Setup of the PMIP3 paleoclimate experiments conducted using an Earth System Model, MIROC-ESM

T. Sueyoshi\textsuperscript{1}, R. Ohgaito\textsuperscript{1}, A. Yamamoto\textsuperscript{2}, M. O. Chikamoto\textsuperscript{3}, T. Hajima\textsuperscript{1}, H. Okajima\textsuperscript{2}, M. Yoshimori\textsuperscript{2}, M. Abe\textsuperscript{4}, R. O’ishi\textsuperscript{2}, F. Saito\textsuperscript{1}, S. Watanabe\textsuperscript{1}, M. Kawamiya\textsuperscript{1}, and A. Abe-Ouchi\textsuperscript{1,2}

\textsuperscript{1}Japan Agency for Marine-Earth Science and Technology, 3173-25, Showa-machi Kanazawa-ku, Yokohama City, 236-0001, Japan
\textsuperscript{2}Atmosphere and Ocean Research Institute, 5-1-5, Kashiwanoha, Kashiwa-shi, Chiba 277-8564, Japan
\textsuperscript{3}International Pacific Research Center, Pacific Ocean Science and Technology Bldg., 1680 East-West Road, University of Hawaii, Honolulu, Hawaii 96822, USA
\textsuperscript{4}National Institute for Environmental Studies, 16-2 Onogawa, Tsukuba-City, Ibaraki, 305-8506, Japan

Received: 27 July 2012 – Accepted: 9 August 2012 – Published: 3 September 2012
Correspondence to: T. Sueyoshi (sue@jamstec.go.jp)
Published by Copernicus Publications on behalf of the European Geosciences Union.
Abstract

The importance of climate model evaluation using paleoclimate simulations for better future climate projections has been recognized by the Intergovernmental Panel on Climate Change. In recent years, Earth System Models (ESMs) were developed to investigate carbon-cycle climate feedback, as well as to project the future climate. Paleoclimate events, especially those associated with the variations in atmospheric CO$_2$ level or land vegetation, provide suitable benchmarks to evaluate ESMs. Here we present implementations of the paleoclimate experiments proposed by the Coupled Model Intercomparison Project phase 5/Paleoclimate Modelling Intercomparison Project phase 3 (CMIP5/PMIP3) using an Earth System Model, MIROC-ESM. In this paper, experimental settings and procedures of the mid-Holocene, the Last Glacial Maximum, and the Last Millennium experiments are explained. The first two experiments are time slice experiments and the last one is a transient experiment. The complexity of the model requires various steps to correctly configure the experiments. Several basic outputs are also shown.

1 Introduction

For better future climate projection, the value of evaluating of climate models using experiments over different time scales is gaining recognition. As an experimental design, paleoclimate simulations provide unique opportunities to test models. Recent periods that are climatologically quite different to present day conditions are targeted for the Coupled Model Intercomparison Project phase 5 (CMIP5; Taylor et al., 2009)/Paleoclimate Modelling Intercomparison Project phase 3 (PMIP3). Here we describe three paleoclimate experiments using an Earth system model (ESM) MIROC-ESM, which are conducted as a series of experiments for the Intergovernmental Panel on Climate Change 5th Assessment Report (IPCC AR5).
Model simulations using ESMs, which are Atmosphere-Ocean general circulation models (AOGCMs) coupled with biogeochemical components, are now available for paleoclimate simulations. These models allow evaluation of more components compared to the previous AGCM work in PMIP phase 1, or model inter-comparison using AOGCMs in PMIP phase 2. With ESMs, the carbon cycle and biogeochemical processes can be calculated on-line as a model component. As the carbon dioxide plays a crucial role in controlling radiation, ESMs are being increasingly used in global climate change projection.

For IPCC AR5, three paleoclimate experiments were officially listed in the CMIP5/PMIP3 project (Taylor et al., 2009; Otto-Bliesner et al., 2009): exp. 3.4, Mid-Holocene (6000 yr before present time-slice, hereafter 6 ka), exp. 3.5, Last Glacial Maximum (21 000 yr before present time-slice, hereafter LGM) from Tier 1, and exp. 3.6 Last Millennium (850–1850 AD transient, hereafter LM) from Tier 2. We briefly review the type of scientific questions that draw the attention of the paleoclimate community.

The pronounced characteristics in the climate reconstruction during 6 ka are the enhancement of African and Asian summer monsoons and the associated enhanced vegetation activity over the Sahara desert (Hoeltzmann et al., 1998; Jolly et al., 1998; Peyron et al. 2006). The enhancement and northward shift of the summer African and Asian monsoons were reported in PMIP phase 1 (Joussaume et al., 1999) and PMIP phase 2 (Braconnot et al., 2007). These changes were consistent in sign with paleoclimate proxy records, mostly from pollen (Kohfeld and Harrison 2000; Prentice et al., 2000; Yu et al., 2000; Harrison et al., 2001, 2003; Bigelow et al., 2003; Pickett et al., 2004; Bartlein et al., 2010), but most of the models failed to simulate enough precipitation enhancement to maintain a vegetated Sahara (e.g. Doherty et al., 2000; Chikira et al., 2006). Many modeling works have investigated the vegetation feedback effect on the African monsoon change in 6 ka, and most have described the presence of positive feedback between vegetation and precipitation (Ganopolski et al., 1998; Doherty et al., 2000; Levis et al., 2004). Several studies investigated the role of the ocean on the summer monsoons using AOGCMs, and most suggested that there is further enhancement...
of the African monsoon when the ocean is coupled, contrasted with attenuation of the
Asian monsoon (Liu et al., 2003; Zhao et al., 2005, 2007; Ohgaito and Abe-Ouchi,
2007, 2009; Zhao and Harrison, 2011). Dallmeyer et al. (2010) used an ESM to in-
vestigate the effect of the ocean and vegetation component on the total change of the
Asian monsoon, and found that the ocean plays a major role in the enhancement of
precipitation. Experiments with ESM coupled with dynamic global vegetation models
provide an opportunity to evaluate the comprehensive influence on Asian and African
monsoon change resulting from dynamic vegetation and oceans.

The LGM was characterized by a very cold and dry climate. The Laurentide and
Fennoscandian ice sheets covered North America and Northern Europe (e.g. Denton
and Hughes, 1981), atmospheric greenhouse gas (GHG) levels were reduced, and
dust transportation was enhanced. (e.g. Barnola et al., 1987; Petit et al., 1981). Land
vegetation responded as well: subtropical deserts expanded and forests generally re-
gressed (Colinvaux et al., 1996, 2000; Marchant et al., 2009). At high latitudes, boreal
forests migrated southward, replaced by tundra and grassland (Prentice et al., 2000;
Tarasov et al., 2000; Ray and Adams, 2001; Harrison and Prentice, 2003). Due to these
significant differences from today, various aspects of the climate can be investigated.
For example, attempts to constrain climate sensitivity using LGM cooling (Annan et
al., 2005; Schmittner et al., 2011), the mechanism of the recorded weakening of the
Atlantic Meridional Overturning Circulation (AMOC) and the associated cooling in the
North Atlantic Ocean (Weber et al., 2007; Murakami et al., 2008), and southward shift of
the Intertropical Convergence Zone (ITCZ) and weakened monsoon activities (Rojas et
al. 2009) have been studied. Temperature changes have been discussed in comparison
with archives of proxy records (Otto-Bliesner et al., 2009; MARGO project members,
2009; Bartlein et al., 2010). With respect to the carbon cycle, the glacial-interglacial
atmospheric CO$_2$ variations remain an open issue (Kohfeld and Ridgwell, 2009). LGM
climate experiments using ESM with an interactive carbon cycle, will contribute to the
understanding of the mechanisms controlling CO$_2$ variations.
The LM includes two key climatic epochs: the “Medieval Climate Anomaly” (MCA, ca. 1000–1200 AD) and the “Little Ice Age” (LIA, ca. 1500–1850 AD), and has relatively rich proxy data compared to earlier times. Comparison of modelled climate with such observed records is useful in verifying the ability of a climate model, and assists the understanding of the mechanisms contributing to the climate variability. More specifically, the need to distinguish anthropogenic climate change from natural climate variability at centennial to millennial time scales is one of the key issues of climatology, addressing the social demand for climate research. LM experiments with fully coupled ESM allow discussion of the evolution of the carbon cycle interacting with the climate variation over the last 1200 yr.

LM simulations appearing in IPCC 4th Assessment Report (2007) were performed mostly by Earth-system Models of Intermediate Complexity (EMICs) (e.g. Gerber et al., 2003; Goosse et al., 2005). On the other hand, recent progress on proxy-based climate reconstruction has slowly revealed the spatial and temporal extent of various climatic epochs (e.g. Trouet et al., 2009; Mann et al., 2009), which allows better comparison between data and GCM simulations. Following some works with coupled models without a carbon cycle (e.g. González-Rouco et al., 2003; Ammann et al., 2007; Servonnat et al., 2010), a LM experiment using an ESM has been reported recently (Jungclaus et al., 2010), and this study follows the same direction in evaluating the strength of natural variability using up-to-date forcing reconstructions.

This paper presents the choice of settings for technical aspects of CMIP5/PMIP3 paleoclimate experiments using MIROC-ESM, including model settings, choice of boundary conditions, special treatment for model spin-up, and the initial data preparation. It is not our intention to present scientific results in detail here, and these will be discussed in subsequent papers.

Model and common settings are described in Sect. 2, specific settings for each paleoclimate experiment are described in Sect. 3 together with preliminary results, and concluding remarks follow in Sect. 4.
2 Model and common settings

MIROC-ESM is an ESM developed at the Japan Agency for Marine-Earth Science and Technology in collaboration with the University of Tokyo and the National Institute for Environmental Studies based on a global climate model MIROC (Model for Interdisciplinary Research on Climate) AOGCM (K-1 Model Developers, 2004; Nozawa et al., 2007). An atmospheric general circulation model (MIROC-AGCM, 2010), including an on-line aerosol component (SPRINTARS 5.00), an ocean GCM with sea-ice component (COCO 3.4), and a land surface model (MATSIRO), are interactively coupled in MIROC-ESM, as illustrated in Fig. 1. These atmosphere, ocean, and land surface components, as well as a river routine, are coupled by a flux coupler (K-1 Model Developers, 2004). As an ESM, MIROC-ESM has a newly introduced atmospheric chemistry component (CHASER 4.1) and carbon-cycle components for the land and ocean ecosystems. The carbon cycle is calculated by a marine biogeochemical component coupled with COCO and a terrestrial ecosystem component dealing with dynamic vegetation (SEIB-DGVM) coupled with MATSIRO. Watanabe et al. (2011) gives detailed descriptions of each component and the model’s performance against 20th century climate variations.

The atmosphere and land components have a spatial resolution of T42 (equivalent grid interval is approximately 2.8° in latitude and longitude). The atmosphere has 80 vertical layers between the surface and about 0.003 hPa up to the stratosphere (Watanabe et al., 2008a). Due to the limited computational resources, part of the CMIP5 experiments were performed with the full version of MIROC-ESM, while other experiments, including the paleoclimate experiments described in this paper, were performed using a version without the coupled atmospheric chemistry component (CHASER).

The ocean component COCO (CCSR Ocean COmponent model) has a resolution of 1.4° (longitude) by variable 0.56–1.4° (latitude) in the horizontal and 44 levels in the vertical (K-1 Model Developers, 2004; Hasumi, 2000). No flux correction is applied in exchanging heat, water, and momentum flux between the atmosphere and the ocean.
The sea ice is based on a two-category thickness representation, zero-layer thermodynamics (Semtner, 1976), and dynamics with elastic-viscous-plastic rheology (Hunke and Dukowicz, 1997).

The land surface component MATSIRO (Minimal Advanced Treatments of Surface Interaction and RunOff) consists of a single layer canopy, three layers of snow, and six layers of soil to a depth of 14 m (Takata et al., 2003). It is coupled to a river routing model, TRIP (Total Runoff Integrating Pathways) (Oki and Sud, 1998), for calculating river discharge. The aging effect on snow albedo and the effects of dirt in snow have been taken into account (Yang et al., 1997; Aoki et al., 2006). The dirt concentration in snow is calculated from the deposition fluxes of dust and soot calculated in the aerosol module, SPRINTARS (Spectral Radiation-Transport Model for Aerosol Species) (Takemura et al., 2000, 2002, 2005).

MATSIRO is coupled with the terrestrial ecosystem model SEIB-DGVM (Spatially Explicit Individual-Based Dynamic Global Vegetation Model) (Sato et al., 2007; Ise et al., 2009). SEIB-DGVM adopts an individual-based simulation scheme that explicitly captures light competition among trees, avoiding parameterized competition. Land vegetation is classified into 13 plant functional types (PFTs), and each PFT has different ecophysiological parameters and allometry relationships. This results in differential growth patterns and competition among PFTs under the environmental conditions in each grid cell. Leaf Area Index (LAI) is predicted by computing plant physiological processes such as photosynthesis, respiration, growth, and mortality, at daily time steps under the environment internally reproduced by MIROC-ESM. The predicted LAI is then used for calculation of the biogeophysical processes in MATSIRO: the processes for radiation transfer including surface albedo, interception of precipitation, and energy transfer by latent/sensible heat. In the current configuration, MATSIRO considers the dynamic variations in LAI, while the ground surface type in MATSIRO is fixed, not referring the PFT predicted by SEIB-DGVM. Thus, the terrestrial ecosystem in MIROC-ESM affects climate only through the carbon cycle and the LAI feedback.
The oceanic biogeochemical model is based on a simplified ecosystem of nutrient-phytoplankton-zooplankton-detritus (NPZD) ecosystem model (Oschlies, 2001; Oschlies and Garçon, 1999). The model includes two plankton functional groups of phytoplankton and zooplankton, suspended and sinking particulate detritus, four dissolved inorganic components of alkalinity, dissolved inorganic carbon (DIC) and nitrate. A constant Redfield ratio \( C / N = 6.625 \) is used to estimate the carbon and calcium flow. The sea-air \( CO_2 \) flux is calculated by multiplying the difference of ocean-atmosphere \( CO_2 \) partial pressures by the ocean gas solubility.

The basic settings of the experiments are common with other CMIP5 historical simulations by MIROC-ESM (Watanabe et al., 2011). Among our paleoclimate experiments, 6 ka and LGM are time-slice experiments conducted to obtain an equilibrium state, while LM is a transient experiment to conduct a time-evolving simulation under the reconstructed time series of boundary conditions.

### 3 Experimental settings

#### 3.1 Pre-industrial control experiment

**3.1.1 Experimental setup**

The Pre-Industrial control experiment (PI) (exp 3.1 in CMIP5) is firstly described as a counterpart of the following paleoclimate simulations. PI is forced by the climatic conditions at 1850 AD. The greenhouse gas levels (GHGs) are set to 285 ppm \( CO_2 \), 0.3 ppm \( N_2O \) and 1.7 ppm \( CH_4 \) as listed in Table 1. The land index for MATSIRO is shown in Fig. 2a, based on modern vegetation types with less croplands corresponding to the plausible distribution in 1850. The orbital parameters are set to 0.01672 for the eccentricity, 23.45° for the obliquity, and 102.04° for the angular precession.
3.1.2 Spin-up of the PI data

Initial data for PI experiment was prepared in stages. The procedure is shown in Fig. 3a: starting with the version of MIROC-ESM with 20 vertical layers in the atmosphere, a 380-yr integration was performed as an atmosphere-ocean coupled system, followed by a 50-yr integration with an atmosphere with 80 vertical layers. In the course of the spin-up runs, representative states and fluxes in the physical climate and carbon cycle components were monitored: surface air temperatures, radiation fluxes at the top of the atmosphere, strength of the thermohaline circulation, sea-ice extent, soil and vegetation carbon storage, etc. The spin-up run was continued until linear trends in the last 50 yr of these quantities became insignificant. After these spin-up procedures, a 630-yr integration was carried out as the PI experiment. The last 100 yr are used for the analyses in this paper.

3.1.3 Model performance

The climate of the PI experiment is briefly presented. The sea surface temperature (SST) is shown in Fig. 4 in comparison with the World Ocean Atlas Temperature data (WOA, 1998; Levitus et al., 1998). PI simulated reasonable SST distribution. The global averaged SST in PI is slightly cooler than that in the observational data, because the PI experiment is set to pre-industrial times whilst the observational data were obtained during the 20th century.

The distribution of simulated precipitation in comparison with observational data (Xie and Arkin, 1996) for boreal summer (JJAS) and winter (DJF) are shown in Fig. 5. PI simulates a reasonable representation of the ITCZ and precipitation distribution of the monsoon area. The other basic climate variables (not shown) are generally well represented.

Although the climatic trends were thought to be insignificant, a slight warming drift (0.1°C century⁻¹ in global average surface air temperature) still existed in the PI experiment, which was overlooked during the spin-up procedure. It is considered to be
due to the slow reduction in sea-ice extent (of 3% century\(^{-1}\)) in the Southern Ocean, which caused a similar warming drift in LM experiment (Sect. 3.4). Because of this drift, additional care is needed in analysis.

3.2 6 ka

3.2.1 Forcing data and boundary conditions

The 6 ka time-slice experiment was performed following the PMIP3 protocol (Table 1). The boundary conditions are kept unchanged from the PI run, except for the orbital parameters (eccentricity: 0.018682, angular precession: 0.87°, obliquity: 24.105°) and GHG concentrations (CO\(_2\): 280.0 ppm, N\(_2\)O: 0.65 ppm, CH\(_4\): 0.27 ppm). Volcanic aerosols changes are not considered. Since 6 ka GHG levels are almost the PI condition, this 6 ka condition essentially tests the change in the seasonal incoming shortwave radiation pattern (Fig. 6a) by the orbital parameter settings.

3.2.2 Preparation of the initial data

Figure 3b shows the procedure to prepare the initial data for the 6 ka experiment. We have made a “branch” for 6 ka, deviated from the 250th year in PI with replacing GHG levels and the orbital parameters.

As the land surface condition in the PI experiment includes impacts from anthropogenic land use changes, in 6 ka we initialized the terrestrial ecosystem to a steady state without anthropogenic land usages. To shorten the spin-up time, the initialization was conducted with the off-line terrestrial ecosystem model that is used in MIROC-ESM. Firstly, 100 yr of integration was performed using the 6 ka GHG concentrations and orbital forcing using MIROC-ESM, then surface physical quantities obtained in the last 25 yr (i.e. last 25 yr of “6 ka (1850 land use)” in Fig. 3b) were recursively adapted to the off-line model for 2000 simulation years. After this procedure, the quantities of
terrestrial ecosystems were merged back into the initial conditions for MIROC-ESM, followed by a spin-up of 200 simulation years with all components coupled (Fig. 3b).

### 3.2.3 Results

The results of the 6 ka experiment are also influenced by the warming drift identified in the PI experiment. In order to minimize the effect of drift, the following analyses are based on the anomalies between 6 ka and PI, and the integration time from the point of initialization of 6 ka (i.e. “branch” time) was kept the same for both experiments.

The changes in simulated temperature 6 ka-PI for JJAS and DJF are shown in Fig. 7. Pronounced changes are the warming over most of the boreal continents by 1–3°C in JJAS in response to the strengthened solar radiation in the boreal summer. The tropical oceans are slightly cooled. This corresponds to the net negative radiation changes over the tropics compared to PI.

The changes in simulated precipitation are shown in Fig. 8. In JJAS, enhanced precipitation over the Sahel and northern India by 1–3 mm day\(^{-1}\) and the reduction of precipitation south of these areas in similar magnitude are simulated. These precipitation changes and the circulation changes suggest an enhancement of the African and Asian monsoon. However, whilst these enhanced monsoons are consistent with the proxy records, the amount of enhancement of precipitation over most of the Sahara Desert is not enough to explain the proxy records. In monitoring the daily-predicted PFTs over the Sahara, SEIB-DGVM predicts non-desert vegetation more often than in the PI case, though they are still much less frequent than desert vegetation. The monsoon activities for the boreal summer are investigated in another work (Ohgaito et al., 2012).

The overall cooling for the boreal winter is seen in Fig. 7, as expected from the radiation change. The precipitation change (Fig. 8) in DJF shows a weakening of precipitation over the continents in the Southern Hemisphere by 0–2 mm day\(^{-1}\), which suggests the weakening of monsoon activities.
3.3 LGM

3.3.1 Forcing data and boundary conditions

The LGM experiment is an equilibrium experiment using the boundary conditions defined by the PMIP3 protocol (Table 1), except the salinity setting. The ocean salinity was assumed to be unchanged from the PI condition.

Atmospheric CO$_2$/N$_2$O/CH$_4$ concentrations, as well as orbital parameters are set to the values shown in Table 1. The shortwave solar radiation deviation at the top of the atmosphere shown in Fig. 6b is much smaller compared to the 6 ka orbital change.

Another major difference in the LGM boundary condition is the topography. The Laurentide and Fennoscandian ice sheets covered wide areas of North America and the northern half of Europe. In the ESM, ice sheets are expressed as mountain ranges with albedo of ice (i.e. flow is not considered). Three ice-sheet reconstructions are currently available: ICE-6G (Peltier, 2009; Argus and Peltier, 2009; Peltier and Drummond, 2008), ANU Ice Model (Lambeck and Chappell, 2001; Lambeck et al., 2002; Lambeck et al., 2003), and GLAC-1 (Tarasov and Peltier, 2002, 2003). Since those reconstructions vary in topography, the settings of our experiment follow the “PMIP standard ice sheet” recommended by the PMIP3 committee, which averages the three reconstructions (PMIP3, 2010). The horizontal resolution of the data provided in PMIP3 ($1^\circ \times 1^\circ$) was regridded to T42 resolution and to the spectral wave number space as other boundary data. The orographic difference for LGM-PI is shown in Fig. 9.

Sea level was lowered by about 90 m, which closes the Bering Strait, and joins together the Indonesian maritime continent. The total mass of the atmosphere was not adjusted. The vegetation boundary condition of the LGM simulation for MATSIRO was created based on the PI vegetation distribution. At first, the vegetation types in the ice sheet area given by the PMIP3 protocol are changed to ice sheet, and the rest of the land areas are basically not changed from PI. However, due to the lower sea level in the glacial climate, the area of the continental shelves should have some vegetation type specified, and the same vegetation types are given as the surrounding grids (Fig. 2).
3.3.2 Preparation of the initial data

The initial data was prepared following the procedure shown in Fig. 3c. The initial condition of the ocean physical field is inherited from the result of LGM experiments by an earlier version of MIROC, MIROC4m (a bug fixed version of MIROC3.2 (medres)), having the same resolution. This experiment has 1900 yr integration under the PMIP2 LGM condition and is regarded as the equilibrium. Since the PMIP3 LGM condition slightly differs from the PMIP2 condition in the ice sheet topography, a spin-up time for the ocean is necessary.

As the atmospheric initial condition, the 250th year of the PI experiment by MIROC-ESM is used, as is the case for the other two experiments. Since MIROC-ESM has a different vertical coordinate system from previous versions, it is difficult to initialize the model by applying the initial condition of these old versions. We thus newly produced the atmospheric initial condition of LGM by starting with PI conditions and conducted the following spin-up for 150 model years. Since the adjustment time of the atmospheric physical field is considerably shorter than that of the ocean, the atmospheric initial condition does not affect the length of the spin-up.

Meanwhile, special care is needed to start an integration under the new topography, which has the large ice sheets in the Northern Hemisphere. The surface pressure field must be adjusted to the change in surface elevation over the continents. As MIROC-ESM has high vertical resolution in the stratosphere, the model was sensitive to the topography change, generating gravity waves and breaking the Courant–Friedrichs–Lewy condition by generating unrealistic high wind velocities. As recommended in the PMIP3 protocol, integration under the new topography can be achieved through two possible methods: either by gradually changing the surface elevation in order to avoid generating gravity waves, or by adjusting the initial pressure field to the LGM surface elevation. In MIROC-ESM LGM, both options were applied.

Since the atmospheric CO₂ concentration in the LGM experiment needs to be reduced by ca. 100 ppm compared to the PI condition, the carbon budget in the initial
data requires long integration to reach a steady state. After the physical field of ESM (i.e. atmosphere-ocean-land) achieved a quasi-equilibrium state, land and ocean biogeochemical components were run separately in an off-line experiment, using the ESM outputs as the required inputs.

The terrestrial ecosystem was spun-up using the offline land ecosystem model, as in the initialization for the 6 ka experiment. The vegetation and soil variables were calculated for 2000 model years under the physical field of the LGM climatology to obtain a steady state. During this initialization process, the influence of anthropogenic land-use was also eliminated. In addition, the prescribed distribution of the ice sheets was used for the processes that prohibit the establishment of new vegetation and accumulations of soil carbon over the ice sheets.

The spin-up of the oceanic ecosystem was performed in a similar manner. Starting from the PI condition, LGM climatology was given recursively to the off-line model (Chikamoto et al., 2012) as boundary conditions for 3500 model years. The partial pressure of the atmospheric CO₂ was set to 185 ppm, to calculate air-sea CO₂ gas exchange. This setting forces the near-surface oceanic pCO₂ to be 189 ppm and achieved a new equilibrium of marine carbon distribution. It results in the removal of the oceanic CO₂, which is equivalent to an increase of ca. 550 ppm CO₂ in the atmosphere.

After these offline experiments had achieved equilibrium, variables are merged back into the experiment using MIROC-ESM and integrated over 200 yr to obtain an equilibrium state for the whole system under the LGM condition.

### 3.3.3 Results

The change in simulated mean annual temperature for LGM-PI is shown in Fig. 10. More than 20° cooling over the Laurentide and Fennoscandian ice sheets is seen. The rest of the globe cooled much less. Tropical cooling ranges about 1 to 3°, which is consistent with the proxy records (MARGO Project Members, 2009).
The changes in simulated annual precipitation are presented in Fig. 11. The precipitation was reduced for most of the simulated area, suggesting that a drier climate is associated with the cooling.

Although some proxy records suggest the weakening of the AMOC in LGM (McManus et al., 2004), the LGM experiment simulated ca. 32 Sv, which is about 15 Sv stronger peak value of AMOC than that of PI. However, about half of the modelling studies have had a problem representing the weaker AMOC (Weber et al., 2007). Oka et al. (2012) suggested that slight differences in surface cooling or wind stress forcing could lead to the different response of the AMOC. More sensitivity tests using various forcing conditions are required to understand its behavior.

The response of the carbon cycle and predicted vegetation in SEIB-DGVM are investigated in another work.

3.4 LM

The LM experiment is a transient experiment using time-dependent boundary data, unlike the other experiments described in this paper. In addition, the experiment was carried out as a “CO₂ prognostic” run, so that the atmospheric CO₂ concentration was calculated by the model itself. It requires the additional procedure in preparing initial data in terms of atmospheric CO₂ concentration. In this section, the preparation/selection of these data is described.

3.4.1 Forcing data and boundary conditions

The forcing data and boundary conditions for the LM experiment include: (a) orbital forcing, (b) solar forcing, (c) volcanic forcing, (d) non-CO₂ GHGs, and (e) anthropogenic land use.

Orbital forcing is based on a provided time series of parameters (Berger, 1978). As for solar forcing, multiple reconstructions of annual total solar irradiance (TSI) are available, and five reconstructions with different coverage periods are recommended in
the PMIP3 protocol (Schmidt et al., 2012). Amongst these, we used Wang et al. (2005) (hereafter, WLS) for 1610–2000 AD and Delaygue and Bard (2010) (hereafter, DB) for 850–1609 AD (Fig. 12b). WLS is based on a spectral reconstruction and on a flux transport model of the open and closed flux using the observed sunspot record as the main input, while DB is based on an Antarctica stack of $^{10}$Be records scaled linearly to the modern-to-Maunder Minimum TSI in the WLS reconstructions. The radiation scheme in MIROC-ESM considers 29 spectral bands, in which spectrally-resolved data provided by WLS can be used directly. The TSI data provided by DB is distributed consistently so that the integrated TSI over the spectra agrees with the reconstruction. No enhanced variations in the UV region proposed by Lean et al. (2000) and used by Shindell et al. (2001) were applied. The parameterization for solar-derived ozone variations suggested by the PMIP3 protocol was not used either.

Two reconstructions of the volcanic forcing: Gao et al. (2008) and Crowley et al. (2008, 2012) are provided by the PMIP3 protocol. Both datasets are based on polar ice cores, but differ in their selection of ice cores, and the Gao data show stronger forcing in general. In our experiment, Crowley data were used (Fig. 12a). The volcanic forcing is calculated using time series of aerosol optical depth (AOD) at 0.55 µm and the effective radius calculated in MIROC-ESM. The data are provided at four latitude bands of equal area. AOD estimates are based on a correlation between sulphate in the Antarctic ice cores and satellite AOD data, which was calibrated with the 1991 eruptions of Mt. Pinatubo and Cerro Hudson (Sato et al., 1993). Aerosol size estimation is based on Pinto et al. (1989).

CO$_2$, CH$_4$ and N$_2$O are considered as GHG forcing in the PMIP3 protocol, and a data table is provided. For the atmospheric CO$_2$ concentration, however, we decided to perform LM as a prognostic CO$_2$ experiment in which CO$_2$ is calculated online (i.e. CO$_2$ concentration predicted by the carbon cycle component is used for radiation transfer calculation). Thus the model predicts CO$_2$ using emission data. Anthropogenic emissions are considered only after 1850 AD, while emissions are assumed to be zero for the 850–1850 AD period. CH$_4$ and N$_2$O are given as per the PMIP3 data table.
As a difference from the PMIP3 protocol, the anthropogenic land use was assumed to be unchanged. The 1850 condition was applied throughout the experiment including the spin-up period.

### 3.4.2 Preparation of initial data

As is the case with the 6 ka experiment, pre-industrial physical fields (i.e. from PI experiment) of the atmosphere and ocean were used for the initial data of LM spin-up. Since the forcing conditions: orbital parameters, solar irradiance and GHG concentrations, are quite similar between the pre-industrial period and 800s, PI can be regarded as a spin-up for the 850 AD condition.

Figure 3d shows the procedure of the spin-up. The 250th year of the PI experiment was used as the initial state, then 56 yr of integration was completed under the year-850 condition, with fixed GHG (including CO$_2$) concentration. Then CO$_2$ concentration was set to be free, so that the model-calculated value is used in the radiation process of the atmosphere component. At the same time, CO$_2$ concentration was reset from 298.71 ppm to the level of 279.3 ppm (i.e. value at year 850; Joos, 2007), then the transient simulation was started. This resetting procedure was performed to maintain consistency with the reconstructed value, and also to avoid unnecessary CO$_2$ feedback. It breaks the conservation of the total carbon in a precise sense, but the effect is small enough to neglect in evaluating the total budget.

### 3.4.3 Results and time series

Figure 12 shows the time series of (a) volcanic and (b) solar forcing data, (c) the anomaly of simulated annual mean surface air temperature in Northern Hemisphere, and (d) the simulated atmospheric CO$_2$. Panel (e) shows the same temperature time series as (c), but with the linear trend of warming drift observed in the PI experiment removed (0.094°/100 yr). The blue shade in (e) represents the range of temperature reconstruction between the 20th and 80th percentile by Frank et al. (2010). The air
temperature represents anomalies (°C) from the 1961 to 1990 mean, and is averaged over the Northern Hemisphere for the comparison with proxy records.

The warming drift is clear, especially during the period until the 19th century, while anthropogenic warming is more pronounced during the 20th century. The drift is also visible in the PI experiment, which is caused by a decreasing trend in the Southern Ocean sea ice. It is likely due to an insufficiency of the initialization. Removing the trend is thus necessary to discuss the long-term climate variations; the time series of temperature in panel (e) is obtained by subtracting the linear trend from (c), the slope of which was calculated from linear regression of the PI experiment results for the period between year 250 to year 630 (i.e. after the “branching”).

Considering that the proxy data have low-pass characteristics, the time series of the simulated temperature roughly follows the reconstructions for most of the period. However, contrary to the good agreement after the year 1700, some discrepancies can be seen in the earlier periods. For example, the simulated values do not show a clear indication of the LIA, while the reconstruction shows relatively low temperature trends in the 16th to 17th centuries. The mismatch is also large during the 12th century, where the simulated temperature tends to be higher than the reconstruction. The model also seems to overestimate the responses to the huge volcanic events (e.g. year 1229, 1258, 1809, 1816), but this could be due to the limitation of detection capability of the reconstruction method (Mann et al., 2012).

Comparing the simulated temperature anomaly with the forcing data (panel a and b in Fig. 14), it is clear that volcanic aerosols have strong control on the surface air temperature, while the solar irradiance, though visible, has minor effects. Under the current set-up of the model, LM climate variations should depend mainly on those two types of forcing. The faint LIA is considered to be due to the relatively “flat” solar forcing.

The simulated atmospheric CO$_2$ concentration (panel d in Fig. 12) is stable in the model over the pre-industrial period, while it shows a sharp increase after 1850, the period during which anthropogenic emissions are considered. It shows 399–400 ppm near the surface in the year 2001, while the observed value in Mauna Loa was 370.13 ppm.
(Tans and Keeling, 2011) in the same year. Considering that the simulated value at year 1850 was 289.5 ppm, which is ca. 5 ppm higher than the reconstruction, the model still shows a tendency of slight overestimation. This is partly attributable to the warming drift of the global temperature.

4 Conclusions and outlook

The three paleoclimate experiments, 6 ka, LGM, and LM, proposed in CMIP5/PMIP3 were performed using MIROC-ESM. Overall, MIROC-ESM gave qualitatively reasonable results for all three cases. Special care was required when conducting experiments with conditions significantly different from modern simulations, especially using a highly complex model like ESM.

The 6 ka experiment was relatively simple, requiring spin-up only for the land carbon cycle. The results showed reasonable seasonal changes in the temperature distribution in response to the insolation change. The precipitation and the circulation changes suggest the enhancement of the summer African and Asian monsoons. These changes are consistent in sign with the paleo-environmental proxy records, but the enhancement over the Sahara Desert was not large enough, as also seen in previous modelling studies.

The LGM experiment required several steps before starting the simulation. Large change in topography requires a careful adjustment to avoid the numerical instabilities by gravity waves, and very long integration for spin-up is required for the ocean physical state and carbon-cycle variables. In this experiment, ocean values are taken from a LGM experiment using the previous version of MIROC AOGCM, and the spin-up for the carbon cycle was performed separately. In the results, changes in temperature and precipitation are mostly consistent with the SSTs and the pollen proxy records (MARGO Project Members, 2009; Bartlein et al., 2010). AMOC was stronger, which is contrary to the results from some reconstructions. This discrepancy is an issue for many climate models, which should be discussed in another study.
In the LM experiment, the same settings as the PI experiment were used for the fixed boundary conditions. As it is a transient experiment, preparations for time-dependent forcing data are required. Since the experiment was performed as “CO₂ prognostic”, the time series of carbon emission data and initialization of the atmospheric CO₂ are also required. After the removal of the linear-regressed temperature trend, the result shows general agreement with the reconstructed hemispheric temperature variations. Since this experiment depends on solar- and volcanic forcing data, updating and cross checking of the data, both for the forcing and proxy records, are continuously required.

It is again recognized that careful and sufficiently long spin-up is mandatory to achieve an equilibrium state, especially for a long transient simulation as LM. In ESMs, more processes such in land and ocean biogeochemistry are considered, and time constants are different among them. In addition to the physical model part, initialization of carbon cycle part is also necessary.

In this paper, we have documented the implementation of the CMIP5/PMIP3 experimental protocol for paleoclimate simulations in MIROC-ESM, with some overview of the results. We have successfully performed the experiments as described here and the results are available via the CMIP5 database (Taylor et al., 2012). Further analysis of each setting, as well as multi-model ensemble analysis will follow this paper. We hope these results contribute to the CMIP5 modelling activity.

Acknowledgements. This study was supported by the Innovative Program of Climate Change Projection for the 21st century, MEXT, Japan. The numerical simulations in this study were performed using the Earth Simulator, and figures were drawn using GTOOL, GrADS, and R. NODC_WOA98 (World Ocean Atlas 1998) data were provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at http://www.esrl.noaa.gov/psd/. This study is supported in part by the Funding Program for Next Generation World-Leading Researchers by the Cabinet Office, Government of Japan (GR079), by the Green Network of Excellence Arctic Climate Change Research Project, MEXT, Japan, and by the Global Environment Research Fund (S-10) of the Japanese Ministry of the Environment.
References


### Table 1. The boundary conditions for the each experiment.

<table>
<thead>
<tr>
<th>abbr. name</th>
<th>PI</th>
<th>6 ka</th>
<th>LGM</th>
<th>LM</th>
</tr>
</thead>
<tbody>
<tr>
<td>period</td>
<td>pre-industrial 1850 AD</td>
<td>mid-Holocene 6000 yr before present</td>
<td>Last Glacial Maximum 21 000 yr before present</td>
<td>Last Millennium 850–1850 AD</td>
</tr>
<tr>
<td>Experiment no. in CMIP5</td>
<td>Exp. 3.1</td>
<td>Exp. 3.4</td>
<td>Exp. 3.5</td>
<td>Exp. 3.6</td>
</tr>
<tr>
<td>Type</td>
<td>time slice</td>
<td>time slice</td>
<td>time slice</td>
<td>Transient</td>
</tr>
<tr>
<td>land boundary condition</td>
<td>pre-industrial</td>
<td>pre-industrial</td>
<td>LGM topography</td>
<td>pre-industrial</td>
</tr>
<tr>
<td>GHG : CO₂ (ppm)</td>
<td>285.0</td>
<td>280.0</td>
<td>185.0</td>
<td>Prognostic</td>
</tr>
<tr>
<td>GHG : N₂O (ppm)</td>
<td>0.31</td>
<td>0.27</td>
<td>0.20</td>
<td>given, time variable</td>
</tr>
<tr>
<td>GHG : CH₄ (ppm)</td>
<td>1.7414</td>
<td>0.65</td>
<td>0.35</td>
<td>given, time variable</td>
</tr>
<tr>
<td>orbit: eccentricity</td>
<td>0.01672</td>
<td>0.018682</td>
<td>0.018994</td>
<td>given, time variable</td>
</tr>
<tr>
<td>orbit: obliquity (°)</td>
<td>23.45</td>
<td>24.105</td>
<td>22.949</td>
<td>given, time variable</td>
</tr>
<tr>
<td>orbit: angular precession (°)</td>
<td>102.04</td>
<td>0.87</td>
<td>114.42</td>
<td>given, time variable</td>
</tr>
</tbody>
</table>
Fig. 1. The structure of MIROC-ESM.
Fig. 2. The vegetation types for the land model MATSIRO for (a) PI and (b) LGM. #0 – sea surface, #1 – continental ice, #2 – broadleaf evergreen forest, #3 – broadleaf deciduous forest and woodland, #4 – mixed coniferous and broadleaf deciduous forest and woodland, #5 – coniferous forest and woodland, #6 – high latitude deciduous forest and woodland, #7 – wooded C4 grassland, #8 – shrubs and bare ground, #9 – tundra, #10 – C3 grassland, #11 – cultivation.
Fig. 3. The procedure of initial data preparation: (a) Pre-industrial control: PI, (b) Mid Holocene: 6 ka, (c) Last Glacial Maximum (LGM): 21 ka, (d) Last Millennium: LM.
Fig. 4. The global annual mean sea surface temperature representation (°C) of PI in comparison with World Ocean Atlas SST (WOA, 2009).
Fig. 5. The global precipitation distribution (mm day$^{-1}$) of PI for JJAS and DJF (b, d) in comparison with re-analysis data (Xie and Arkin, 1996) (a, c).
Fig. 6. The short wave radiation change (W m$^{-2}$) at the top of the atmosphere for 6ka-PI, LGM-PI.
Fig. 7. The surface temperature change 6 ka-PI (°C) for (a) JJAS and (b) DJF.
Fig. 8. The precipitation change 6 ka-PI (mm day\(^{-1}\)) for (a) JJAS and (b) DJF.
Fig. 9. The orographic change for LGM-PI (m).
Fig. 10. The annual mean temperature change LGM-PI (°C).
Fig. 11. The annual mean precipitation change LGM-PI (mm day$^{-1}$).
Fig. 12. The (a) solar and (b) volcanic forcing, (c) annual mean surface air temperature in the Northern Hemisphere, (d) model-predicted atmospheric CO$_2$ concentration, (e) detrended time series of mean annual surface air temperature anomaly (anomaly from the 1970–2000 mean). The blue shading represents the range of temperature reconstruction between the 20th and 80th percentile by Frank et al. (2010).