Global-Scale Energy and Freshwater Balance in Glacial Climate: A Comparison of Three PMIP2 LGM Simulations

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ABSTRACT

Three coupled atmosphere–ocean general circulation model (AOGCM) simulations of the Last Glacial Maximum (LGM: about 21 000 yr before present), conducted under the protocol of the second phase of the Paleoclimate Modelling Intercomparison Project (PMIP2), have been analyzed from a viewpoint of large-scale energy and freshwater balance. Atmospheric latent heat (LH) transport decreases at most latitudes due to reduced water vapor content in the lower troposphere, and dry static energy (DSE) transport in northern midlatitudes increases and changes the intensity contrast between the Pacific and Atlantic regions due to enhanced stationary waves over the North American ice sheets. In low latitudes, even with an intensified Hadley circulation in the Northern Hemisphere (NH), reduced DSE transport by the mean zonal circulation as well as a reduced equatorward LH transport is observed. The oceanic heat transport at NH midlatitudes increases owing to intensified subpolar gyres, and the Atlantic heat transport at low latitudes increases in all models whether or not meridional overturning circulation (MOC) intensifies. As a result, total poleward energy transport at the LGM increases in NH mid- and low latitudes in all models. Oceanic freshwater transport decreases, compensating for the response of the atmospheric water vapor transport. These responses in the atmosphere and ocean make the northern North Atlantic Ocean cold and relatively fresh, and the Southern Ocean relatively warm and saline. This is a common and robust feature in all models. The resultant ocean densities and ocean MOC response, however, show model dependency.

1. Introduction

Energy transports by the atmosphere and ocean are important components of the earth’s climate system that compensate the local disequilibrium of radiative energy. In the annual mean, the energy of incoming solar radiation exceeds that of outgoing earth radiation at low latitudes, and vice versa at high latitudes. An energy flow from the equatorial regions toward the polar regions is required for equilibrium of the climate system. This energy flow is achieved by large-scale motions of the atmosphere and ocean (the general circulations). The general circulation affects the temperature field through energy transport, and the temperature field also affects the local energy balance through longwave radiation. The chain of these processes ultimately determines the climatological features of temperature and circulation.

In the atmosphere, a considerable part of the energy is transported as latent heat (LH) of water vapor. Since
water vapor in the atmosphere is concentrated in the lower troposphere and its content (specific humidity) is sensitive to the surface temperature, the latitudinal profile of LH transport in the earth climate system differs from that of total energy transport. The energy and water vapor transport by the atmosphere affects the ocean temperature and salinity fields through surface fluxes and, thus, the ocean thermohaline circulation (THC). Comprehending the processes responsible for energy and freshwater transport by the atmosphere and ocean is of great importance, not only for understanding the present-day climate but also for the study of past and future climate change.

At geological time scales, climates of the past are reconstructed on the basis of paleoenvironmental proxies. Unfortunately, none of these proxies may be immediately interpreted as an indicator of the energy transport. In this context, paleoclimate simulations with general circulation models (GCMs) provide useful information for the physical understanding of the climate of specific ages. However, it is also true that existing models are not perfect. Each model has some bias, uncertainty, and processes that are not yet implemented. Computational resources can also restrict the length of the simulation. This can be especially problematic for paleoclimate simulations because the boundary conditions required are far different from those of the present day, and long-term integration is needed to approach steady state. Results from such simulations sometimes contradict each other. For example, atmosphere–ocean general circulation model (AOGCM) studies of the Last Glacial Maximum (LGM) disagree on the response of the Atlantic meridional overturning circulation (MOC) (e.g., Hewitt et al. 2001a; Kitoh et al. 2001; Kim et al. 2003; Shin et al. 2003a; Otto-Bliesner et al. 2006), even though many paleoceanographic proxies suggest a shallow and probably weakened Atlantic MOC during the LGM (e.g., Duplessy et al. 1988; Curry and Oppo 2005; McManus et al. 2004).

The Paleoclimate Modelling Intercomparison Project (PMIP) was launched in 1994 with two aims: 1) to understand the mechanisms of climate change by examining such changes in the past, that is, when the external forcing was large and relatively well known and when various kinds of geological record provide evidence of what actually happened, and 2) to provide a framework for the evaluation of climate models so as to determine how well they are able to reproduce climatic states very different from the present day (Joussaume and Taylor 2000; http://www.pmip2.cnrs-gif.fr/). For these purposes, the PMIP has defined a standard set of forcing and boundary conditions for the LGM, the mid-Holocene, and the preindustrial control climate (CTL). It also provides useful datasets synthesized from various kinds of geological evidence for those periods. The PMIP started with simulations using atmospheric GCMs and is now in the second phase (PMIP2)—targeting AOGCM or atmosphere–ocean–dynamic vegetation coupled general circulation model (AOVGCM) simulations.

In this paper, details of the energy and freshwater transport by the atmosphere and ocean in three PMIP2 LGM simulations are investigated in connection with changes in the general circulation. As shown in section 3b, the total energy transport at the LGM increased in the NH in all three models, except at high latitudes. The main subject of this paper is the detection of the processes responsible for this enhancement. The implications for the responses of the atmosphere and ocean MOCs are discussed. This is also a companion paper to Otto-Bliesner et al. (2007) and Weber et al. (2007).

2. Models and experiments

Climate models ranging from earth system models of intermediate complexity (EMIC) to AOVGCMs participated in PMIP2. Among those models, the Third Hadley Centre Coupled Ocean–Atmosphere GCM (HadCM3), with the second Met Office Surface Exchange Scheme (MOSES2); hereafter HadCM; the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 3 (CCSM3, hereafter CCSM); and the Center for Climate System Research (CCSR)/National Institute for Environmental Studies (NIES)/Frontier Research Center for Global Change (FRCGC) coupled GCM named Model for Interdisciplinary Research on Climate3.2.2 (MIROC3.2.2, hereafter MIROC) are presented in this paper. Only these three AOGCMs provide all of the variables required by our analysis for the LGM and CTL. The other PMIP2 submissions for LGM simulation [L’Institut Pierre-Simon Laplace Coupled Model, version 4 (IPSL CM4, hereafter IPSL); the National Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG) coupled model [Flexible Global Ocean–Atmosphere–Land System Model gridpoint version 1.0 (FGOALS-g1.0, hereafter FGOALS)]; and the Royal Netherlands Meteorological Office (KNMI) intermediate complexity model (ECBilt) coupled with the Louvain-la-Neuve Coupled Large-Scale Ice–Ocean (CLO) model (ECBilt-CLIO, hereafter ECBILT)] did not contribute some key variables (e.g., atmospheric or oceanic three-dimensional velocity fields) to the PMIP2 database. When it was possible, more limited analyses of these latter three models were used to check the robustness of the results. The AOGCMs contributing
to PMIP2 are basically the same model versions that were used in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) at each model center. Detailed descriptions of the models can be found on the PMIP2 Web site or the Program for Climate Model Diagnosis and Intercomparison (PCMDI) Web site (http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php).

Under the PMIP2 protocol, the LGM experimental conditions are summarized as follows: 1) reduced greenhouse gas (GHG) concentration, 2) insolation change due to 21 kyr BP orbital parameters, 3) surface albedo changes due to prescribed ice sheets, 4) orography changes due to prescribed ice sheets, and 5) changes in land–sea distribution and altitude due to LGM sea level drop (about 120 m). Detailed explanations of the experimental design can be found on the PMIP2 Web site.

Conditions 4 and 5 are based on ICE-5G data (Peltier 2004). Dynamic ice sheet models are not coupled to the AOGCMs under the PMIP2 protocol. Dynamic vegetation models and carbon cycle models are also not implemented in the versions of those models used in this study. Vegetation is prescribed to be the same as in the CTL simulations except for the regions covered by ice sheets or exposed due to the LGM sea level drop.

The conditions above are basically the same for all three models—CCSM, HadCM, and MIROC—but with some variation in the oceanic part. Bering Strait at the LGM is closed for the CCSM and HadCM simulations but is open for the MIROC simulation. The initial state of the ocean is taken from a previous LGM simulation for the CCSM (Shin et al. 2003a) and HadCM (Hewitt et al. 2003). The initial ocean state in the MIROC LGM simulation is taken from a preindustrial spinup run and integrated for about 500 years. After each spinup process, each model integrated more than 100 years and 50 or 100 yr means of the simulations are used as the model climatology, except for the 3D ocean variables in the CCSM simulations. The model climatology of 3D ocean variables (potential temperature, salinity, and velocity) in the CCSM simulations in the PMIP2 database are 10-yr monthly means. We may not be free from the influence of interdecadal variability in interpreting the results of the CCSM.

In the previous CCSM LGM simulation, the CCSM version 1.4 (CCSM1.4), a relatively coarse resolution model when compared to CCSM3, was used. A 1-psu salinity rise was applied to the CCSM1.4 LGM simulation to account for the saltier glacial ocean owing to large amount of water stored in the ice sheets. It is implemented in the CCSM3 LGM simulation by taking the ocean initial state from that simulation. No corresponding salinity rise is applied in the HadCM and MIROC simulations according to the PMIP2 protocol. It must be noted that an additional freshwater flux correction is applied to the HadCM simulations north of 40°N and south of 50°S to compensate for the long-term accumulation of snow over the adjacent ice sheets (see Weber et al. 2007). The total amount of corrected flux for the Atlantic/Arctic region is 0.033 Sv (Sv = 10⁶ m³ s⁻¹) in the CTL, 0.054 Sv for the LGM and around Antarctica it is 0.094 Sv for the CTL, and 0.054 Sv for the LGM.

3. Some basic results

In this section, we briefly summarize several basic results as a background for the subsequent discussions.

a. Global mean responses

Responses in global mean properties are given in Table 1. Differences in annual mean values of globally averaged insolation at the top of the atmosphere (TOA) between the LGM and CTL are negligible (less than 0.1 W m⁻²), but expansion of the cryosphere increases the planetary albedo by 1.6% as an average for the three models. Reduction in net downward shortwave radiation at the TOA is 5.6 W m⁻², and the decrease in the emission temperature of the earth is about 1.5°C. This is less than the decrease in surface air temperature of about 4.7°C, which means that GHG reduction and internal feedbacks have a large effect on LGM cooling. One feedback is reduced water vapor in the atmosphere. The reduction of water vapor content (precipitable water) is about 22%. This value is nearly equal to the product of near-surface relative humidity and the decrease rate of saturated water vapor pressure at the near-surface temperature. Reduction in precipitation is about 9%. This value is, of course, consistent with a decrease in surface latent heat flux, but the precipitation decrease is less than that of the total water vapor content in the atmosphere. The relationship between the responses of mixing ratio and precipitation is similar to the case of global warming experiments (e.g., Held and Soden 2006). A ratio of net surface radiation and latent heat flux (radiative dry index) calculated from the global mean properties increases by 0.04, suggesting a globally dry condition of the LGM climate. These features are common to all three models.

Table 2 gives responses of global mean ocean variables in the three models together with the values for the CTL. The decrease in sea surface temperature (SST) is shown to be about 1.9°C. This is nearly equal
to the decrease in the entire ocean volume-mean potential temperature. It must be noted that the MIROC shows a warm bias for the ocean interior in the CTL when compared with the other two models. However, the responses of volume mean temperature are similar among all three models. On the other hand, the three models differ in salinity response. The volume mean salinity for the LGM in the CCSM simulation is 1.8 psu higher than that in the CTL. This salinity rise is already seen in the previous CCSM LGM simulation and is partly attributed to the 1-psu salinity rise applied to that simulation (see section 2). The mean salinity for the LGM in the HadCM is 0.26 psu higher than that in the CTL. This discrepancy results from an incidental procedure taking the initial ocean state from the previous simulation. The mean salinity change in the MIROC is small. These differences in volume mean salinity between the two climate states are excluded from the comparisons in this paper except in the density calculation for seawater.

Another notable feature in the ocean response is sea ice expansion. As mentioned in section 2, the ice sheet distribution for the LGM is prescribed as a boundary condition. However, sea ice extent is calculated in each model and it acts as an internal feedback through the albedo effect. The expansion of the sea ice area in the three models is greater in the SH and less in the NH. In particular, the sea ice expansion in the SH for the CCSM is large when compared to the other two models.

b. Zonal mean responses

Figure 1 displays latitudinal profiles of zonal mean net shortwave and longwave radiation at the TOA, planetary albedo, and surface air temperature ($T_{sa}$). Changes in precipitation and precipitable water are shown as the ratio to CTL values (1: 30-yr mean).

| Table 1. Global mean responses in three AOGCM simulations. Changes in shortwave and longwave radiation fluxes at the TOA, planetary albedo, surface energy fluxes, surface air temperature ($T_{sa}$), precipitation, water vapor content (precipitable water), relative humidity at 925 hPa, and radiative dry index calculated from global mean values are shown with the CTL values averaged over the three models. Changes in precipitation and precipitable water are shown as the ratio to CTL values (1: 30-yr mean). |
|-----------------|--------|--------|--------|--------|--------|
|                 | CCSM   | HadCM  | MIROC  | CTL    | Units  |
| Downward SW at TOA | —      | —      | —      | 341.5  | W m$^{-2}$ |
| Upward SW at TOA  | 5.2    | 6.2    | 5.4    | 103.7  | W m$^{-2}$ |
| Upward LW at TOA  | −5.2   | −6.0   | −5.4   | 237.1  | W m$^{-2}$ |
| Planetary albedo   | 1.5    | 1.8    | 1.6    | 30.4   | %       |
| Downward SW at surface | 5.9    | 6.3    | 6.5    | 185.4  | W m$^{-2}$ |
| Upward SW at surface | 8.9    | 9.1    | 8.3    | 23.7   | W m$^{-2}$ |
| Downward LW at surface | −23.2  | −27.6  | −24.5  | 331.3  | W m$^{-2}$ |
| Upward LW at surface | −20.8  | −23.6  | −21.2  | 392.8  | W m$^{-2}$ |
| Latent heat flux ($T$) | −7.3   | −7.8   | −6.4   | 80.6   | W m$^{-2}$ |
| Sensible heat flux ($T$) | 1.6    | 1.1    | 1.1    | 19.0   | W m$^{-2}$ |
| $T_{sa}$          | −4.5   | −5.1   | −4.5   | 13.5   | °C      |
| Precipitation     | −9.2%  | −9.3%  | −8.2%  | 3.2 × 10$^{-5}$ | kg m$^{-2}$ s$^{-1}$ |
| Precipitable water| −18.4% | −21.6% | −22.2% | 23.0   | g m$^{-2}$ |
| Humidity at 925 hPa| −2.1   | −1.9   | −2.1$^1$ | 76.1   | %       |
| Radiative dry index| 0.05  | 0.04   | 0.04   | 1.25   |         |

Table 2. Ocean mean responses in three AOGCMs simulations. SST, potential temperature ($T$), sea surface salinity (SSS), salinity ($S$), sea ice area (SIA) are shown. The area of ice sheet covered region at LGM is excluded from SIAs with both states (1: 10-yr mean).

| Table 2. Ocean mean responses in three AOGCMs simulations. Changes in potential temperature ($T$), sea surface salinity (SSS), salinity ($S$), sea ice area (SIA) are shown. The area of ice sheet covered region at LGM is excluded from SIAs with both states (1: 10-yr mean). |
|-----------------|--------|--------|--------|--------|--------|
| LGM – CTL       | CCSM   | HadCM  | MIROC  | CTL    | Units  |
| SST             | −1.88$^1$ | −1.89  | −1.93  | 17.05$^1$ | °C    |
| $T$             | −2.06$^1$ | −1.88  | −1.96  | 2.81$^1$  | °C    |
| SSS             | 1.08$^1$  | 0.17   | −0.08  | 34.22$^1$ | psu   |
| $S$             | 1.87$^1$  | 0.26   | 0.01   | 34.72$^1$ | psu   |
| SIA             | 18.1    | 11.3   | 11.4   | 29.3   | 10$^{12}$ m$^2$ |
| SIA (NH)        | 1.8     | 3.3    | 3.5    | 13.3   | 10$^{12}$ m$^2$ |
| SIA (SH)        | 16.3    | 7.5    | 7.9    | 16.0   | 10$^{12}$ m$^2$ |
longwave radiation. This situation increases the meridional temperature gradient of the midlatitudes in the NH and can be expected to change the poleward transport of energy. On the other hand, radiation changes in the SH are not so large when compared with the changes at NH midlatitudes except near 60°S in the CCSM and around 20°S in the MIROC. In the CCSM, expansion of the sea ice area increases the planetary albedo around 60°S and increases the surface temperature gradient near that latitude.
An estimate of the total energy transport by the climate system is given by an area integration of net radiative energy flux at the TOA from the pole to each latitude. Similarly, an integration of net energy flux at the bottom of the atmosphere (BOA) gives an estimate of oceanic heat transport, and the balance of those two quantities gives an estimate of atmospheric energy transport. Figure 2 displays latitudinal profiles of these energy transports for the LGM and CTL (solid and dotted lines in left panels) and the differences between the two states (right panels) for each model. The figure clearly shows that the total poleward energy transport
increases in the NH in all three models except at high latitudes, and the response of the atmospheric transport roughly follows the total response.

Transport responses in the SH are small, as expected, from the radiation changes. The MIROC and HadCM show reductions of southward energy transport in the SH, but the CCSM shows enhancement.

Similar zonal mean responses in the NH (decrease in radiative fluxes at high latitudes and enhanced poleward energy transport, except at high latitudes) are seen in the other PMIP2 LGM simulations (IPSL, FGOALS, and ECBILT), but the responses in the SH show model dependency, such as weakened energy transport in IPSL and ECBILT and intensification in FGOALS (not shown).

4. Atmospheric energy transport and circulation

a. Decomposition of atmospheric energy transport

The atmospheric (static) energy transport is divided into two components: dry static energy (DSE) and LH transport. DSE transport consists of enthalpy (temperature) transport and potential energy transport (e.g., Peixoto and Oort 1992). Therefore, the meridional components of the atmospheric DSE and LH transport are defined as the following:

\[
\text{merid. DSE transp.} = \frac{2\pi a \cos \phi}{g} \int_0^{\text{ps}} \left[ C_p \bar{T} + g \bar{z} \bar{v} \right] dp,
\]

\[
\text{merid. LH transp.} = \frac{2\pi a \cos \phi}{g} \int_0^{\text{ps}} \left[ L \bar{q} \bar{v} \right] dp,
\]

where \( v \) is northward wind speed, \( T \) air temperature, \( z \) geopotential height, \( q \) specific humidity, \( p \) air pressure, \( \rho_s \) surface air pressure, \( g \) the acceleration due to gravity, \( a \) radius of the earth, \( \phi \) latitude, \( C_p \) atmospheric specific heat at constant pressure, and \( L \) is the latent heat of evaporation. The square brackets and overbar denote zonal and temporal average operators. Figure 3 displays latitudinal profiles of these quantities for the LGM (solid lines) and CTL (dashed lines) in the HadCM and MIROC. The DSE transport for the CCSM in this figure is derived as the residual of LH transport calculated from the climatology of surface fluxes because the CCSM lacks the climatology of appropriate variables (time averages of second-degree quantities) on the PMIP2 database.

The figure shows a decrease of LH transport over most latitudes and a southward shift of the peak in the NH. The decrease of LH transport is attributed to a 20% reduction in water vapor content in the lower troposphere. Conversely, DSE transport increases by about 20% at the NH midlatitudes for all three models. A similar response is also seen in the SH midlatitudes, although the increase in the MIROC is slight. At low latitudes, DSE transport is similar to that in the CTL in the NH, but decreased in the SH. The decrease in magnitudes of the LH and DSE transport in the SH low latitudes in the MIROC reaches about 20% of the CTL.

DSE transport can be divided further into three components: transient eddy transport, stationary eddy transport, and the transport by the mean meridional circulation (MMC). Figure 4 displays latitudinal profiles of these quantities. The MMC and stationary eddy components are calculated from the climatological monthly means of the 3D atmospheric variables, and the transient eddy component is calculated as the residual. The stationary eddy transport around 50°N
within the CCSM and MIROC (light blue lines) exceeds twice that of the CTL. The increase in the stationary eddy component in the HadCM around that latitude reaches about 40% of the CTL. The transient eddy transport decreases around that latitude with a shift in the peak southward in the NH (orange lines). MMC transport decreases at low latitudes in both hemispheres in all three models (purple lines). These features are investigated in further detail in the following two subsections.

b. Midlatitude waves and transport

In this subsection, the horizontal features of DSE transport in the CTL and LGM simulations are investigated. Figure 5 displays geographical maps of the vertically integrated northward DSE flux transported by stationary eddy waves in the CTL (left panels) and LGM (right panels). The DSE flux is largest over the east coast of the Eurasian continent in the CTL. In contrast, the DSE flux during the LGM decreases over the Eurasian region and increases over the North American continent or the North Atlantic in the three models. This feature implies an enhancement of stationary waves over these regions for the LGM. Figure 6 displays geographical maps of the annual mean geopotential height at 500 hPa (contour) and the anomaly from the zonal mean (shaded) for both climate states. The figure clearly shows an enhancement of stationary waves across or downstream of the Laurentide ice sheet during the LGM in the three models, which was also observed in many previous glacial climate simulations (e.g., Manabe and Broccoli 1985; Kutzbach and Guetter 1986). The increased DSE flux over the North American continent or the Atlantic for the LGM is maintained by the enhanced stationary waves due to the existence of large ice sheets. This response is common to all the models (including IPSL and FGOALS) except for a different response pattern between the HadCM and the other two models. These features well explain the response in the stationary eddy component seen in Fig. 4.

A similar analysis has been carried out for the transient eddy component of DSE transport for the MIROC using a 30-yr time series of 6-hourly instantaneous output. Zonal mean profiles of the three components of DSE transport calculated from these outputs show basically the same feature as in the bottom panels of Fig. 4 (not shown). The top panels of Fig. 7 display geographical maps of the vertically integrated northward DSE flux transported by the transient eddies. The transient eddy DSE flux is largest over the east coast of the Eurasian continent for the CTL, but weakens during the LGM. In contrast, the transient eddy transport intensifies over the North Atlantic. The other panels of Fig. 7 display geographical maps of the root-mean-square fluctuation of the geopotential height field at 250 hPa (Z250) as an indicator of transient wave activity. The middle panels show a rms anomaly from the climatological annual mean, and the bottom panels show the rms of synoptic-scale waves for each climate state. The latter is often referred to as the storm tracks. The figure shows a weakening of transient wave activity over the east coast of the Eurasian continent and an intensification over the Atlantic, consistent with the responses of the energy transport. This feature is seen both in total and synoptic-scale components. During the LGM, transient waves move the center of activity from the east coast of the Eurasian continent or the Pacific to the Atlantic.

In the present-day condition, it is well known that
atmospheric energy transport in the midlatitudes is mainly maintained by waves. This structure is basically the same for the LGM. However, the relative contribution from stationary waves increases during the LGM, and the contrast between the Pacific and Atlantic regions changes.

c. Hadley circulation and tropical atmosphere

In this subsection, the response of the MMC component in DSE transport at low latitudes is investigated in relation to the Hadley circulation response. Figure 8 displays mass streamfunctions of the mean
meridional circulation for June–August (JJA) and December–February (DJF) climatology during the LGM (contour) and deviations from the CTL (color shade). It is seen that the NH winter Hadley cell intensifies in all three models and the SH winter Hadley cell intensifies except in the MIROC. The figure also shows that the Hadley cell reduces its vertical and latitudinal extent during the LGM, and the NH winter Hadley cell seems to shift southward slightly. These features are also seen for the annual mean (not shown). As seen in Fig. 3, the magnitudes of the LH and DSE transport in low latitudes decrease, except for the DSE transport in low latitudes.
the NH. In addition, Fig. 4 shows that the MMC component of DSE transport decreases in the NH low latitudes. These facts, at first, appear to indicate a weakening of the Hadley circulation. However, the Hadley circulation during the LGM intensifies except in the SH for the MIROC. This issue is discussed next.

The left column of Fig. 9 displays vertical profiles of the annual mean northward wind velocity near 10°N

Fig. 7. Geographical maps of (top) vertically integrated northward DSE flux transported by transient eddies, (middle) rms 250-hPa height anomaly from the annual mean climatology, and (bottom) synoptic-scale waves using the MIROC for the (left) CTL and (right) LGM simulations. Contour intervals are $0.05 \times 10^8$ W m$^{-1}$, 30 m, and 30 m, respectively.
(thick lines) and 10°S (thin lines) for the LGM (solid lines) and CTL (dotted lines). Clear peaks of north–south wind speed can be seen at 200-hPa and near-surface levels. These two peaks correspond to the upper and lower limbs of the Hadley circulation. Vertical integration of this quantity from the TOA to each pressure level gives a vertical section of mass streamfunction of the annual mean MMC at each latitude. The middle column of Fig. 9 displays vertical profiles of potential temperature (equivalent to DSE) averaged over the 20°N–20°S tropical zone for both climate states. Vertical integration of the product of these two quantities gives the vertically integrated DSE transport by the MMC at each latitude.
The Hadley circulation carries DSE poleward in the upper limb and equatorward in the lower limb. Since the atmospheric DSE is larger in the upper troposphere, the Hadley circulation carries net DSE poleward. The situation is not changed for the LGM; however, the decrease in potential temperature is larger in the upper troposphere, and the upper limb of the MMC shifts somewhat downward. This situation causes a decrease of DSE transport during the LGM. To quantify this effect, we divide the change of DSE transport by the MMC into three contributions: a thermal component owing to DSE change, a kinematic component...
owing to velocity change, and a component due to both changes. The right-hand column of Fig. 9 displays latitudinal profiles of the vertically integrated thermal component (thick dashed lines) and the kinematic component (thick dotted lines) over low latitudes. The figure shows that the contribution from the thermal component is symmetrical in both hemispheres and acts to decrease the DSE transport in all three models. In particular, changes around 10°N roughly follow the thermal component.

The contributions from the kinematic component are more complicated. Near 30°N, the kinematic component mainly contributes to a decrease of DSE transport in all three models. This feature is attributed to the decrease of the latitudinal extent of the NH Hadley cell, seen in Fig. 8. Similarly, a peak of the kinematic component near 30°S in the HadCM is attributed to the decrease in the latitudinal extent of the SH Hadley cell. Near 10°S the kinematic component contributes to the decline in DSE transport for the CCSM and the MIROC. Particularly, in the MIROC the weakening of the SH Hadley cell, attributed to the reduced upper limb, Fig. 9 (c-1), also intensifies the weakening of DSE transport by the MMC. Finally, it is important to note that the kinematic component acts to weaken the cross-equatorial DSE transport. It is partly attributed to the decrease in the vertical extent of the SH Hadley cell and partly attributed to the southward shift of the NH Hadley cell. The southward shift of the Hadley cells and the asymmetric transport response induced by expansion of the NH cryosphere in mixed layer ocean coupled GCMs were already reported by Chiang and Bitz (2005) and Broccoli et al. (2006). These features consistently explain the response of DSE transport in the low latitudes seen in Figs. 3 and 4.

The intensification of the NH Hadley cell and decreased static stability in the tropical atmosphere of the models are in contrast to the results for the global warming experiments seen in some GCMs. However, weakened Hadley cells were found in some previous LGM simulations (e.g., Kitoh and Murakami 2002; Otto-Bliesner and Clement 2005). Furthermore, some models produce enhanced Hadley cells in a global warming experiment (e.g., Mitas and Clement 2006). In addition, observations of several reanalysis products suggest an intensification trend in the NH winter Hadley cell with respect to ongoing global warming (e.g., Tanaka et al. 2004; Mitas and Clement 2005). We must, therefore, be careful in interpreting the results. However, as shown in this paper, it is important to note that intensification of the Hadley circulation does not necessarily cause enhancement of the energy transport.

5. Ocean circulation and transport

a. Ocean circulations

The response of the Atlantic MOC in the PMIP2 LGM simulations has already been discussed by Otto-Bliesner et al. (2007) and Weber et al. (2007). For convenience to our later discussion, we briefly summarize the response of ocean circulation in the three PMIP2 LGM simulations. Figure 10 displays geographical maps of depth-integrated ocean circulation for the LGM (contour), and the differences from those in the CTL (color shade) for the Pacific and Atlantic Oceans. The figure clearly shows intensification and equatorward shifts of the gyre circulations in both the North Pacific and North Atlantic. These are the direct result of the intensified wind stress and are common features to the three models. The response of the subpolar gyre in the North Atlantic is somewhat complicated due to sea ice expansion. Figure 11 displays latitude–depth cross sections of the global ocean MOC streamfunctions for the LGM (contour) and the differences from those in the CTL (color shade). As mentioned by Otto-Bliesner et al. (2007), the LGM Atlantic MOC related to the formation of the North Atlantic Deep Water (NADW) is clearly shallower and weaker in the CCSM experiment, compared to the CTL, and is somewhat weaker and shallower in the HadCM experiment, but intensified and deepened in the MIROC. It is also noted that the MOC cell associated with the Antarctic Bottom Water (AABW) overflow intensifies in the CCSM and weakens in the MIROC and HadCM, compared to their CTLs. The NH MOC cell in the Pacific near the surface at low latitudes weakens in the CCSM and HadCM during the LGM but intensifies in the MIROC.

b. Oceanic heat transport

The left panels of Fig. 12 display latitudinal profiles of the northward oceanic heat transport for each ocean basin calculated from 3D ocean variables. Because the calculations have been made from climatological monthly means, these profiles do not contain transient eddy components. However, a comparison with area-integrated surface flux shows an overall good agreement between the two types of quantities for each ocean basin and in each model north of 30°S (not shown).

The right panels of Fig. 12 display the heat transport differences between the two climate states. The response of the heat transport by the global ocean (black line in Fig. 12) is essentially equivalent to the response of atmospheric bottom flux (blue line in Fig. 2). The oceanic heat transport increases in all three models between 40° and 60°N. South of 40°N, the northward
transport by the global ocean decreases in the CCSM, is virtually unchanged in the HadCM, and increases in the MIROC. However, considered separately, the response in each ocean basin is robust across the three models: the northward heat transport by the Atlantic generally increases south of 30°N and that by the Pacific decreases south of 40°N. These features are common to all three models, but the large decrease in the Pacific in the CCSM makes the total response for that model negative.

Oceanic heat transport shown in Fig. 12 can be divided into a component due to the mean meridional circulation and one by stationary eddies. These two categories are often referred to as the overturning and gyre components (e.g., Hewitt et al. 2001a). Hereafter, we follow this convention. We can also divide the difference of heat transport into three terms:

\[ \Delta(\bar{T}V) = V\Delta\bar{T} + \bar{T}\Delta V + \Delta\bar{T}\Delta V. \]
where $V$ is the northward velocity of ocean current, $\bar{T}$ is the temperature deviations from the entire ocean volume mean in each climate state, and $\Delta$ denotes the operation that takes the difference between the two states. The terms $V\Delta \bar{T}$, $\bar{T}\Delta V$, and $\Delta T \Delta V$ represent the contributions from thermal, kinematic, and both changes. Figure 13 shows these decompositions of the transport responses using thick lines in the Atlantic (right) and the Pacific (left). The former decomposition is represented by purple (overturning) and light blue (gyre) lines, and the latter is represented by green (thermal) and orange (kinematic) lines. The former decomposition for the heat transport itself is also displayed in the same panels with thin lines.

North of 25°N in the Pacific and north of 45°N in the Atlantic, the gyre component dominates the overturning component (thin light blue line). Decomposition of the transport response (thick lines) shows that the intensification of the Pacific heat transport north of 40°N is caused by an intensification of the subpolar gyre and
a southward shift of the polar front. In the Atlantic, the peak related to the subtropical gyre is relatively small, but a clear intensification and southward shift of the subpolar gyre component are seen (thin light blue line). In both regions, the intensification and southward shift of the subpolar gyre play the main role for enhancement of the midlatitudes ocean heat transport during the LGM.

Elsewhere in the Pacific and Atlantic, the overturning component dominates the gyre component. This feature is especially clear in the Atlantic. The northward heat transport generally increases in the low latitudes of the Atlantic. The decomposition into the thermal and kinematic components shows that this enhancement of heat transport is caused by temperature differences between the LGM and CTL for the CCSM.
and HadCM and is caused by velocity differences for the MIROC.

Figure 14 displays the streamfunction in the Atlantic for the LGM (contour) and the differences of zonal mean temperature deviations from the global volume mean in each climate state \( \Delta \bar{T} \) (color shades). Strong cooling in the lower limb of the Atlantic MOC can be seen in the CCSM and HadCM results. Depth integration of the product of \( \Delta \bar{T} \) and zonal mean \( V \) in the CTL gives the overturning component of the thermal contribution of the heat transport response at each latitude. The figure well explains how temperature contrasts between the upper and lower limbs of the Atlantic MOC cause the increases of heat transport in the CCSM and HadCM. Contrarily, in the MIROC the change in temperature contrast be-
tween the upper and lower limbs is indistinctive, and the velocity change is responsible for the increase in heat transport. An increased Atlantic heat transport for the LGM owing to enhanced MOC is also seen in the FGOALS and ECBILT results.

The situation to the south of 40°N in the Pacific is different. In that region, the response of the heat transport essentially follows the change in overturning and kinematic components, except near the equator in the MIROC. Some other mechanisms, related to wind stress, may act in that region. It is notable that all three models show the same transport response in the North Pacific and Atlantic, respectively, although the MOC responses are different.

c. Oceanic freshwater transport

Figure 15 displays latitudinal profiles of the northward freshwater transport by each ocean basin for the
LGM and CTL and the differences of those quantities between the two states in the three models. Oceanic freshwater transport must balance with the atmospheric water vapor transport in a steady state. During the LGM, freshwater transport by the global ocean decreases at most latitudes, consistent with the response of the atmospheric water vapor transport. Comparison with the latitudinal profiles of area-integrated surface freshwater flux in each ocean basin, including river run-off and the flux from sea ice, shows good agreement between the two profiles for each model except south of 30°S (not shown).

Figure 16 shows the decomposition of freshwater transport for each climate state (thin line) and differences between the two states (thick line), similar to Fig. 13. In the high latitudes (north of 50°N) of the Atlantic, a reduction of the gyre component mainly contributes to the decrease in freshwater transport. In the other

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**Fig. 15.** As in Fig. 12 but for freshwater transport (Sv).
parts of the Atlantic, changes in the overturning component dominate changes in the gyre component, except at some specific latitudes. The division into haline (green) and kinematic (orange) components shows a dominance of the salinity changes in the freshwater transport response. The decompositions of the responses in the Pacific are more complicated, but the salinity changes remain the principal control.

Contributions from the salinity changes to the response of Atlantic freshwater transports are also illustrated in Fig. 17 in which streamfunctions of the Atlantic MOC during the LGM and the differences in zonal mean salinity deviation from the global mean \( ([\Delta S] = \tilde{S}_{\text{LGM}} - \tilde{S}_{\text{CTL}}) \) are displayed. The upper limb of the Atlantic THC becomes relatively fresh during the LGM in comparison with the lower limb, and freshwater transport decreases in the North Atlantic. This feature is common to all three models.
Finally, we consider how the transport response of the atmosphere and ocean changes the heat and salt content of the ocean.

Figure 18 displays meridional profiles of the vertically averaged zonal mean deviations of ocean potential temperature (top), salinity (middle), and density (bottom) from the entire ocean volume means for the LGM (thick line) and CTL (thin line). The vertical average is taken from 2000-m depth to the surface. The lower limit of averaging was chosen based on the boundary of two MOC cells in the Atlantic during the LGM in the CCSM, but the results are insensitive to the choice of a lower limit ranging between 1500 and 3000 m. The profiles in Fig. 18 essentially represent the latitudinal profiles of heat and salt content from 2000-m depth to the surface and profiles of the pressure at 2000-m depth. Note that these are deviations from the global means in each climate state.

The top panels clearly show strong cooling in the Atlantic and relative warming in the Antarctic. This feature can also be seen in Fig. 14 for the Atlantic.
Temperature changes in the Pacific are relatively small in comparison with the other ocean regions. The middle panels clearly show a relative freshening of the Atlantic and a salinization of the Antarctic. The salinity increase in the Antarctic is particularly large for the CCSM compared to the other models. The Pacific also shows a clear freshening above 2000 m, which is particularly large in the CCSM. These facts imply that a considerable amount of salt must be moved to the Antarctic or the deep ocean from other ocean regions under the LGM climate. Salty bottom water at the time of the LGM can be seen in biogeochemical evidence (e.g., Adkins et al. 2002).

The density responses shown in the bottom panels result from the temperature and salinity responses discussed above. In the Atlantic and Pacific, the effects of
temperature and salinity changes essentially cancel each other and the latitudinal density profiles do not change much from the CTL, except for a slight intensification of the meridional density gradient around 40°N. However, density in the Antarctic greatly increases within the CCSM experiment owing to a large salinity rise and decreases in the HadCM. It is of interest that the north–south density contrast during the LGM increases in all three models, although the intensity of the Atlantic MOC differs among them.

It is notable that the indicated directions of change in temperature and salinity in the Antarctic and northern North Atlantic are common to all three models and seem to be consistent with the transport responses of the atmosphere and ocean (including interbasin-scale transport). Similar ocean responses in temperature and salinity are also seen in the other PMIP2 models (IPSL, FGOALS, and ECBILT). However, the degree of change and resultant density fields in those regions differ among the models.

6. Summary and discussion

We investigated the details of meridional energy and freshwater transports by the atmosphere and ocean in the three PMIP2 LGM simulations.

Atmospheric latent heat transport decreases at most latitudes owing to reduced water vapor content in the lower troposphere. Dry static energy transport greatly increases over the NH midlatitudes mainly due to enhanced stationary waves over the Laurentide ice sheet. The contributions from stationary waves in the midlatitude energy transport relatively increase, and the contrast between the Pacific and Atlantic regions changes under LGM conditions. At low latitudes, in spite of enhanced Hadley circulation in the NH, there exists a slight reduction of poleward DSE transport by the MMC as well as a reduction of equatorward LH transport. Changes in vertical structure of the tropical atmosphere play the main role for this response.

Oceanic heat transport increases from 40° to 60°N owing to intensification and southward shift of the subpolar gyres. South of 40°N, the northward heat transport decreases in the North Pacific and increases in the Atlantic. In some of the PMIP2 LGM simulations, an increased temperature contrast between the upper and lower limbs of the Atlantic MOC due to a strongly cooled NADW during the LGM contributes to the enhancement of Atlantic heat transport in low latitudes, even though the volume transport decreases in those models. In the other simulations, the change in the temperature structure is small, and enhanced volume transport is responsible for the increased heat transport. These two types of response explain why all models show a common heat transport response, although the MOC response shows model dependency.

As a result, the total energy transport increases from the equator to 50°N in all simulations. An enhanced energy transport in the NH at the time of the LGM can also be found in several previous simulation studies (e.g., Ganopolski et al. 1998; Hewitt et al. 2003; Shin et al. 2003a; Murakami and Kitoh 2005). As suggested by Murakami and Kitoh (2005), the PMIP2 LGM boundary conditions require an enhanced poleward energy transport in the NH, simultaneously with the increased meridional temperature gradient. It is of interest that the atmosphere and ocean seem to act together to satisfy this requirement, with an enhancement of stationary eddies in midlatitudes and vertical structure change at low latitudes. In some models, however, the ocean response in this context seems to be not yet achieved.

The response of the freshwater transport is rather simple except for the Arctic and Antarctic. Reduced surface evaporation makes the upper limb of the Atlantic MOC relatively fresh and the freshwater transport weak, consistent with the atmospheric response. This mechanism is common to all three models. At large scale, the water cycle must be closed within the atmosphere–ocean system; hence, the oceanic freshwater transport must be a direct reflection of the atmospheric water vapor transport in a steady state.

The response in the atmosphere and ocean also makes the northern North Atlantic colder and relatively fresh and the Antarctic relatively warm and saline during the LGM in comparison with changes in other ocean regions. These responses are common to all three models. However, the degrees of change in both regions, especially salinity change in the Antarctic region and the resultant density contrast between NADW and AABW, is different among the models.

As concluded by Weber et al. (2007), this difference in the simulation of density contrast between NADW and AABW among the models may be responsible for the difference in the ocean MOC response in the PMIP2 LGM simulations. However, the important issue is why the models differ in the extent of the salinity (or density) response. A possible cause for this difference in salinity change in the Antarctic (or the northern North Atlantic) is the response of sea ice amount, as suggested by Shin et al. (2003b) and Otto-Bliesner et al. (2007). This fact, however, begs the question why the models differ in sea ice response and questions the reliability of sea ice models under conditions quite different from the present day. Differences in the CTL climate (temperature and salinity fields or sea ice extent) may play some role (as shown by Hewitt et al. 2001b).
Cloud representation over ice-covered regions in the atmospheric models also may be responsible. Moreover, some ocean parameters may affect the results.

Regarding the ocean MOC response, we can find an important fact in previous AOGCM simulations of the LGM climate. In most models that provide a time series of MOC intensity approaching quasi equilibrium, the Atlantic MOC shows a clear intensification of volume transport within the early stages of the experiment, even in the models that finally produce a weakened Atlantic MOC (Hewitt et al. 2001c; Kitoh et al. 2001; Kim et al. 2003; Shin et al. 2003a). This fact, and the analysis in this paper, allows a hypothetical interpretation about the mechanism of the Atlantic MOC response to the LGM climate. The Atlantic MOC responds to the LGM boundary conditions by enhancing volume transport as an initial response so as to increase the northward heat transport. However, reduced water vapor transport in the atmosphere causes a reduced interbasin-scale freshwater transport in the ocean. It makes the North Atlantic relatively fresh and stabilizes the northern North Atlantic. This condition and the strong cooling over the northern North Atlantic contribute to form more cold NADW, and changes the vertical temperature structure of the Atlantic MOC cell. These factors provide another way to enhance the ocean heat transport and allow the MOC to reduce the volume transport. Synchronously, decreased water vapor transport in the atmosphere and enhanced sea ice processes in the Antarctic cause a saltier and denser AABW. The balance of these two processes may ultimately determine the water mass distribution at the time of the LGM.

As is shown in this paper, we can find several robust features in the transport response of the atmosphere and ocean in AOGCMs under the LGM boundary conditions. These provide a certain synthetic view of the LGM climate. However, many uncertainties still remain about circulation responses during the LGM, particularly about the MOC response in the atmosphere and ocean. More studies will be needed for a better understanding of the LGM climate and the climate system itself.

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