Improved Climate Simulation by MIROC5: Mean States, Variability, and Climate Sensitivity

Masahiro Watanabe,* Tatsuo Suzuki,+ Ryouta O'ishi,* Yoshiki Komuro,+ Shingo Watanabe,+ Seita Emori,*,+,# Toshihiko Takemura,@ Minoru Chikira,+ Tomoo Ogura,# Miho Sekiguchi,& Kumiko Takata,+ Dai Yamazaki,** Tokuta Yokohata,+ Toru Nozawa,# Hiroyasu Hasumi,* Hiroaki Tatebe,+ and Masahide Kimoto*

* Center for Climate System Research, ++ The University of Tokyo, Kashiwa, Japan

⁺ Japan Agency for Marine-Earth Science and Technology, Yokohama, Japan

[#] National Institute for Environmental Studies, Tsukuba, Japan

[@] Kyushu University, Fukuoka, Japan

[&] Tokyo University of Marine Science and Technology, Tokyo, Japan ** Institute of Industrial Science, The University of Tokyo, Tokyo, Japan

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ABSTRACT

A new version of the atmosphere–ocean general circulation model cooperatively produced by the Japanese research community, known as the Model for Interdisciplinary Research on Climate (MIROC), has recently been developed. A century-long control experiment was performed using the new version (MIROC5) with the standard resolution of the T85 atmosphere and 1° ocean models. The climatological mean state and variability are then compared with observations and those in a previous version (MIROC3.2) with two different resolutions (medres, hires), coarser and finer than the resolution of MIROC5.

A few aspects of the mean fields in MIROC5 are similar to or slightly worse than MIROC3.2, but otherwise the climatological features are considerably better. In particular, improvements are found in precipitation, zonal mean atmospheric fields, equatorial ocean subsurface fields, and the simulation of El Niño–Southern Oscillation. The difference between MIROC5 and the previous model is larger than that between the two MIROC3.2 versions, indicating a greater effect of updating parameterization schemes on the model climate than increasing the model resolution. The mean cloud property obtained from the sophisticated prognostic schemes in MIROC5 shows good agreement with satellite measurements. MIROC5 reveals an equilibrium climate sensitivity of 2.6 K, which is lower than that in MIROC3.2 by 1 K. This is probably due to the negative feedback of low clouds to the increasing concentration of CO_2 , which is opposite to that in MIROC3.2.

1. Introduction

A comprehensive climate model that couples the atmosphere and ocean general circulation models together with the land and sea ice modules is called the coupled general circulation model (CGCM), or the global climate model. The development of CGCMs has a history of several decades, and they provide a unique way of physically based modeling the global climate and its variability (cf. Meehl et al. 2007; Reichler and Kim 2008). As human-induced climate change has attracted wider societal attention, CGCMs have become more important tools than ever (Solomon et al. 2007).

Despite the usefulness of CGCMs, it is well known that such models still contain errors in various fields, such as precipitation and sea surface temperature (SST), and reveal considerable disagreement, for example, in the cloud feedback when performing climate change simulations (e.g., Bony and Dufresne 2005). Some aspects of the model errors can certainly be reduced by increasing the resolution for either the atmosphere or ocean component (e.g., Shaffrey et al. 2009). High-resolution atmosphere models produce an improved precipitation distribution arising from higher-resolution orography

⁺⁺ Atmosphere and Ocean Research Institute, Tokyo, Japan, as of April 2010.

Corresponding author address: M. Watanabe, Atmosphere and Ocean Research Institute, The University of Tokyo, 5-1-5 Kashiwanoha, Kashiwa, Chiba 277-8568, Japan. E-mail: hiro@aori.u-tokyo.ac.jp

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(Pope and Stratton 2002; Jung et al. 2006; Gent et al. 2009) and more realistic tropical cyclone frequency (Oouchi et al. 2006). Similarly, the SST and ocean surface fields are better simulated by partly resolving oceanic eddies (Semtner and Chervin 1992; Sakamoto et al. 2004). However, other errors attributed to complicated feedback processes are not necessarily reduced by increasing the resolution without changing the parameterization schemes. In this regard, not only the use of higher resolution models but also continuously developing the model itself is clearly a key issue for better reproducing the past climate variability, projecting future climate change, and understanding their mechanisms.

In this paper, we present the basic results obtained from a new version of our coupled model, the Model for Interdisciplinary Research on Climate (MIROC), developed jointly at the Center for Climate System Research (CCSR), University of Tokyo; National Institute for Environmental Studies (NIES); and Japan Agency for Marine-Earth Science and Technology. This new version, called MIROC5, will be used for the forthcoming Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (AR5). The previous version, MIROC3.2 (Hasumi and Emori 2004), included a standard physics package and was well tuned at the time of the Fourth Assessment Report (AR4). The model showed mean states that were relatively close to the average of CGCMs participating in the Coupled Model Intercomparison Project phase 3 (CMIP3). There were, however, deficiencies in the natural variability and climate sensitivity (e.g., see Santer et al. 2009). One important shortcoming was the extremely weak El Niño-Southern Oscillation (ENSO), even though the equatorial mean states were good (Guilyardi et al. 2009b). In addition, the cloud representation in MIROC3.2 was crude; the prognostic variable was the total water mixing ratio, from which the cloud fraction was diagnosed using a conventional large-scale condensation (LSC) scheme, and the water/ice separation depended simply on temperature. It has been reported that the climate sensitivity heavily depends on the response of mixed-phase clouds to radiative forcing (Tsushima et al. 2006), so the cloud representation had to be totally reconsidered.

As will be described in the next section, most parts of the model, except for the atmospheric dynamical core, were updated or even replaced with new parameterization schemes in MIROC5. A century-long, preindustrial simulation was performed with the standard resolution of T85L40 atmosphere and approximately 1° ocean component models. The atmospheric resolution is between the two MIROC3.2 products included in AR4—that is, "MIROC3.2(medres)" and "MIROC3.2(hires)" which adopted the T42L20 and T106L56 atmosphere components, respectively. The horizontal resolution for the ocean components in MIROC5 is the same as that used in MIROC3.2(medres), and coarser than that of MIROC3.2(hires), which used an eddy-resolving ocean model with a $\frac{1}{4^{\circ}} \times \frac{1}{6^{\circ}}$ resolution. A comparison of the climatology between MIROC5 and the two datasets from the MIROC3.2 runs makes it possible to evaluate the effect of the new model configuration relative to the effect of increasing the model resolution. Overall, the new standard resolution is close to that of MIROC3.2(medres), but it is demonstrated that the difference between the two model versions is greater than that between MIROC3.2(medres) and MIROC3.2(hires). It may have been better to run MIROC5 with the same resolutions as MIROC3.2 to thoroughly investigate the resolution versus parameterization issue, but the standard resolution of MIROC5 was chosen mainly because of the limitation of computational resources.

This paper is organized as follows. In the next section, the framework of MIROC5 is described without the details of the individual parameterization schemes, which have been presented elsewhere. In section 3, the timemean states and interannual variability are compared with observations and those obtained from MIROC3.2. The results demonstrate that deficiencies observed in MIROC3.2 are greatly reduced in several respects. The investigation of model behavior, in response to an abrupt quadrupling $CO_2 (CO_2 \times 4)$, is described in section 4. The climate sensitivity of MIROC5 is then briefly examined and is found to be lower than that of the previous model version. Section 5 presents the concluding discussion.

2. Model description

As described by Hasumi and Emori (2004), MIROC3.2 couples the following component models. The atmosphere model is the CCSR–NIES–Frontier Research Center for Global Change (FRCGC) AGCM (Numaguti et al. 1997), which is based on a global spectral dynamical core and includes a standard physics package. The ocean model is the CCSR Ocean Component Model (COCO; Hasumi 2006), which includes a sea ice model. A land model that includes a river module is also coupled. MIROC5 was developed based on MIROC3.2, but many of the schemes have been replaced as follows.

a. Atmosphere component

1) DYNAMICAL CORE

In recent years, atmospheric dynamical cores tend to be represented either by finite volume or finite difference schemes that are favorable for high-resolution computing (e.g., Tomita and Satoh 2004; Lin 2004). In this

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regard, the spectral dynamical core that is used in MIROC5, as in the previous version, may be outdated and will be replaced in the next stage.

We used a vertical σ coordinate in MIROC3.2, in which the model top was about 8 hPa in the medium resolution. In MIROC5, a hybrid σ -p coordinate (cf. Arakawa and Konor 1996), which is shared by our Earth system model (ESM), is adopted with the model top at around 0.003 hPa (Watanabe et al. 2008). The standard vertical resolution of the MIROC5 atmosphere is 40 levels up to 3 hPa. This resolution is between MIROC3.2(medres) and MIROC3.2(hires), which had 20 and 56 σ levels, respectively. The Asselin time filter has also been modified following the method of Williams (2009).

2) RADIATION

The radiative transfer in MIROC5 is calculated by an updated version of the *k*-distribution scheme used in MIROC3.2 (Sekiguchi and Nakajima 2008). This scheme results in improvements in the line absorption and continuum absorption, with an increase in the number of absorption bands from 18 to 29. These changes all contribute to a more accurate calculation of radiative heating and toward reducing a cold bias near the tropopause, which was found in MIROC3.2 (cf. section 3b).

3) CUMULUS CONVECTION

The cumulus scheme employed in MIROC5 is one that was recently developed by Chikira and Sugiyama (2010). It is an entraining plume model, in which the lateral entrainment rate varies vertically depending on the surrounding environment. Its formulation is similar to the scheme by Gregory (2001), who assumed that a certain fraction of buoyancy-generated energy is consumed by the entrainment process. Multiple cloud types having different cloud tops are considered, as was done by Arakawa and Schubert (1974), except for the representation according to the updraft velocity at the cloud base. The cloud base mass flux is determined with a prognostic convective kinetic energy closure (Xu 1993; Pan and Randall 1998), which has been employed in MIROC3.2.

The sensitivity of the scheme to temperature and humidity profiles and the scheme's influence on model climatology have been documented by Chikira and Sugiyama (2010) and Chikira (2010). The scheme tends to produce larger entrainment rates near the cloud base as compared with Pan and Randall's scheme. By incorporating a state-dependent entrainment rate, deep convection tends to be effectively suppressed when the environment is dry in the free troposphere. This can eliminate an artificial triggering function for the deep convection, which was used in the previous version (Emori et al. 2005).

4) CLOUD AND CLOUD MICROPHYSICS

In MIROC3.2, the formation and dissipation of clouds were represented by a diagnostic LSC scheme proposed by Le Treut and Li (1991) and a simple microphysics scheme. Ogura et al. (2008) found that the climate sensitivity in MIROC3.2 is primarily controlled by the crude representation of clouds and hence may not be realistic.

To better represent cloud and cloud-radiative feedback, two major changes were made in MIROC5: the development of a prognostic LSC scheme (Watanabe et al. 2009) and the implementation of a bulk microphysical scheme (Wilson and Ballard 1999). The new LSC scheme solves prognostic equations for the subgridscale variance and skewness of a conservative quantity associated with temperature and total water, and hence represents various cloud regimes having different optical properties. The cloud microphysical scheme explicitly deals with the warm and cold rain processes: nucleation, deposition and sublimation, riming, ice melting, and raindrop capturing by falling ice, among others.

Because the original scheme by Watanabe et al. (2009) did not consider the cloud ice, it was modified when coupled with the Wilson and Ballard (1999) scheme. Since the LSC scheme employs a "fast condensation" assumption that is not relevant to ice, the ice mixing ratio is conserved in the LSC process by assuming that the cloud ice exists preferentially in a subgrid area having the largest amount of total condensate. Specifically, mixed-phase cloud is generated when the condensate amount is more than the ice content, whereas the cloud fraction and vapor amount are adjusted in the case of a pure ice cloud when the condensate amount is less than the ice content. By combining the two schemes, the fraction of cloud liquid and ice is no longer just a simple function of temperature, and the processes controlling climate sensitivity will also be qualitatively different from those in the previous version (cf. section 4).

5) TURBULENCE

The vertical diffusion scheme is based on Nakanishi (2001) and Nakanishi and Niino (2004). It is a Mellor–Yamada (Mellor and Yamada 1974, 1982)-type scheme with a closure level of 2.5 but improved in several respects. The master length scale, L, is newly devised for large-scale models and determined by the harmonic mean of three length scales— L_S , L_T , and L_B —which characterize the surface layer, convective boundary layer, and stably stratified layer, respectively. Shear and buoyancy effects on the pressure covariance terms have been

added, and the closure constants were reevaluated by large eddy simulation outputs (cf. Nakanishi 2001; Nakanishi and Niino 2004). The effect of the vapor– liquid transition on buoyancy is considered by using the new LSC scheme. The improved turbulence scheme reduces some common deficiencies of the Mellor–Yamada scheme (cf. section 3b).

The formulation of L in the original scheme was not necessarily adequate for the free atmosphere when its stability was reduced by radiative cooling because of clouds. Therefore, the master length scale of the free atmosphere is given by the harmonic mean of L_S , L_A , and L_{max} , where L_A , expressed in terms of the turbulence kinetic energy (TKE) and the Brunt–Väisälä frequency, represents a length scale on which an air parcel with a given TKE can be vertically displaced in a stably stratified layer. A constant of $L_{\text{max}} = 500$ m gives the upper limit.

6) AEROSOLS

An aerosol module in MIROC, called Spectral Radiation-Transport Model for Aerosol Species (SPRINTARS), predicts the mass mixing ratios of the main tropospheric aerosols: carbonaceous (black carbon and organic matter), sulfate, soil dust, and sea salt, as well as the precursor gases of sulfate, that is, sulfur dioxide and dimethylsulfide (DMS). The aerosol transport processes include emission, advection, diffusion, sulfur chemistry, wet deposition, dry deposition, and gravitational settling. The emissions of soil dust, sea salt, and DMS are calculated using the internal parameters of the model, and external emission inventories are used for the other aerosol sources.

SPRINTARS is coupled with the radiation and cloud microphysics schemes to calculate the direct and indirect effects of the aerosols. In the calculation of the direct effect, the refractive indices depending on wavelengths, size distributions, and hygroscopic growth are considered for each kind of aerosol. The aerosol semidirect effect is also included as a consequence of the combination of the aerosol module and other schemes. A prognostic scheme for determining the cloud droplet and ice crystal number concentrations is introduced for calculating the aerosol indirect effect and cloud nucleation process. Changes in their radii and precipitation rates due to the indirect effect affect the radiation and cloud processes. Readers may refer to Takemura et al. (2005, 2009, and references therein) for further details on the present version of SPRINTARS.

b. Ocean component

1) GENERAL FEATURES

The ocean general circulation model used for MIROC5 is COCO version 4.5. The primary update from the

previous version includes a change in the coordinate system. The governing equations in COCO4.5 are formulated on a generalized curvilinear horizontal coordinate. The generalization is made by transforming the longitude-latitude coordinate system and its meridians and latitude circles using the polar stereographic projection and conformal mapping, following Bestsen et al. (1999). The North Pole (South Pole) of the model coordinate system is moved to 80°N, 40°W over Greenland (80°S, 40°W over Antarctica). The zonal resolution is a fixed 1.4° , whereas the meridional resolution is 0.5° at latitudes equatorward of 8°, 1.4° at higher latitudes (poleward of 65°), with a smooth transition in between $(256 \times 224 \text{ grid points for zonal and meridional di$ rections). This horizontal resolution is approximately the same as the ocean model in MIROC3.2(medres); however, the number of vertical levels has been increased from 43 to 49, excluding the bottom boundary layer. The vertical grid spacing varies with depth: 2.5 m at the surface, 20 m at the depth of 100 m, 100 m at the depth of 1000 m, and 250 m below the 2000 m depth. The other features of the discretization follow the previous version (cf. Hasumi 2006).

In the model bathymetry, the Bering Strait is represented by a two-grid point gap, so that there is only one velocity grid point at the strait. The water pathway through the Canadian Archipelago is also represented by artificially excavating a channel. The Mediterranean Sea is represented as an isolated basin. At the strait of Gibraltar, the sea surface elevation and tracers are artificially exchanged by two-way linear damping with the time constants of 100 and 300 days, respectively, at depths above 1260 m (30th level).

The numerical scheme for the tracer advection is replaced by a second-order moment method (Prather 1986). The vertical mixing of momentum and tracers uses a harmonic formulation. To eliminate a checkerboard noise in the sea surface elevation field, a weak horizontal diffusion is applied with the coefficient of $50 \text{ m}^2 \text{ s}^{-1}$, which does not violate the conservation of tracer quantities.

2) PHYSICAL PARAMETERIZATION

Some of the physical parameterization schemes employed in COCO4.5 have been updated. The treatment of the vertical convection, bottom boundary layer, background diffusivity, and penetration of shortwave radiation remains unchanged [see Hasumi (2006) for the references]. To reproduce the formation of North Pacific Intermediate Water (Tally 1993; Yasuda 1997), the background diffusivity is raised to 2.0×10^{-2} m² s⁻¹ below the 100-m depth along the Kuril Islands (from the tip of the Kamchatka Peninsula to Hokkaido, Japan), as suggested by a direct calculation of tidal effects (Nakamura et al. 2000).

The turbulent mixing process is represented by the parameterization of Noh and Kim (1999), as in the previous version. However, the definition of the turbulent Prandtl number has been modified following Noh et al. (2005). The lateral mixing process as represented by harmonic viscosity has also been revised. Its coefficient is $3.0 \times 10^4 \Delta x / \Delta x_{\rm max} {\rm m}^2 {\rm s}^{-1}$, where Δx and $\Delta x_{\rm max}$ are the local and maximum longitudinal grid intervals, respectively, and it is reduced to $2.0 \times 10^4 \Delta x / \Delta x_{\text{max}} \text{ m}^2 \text{ s}^{-1}$ at the equator, with a Gaussian distribution between 15°S and 15°N to reproduce a realistic equatorial undercurrent (EUC), as suggested by Large et al. (2001). The harmonic horizontal diffusion, isopycnal diffusion, and horizontal diffusion of the isopycnal layer thickness (Gent et al. 1995) are also applied with coefficients of 1.0×10^2 , 1.0×10^3 , and 3.0×10^2 m² s⁻¹, respectively.

The bottom boundary layer is applied at high latitudes to the north of 49°N and to the south of 56°S following Nakano and Suginohara (2002). The Rayleigh drag coefficient is taken to be the same as the Coriolis parameter above 2000 m for the Northern Hemisphere (NH) and above 4000 m for the Southern Hemisphere (SH), with a value of zero below, as suggested by Nakano and Suginohara (2002).

c. Sea ice component

In MIROC3.2, the sea ice is treated as a twodimensional continuum in terms of dynamics, with which the concentration, thickness, and horizontal velocity components are predicted at each grid. In MIROC5, the sea ice concentration, ice thickness, snow thickness, and energy of ice melting are predicted for multiple categories in a grid cell. The sea ice model calculates the evolution of the subgrid-scale sea ice thickness distribution following the governing equation by Thorndike et al. (1975). The thickness distribution and evaluation of the mechanical redistribution term are discretized according to Bitz et al. (2001). The sea ice at each horizontal grid is divided into five categories, plus open water. The lower bounds of the ice thickness for these categories are 0.3, 0.6, 1, 2.5, and 5 m.

1) THERMODYNAMICS

The heat capacity of sea ice is considered in the new sea ice module. The growth, melting, and temperature change of sea ice are computed based on the energyconserving thermodynamic scheme of Bitz and Lipscomb (1999). We use this scheme with only one layer for sea ice. The temperature of snow is not considered since snow is assumed to have no heat capacity. The salinity of sea ice is fixed at 5 psu. The penetration of solar radiation into the snow or ice is not taken into account. The albedo values for a bare ice surface are fixed at 0.8 and 0.65 for the visible and near-infrared bands, respectively. The surface albedo over a snow-covered area depends on temperature by considering the existence of partial snow cover at a relatively high temperature. It is 0.9 (0.8) for a temperature lower than -5° C, 0.8 (0.65) at 0°C, and changes linearly in between for the visible (near infrared) radiation.

In open water, including that of a partially ice-covered grid, cooling forms new ice if the seawater temperature is at the freezing point. The newly formed ice is added to the thinnest category with the same thickness when it already exists. The new ice thickness is otherwise equal to the lower limit (i.e., 0.3 m). On the other hand, the warming of open water melts the sea ice. This occurs from the bottom of the sea ice.

Once the thermodynamic growth rates for each category are determined, the linear remapping scheme of Lipscomb (2001) is applied. It evaluates the thermodynamic transfer of the ice between categories by assuming a linear ice thickness distribution within the categories.

2) Dynamics

In the multicategory sea ice model, thermodynamic variables, such as the sea ice concentration for each category, are advected by the prognostic ice velocity, which is common for all the categories in a grid. A simple first-order upstream scheme is employed for computing the advection term. The dynamical schemes are otherwise the same as the previous version, except for changes in the parameter values: the strength of the ice per unit thickness and concentration is set at 2.0×10^2 N m⁻¹, and the ice–ocean drag coefficient has been increased to 0.02.

d. Land component

MIROC5 adopts an updated version of the land surface model called Minimal Advanced Treatments of Surface Interaction and Runoff (MATSIRO; Takata et al. 2003), which predicts the temperature and water in six soil layers down to a 14-m depth, one canopy layer, and three snow layers. In this version of MATSIRO, a tile treatment of the land surface has been introduced to represent the subgrid fraction of land surface types. One land surface grid is divided into three tiles in the control run: potential vegetation, cropland, and lake. All the prognostic and diagnostic variables are calculated in each tile, and the fluxes at the land surface are averaged using their fractional weights. Other modifications from the previous version are briefly described below.



FIG. 1. Time series of global mean SST (thick black), VAT700 (thin black), and VAT (gray) for control integration of MIROC5.

1) LAKE SUBMODEL

We calculate the surface heat and water fluxes over lakes as one of the tiles in a grid. The water temperature and mass are predicted for the surface layer (minimum thickness of 1 m) and four subsurface layers, based on the thermal diffusion and mass conversion, considering vertical overturning, evaporation, precipitation, and inflow from and outflow to rivers. The distribution and fraction of lakes are fixed in time.

2) RIVER ROUTING

The river routing model is basically the same as that in MIROC3.2, except for the updated river network for the T85 resolution (Yamazaki et al. 2009) and the method for calculating the river discharge (Oki et al. 2003).

3) SNOW AND ICE ALBEDO

The effect of snow aging on surface albedo is considered in MATSIRO following Yang et al. (1997). Among the three coefficients that affect the increment in the nondimensional age of snow, the one representing the effect of dirt increases according to its concentration in the surface snow layer. This mimics the observed relation between snow albedo and dirt concentration (Aoki et al. 2006). The dirt concentration is calculated from the deposition fluxes of dust and soot in SPRINTARS. Since the absorption coefficients of dust and soot are very different, the deposition fluxes are multiplied by their relative weights (0.012 for soil dust and 0.988 for black carbon). The sum is used as the radiatively effective deposition of dust and soot.

The previous version of MATSIRO assumed constant values for the surface albedo over an ice sheet. This has been changed in the present version following Bougamont et al. (2005), who proposed that the ice sheet albedo be expressed as a function of the water content above the ice. This scheme is applicable for both visible and nearinfrared radiation, with a fixed value of 0.05 being used for the infrared band.

e. Control experiment

1) BOUNDARY CONDITIONS

The historical changes in the total solar irradiance and in volcanic aerosol optical depth in the stratosphere are

TABLE 1. Global mean radiative and energy budgets from observational estimates, MIROC3.2, and MIROC5. Radiative budgets given in W m^{-2} are the values at TOA. References for the observational estimates are shown at the bottom. All the quantities are the annual averages. The values for net SW are downward positive, while those for net LW and net are upward positive.

Parameter	Observed	MIROC3.2(medres)	MIROC3.2(hires)	MIROC5
Incoming solar	341.5 ^a /341.8 ^b	341.6	341.5	341.5
Net SW (all sky)	235.8 ^a /240.5 ^b	235.7	240.0	237.6
Net LW (all sky)	236.3 ^a /239.6 ^b	234.8	239.4	236.5
Net (all sky)	$0.5^{\rm a}/0.85^{\rm b}$	-0.9	-0.6	-1.1
SW CRF	$-51.0^{\rm a}/-46.6^{\rm b}$	-53.8	-48.7	-53.8
LW CRF	26.5 ^a /29.5 ^b	27.6	27.0	26.3
Net CRF	$-24.5^{\rm a}/-17.1^{\rm b}$	-26.2	-21.7	-27.5
SAT (°C)	14.0 ^c /14.5 ^d	13.4	13.9	14.5
SST (°C)	18.2 ^e	17.4	17.8	17.9
Precipitation (mm d^{-1})	2.61^{f}	2.72	2.96	3.2
Evaporation $(mm d^{-1})$	2.89 ^g	2.72	2.96	3.2
Precipitation – evaporation (mm d^{-1})	-0.28	0.0	0.0	0.0
Cloud cover (%)	60 ^h	51.9	51.8	56.3

^a ISCCP FD dataset (Zhang et al. 2004).

^b Adjusted CERES (Loeb et al. 2009).

^c CRU 1961–90 mean (Jones et al. 2001).

^e 1945–2006 mean (Ishii et al. 2006).

^f CMAP 1979–2007 mean (Xie and Arkin 1997).

^g Ocean Model Intercomparison Project (OMIP; Röske 2001).

^h Kiehl and Trenberth (1997).

^d ERA-40 (Uppala et al. 2005).



FIG. 2. (a) Annual mean climatology of the net LW in MIROC5, and biases in the annual mean LW climatology in (b) MIROC3.2(medres) and (c) MIROC5. (d)–(f) As in (a)–(c), but for the net incoming SW. The unit is W m⁻².

given by Lean et al. (2005) and Sato et al. (1993), respectively. The former is set to 1365.7 W m⁻², and the latter, including its seasonal change in latitude and height, is fixed at the value for the year 1850 for the control simulation. The atmospheric concentrations of wellmixed greenhouse gases and the surface emissions of tropospheric aerosols are provided by the international task group of the Representative Concentration Pathways (RCP) Concentration Calculations and Data (available online at http://www.pik-potsdam.de/~mmalte/rcps/ index.htm). The concentrations of CO₂, CH₄, and N₂O are set to 284.725 ppm, 790.979 ppb, and 275.425 ppb, respectively. The three-dimensional atmospheric concentrations of ozone are precalculated by a chemical AGCM for the study of atmospheric environment and radiative forcing (Sudo et al. 2002), driven with emissions of its precursors in the 1850 condition given by the task group.

The boundary conditions for the land module consist of the fractions of three tiles (potential vegetation, cropland, and lake), the distribution of potential vegetation, and the leaf area index (LAI). The historical land use change is given by the land use harmonization data (Hurtt et al. 2009), in which the cropland fraction is fixed at the value in 1850 for the control simulation. The lake fraction is based on the U.S. Geological Survey global land cover characterization dataset. The potential vegetation data are prepared on the basis of the Stratospheric Aerosol and Gas Experiment (SAGE) dataset (Ramankutty and Foley 1999) and interpreted for the MATSIRO vegetation types. Since the SAGE data do not include C4 vegetation and permanent ice distribution, they are estimated from International Satellite Land Surface Climatology Project Initiative II (ISLSCPII) C4 vegetation data (Still et al. 2003) and Moderate Resolution Imaging Spectroradiometer (MODIS) snow cover data for 2001–07 (Hall et al. 2006), respectively. The LAI data are prepared from the MODIS LAI products (Shabanov et al. 2005) by separating the LAI into two tiles using the Hundred Year Database of the Global Environment version 3 (HYDE3) land use data (Goldewijk et al. 2007) to detect the cropland LAI in the MODIS data. The effect of the historical land use change on LAI is taken into account by changing fractions between the potential vegetation and cropland in the tile treatment.

2) TUNING AND SPINUP PROCEDURES

The atmospheric component is run for 10 yr from an initial state obtained from the MIROC3.2 control simulation. The ocean and sea ice components are spun up for 530 yr from the initial states provided by the Polar Science Center hydrographic climatology (PHC3.0; Steele et al. 2001). Acceleration (cf. Bryan 1984) is applied to the abyssal dynamical fields, except for the last 20 yr, to obtain quasi equilibrium during the spinup. After the component models are coupled, MIROC5 was further integrated for about 1000 years, during which parameters in cloud, convection, turbulence, aerosol, and sea ice schemes are perturbed to find the best set that results in a realistic climate. The number of parameters (fewer than 20) and the range of their perturbations were subjectively determined by experts as in the physics ensemble experiment (Yokohata et al. 2010).

The strategy of the model tuning follows that adopted for developing MIROC3.2 and is conventional. Specifically, the model's time-mean states are compared at every 5 yr in addition to monitoring the global mean time series of several important quantities [e.g., SST, surface air temperature (SAT), and the top-of-atmosphere (TOA) radiative fluxes]. While we have not employed a quantitative metric to rate the model's performance (Reichler and Kim 2008), the biases of many atmosphere and ocean variables are evaluated in terms of their global means and spatial patterns at each tuning cycle. This type of tuning is sometimes criticized because the same set of observational data is used for tuning and validation. However, the bias structure is complex, as will be shown later, and the parameter values are uniform both in time and space. Thus, we cannot control the model biases in an artificial manner when the global means of the primary variables (i.e., radiative budgets and temperature) are tuned to be close to the observations.



FIG. 3. Annual mean climatology of SST from (a) observations and (b) MIROC5. The unit is $^{\circ}$ C and the contour interval is 1° C (drawn only above 26 $^{\circ}$ C).

Figure 1 shows the global mean SST, along with the ocean temperature [vertically averaged temperature (VAT)] averaged for the upper 700 m and for the entire depth level (denoted as VAT700 and VAT, respectively) during a 300-yr period following the spin-up and parameter tuning phases, in which the parameters were held fixed. The years are arbitrarily labeled 2001–2300 for this period. SST and VAT700 time series indicate that the upper ocean is close to equilibrium at least after 2100; however, the deep ocean still warms very slowly, as confirmed by VAT. While another several hundred years may be required to obtain a fully equilibrated state, the climatological mean fields obtained from the 100-yr average during 2101 and 2200 are used in this paper.

Several observational datasets are used to validate the model climatology and variability. The primary data are derived from Ishii et al. (2006), who provide the SST, sea surface salinity (SSS), their subsurface fields, and sea ice concentration for 1945–2006. The climatology of the atmospheric and the precipitation fields are obtained from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005) and the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997), respectively. Several satellite products are also used for validating the radiative budgets,



FIG. 4. Biases in the annual mean climatology of (a) SST and (b) SSS in MIROC3.2(medres). (c),(d) As in (a),(b), but for MIROC5. The units are °C and psu, respectively.

precipitation, and cloud fields. These data are described when referred to in the subsequent sections.

3. Mean states and variability

a. Global mean budgets

Care should be taken in using the instrumental measurements for recent decades to validate the preindustrial climate simulated in the control run (CTL) because of the difference in radiative forcing. Ideally, the model climate should be compared with observations by performing a preliminary twentieth-century historical experiment (20C run) with an initial state available from recent data, and then the initial state for CTL (i.e., 1850) should be calculated using a method proposed, for example, by Stouffer et al. (2004). However, because such a method is computationally expensive, a preindustrial experiment was first carried out. This is justified by the fact that the differences in the mean state between CTL and the 20C run are smaller than the model biases identified in CTL. Indeed, we found that the root-mean-square (RMS) differences in precipitation, SAT, and SST between the CTL climatology and the 1961-90 average obtained from the 20C run using MIROC5 were 0.19 mm dy^{-1} , 0.4 K, and 0.32 K (accounting for 13.1%, 20.2%, and 19.8% of the RMS biases in the corresponding variables in CTL), respectively. These differences may not be negligible, but the horizontal patterns of the biases in the two runs are of great similarity (not shown), justifying the comparison of the CTL climatology with recent observations.

A brief comparison is made for the global mean radiative budgets at the TOA, temperature, and hydrological quantities between the observed estimates and two models (Table 1). The observed radiative budgets contain uncertainties, so two different estimates are listed. The solar insolation depends only on the solar constant, and thus the values from the observations and models are almost identical. On the other hand, the TOA net incoming shortwave (SW) and outgoing longwave (LW) fluxes vary among models, and even between two observational estimates. The SW flux in MIROC3.2(medres) (235.7 W m^{-2}) is similar to the estimate from the First International Satellite Cloud Climatology Project (ISCCP; 235.8 W m⁻²; Zhang et al. 2004). Both MIROC3.2(hires) (240.0 W $m^{-2})$ and MIROC5 (237.6 W $m^{-2})$ show a larger net SW flux closer to the recent Clouds and the Earth's Radiant Energy System (CERES) estimate (240.5 W m⁻²; Loeb et al. 2009). It should be noted that the net flux in MIROC5 denotes a negative value slightly larger than that in MIROC3.2. This mostly comes from the spinup of the abyssal ocean as it gradually approaches equilibrium (the linear trend is 0.07 \pm $0.04 \text{ W m}^{-2} \text{ century}^{-1}$ for the last 100 yr).



FIG. 5. Annual mean climatology of precipitation from (a) CMAP and (b) MIROC5. The unit is mm d^{-1} .

Global cloud radiative forcing (CRF) reveals a different property from the net fluxes. The cooling effect due to SW CRF in MIROC3.2(medres) and MIROC5 (both -53.8 W m^{-2}) is stronger than that in MIROC3.2(hires) (-48.7 W m^{-2}) by about 5 W m⁻², whereas the warming due to LW CRF is slightly weaker in MIROC5 (26.3 W m⁻² against 27.0–27.6 W m⁻² in MIROC3.2). The net CRFs in MIROC3.2(medres) (-26.2 W m^{-2}) and MIROC5 (-27.5 W m^{-2}) are similar to each other, whereas that in MIROC3.2(hires) is weaker (-21.7)W m⁻²). The net CRF difference between the two observational estimates (more than 7 W m^{-2}) shows that the difference between the three models may not be large. As will be shown in section 4, a similar magnitude for the net CRF does not imply a similar cloud feedback in a climate change simulation.

The global mean SAT and SST are also presented in Table 1. As compared with observations by Jones et al. (2001), the mean temperature is cold by 0.6–0.8 K in MIROC3.2(medres), while close but slightly cooler, by 0.1–0.4 K, in MIROC3.2(hires). In contrast, the MIROC5 climate is somewhat warm with the SAT 0.5 K warmer than the observed value in Jones et al.'s data, which is nearly identical to ERA-40.

The atmospheric water budgets (precipitation – evaporation) are strictly balanced in all the models. The global mean precipitation is excessive in MIROC5,



FIG. 6. Biases relative to CMAP data in annual mean precipitation in (a) MIROC3.2(medres), (b) MIROC3.2(hires), and (c) MIROC5. The unit is mm d^{-1} .

suggesting a hydrological cycle that is too active. This may partly be due to a warmer mean state, which will also result in more cloud production. The total cloud cover is the largest in MIROC5 (56.3%), which is closer to observation (60%).

b. Climatological fields

The climatological radiative budgets at TOA in MIROC5 are displayed in Figs. 2a,d. Unless otherwise stated, the model climatological states are compared with observations on an annual mean basis. The horizontal



FIG. 7. Climatological precipitation and 850 hPa winds in JJA from (a) TRMM PR and ERA-40, (b) MIROC3.2(medres), (c) MIROC3.2(hires), and (d) MIROC5. The units are mm mo⁻¹ and m s⁻¹.

average values of the net outgoing LW (Fig. 2a) and incoming SW (Fig. 2d) are shown in Table 1. Overall, the distribution is realistic; yet, their differences from the Earth Radiation Budget Experiment (ERBE; Barkstrom 1984) climatology indicate a systematic radiation bias (Figs. 2c,f). Both SW and LW are excessive in high latitudes between about 40° and 65°, whereas the solar insolation is insufficient over the tropical oceans. For comparison, similar bias maps for MIROC3.2(medres) are shown in Figs. 2b,e. On one hand, the LW bias is reduced in MIROC5, especially over the tropics. This is due to an improvement in high clouds in association with the deep convections. On the other hand, the SW bias in MIROC5 is generally larger than that in MIROC3.2(medres). In particular, the deficient insolation over the tropical oceans reflects that low clouds are overrepresented. The large bias near the equator in MIROC3.2(medres) (Fig. 2e) is less in MIROC5, but a negative bias over the subtropical western Pacific is worse. We tested a parameterization for cloud-top entrainment instability (CTEI), which works to remove the excessive low clouds. However, it is doubtful that CTEI is in reality so active as to dissipate the boundary layer cloud over a wide area of the tropics, so this parameterization was not included in the control experiment.

Figure 3 shows the mean SST from observations and MIROC5. The warm pool in the Indo-Pacific region extends well to the west and east of the Maritime Continent. As compared with the observations and MIROC3.2 (not shown), the meridional width of the warm pool in MIROC5 is somewhat narrower; however, the SST around the date line is still high enough to affect the ENSO simulation. The modeled zonal SST gradient in the equatorial Atlantic is opposite of the observations, which is a common error of the current generation CGCMs (Richter and Xie 2008).

The SST and SSS biases are compared between MIROC3.2(medres) and MIROC5 (Fig. 4). In MIROC3.2 (medres), a cooling bias is found in a wide area of the ocean, except for the eastern periphery of each ocean basin and the Antarctic Ocean, where a warming bias dominates (Fig. 4a). The bias in SSS is positive (negative) in the western equatorial Pacific (eastern Indian Ocean and the tropical Atlantic), and especially large in the Arctic Ocean (Fig. 4b). The SST bias in MIROC5 is reduced in low latitudes, but it is amplified in the North Pacific and in the Antarctic Ocean (Fig. 4c). The bias in the North Pacific is likely due to surface westerlies shifted southward in the atmosphere model. The warming bias in the southern polar region may be partly reduced when

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FIG. 8. Biases in annual and zonal mean climatology of the zonal wind in (a) MIROC3.2(medres) and (b) MIROC5. The unit is m s⁻¹, and the model climatology is shown by contours. (c),(d) As in (a),(b) but for temperature (K). (e),(f) As in (a),(b) but for water vapor specific humidity (g kg⁻¹).

the deep ocean is fully spun up. While the root-meansquare error of the SST in MIROC5 is 1.5 K, which is only slightly smaller than the value in MIROC3.2(medres) (1.6 K), the error reduction is more evident for the tropics (20°S–20°N), where the error is 1.2 K in MIROC5 and 1.5 K in MIROC3.2(medres). Although the SSS biases in the two models are relatively similar, the bias is slightly smaller in MIROC5 (Fig. 4d).

As compared with the CMAP climatology (Fig. 5a), the precipitation pattern simulated in MIROC5 appears to have a sharp contrast between the heavy precipitation regions, such as the intertropical convergence zone (ITCZ) and the surrounding areas. The mean precipitation in MIROC5 is realistic overall, but a deficiency is seen by taking the difference from CMAP (Fig. 6). It is known that MIROC3.2 fails to produce sufficient precipitation along the South Pacific convergence zone and over the western and eastern sides of the Maritime Continent (Figs. 6a,b). It also shows a double-ITCZ structure, as revealed by the positive bias to the south of the equator over the eastern Pacific. These shortcomings are greatly diminished in MIROC5, which, instead, shows the ITCZ that is too strong, accompanied by weak precipitation over the western equatorial Pacific (Fig. 6c). A positive bias around 60°S is also amplified. It is noticeable that the bias pattern in MIROC5 is considerably different from that in MIROC3.2(medres), which is similar to the bias in MIROC3.2(hires) (Fig. 6b). This clearly indicates that replacing the atmospheric physics package (convection scheme, in particular), as well as the changes in the other component models, had a drastic effect on the model bias. In contrast, increasing the horizontal resolution, at least from T42 to T106, does not alter the large-scale bias pattern in MIROC3.2.

Another feature of the precipitation in MIROC5 is topographically generated precipitation over continents (e.g., south of the Himalayas, Fig. 5b). This, however, may not be well resolved by the CMAP data. To validate the small-scale features in the precipitation climatology, a comparison is made of the June–August (JJA) mean precipitation over the Asian monsoon region between the Tropical Rainfall Measuring Mission (TRMM) satellite data (available online at http://www.eorc.jaxa.jp/ TRMM/index_e.htm) and MIROC (Fig. 7). The TRMM precipitation radar (PR) data averaged for 1998–2008 provide a detailed picture of the monsoon rainfall concentrated along narrow mountains (Fig. 7a). MIROC3.2 (medres), which does not resolve such a narrow topography, yields a diffuse precipitation climatology broadly representing the centers of monsoon rainfall, while additional heavy rainfall is observed over oceans (Fig. 7b). The precipitation pattern in MIROC3.2(hires) is more confined to narrow regions along the Western Ghats, Himalayas, and the western Indochina peninsula (Fig. 7c). The amount of orographically anchored precipitation is much less than the TRMM precipitation. The mean distributions are qualitatively similar in MIROC3.2(medres) and MIROC3.2(hires), but are they different from that in MIROC5 (Fig. 7d). Despite the coarser resolution of MIROC5 when compared to MIROC3.2(hires), it is better at representing the orographic rainfall over the monsoon region-even overestimated in association with the strong hydrological cycle (cf. Table 1). The shortcomings of the precipitation pattern in MIROC5 are too heavy rainfall over southern China and insufficient rainfall to the west of the Philippines, which is also seen in MIROC3.2(hires).

The zonal mean climatological fields of the zonal wind, temperature, and specific humidity, along with their biases, are presented in Fig. 8. In MIROC3.2(medres), the SH jet is shifted southward, and a cold (dry) bias near the tropopause (above the boundary layer) is conspicuous (Figs. 8a,c,e). These are all improved considerably in MIROC5 (Figs. 8b,d,f). One may wonder if this improvement is due to the doubled resolution of the atmosphere model. However, the biases in MIROC3.2(medres) are similarly found in MIROC3.2(hires), indicating that they are associated with the parameterization schemes. We found that the updated radiation, turbulence, and cloud schemes all act to reduce the temperature and moisture biases.

In addition to the SST (Fig. 3), the zonal wind stress (τ_x) and ocean subsurface states, which play a vital role in the ENSO simulation, should be validated. The annual means of τ_x and SST along the equator are plotted in Figs. 9a,b. It is apparent that the Pacific trade winds are underestimated in MIROC3.2(medres), but they are very close to the observations in MIROC5. The mean SST in the central equatorial Pacific is underestimated in both models; yet, the SST gradient is slightly larger in MIROC5. As also found in Fig. 3, the zonal SST gradient in the models is reversed in the Atlantic.

The observed subsurface temperature climatology is obtained from Ishii et al. (2006). Because there is no gridded subsurface current dataset, we use the assimilation products of Simple Ocean Data Assimilation (SODA) data for the 1958-99 climatology (Carton and Giese

FIG. 9. Annual mean climatology along equator of (a) zonal wind stress (N m⁻²) and (b) SST (°C) from observations and models. (c) Annual mean climatology along equator of observed ocean temperature (shading, °C) and ocean zonal current (contour interval: 10 cm s⁻¹) estimated by SODA analysis. (d),(e) As in (c), but for the MIROC3.2(medres) and MIROC5, respectively.

2008). The observations in the Pacific are characterized by a contrast between the warm pool extending down to 100 m in the west, and an eastern Pacific cold region where upwelling cools the subsurface as well as by a strong EUC having an eastward velocity greater than 80 cm s^{-1} (Fig. 9c). The subsurface temperature climatology in both models appears similar to each other. However, a careful comparison shows that the temperature in the central Pacific at around 100 m is warmer in MIROC5, resulting in the larger zonal gradient to the east (Figs. 9d,e). This difference is consistent with the EUC being shifted westward and weaker (\sim 30 cm s⁻¹) in MIROC3.2(medres), while in MIROC5 it is located at around 140°W, as in the SODA climatology, with an





FIG. 10. Climatological Arctic sea ice concentration in (a) JFM and (b) JAS derived from observations. (c),(d) As in (a),(b), but for the Antarctic sea ice. (e)–(h) As in (a)–(d), but for MIROC5.



FIG. 11. (a) Annual mean climatology of AMOC (contours, Sv) and bias of zonal mean temperature in the Atlantic Ocean in MIROC3.2(medres) (shading, K). (b) Time series of maximum (black) and equatorial (red) transport of the AMOC in MIROC3.2(medres). (c),(d) As in (a),(b), but for MIROC5.

increase in intensity (\sim 40 cm s⁻¹). Since the horizontal resolutions of the ocean models are identical, this change is attributed mainly to the different advection schemes used in the ocean model and indirectly to the atmosphere model. The EUC in MIROC5 is still too weak, which should be closer to the SODA data if a higher-resolution ocean component is used.

The Arctic and Antarctic sea ice concentrations are shown in Fig. 10. In the NH, the sea ice fraction in MIROC resembles the observations in both the winter [January–March (JFM)] and summer [July–September (JAS)] seasons, except for an underestimation over the Okhotsk Sea and off Newfoundland in winter and off the shores of Eurasia and Alaska in summer. A large bias is found in the SH, where the sea ice is always less than observed around Antarctica. This bias is worse than MIROC3.2 and appears to be related to the amplified warm bias in SST (Fig. 4). Since the deep ocean is still weakly drifting, a slow warming of the abyssal layer may eventually stimulate convection, which works to reduce the SH bias. We also compared the seasonal cycle of the sea ice concentration (not shown). Again, the NH



FIG. 12. Zonal wavenumber–frequency power spectra of the symmetric component of OLR divided by background power, obtained from (a) NOAA satellite, (b) MIROC3.2(medres), (c) MIROC3.2(hires), and (d) MIROC5, based on 30-yr daily data. Dispersion curves of equatorial waves for the three equivalent depths of 12, 25, and 50 m are indicated by red lines. Signals corresponding to the westward and eastward inertio-gravity (IG) waves [westward IG (WIG) and eastward IG (EIG)], ER waves, equatorial Kelvin waves, and MJO are labeled in (a).

sea ice extent is remarkably well simulated, but both the annual mean and amplitude of the seasonal cycle are underestimated in the SH.

The Arctic sea ice thickness field in MIROC5 has a maximum along the Canadian coast and the northern coast of Greenland (not shown). This spatial pattern, also found in MIROC3.2(hires), is improved from MIROC3.2(medres), in which the sea ice along the Siberian coast was thicker.

c. Variability

Among the various aspects of the natural variability, several phenomena are selected as examples to discuss the simulated perturbations arising from the atmosphere and ocean, and their coupling, namely, the Atlantic meridional overturning circulation (AMOC), equatorial waves, and ENSO. A full description of these modes of variability in MIROC5 and their sensitivity to parameters will be reported elsewhere (Watanabe et al. 2011; Chikira and Sugiyama 2010).

Figure 11 compares the mean AMOCs and their fluctuations. The AMOC intensity is measured by the maximum transport in the North Atlantic and the transport at the equator (Figs. 11c,d). MIROC5 generates a somewhat strong AMOC, which has a maximum transport of about 22 Sv (1 Sv $\equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) and a transport of about 17 Sv at the equator. These are stronger by about 2 Sv than those in MIROC3.2(medres). Because of observational uncertainty, it is not clear which is more realistic. The AMOC in MIROC5 shows a slight weakening tendency during this period, but it again gradually strengthens in the next 100 yr (not shown). In association with the AMOC variability, we also analyzed the time series and spatial pattern of the Atlantic multidecadal oscillation (AMO). When compared with observations (Trenberth and Shea 2006), both MIROC3.2(medres) and MIROC5 reproduced well a horseshoe pattern in the annual mean SST anomalies in the North Atlantic (not shown). The variances of the AMO time series were similar between MIROC3.2(medres) and MIROC5 (0.14 and 0.16 K); however, the time scale of the variability was somewhat shorter in MIROC3.2(medres) than in MIROC5, with the latter comparable to the observations.

The property of equatorial waves in the atmosphere is examined by calculating zonal wavenumber–frequency power spectra for the symmetric component of the outgoing longwave radiation (OLR), following the procedure proposed by Wheeler and Kiladis (1999). The daily mean OLR data derived from the Advanced Very High Resolution Radiometer (AVHRR) of the National Oceanic and Atmospheric Administration (NOAA) satellites for 1979–2005 (Liebmann and Smith 1996) are



FIG. 13. Time series of Niño-3 SST anomaly: (a) observations, (b) MIROC3.2(medres), (c) MIROC3.2(hires), and (d) MIROC5. The standard deviation in K is indicated in each panel.

used for observational references, which show wellseparated signals corresponding to the Madden–Julian oscillation (MJO) as well as the equatorial Rossby (ER) and Kelvin waves (Fig. 12a). The overall spectra are similar between MIROC3.2 and MIROC5 (Figs. 12b–d), but the equatorial Kelvin waves are overrepresented (underrepresented) in MIROC3.2 (MIROC5). In particular, both MIROC3.2(medres) and MIROC3.2(hires) simulate too much of the high-frequency Kelvin waves having a deep equivalent depth. It is noticeable that the power spectrum in MIROC5 exhibits a distinct peak akin to the observed MJO, even though it includes perturbations with smaller zonal scales (Fig. 12d). Further exploration of the modeled intraseasonal variability is beyond the scope of this paper, but Fig. 12



FIG. 14. Monthly anomalies in observed (a) SST (K), (b) precipitation (mm d⁻¹), and (c) 500-hPa height (m) regressed upon the Niño-3 SST time series. The contour intervals are 0.2 K, 0.5 mm d⁻¹, and 5 m (zero contours omitted), respectively. (d)–(f) As in (a)–(c), but for MIROC5. The shading indicates the correlation coefficient.

strongly suggests that the MJO is better reproduced in MIROC5.

As mentioned in the introduction, MIROC3.2 was not able to simulate sufficient ENSO amplitude, although the zonal gradient of the mean thermocline was realistic (Guilyardi et al. 2009b). Fortunately, this deficiency is greatly improved in MIROC5. Figure 13 compares the ENSO amplitudes as measured by the Niño-3 SST anomalies between the observations and three models. As evident from the time series and the standard deviation (SD), the ENSO amplitude in MIROC3.2(medres) is roughly half that observed; it is even weaker in MIROC3.2(hires) (Figs. 13b,c). In MIROC5, ENSO is much more realistic in terms of the amplitude and asymmetry between El Niño and La Niña (Fig. 13d). The SD is nearly identical to the observations in the 100-yr period shown, but this is partly coincidental because the SD varies from 0.64 to 0.99 K when sampling different 100-yr periods between 2001 and 2300. The ENSO periodicity was also examined using the power spectrum of the Niño-3 SST anomalies (not shown). A comparison of the spectra shows that the ENSO in MIROC5 has clear double peaks at 4.3 and 5.3-6.7 yr, which are close to the observed peaks at 3.7 and 5.2 yr, respectively, for the 1945–2006 period. In contrast, the ENSO in MIROC3.2(medres) has a broad single peak at around 4-10 yr.

Figure 14 illustrates the monthly anomalies regressed on the Niño-3 time series shown in Figs. 13a,d. The global SST anomaly pattern associated with the ENSO in MIROC5 is remarkably similar to the observations (Figs. 14a,d). The negative precipitation anomaly over the western tropical Pacific is somewhat underestimated, but the extratropical response to ENSO is also very realistic (Figs. 14c,f).

A full explanation of why MIROC5 simulates ENSO much better than MIROC3.2 is difficult because of the complexity of the ENSO dynamics. One possible reason is the intensified atmosphere-ocean coupling, as measured by the so-called coupling feedback parameter, μ , which is defined by the regression slope of τ_x anomalies over the Niño-4 region upon the Niño-3 SST anomalies (Guilyardi et al. 2009a). A scatterplot of these two quantities shows that in MIROC5, μ is twice as large as in MIROC3.2(medres) (Fig. 15), and it lies within the observational estimate of $\mu = 8.6-12.8 \times 10^{-3} \text{ N m}^{-2} \text{ K}^{-1}$ (cf. Guilyardi et al. 2009a). The realistic intensity of the wind stress response to the SST anomaly is related to the improved zonal profile of mean τ_x (Fig. 9a). We performed several additional experiments in which the ENSO amplitudes were found to be sensitive to the change in a parameter that affects the efficiency of cumulus entrainment. This convective control of ENSO



FIG. 15. Scatterplot of Niño-4 τ_x anomaly against Niño-3 SST anomaly in (a) MIROC3.2(medres) and (b) MIROC5. The value of the regression slope that defines μ (10⁻³ N m⁻² K⁻¹) is denoted at the bottom-right corner.

has been fully investigated and reported in a separate paper (Watanabe et al. 2011).

4. Climate sensitivity

One of the lessons from the CMIP3 is the necessity of using a variety of metrics to evaluate the errors in CGCMs. This implies that a model representing reasonable climate mean states ensures neither realistic internal variability nor reliable climate sensitivity (e.g., Knutti and Hegerl 2008). Therefore, in this section we perform a preliminary examination of the equilibrium climate sensitivity in MIROC5.

As a prelude to the climate sensitivity analysis, cloud properties are compared between MIROC and satellite estimates that have recently become available, as they are crucial in determining the climate sensitivity to radiative forcing. First, the upper-tropospheric cloud ice contents obtained from the control runs in MIROC3.2(medres) and MIROC5 are presented together with the annual mean of two satellite estimates (Fig. 16, partly reproduced from Waliser et al. 2009). Since the Aura/Microwave Limb Sounder (MLS) and CloudSat products are available only for 2007 and from August 2006 to July 2007, respectively, full comparisons with satellite climatology are not possible. Yet, a tendency for MIROC3.2(medres) to produce excessive cloud ice is clearly seen, whereas MIROC5 generates a more reasonable amount of ice content (Figs. 16c,d). The lack of cold rain processes in MIROC3.2 might explain why the old model overestimates cloud ice.

Representation of the cloud liquid/ice partitioning is a more severe test of a model's cloud scheme. The fraction of cloud liquid to total cloud condensate (liquid + ice), denoted as F_{liq} , may follow the temperature in the environment, as has been assumed in MIROC3.2; however, it will also depend on microphysical processes. The observed F_{liq} is estimated from the Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) derived from algorithms by Yoshida et al. (2010) and is compared with the MIROC5 outputs. Because the period of satellite data is too short, more significance is obtained via latitudinal distributions of $F_{\rm liq}$ as a function of temperature rather than geographical maps (Fig. 17). As anticipated, the F_{liq} in CALIPSO gradually decreases for lower temperatures, but is high over the subtropics and polar latitudes (Fig. 17a). The average temperature when the amount of cloud liquid and ice is equal ($F_{\text{liq}} = 0.5$) is about -10° C. While the cause is not yet clear, MIROC5 does well reproducing the latitudinal distribution of F_{lig} (Fig. 17b). A wavy pattern near the equator is due to the insufficient number of samples for $T < 0^{\circ}$ C and is of no concern. It should be noted that a given function of F_{liq} in MIROC3.2 was set so that $F_{\text{liq}} = 0.5$ at $T = -7.5^{\circ}\text{C}$, and it was tuned from a standard function of $F_{\text{lig}} = 0.5$ at $T = -15^{\circ}$ C, which results in a very high climate sensitivity. The realistic F_{liq} as a function of T in MIROC5 appears to lead to a better response of the mixed-phase cloud to radiative forcing.

Given that some of the cloud properties in MIROC5 are comparable with satellite data, the equilibrium climate sensitivity, denoted as ΔT_{2x} , is then evaluated by using



FIG. 16. (a),(b) Annual mean cloud ice mixing ratio at 215 hPa derived from satellite estimates (adapted from Fig. 16 of Waliser et al. 2009). The unit is 10^{-6} kg m⁻³. (c),(d) As in (a),(b), but for MIROC3.2(medres) and MIROC5, respectively.

initial 20-yr products of the $CO_2 \times 4$ experiment. For the initial states taken from the control integration, the model was rerun with an abruptly increased CO₂ concentration of 1138.8 ppm, which is 4 times the value used in the control run. Because the ensemble $CO_2 \times 4$ experiment using MIROC5 is still ongoing, the full analysis will be done in future work. To estimate ΔT_{2x} without an equilibrium calculation, which requires a very long integration, a method proposed by Gregory et al. (2004) is adopted. Given the fact that the relationship between the annual and global mean radiative flux change (CO₂ \times 4 – control) at TOA and the SAT change is well fitted by a straight line, ΔT_{2x} can be obtained at the intersection of the regression curve with the horizontal axis (i.e., extrapolation to the equilibrium state), divided by 2. Gregory et al.'s method is very convenient for estimating ΔT_{2x} using a transient time evolution in the $CO_2 \times 4$ experiment with full CGCMs.

The regression of the TOA net radiation change on the SAT change is presented in Fig. 18 (black line), where ΔT_{2x} is estimated at 3.6 and 2.6 K in MIROC3.2(medres) and MIROC5, respectively. The reduction of the climate sensitivity in MIROC5 arises from a weakly negative SW feedback due to clouds; it has an opposite sign to that in MIROC3.2(medres) (red lines). This is consistent with the previous works showing that the cloud– SW feedback varies the most among CGCMs (e.g., Bony and Dufresne 2005).

The cloud-SW feedback in MIROC3.2 is determined mostly by low clouds (Yokohata et al. 2010). Therefore, the signs of the cloud-SW feedback suggest that low clouds decrease by quadrupling the CO_2 in MIROC3.2(medres), whereas they slightly increase in MIROC5. The changes in low cloud cover indeed reveal such differences (Fig. 19). The polar cloud is found to increase in both models, and the overall patterns are not drastically different. However, the low clouds decrease in low latitudes, except for the off-equatorial Pacific in MIROC3.2(medres) (Fig. 19a), whereas they increase over the tropical oceans in MIROC5 (Fig. 19b). The tropical $(30^{\circ}\text{S}-30^{\circ}\text{N})$ mean change is +0.84% in MIROC5, compared to -0.62% in MIROC3.2(medres). It is interesting that the subtropical subsidence has weakened in the $CO_2 \times 4$ run because of increased static stability; nevertheless, the boundary layer is thinner in the MIROC5 $CO_2 \times 4$ experiment. This appears to be consistent with the prediction of a simple model by Caldwell and Bretherton (2009), but further investigation of the mechanism responsible for this low cloud increase and the resultant negative cloud-SW feedback is beyond the scope of this paper.

5. Concluding discussion

A new version of the global climate model MIROC was developed for better simulation of the mean climate,



FIG. 17. Latitudinal distribution of annual mean F_{liq} as a function of temperature: (a) 2006/07 mean of the CALIPSO data (H. Okamoto 2010, unpublished manuscript) and (b) climatology in MIROC5.

variability, and climate change due to anthropogenic radiative forcing. A century-long control experiment was performed using the new version (MIROC5) with the standard resolution of the T85 atmosphere and 1° ocean models. The climatological mean state and variability were then compared with observations and those from a previous version (MIROC3.2) with two different resolutions, coarser and finer than the resolution of MIROC5.

Climatological precipitation and SST improved in MIROC5 in several respects: a single ITCZ, more realistic zonal SST gradient on the equator, and topographically anchored precipitation associated with the Asian monsoon (Figs. 3, 5, 7). A new cumulus convection scheme and a more accurate advection scheme for ocean currents may be the major contributors to these improvements. Updated schemes for radiation and turbulence also work to reduce the zonal mean biases in temperature and moisture (Fig. 8). It is noticeable that MIROC5 simulates ENSO more realistically than the previous models, which produced a weak ENSO even with a higher resolution (Figs. 13, 14). The new MIROC employs a prognostic treatment for the cloud water and ice mixing ratio, as well as the cloud fraction, considering both warm and cold rain processes. Validation of the model cloud fields using recent satellite data shows that



FIG. 18. Gregory plots obtained from abrupt $CO_2 \times 4$ experiment in (a) MIROC3.2(medres) and (b) MIROC5 (see text for the experiment). The intersection between the net radiative flux at TOA (black line) and the horizontal axis (divided by 2) indicates the equilibrium climate sensitivity.

they are better simulated in MIROC5 than in MIROC3.2 (Figs. 15, 16). MIROC5 reveals an equilibrium climate sensitivity of 2.6 K, which is 1 K lower than that in MIROC3.2(medres) (Fig. 17). This is probably because in the two versions, the response of low clouds to an increasing concentration of CO_2 is opposite; that is, low clouds decrease (increase) at low latitudes in MIROC3.2 (medres) (MIROC5).

The comparison of the two versions of MIROC presented here indicates that the overall effect on the model climatology of updating the parameterization schemes is greater than the effect of increasing the model resolution (at least for T106 versus T85). This may not be



FIG. 19. Changes in annual mean low cloud fraction (%) obtained from control run and abrupt $CO_2 \times 4$ experiment in (a) MIROC3.2(medres) and (b) MIROC5.

surprising because the high-resolution model used here does not explicitly resolve some key phenomena, such as the convective systems. However, a part of the model bias will certainly be improved by MIROC3.2(hires). For example, a substantial cooling bias in the SST is found over the North Atlantic in both MIROC3.2(medres) and MIROC5 (Figs. 4a,c). This bias corresponds to a slight displacement of the sharp zonal SST gradient associated with the Gulf Stream and is reduced in MIROC3.2(hires) that adopts the high-resolution ocean model (not shown). The new physics package in MIROC5 is also not capable of removing several biases in mean states. In particular, the lack of heavy precipitation in the west of the Philippines in boreal summer (cf. Fig. 7) may be crucial for the simulation of the tropical cyclone tracks associated with the subtropical high in the western Pacific.

The simulation of the climate variability and climate change in MIROC5 are only briefly described in the present paper. The mechanisms and their sensitivity to the details of the parameterizations will be elaborated upon in subsequent papers. We have performed several sensitivity experiments, which strongly suggest that the model ENSO is primarily controlled by the cumulus convection (Watanabe et al. 2011). It is also suggested from a series of aqua planet experiments that the equilibrium climate sensitivity, which is qualitatively different from that in the previous model version, can be attributed to the different treatment of clouds and cloud microphysics.

A possible extension of MIROC5 in the next stage may be the incorporation of modules-for example, carbon and chemistry cycles, dynamic vegetation-required for the ESM. Before such an extension, however, we plan to extensively use MIROC5 for mechanism studies to understand natural climate variability and for a series of the near-term climate prediction experiments designed in the Coupled Model Intercomparison Project phase 5 (CMIP5; details of the CMIP5 experiments are available online at http://cmip-pcmdi.llnl.gov/cmip5/experiment_ design.html). The better simulation of the ENSO and other modes of variability, as well as improved mean states, should increase the fidelity of near-term prediction, which is affected not only by anthropogenic radiative forcing but also intrinsic fluctuations in the climate system.

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