback in the Arctic between the LGMGHG and LGM experiments. More equatorward extension of the positive lapse-rate feedback region in the Southern Hemisphere in the LGMGHG and LGM experiments than in the $2 \times CO_2$ experiment coincides with sea ice changes.

e. Surface albedo feedback

Figure 10 shows the zonal and annual mean strength of the albedo feedback. In the Northern Hemisphere, the maximum in zonal mean albedo feedback strength occurs around 80°N in the $2 \times CO_2$ experiment, whereas it is around 60° and 43°N, much farther south, in the LGMGHG and LGM experiments, respectively. In the Southern Hemisphere, the maximum in zonal mean
The albedo feedback strength occurs around 66°S in the 2 × CO₂ experiment, whereas it is around 57°S, farther north, in both the LGMGHG and LGM experiments. This pattern is expected qualitatively because snow and ice retreat poleward in warming experiments, whereas they extend equatorward in cooling experiments in general. The smaller albedo feedback around 60°N in the LGM experiment than in the LGMGHG experiment is likely due to the fact that the LGM ice sheets and snow cover on their tops is treated as forcing, rather than feedback. Farther south, the albedo feedback essentially reflects snow cover changes on the continents (Fig. 11). In the Southern Hemisphere, it essentially reflects the expansion and retreat of Southern Ocean sea ice cover.

Figure 12 shows the annual cycle of the albedo feedback strength. In the 2 × CO₂ experiment, the albedo feedback is most pronounced in summer when the insolation is strong. There is a general tendency for the albedo feedback to peak in earlier melting seasons on the continents as one goes to lower latitudes. Indeed, both the LGMGHG and LGM experiments exhibit that the Northern Hemisphere maxima occur in spring. The length when the albedo feedback operates effectively also differs among experiments.

f. Cloud feedback

Figure 13a shows the zonal and annual mean strength of the cloud feedback. All three experiments exhibit the

![Cloud Feedback Diagram](image-url)
largest negative cloud feedback near 60°S, with more equatorward extension in the LGMGHG and LGM experiments than in the 2 × CO₂ experiment. On the contrary, the cloud feedback is positive in low latitudes in all of the experiments, and it is divergent near 60°N in all of the experiments. The difference between the LGM and 2 × CO₂ experiments is seen mainly in the Southern Hemisphere midlatitudes, north of the equator, and around 60°N (Fig. 13b). These differences mostly reflect differences in the shortwave component. Therefore, we focus on the shortwave cloud feedback. The differences in the shortwave cloud feedback between the LGMGHG and 2 × CO₂ experiments are seen also in similar latitudinal bands. In the shortwave cloud feedback, low latitudes (30°S–30°N) contribute to 50.8% and 49.5% of the global differences between the LGM and 2 × CO₂ experiments, and between the LGMGHG and 2 × CO₂ experiments, respectively.

The spatial patterns of the annual mean shortwave cloud feedback strength are shown in Fig. 14. As stated above, the gross tendency of negative and positive feedbacks, respectively, over the Southern Ocean and low-latitude oceans is a robust feature. The more equatorward extension of the negative cloud feedback in the Southern Ocean in the LGMGHG and LGM experiments than in the 2 × CO₂ experiment appears consistent with the more equatorward extension of positive albedo feedback in the corresponding experiments. It is
noticeable, however, that the latitudes where signs of
the cloud feedback switch do not exactly coincide with
the latitudes where the albedo feedback becomes zero.
This indicates that the cloud feedback in these latitudes
is not determined solely by changes in surface condi-
tions. There are relatively large longitudinal variations
in the shortwave cloud feedback in the cooling experi-
ments near the equator. Somewhat similar to the albedo
feedback, there is only a weak cloud feedback in the
Arctic in the LGMGHG and LGM experiments.

Changes in cloud amount normalized by the global
mean temperature change show a similar pattern in the
three perturbation experiments in which there is an
increase near the tropopause and a general reduction
below it (Figs. 15a–d). The latter contributes to the
positive shortwave cloud feedback in low latitudes, as
discussed in Wetherald and Manabe (1988) and Mitchell
and Ingram (1992). There is either a slight increase or a
neutral change in the cloud amount in the tropical
midtroposphere. This is accompanied by an increase in
relative humidity (not shown), but the changes in cloud
amount are relatively moderate because the relative
humidity is climatologically small there, and does not
readily reach saturation. The increased relative humidity
seems to be induced by the increased vertical transport
of moisture (Senior and Mitchell 1993). The earlier
models that parameterize the cloud amount as a func-
tion of only relative humidity also exhibit an increase
in low-level clouds in high latitudes in the $2 \times CO_2$
experiment, resulting from the confinement of in-
creased water vapor in the strongly stratified environ-
ment (Wetherald and Manabe 1988; Mitchell et al. 1989).
A stronger and much more similar response structure to
our three experiments is seen in models when tempera-
ture-dependent conversion between both cloud liquid
and ice waters are included (Mitchell et al. 1989; Senior
and Mitchell 1993).

The warming (cooling) leads to an increase (decrease) of
cloud liquid water at the expense of cloud ice water, and an
increase (decrease) of total cloud water because ice crystals
grow and precipitate more efficiently than cloud droplets
(Le Treut et al. 1994; Tsushima et al. 2006; Ogura et al.
2008). The shortwave cloud feedback associated with
this mechanism appears to be stronger in the tropical
midtroposphere and displaced equatorward near $60^\circ N$
in the cooling experiments (Figs. 15e,f, and 16). The
cooling leads to a displacement of the region of mixed-
phase clouds, between $-15^\circ$ and $0^\circ$, as parameterized
in the model, to lower latitudes and altitudes, where
more cloud liquid water exists climatologically; hence,
the feedback associated with the conversion is stronger.

The largest difference in the shortwave cloud feed-
back between the experiments occurs near $60^\circ N$. The
anomalous distributions of cloud liquid water shown in
Figs. 16e,f do not account for these differences. Note
that changes in cloud liquid water in polar regions
probably have a relatively small effect on feedback over
a snow/ice-covered reflective surface with small insola-
tion. In fact, a good geographical correspondence is
found in the anomalous distributions of cloud ice water
shown in Figs. 17e,f, in which the normalized cloud ice
water increases near $60^\circ N$ in the LGMGHG and LGM
experiments relative to the $2 \times CO_2$ experiment.
Changes in cloud ice water follow temperature changes
in the atmosphere, that is, there is a migration to higher
latitudes and altitudes with warmings (Figs. 17a–d). The
intensified negative shortwave cloud feedback near $60^\circ N$
in the LGM experiment with respect to the LGMGHG
experiment in Fig. 14 coincides with reduced low-level
clouds over and downstream of the Laurentide and

![Fig. 13. Zonal and annual mean cloud feedback strength (a)
with respect to the control experiment; and (b) difference from the
$2 \times CO_2$ experiment. “SW” denotes the shortwave component.](image-url)
Fennoscandian ice sheets (Fig. 18), where temperature and evaporation decrease substantially. Note that changes in cumulus cloud water are too small to account for differences in cloud feedbacks in all of the experiments. The difference in longwave cloud feedback between the experiments is mainly concentrated south of the equator, and appears to coincide with differences in the cloud ice water distribution.

Taken together, we propose as a hypothesis that the primary difference in shortwave cloud feedback strength in the cooling experiments from the warming experiment is due to the displacement of the region of mixed-phase clouds to lower latitudes and altitudes where cloud water is more abundant climatologically and there is a larger insolation for the case of lower latitudes.

6. Summary and discussion

Three different perturbation experiments—2×CO₂, LGMGHG, and LGM—are carried out in addition to a reference control experiment (CTRL). The climate feedback strength is quantitatively diagnosed for these experiments, and similarities and differences are examined. Although the results are based on only a single model, it is important to stress that computations of radiative forcing, climate simulation, and feedback analysis for forcings relevant to the past and the future are completed in a consistent way. This approach enables us to understand different climate sensitivities for different forcing constituents in terms of the feedback processes.

The water vapor feedback is weakest in the LGM experiment because it is more sensitive to forcing in lower latitudes, and the LGM has a fractionally larger high-latitude forcing due to the ice sheets. The lapse-rate feedback is positive in the LGM experiment, while it is negative in the 2×CO₂ and LGMGHG experiments. The strength of the global mean lapse-rate feedback is determined by the competition between the negative feedback in low latitudes and the positive feedback in high latitudes, and again the LGM experiment has relatively large high-latitude forcing due to ice sheets. The weaker positive water vapor feedback in low latitudes is compensated for by the stronger positive lapse-rate feedbacks in the mid- and high latitudes in the LGM experiment. As a result, neither the sign nor the meridional structure of the forcing substantially alter the combined global mean feedback strength. The albedo feedback is also shown to be nearly the same strength in all of the experiments. The albedo feedback strength varies, however, in latitudes and seasons.
among the experiments. The difference in the climate sensitivity and total feedback strength among experiments is attributed primarily to the shortwave cloud feedback, in which there is a tendency for the shortwave cloud feedback to become weaker or even negative with stronger cooling. This asymmetry in cloud response between the cooling and warming experiments is likely explained by a direction of displacement of the region of mixed-phase clouds.

The use of the slab ocean model rather than the ocean GCM excludes the effect of the ocean dynamical feedback. In addition, the thermodynamic-only sea ice model used in this study is simpler than the sea ice model of the corresponding atmosphere–ocean GCM, MIROC3.2(medres). Nevertheless, Crucifix (2006) also identified the shortwave cloud feedback as a primary factor for the difference between the $2 \times CO_2$ and LGM experiments with the MIROC3.2(medres), consistent...
with the current study. Although Crucifix (2006) estimated the longwave cloud feedback for $2 \times CO_2$ as being negative in MIROC3.2(medres), we find that the longwave cloud feedback is positive for both the $2 \times CO_2$ and LGM experiments. Because the application of the cloud forcing method to our $2 \times CO_2$ experiment also results in negative feedback, the difference is likely due to the different analysis methods used. Hewitt et al. (2001) investigated the effect of sea ice dynamics and found that the $2 \times CO_2$ climate sensitivity is reduced by 15% when it is included, and there is virtually no difference for the LGM climate sensitivity. Given that our LGM climate sensitivity is about 14% smaller than that of $2 \times CO_2$, the inclusion of sea ice dynamics might lead to the two climate sensitivities becoming closer. The effect of both sea ice dynamics

![Fig. 16. Same as in Fig. 15, but for cloud liquid water: $-15^\circ$ (black line) and $0^\circ$C (red line) isothermal lines. Perturbation (solid line) and reference experiments (dashed line). [For example, the solid and dashed lines in (b) represent $2 \times CO_2$ and CTRL experiments, respectively; and in (c) LGMGHG and $2 \times CO_2$ experiments, respectively.]]
and changes in ocean heat transport on the fast feedback process should be investigated in the future.

The current study suggests that the intermodel spread in the $2 \times \text{CO}_2$ climate sensitivity does not necessarily indicate a similar spread in the LGM climate sensitivity. Nevertheless, the difference in climate sensitivity and feedback strength between the $2 \times \text{CO}_2$ and LGM experiments in this particular model is much smaller, and it would be still useful to make use of paleoclimatic information to constrain the uncertainty.

The current study would enhance its value if a similar analysis was conducted for different models. In addition, this type of study would contribute to the interpretation of the climate sensitivity of more simplified models or perturbed physics ensemble simulations by GCMs or earth system models of intermediate complexity. For the assessment of the climate sensitivity during time periods other than the LGM (e.g., Barron et al. 1995; Covey et al. 1996; Pagani et al. 2006; Higgins and Schrag 2006; Royer et al. 2007), the current approach would also be
useful because it quantifies the link between rather complex boundary conditions and climate sensitivity. The difference in feedbacks needs to be taken into consideration in order to understand the link between the past climate and future projections, as well as between paleodata and model results.

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FIG. 18. Annual mean low-level cloud amount with respect to the control experiment: (a) $2 \times \text{CO}_2$, (b) LGMGHG, and (c) LGM experiments. Values are normalized by the global mean surface air temperature changes. Note that low-level clouds are not defined near the peak of the Laurentide ice sheet in the LGM experiment.

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