Improvement in Global Cloud-System-Resolving Simulations by Using a Double-Moment Bulk Cloud Microphysics Scheme

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ABSTRACT

This study examines the impact of an alteration of a cloud microphysics scheme on the representation of longwave cloud radiative forcing (LWCRF) and its impact on the atmosphere in global cloud-system-resolving simulations. A new double-moment bulk cloud microphysics scheme is used, and the simulated results are compared with those of a previous study. It is demonstrated that improvements within the new cloud microphysics scheme have the potential to substantially improve climate simulations. The new cloud microphysics scheme represents a realistic spatial distribution of the cloud fraction and LWCRF, particularly near the tropopause. The improvement in the cirrus cloud-top height by the new cloud microphysics scheme substantially reduces the warm bias in atmospheric temperature from the previous simulation via LWCRF by the cirrus clouds. The conversion rate of cloud ice to snow and gravitational sedimentation of cloud ice are the most important parameters for determining the strength of the radiative heating near the tropopause and its impact on atmospheric temperature.

1. Introduction

After the work by Cess et al. (1989), the importance of cloud feedbacks for climate projection has been revealed using general circulation models (GCMs); cloud feedbacks are the major cause of significant diversity in simulated results. The elaboration of cloud processes can lead to improvements in climate projection; however, GCMs necessarily assume subgrid cloud processes because of their coarse spatiotemporal resolution. The subgrid cloud processes are highly parameterized; the modeling of those processes has been an issue over the last few decades. Another way to contribute to understanding cloud processes is global high-resolution simulations using detailed cloud microphysics schemes.

Recently, global cloud-system-resolving models (GCRMs) and multiscale modeling frameworks (MMFs) have been developed to investigate the roles of convective cloud systems in large-scale atmospheric disturbances and general circulation using cloud microphysics schemes without cumulus parameterizations (Grabowski 2001; Khairoutdinov and Randall 2003; Tomita and Satoh 2004; Khairoutdinov et al. 2008; Satoh et al. 2008; Tao et al. 2009). Although GCRMs and MMFs have been successful in exploiting simulations of multiscale interactions over the tropics, simulation biases exist in precipitation, cloud amount, and cloud radiative forcing (CRF) (Khairoutdinov et al. 2008; Tao et al. 2009; Kodama et al. 2012; Hashino et al. 2013). It is expected that these systematic biases associated with hydrological cycles will be reduced by improving the representation of cloud microphysical processes.

The Nonhydrostatic Icosahedral Atmospheric Model (NICAM; Tomita and Satoh 2004; Satoh et al. 2008, 2014) GCRM uses the single-moment bulk cloud microphysics
schemes proposed by Grabowski (1998) and Tomita (2008). Because global simulations with a horizontal resolution of a few kilometers are quite expensive despite the use of massively parallel computers, the aforementioned simple cloud microphysics schemes are acceptable (e.g., the computational cost of the simulations using a double-moment bulk cloud microphysics scheme described here is approximately twice as large as that using the aforementioned simple cloud microphysics schemes). Only a limited number of studies have used a double-moment bulk cloud microphysics scheme in either MMFs or GCRMs (Wang et al. 2011a, b). Several studies have shown that the cloud optical properties, radiative budget, and structure of convective clouds and associated anvil clouds are sensitive to the assumed values of empirical parameters used in such simple cloud microphysics schemes (Iga et al. 2007; Satoh et al. 2010; Kodama et al. 2012; Hashino et al. 2013). For general use in climate simulations such as climate change conditions, use of empirical formulations with uncertain parameters in cloud microphysics schemes must be avoided.

In general, cloud microphysics schemes used in regional nonhydrostatic models have not been able to be utilized for global simulations because their numerical results have been typically evaluated in specific local regions (e.g., Ikawa et al. 1991). Recently, various observational campaigns and satellite measurements have provided the modeling community with extensive datasets, and then cloud microphysics schemes have been statistically evaluated and improved for the purpose of general use based on these datasets (e.g., Matsui et al. 2009; Lang et al. 2011, 2014; Tao et al. 2014). In particular, general information about hydrometeor particle size distributions (PSDs) can be derived from an analysis of various observational datasets; methods to diagnose PSDs have contributed to the improvements of single-moment bulk cloud microphysics schemes (Field et al. 2005, 2007; Thompson et al. 2008; Abel and Boutle 2012). The diagnosis methods for PSDs have been tested against various types of cloud systems, and their effectiveness to reproduce cloud systems has also been reported (e.g., Lang et al. 2011). Recently, Roh and Satoh (2014) compiled a method to diagnose PSDs and showed improvements in a single-moment bulk cloud microphysics scheme used in NICAM (Tomita 2008; i.e., the NICAM single-moment scheme with six water categories, which is referred to as NSW6). It is promising that the realistic representation of hydrometeor PSDs in numerical models will improve the performance of global cloud-system-resolving simulations. However, the cloud microphysical processes in NSW6 scheme remain the simple formulations that are used in conventional regional nonhydrostatic models and will be improved. Our objective is to examine the change in the global distribution of simulated clouds by replacing the NSW6 scheme with a cloud microphysics scheme that includes different representations of microphysical processes and to investigate the reasons for the differences in the simulated results.

In parallel with the development of single-moment bulk microphysics schemes used in GCRMs, Seiki and Nakajima (2014) developed a double-moment bulk cloud microphysics scheme based on Seifert and Beheng (2006) for NICAM (i.e., the NICAM double-moment scheme with six water categories, which is referred to as NDW6). Double-moment bulk cloud microphysics schemes represent PSDs using the number concentration and mixing ratio of hydrometeors as the prognostic variables; the mode radii are dynamically evaluated. The mode radii of hydrometeors are particularly important for the initiation of precipitating particles through collisional processes (Berry and Reinhardt 1974; Seifert and Beheng 2001). After the initiation of precipitation, gravitational sedimentation further distorts the PSD, and the time evolution of the PSD affects the subsequent chain of particle growth (Wacker and Seifert 2001; Milbrandt and Yau 2005; Milbrandt and McTaggart-Cowan 2010). The NSW6 scheme uses empirical formulations with tuned parameters in these collisional and sedimentation processes. These processes are validated by only the radiation budget at the top of the atmosphere and are problematic from the aspect of the vertical structure of clouds (Kodama et al. 2012). In contrast, Seifert’s scheme has been well evaluated compared with a spectral bin cloud microphysics scheme and observations (Seifert and Beheng 2001; Seifert et al. 2006; Seifert 2008). We expect that a comparison of the NDW6 and NSW6 schemes will provide insight into the improvements of cloud microphysics schemes in GCRMs and MMFs. In addition, some GCMs have begun to implement detailed cloud microphysics schemes (e.g., Gettelman et al. 2008; Morrison and Gettelman 2008; Salzmann et al. 2010). This study will be increasingly informative for long-term simulations using such GCMs. In this study, we concentrate on the change in simulated cloud ice. We compare the distributions of solid hydrometeors and longwave cloud radiative forcing (LWCRF) among simulated results using the NDW6 and NSW6 schemes and retrieved datasets from satellite observations. We then examine the change in the impact of cirrus cloud radiative forcing on atmospheric temperature using the NDW6 scheme.

In section 2, we describe the experimental design of the global simulations. In section 3, we introduce the observational data used for comparisons. Moreover, in section 4, we present the results. We discuss the origin of biases in the atmospheric
2. Experimental designs

a. Control experiment

As a reference, the latest simulated results of the control run in the study by Kodama et al. (2012) are used herein (hereafter referred to as CTL). In the simulation, NICAM was used as a GCRM. The horizontal grid interval was approximately 14 km over the globe. The model top was set to 40-km altitude, and the vertical model domain had 40 layers. The authors analyzed 3-month simulations beginning on 1 June 2004.

Recently, it has been revealed that systematic biases in physical processes (e.g., cloud microphysics) in climate simulations using GCMs appear in short-range forecasts (Phillips et al. 2004; Martin et al. 2010; Williams et al. 2013). Thus, it is expected that 3-month simulations will provide characteristic differences in the performance of cloud microphysics schemes.

Cumulus parameterizations are avoided, and cloud growth is explicitly calculated in each grid box using the NSW6 scheme. The NSW6 scheme predicts the mixing ratio of cloud water, rain, cloud ice, snow, and graupel \( q_{cw}, q_r, q_i, q_s, \) and \( q_g \), respectively), and its microphysical formulations are based on Lin et al. (1983) and Rutledge and Hobbs (1983). Radiative transfer is calculated using a broadband model with 29 spectral bands (Sekiguchi and Nakajima 2008). The cloud optical properties are calculated in advance according to Mie theory, in which the refractive indices between liquid and solid hydrometeors are separated, and determined based on lookup tables as a function of the effective radii (Nakajima et al. 2000). In the CTL experiment, the effective radii of the liquid and solid hydrometeors are assumed to be 8 and 40 \( \mu m \), respectively.

The numerical settings of other physical processes are briefly summarized here; Kodama et al. (2012) should be referred to for additional details. The subgrid turbulent mixing is calculated using level 2 of the second-order closure scheme with moist processes (Nakanishi and Niino 2004; Noda et al. 2010). The land surface model is the Minimal Advanced Treatments of Surface Interaction and Runoff (MATSIRO; Takata et al. 2003), and a single-layer slab ocean model is used. The bulk surface fluxes over the ocean are calculated using the method proposed by Louis (1979). The initial conditions are determined by linearly interpolating the NCEP final operational grid analysis (FNL) ds083.2 product.

Kodama et al. (2012) reported that the CTL experiment captures the characteristic signal of deep convective clouds, which is a boomerang pattern in the radar reflectivity–height histogram, according to the CloudSat satellite observations, whereas the cirrus cloud fraction is overestimated compared to the CALIPSO data. The empirical parameters in the NSW6 scheme in the CTL experiment are selected to adjust the radiation budget at the top of atmosphere: sedimentation of the cloud ice is switched off and the critical value of the mixing ratio \( q_{icrt} \) for autoconversion from cloud ice to snow is set to 5 mg kg\(^{-1}\) (see the appendix).

b. New cloud microphysics scheme

The double-moment bulk cloud microphysics scheme, NDW6, predicts the number concentration of cloud water, rain, cloud ice, snow, and graupel \( N_{cw}, N_r, N_i, N_s, \) and \( N_g \), respectively) in addition to their mixing ratios. We use the number mixing ratio, which is the number concentration divided by the air density, in advection, sedimentation, and turbulent mixing processes. The merits of the use of the number mixing ratio instead of the number concentration in these processes were argued by Mansell (2010). Cloud microphysical processes in the scheme are based on Seifert and Beheng (2001, 2006) and Seifert (2008), and the differences from the original scheme are described in detail by Seiki and Nakajima (2014) and Seiki et al. (2014). Here, we briefly describe the numerical modeling of cloud ice. Non-sphericity of ice hydrometeors is characterized by power-law relationships between the particle mass \( m \) and maximum dimension \( D \) and between the particle mass and projected area perpendicular to the flow \( A \):

\[
D = a_m x^{b_m} \quad \text{and} \quad A = a_A x^{b_A},
\]

where \( a_m, b_m, a_A, \) and \( b_A \) are constant coefficients. We assume that cloud ice can be represented by hexagonal columns, snow as assemblages of polycrystals, and graupel as lump graupel; the coefficient values are those compiled by Mitchell (1996). In the NDW6 scheme, the power-law relationships are used for calculation of the terminal velocity, the collisional cross section is used in the aggregation process, and the capacitance is used in the deposition and sublimation processes. The scheme is coupled to the radiative transfer model for calculating the cloud optical properties from a lookup table of the effective radii \( r_e \). Following Fu (1996), the effective radius is defined as the ratio of the ice water content to the total projected area perpendicular to the flow as follows:

\[
r_{e,j} = \frac{3}{4 \rho_j} \int_0^\infty \frac{\rho_j f_j(x_j) dx_j}{A_j(x_j)} \quad \text{for} \quad j = i, s, g,
\]
where \( \rho_i = 916.7 \text{ kg m}^{-3} \) is the density of ice, \( \rho \) is the air density, and \( f(x) \) is a particle mass distribution function. Using broad look-up table that covers the effective radii from 1 \( \mu \text{m} \) to 1 mm, the scheme can estimate the CRF including precipitating hydrometeors. In addition, we assume nonspherical single scattering properties for ice hydrometeors based on Fu (1996) and Fu et al. (1998). This is the first study to utilize the NDW6 scheme in a global simulation.

Aerosol transport models have not yet been coupled with the NDW6 scheme thus far. This simulation examines the impact of cloud microphysics schemes on global cloud-system-resolving simulations in a stepwise fashion. Global cloud-system-resolving models and MMFs, when not coupled with aerosol transport models, have successfully reproduced interannual and intra-seasonal variabilities (Khairoutdinov et al. 2008; Satoh et al. 2012). These previous studies have shown that global cloud-system-resolving simulations could improve the climate simulations by explicitly calculating cloud growth. In the next stage, we examine the change in CRF and atmospheric conditions by altering the cloud microphysics scheme in an attempt to determine the physical mechanisms of the process chain that originate from CRF changes.

In this study, nucleation processes are calculated by assuming background aerosols. We used a warm cloud nucleation scheme assuming an activated spectrum proposed by Twomey (1959) with a maritime cloud condensation nuclei condition (Seiki and Nakajima 2014). We used the heterogeneous ice nucleation scheme proposed by Philips et al. (2007), which is based on the observational data collected in the free troposphere in the northern midlatitudes (Demott et al. 2003). Hereafter, the simulation using the NDW6 scheme will be referred to as NEW.

### 3. Observational data

To evaluate the simulated hydrometeors, we used the ice water content (IWC) data from the CloudSat level 2B radar-only cloud water content (2B-CWC-RO) product (Austin and Stephens 2001; Austin et al. 2009). In the analysis, we allocated the data into equal-latitude grid boxes of 2.5°, and then calculated the zonal mean values and averaged them over June–August (JJA) from 2007 to 2010. Because CloudSat was not launched prior to the simulation period, we used the climatological zonal mean IWCs as a reference. We assumed that the interannual variability in the zonal mean IWCs from the CloudSat 2B-CWC-RO product is smaller than the difference between the observed values and the simulated values.

We used the daytime cloud fraction data from the ISCCP D1 product (Rossow and Schiffer, 1999). The cloud fraction is defined by the ratio of cloud pixels to the grid boxes with a size of 2.5° × 2.5°; the cloud fraction is categorized by the cloud-top pressure \( p_c \) into three types: high, middle, and low cloud fraction that are defined by cloudy pixels for which \( p_c < 310 \text{ hPa} \), \( 310 < p_c < 680 \text{ hPa} \), and \( 680 \text{ hPa} < p_c \), respectively. In the simulations, the radiative transfer model includes the ISCCP simulator (Klein and Jakob 1999; Webb et al. 2001) to estimate the equivalent cloud fraction of the ISCCP product.

To evaluate the simulated radiative fluxes at the top of the atmosphere and the surface, the radiative fluxes data from the CERES synoptic 1° latitude–longitude gridded edition 3A (SYN1deg_Ed3A) product are used (Wielicki et al. 1996; Kato et al. 2011; Doelling et al. 2013). This product contains integrated observed radiative fluxes from CERES, MODIS, and several geostationary imagers and computed radiative fluxes at the surface based on the Langley Fu–Liou radiative transfer model (Fu and Liou 1992, 1993). The longwave cloud radiative forcing is estimated by calculating the difference between the observed all-sky radiative fluxes and computed clear-sky radiative fluxes. In the comparison, a positive sign indicates a warming effect to Earth’s surface.

To compare the precipitation, the precipitation data from the Global Satellite Mapping of Precipitation moving vector with Kalman filter (GSMaP-MVK), version 5, product are used (Ushio et al. 2009); this product was developed by integrating data from the microwave radiometer from the Tropical Rainfall Measuring Mission (TRMM), various microwave radiometers on polar orbital satellites, and infrared radiometers on geostationary satellites using optimization with a Kalman filter method. Because the TRMM orbit is concentrated over the tropics, the analysis region is limited to between 35°S and 35°N. In the comparison, the observation data are averaged over every grid box with a size of 1.0° × 1.0°.

We used the Japanese 25-year Reanalysis Project (JRA-25) data (Onogi et al. 2007) to reference the atmospheric conditions. For the comparison, the simulated results are linearly interpolated into equal-angle grids and subsequently averaged over each grid box with a size of 2.5° × 2.5°. In addition, the vertical coordinates of the simulated results are converted into pressure coordinates with 23 pressure levels in the same manner as the JRA-25 product.

### 4. Results

#### a. Simulated distributions of hydrometeors and longwave cloud radiative forcing

This section briefly summarizes the characteristic differences in the spatial distribution of simulated hydrometeors
and LWCRF between the CTL and NEW experiments and the differences between the simulated results and the observed data. Figure 1 shows the vertical distributions of the zonal mean IWCs from the observations and the NEW and CTL experiments. Here, the simulated IWCs include cloud ice, snow, and graupel. The IWC differences between the CTL and NEW experiments are large over the tropics and midlatitudes. In these regions, the IWCs in the NEW experiment are approximately twice as large as the IWCs in the CTL experiment. The zonal mean IWCs in the NEW experiment are closer to the observed values, although interannual variability might be an issue because the observed values are climatological. In the NEW experiment, excessively distributed thin cloud ice in the CTL experiment is clearly removed near the tropopause, and the cloud-top height is similar to the observed height. The numerical setting to switch off the sedimentation of cloud ice in the CTL experiment results in the overestimation of cloud ice near the tropopause. The role of the sedimentation of cloud ice is discussed in detail in section 5c.

Figure 2 shows the horizontal distribution of the observed and simulated ISCCP high cloud fraction. The globally averaged values are summarized in Table 1. The globally averaged values of the simulated cloud fractions in the NEW experiment are closer to the observed values.
values than those in the CTL experiment. The difference in simulated high cloud fraction between the CTL and NEW experiments is large over the tropical ocean and mid- to high-latitude regions. Correspondingly, the IWCs above 200 hPa decrease in the NEW experiment.

Figure 3 shows the horizontal distributions of the LWCRF; the globally averaged LWCRFs are summarized in Table 2. The NEW experiment reduces the positive bias over the tropical ocean and Antarctica that is found in the CTL experiment (Figs. 3d,e). The reduction in the LWCRFs is related to the change in the cloud-top height in the NEW experiment, with lower cloud-top heights generally inducing lower LWCRFs.

As discussed by Satoh et al. (2010) and Kodama et al. (2012), the terminal velocity of cloud ice and the conversion process of cloud ice to snow strongly affect the amount of cloud ice. The formulation of ice cloud microphysics in the NEW experiment efficiently removes the cloud ice near the tropopause and reduces the positive LWCRFs over such regions. However, these results lead to a negative effect on the globally averaged radiation budget (Table 2) because positive biases over the western Pacific Ocean, Atlantic Ocean, and Maritime Continent and strong negative biases from the eastern Pacific Ocean to South China Sea are cancelled in the CTL experiment (Figs. 3d,f). Similarly, in the NEW experiment, shortwave cloud radiative forcing is reduced over the tropical ocean and becomes similar to the observations; however, the globally averaged shortwave cloud radiative forcing is less accurate than the CTL experiment (not shown).

Figure 4 shows the horizontal distributions of precipitation and the corresponding zonal means. In the

<table>
<thead>
<tr>
<th></th>
<th>ISCCP</th>
<th>CTL</th>
<th>NEW</th>
</tr>
</thead>
<tbody>
<tr>
<td>High cloud fraction</td>
<td>0.23</td>
<td>0.27</td>
<td>0.25</td>
</tr>
</tbody>
</table>

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NEW experiment, precipitation over the tropical western Pacific Ocean and northern Indian Ocean is improved by reducing the underestimation in the CTL experiment. Although the intertropical convergence zone (ITCZ) in the CTL experiment has a double peak structure, the ITCZ in the NEW experiment is concentrated in the middle and has a similar structure to the observed ITCZ. These improvements may be linked with the improvements in LWCRF and its effect on atmospheric temperature over the tropics (shown in section 4b) because precipitation is constrained by the energy balances of the atmosphere and the surface (Allen and Ingram 2002). However, precipitation over the tropics is strongly influenced by changes in the horizontal transport of enthalpy flux and LWCRF (Muller and O’Gorman 2011). An examination of the effect of cloud microphysics on the general circulation and changes in the horizontal transport of enthalpy flux is beyond the scope of this study. In the NEW experiment, these improvements in precipitation with different signs over various regions result in a slight increase in precipitation averaged over the tropics.

b. Reduction in the warm biases near the tropical tropopause

The zonal mean vertical profile of atmospheric temperature is strongly modified in the NEW experiment (Fig. 5). In the troposphere, the CTL experiment has a warm bias; the maximum value is approximately 7 K near the lapse-rate tropopause (~100 hPa) over the tropical region and the warm bias is distributed down to the freezing level. Here, the lapse-rate tropopause is defined by the altitude where the temperature lapse rate becomes zero. In the NEW experiment, the warm bias is clearly reduced and the biases are nearly less than ±2 K for the entire troposphere. The key factor for this improvement in the NEW experiment is the CRF by cirrus clouds near the tropopause. This subsection examines the origin of the warm bias in the CTL experiment.

The radiative heating rate due to the presence of cirrus clouds is shown. Figure 6 shows the time series of the simulated vertical profiles of the heating rate resulting from radiation, phase changes, turbulent mixing, and the difference of the atmospheric temperature relative to the JRA-25 product averaged over the tropics during JJA in 2004. In addition, the mixing ratio of cloud ice is superimposed on those profiles. Here, the tropical region, 10°S–10°N over all longitudes, is analyzed. Anvil

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**Table 2.** Globally averaged differences between simulated LWCRFs and those from the CERES data. The second and third row in each cell show the average values over the Northern and Southern Hemispheres, respectively.

<table>
<thead>
<tr>
<th></th>
<th>CTL</th>
<th>NEW</th>
<th>NEWFIX</th>
</tr>
</thead>
<tbody>
<tr>
<td>Global</td>
<td>1.31</td>
<td>−2.01</td>
<td>5.16</td>
</tr>
<tr>
<td>NH</td>
<td>−0.56</td>
<td>−1.49</td>
<td>6.51</td>
</tr>
<tr>
<td>SH</td>
<td>3.19</td>
<td>−2.52</td>
<td>3.80</td>
</tr>
</tbody>
</table>

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**Fig. 4.** Horizontal distribution and zonal mean of precipitation (mm h⁻¹) during JJA in 2004 based on (c) the GSMaP_MVK product, and simulations by the (a) NSW6 and (b) NDW6 schemes. The difference between the observed precipitation and simulated precipitation using the (d) NSW6 and the (e) NDW6 schemes; (f) the corresponding zonal means are also portrayed.
cirrus clouds expand and cloud ice is concentrated from 300 to 200 hPa in both the CTL and the NEW experiments (Fig. 6). Strong radiative heating occurs at approximately 100 hPa, which is where small amounts of cirrus clouds are distributed. In contrast, the heating rates resulting from phase changes and turbulent mixing are much smaller near the tropopause.

The aforementioned features are consistent with the observations and show that the atmospheric heating near the tropopause is dominated by CRF from cirrus clouds (Ackerman et al. 1988; Jensen et al. 1996; Hartmann et al. 2001), which occurs because convective heating and absorption of longwave radiation by carbon dioxide and water vapor is relatively smaller than the absorption of longwave radiation by cirrus clouds above the 200 hPa.

In the CTL experiment, the simulated atmospheric temperature continues to increase, particularly around the tropopause, until the simulated atmospheric temperature and radiative fluxes are in approximate balance 30 days into the simulation. In general, the greenhouse effect of cirrus clouds becomes stronger as the cirrus temperature decreases (Liou 2002). Relatively large amounts of cloud ice remains at higher levels in the CTL experiment than in the NEW experiment (Figs. 1 and 6); moreover, weaker radiative cooling is observed in the troposphere in the CTL experiment (Figs. 6a,e). In the middle troposphere, the magnitude of radiative cooling is reduced by an average of 0.2 K day$^{-1}$ during JJA in 2004 compared with the NEW experiment. This reduction results in an overestimation of atmospheric temperature by a few degrees Kelvin during the first 10 days. Therefore, the CTL experiment has a positive atmospheric temperature bias in the troposphere as a result of the cirrus clouds located at approximately 100 hPa, although the total amount of cloud ice is smaller than in the NEW experiment. We discuss the reason why the vertical distribution of cirrus clouds changes because of the alteration of the cloud microphysics scheme in section 5c.

Atmospheric temperature biases remain in the NEW experiment, such as those above the tropopause or in the southern midlatitude region. The alteration of the cloud microphysics scheme changes the global atmospheric temperature and does not necessarily improve the regional-scale atmospheric temperature biases. It is beyond the scope of this study to discuss the origin of the regional-scale biases.

5. Discussion

a. Role of the cloud optical properties

In the NEW experiment, there are two factors that can induce substantial CRF differences based on the CTL experiment: cloud microphysics and cloud optical properties. To isolate the contribution of the two factors to improve atmospheric temperature estimates, an additional
experiment is performed (called NEWFIX). The experiment is the same as the NEW experiment except that the cloud optical properties are calculated in the same manner as in the CTL experiment; that is, the cloud optical properties are calculated according to Mie theory and the effective radii of liquid and solid hydrometeors are assumed to be 8 and 40 μm, respectively (see section 2a).

Figure 7 shows the LWCRFs in the NEWFIX experiment and their differences relative to the CERES product. Table 2 summarizes the globally averaged LWCRFs. The longwave cloud radiative forcings are obviously overestimated on a global scale. The major contributors to the LWCRF are ice hydrometeors. Compared with the observed IWCs, the IWCs in the NEW experiment are not overestimated and are instead slightly underestimated (Figs. 1a,c). The NEWFIX experiment has been suggested to overestimate the LWCRF by assuming effective radii of solid hydrometeors as 40 μm in the calculation of the cloud optical properties. Recent progress in satellite analyses has revealed global vertical distributions of the effective radii of solid hydrometeors (Okamoto et al. 2010; Sato and Okamoto 2011). According to the observed profiles, the
effective radii below 10 km (∼300 hPa) exceed 100 μm over the tropics. Thus, the assumed effective radii in the CTL and NEWFIX experiments are inconsistent with the relatively large observed effective radii, which originate from the presence of the precipitating solid hydrometeors.

Figure 8 shows the zonal mean vertical profile of atmospheric temperature in the NEWFIX experiment and its difference relative to the JRA-25 product. The general features of the atmospheric temperature bias are quite similar to the NEW experiment, although the radiative budget at the top of the atmosphere increasingly differs from the observations in the NEWFIX experiment.

Comparing the CRFs between the NEW and NEWFIX experiments, the CRF change that results from the assumed cloud optical properties in the radiative transfer calculation is found to be 7 W m⁻² for the LWCRF. The extent of the change is greater than the change in the CRF that results from the alteration of cloud microphysics scheme (see Table 2). Careful attention is necessary to investigate process chains involving CRF using such simple cloud microphysics schemes (e.g., the NSW6 scheme) that assume fixed effective radii for calculating the cloud optical properties because tuning globally averaged CRFs (e.g., Kodama et al. 2012) could conceal cloud microphysics biases.

b. Persistence of cirrus clouds

Here, we discuss the contribution of cloud ice to the radiative heating rate near the tropopause over the tropics. The domain extending over 10°S–10°N and all longitudes from 200 to 100 hPa is analyzed. Figure 9a shows the cloud fraction sorted by the cloud ice mixing ratio CF(qi), which is defined by the ratio of the grid-scale air mass for each value of qi against the air mass of the entire domain as follows:

$$\text{CF}(q_i) = \frac{\sum_{q_i < q < q_i + \Delta q} \rho V}{\sum_{q_{\min} < q < q_{\max}} \rho V},$$

where V is the volume of a grid box, q_{min} = 0.1 mg kg⁻¹, q_{max} = 100 mg kg⁻¹, and qi is discretized into 30 bins within a range from q_{min} to q_{max} on a logarithmic scale. In the CTL experiment, there is a turning point at q_i = 5 mg kg⁻¹. In contrast, a moderate distribution of the cloud fraction is produced in the NEW experiment, which has a mode q_i of approximately 1.4 mg kg⁻¹.
which is smaller than that produced in the CTL experiment. Figure 9b shows the domain-averaged total radiative heating rate sorted by cloud ice mixing ratio $Q_{\text{RAD ave}}(q_i)$, which is defined as follows:

$$Q_{\text{RAD ave}}(q_i) = \frac{\sum_{q_{\text{min}} < q_i < q_{\text{max}}} \rho V Q_{\text{RAD}}}{\sum_{q_{\text{min}} < q_i < q_{\text{max}}} \rho V}.$$

where $Q_{\text{RAD}}$ is the radiative heating rate in a grid box. As shown in Fig. 9a, there is a turning point at $q_i = q_{\text{icrt}}$ in the CTL experiment. Because the radiative heating rate of cirrus clouds increases as $q_i$ increases, the large $q_i$ mode in the cloud fraction in the CTL experiment leads to a large radiative heating rate near the tropopause. In the NEW experiment, the cirrus cloud fraction for $q_i$ exceeding 1 mg kg$^{-1}$ is efficiently removed. Therefore, the radiative heating rate near the tropopause decreases.

As mentioned in section 2, the conversion from cloud ice to snow is initiated at an empirical critical value of the cloud ice mixing ratio of $q_{\text{icrt}} = 5$ mg kg$^{-1}$, assuming a time scale for the autoconversion ($\tau_{\text{auto}}$) in the NSW6 scheme (Kodama et al. 2012). The artificial turning points observed in the cloud fraction and radiative heating rate shown in Figs. 9a and 9b correspond to the empirical value of $q_{\text{icrt}}$. In contrast, in the NEW experiment, the autoconversion from cloud ice to snow is resolved using an approximated stochastic collection equation instead of the simple linear dumping, which is used in the NSW6 scheme (see the appendix). This theoretical formulation of the conversion rate provides for the continuous growth of ice particles by aggregation depending on their size and number concentration. As a result, the radiative heating rate near the tropopause is reduced; moreover, the zonal mean profiles of atmospheric temperature are improved in the NEW experiment.

We then speculate as to why the ice particles are persistent near the tropopause in the CTL experiment. Large values of $q_i$ might be removed from the tropopause by gravitational sedimentation or sublimation, although the autoconversion from the cloud ice to snow is not sufficiently efficient for dissipation. Jensen et al. (1996) suggested that the persistence of cirrus clouds near the tropical tropopause is strongly influenced by sublimation and the lofting of cloudy parcels by buoyancy that originates from radiative heating. The former works to decreases the occurrence of cirrus clouds, whereas the latter works to enlarge cirrus clouds. In the CTL experiment, large amounts of thin cirrus clouds near the tropopause can be found, which means that the lofting effect by radiative heating works efficiently to maintain persistent cirrus clouds. The lofting effect can be estimated according to the sum of the updraft and sedimentation. In the CTL experiment, the sedimentation of cloud ice is artificially switched off to adjust for the simulated CRF at the top of the atmosphere to the observed value (Kodama et al. 2012). In the NEW experiment, the average $q_i$ fallout is approximately 720 m h$^{-1}$ (20 cm s$^{-1}$) at $q_i = 5$ mg kg$^{-1}$ (Fig. 10). In the context of Jensen’s suggestion, the lack of sedimentation overestimates the lofting effect and makes it more persistent in the CTL experiment, particularly over the tropics.

The warm bias in the troposphere in the CTL experiment is suggested to originate from two factors in the NSW6 scheme: the empirical formulation of the autoconversion from cloud ice to snow and the sedimentation of cloud ice. Kodama et al. (2012) reported that smaller values of $q_{\text{icrt}}$ and the implementation of cloud ice sedimentation reduce the cloud fraction of thin cirrus clouds and lower the cirrus cloud-top height; therefore, the CRF by ice clouds is underestimated. A sensitivity experiment that is the same as the CTL experiment except that the sedimentation of cloud ice is calculated and $q_{\text{icrt}}$ is set to 0 [the qicrt0–4water experiment in Kodama et al. (2012)] has shown smaller biases in atmospheric temperature during the first 10 days of the simulation (not shown). However, the high cloud fraction is reduced by half and CRFs are underestimated (approximately 13.1 W m$^{-2}$ for the LWCRF). A similar sensitivity of the terminal velocity of cloud ice to the CRF was obtained by Heymsfield and Iaquinta (2000). Rougier et al. (2009) showed that the choice of small values for the terminal velocity of cloud ice increases the
climate sensitivity in the Hadley Centre Slab Climate Model version 3. Thus, the radiation budget and persistence of cirrus clouds are very sensitive to the balance between the autoconversion and sedimentation processes.

Here, the balance between the two factors in the NEW experiment is not necessarily accurate because the two factors are not directly validated by the observations, although atmospheric temperature is well reproduced. We expect that the new earth observation satellite EarthCARE, which has been developed by the European Space Agency, Japan Aerospace Exploration Agency, and National Institute of Information and Communications Technology and is planned to be launched in 2015, will provide valuable data to constrain the terminal velocity of ice particles using Doppler radar.

6. Conclusions

We examined the impact of the alteration of the cloud microphysics scheme on the distribution of cloud fraction, LWCRF, and atmospheric temperature in global cloud-system-resolving simulations over three months. The NDW6 scheme was relatively accurate at reproducing the vertical distribution of IWC and cirrus cloud-top heights. Improvements were also shown in the horizontal distribution of high clouds. However, the globally averaged LWCRF exhibit additional bias despite decreases in the high cloud fraction biases. The improvement in the horizontal distribution of precipitation is similar to that of LWCRF. To reduce the remaining biases, further investigations into other cloud microphysical processes and related physical processes are necessary in the future.

In the CTL experiment, atmospheric temperature has warm biases of up to approximately 7 K that are primarily located near the tropopause. A change in the cloud optical properties works to slightly reduce the warm biases; however, it does not dominantly influence atmospheric temperature despite the substantial change in the radiation budget. The warm biases can be greatly reduced using the NDW6 scheme through the cloud microphysical response and associated cirrus LWCRF. The amount of high clouds quickly responds to the alteration of the cloud microphysics scheme within the first few days of the simulation. In the CTL experiment, the warm trend continues until the atmospheric temperature balances at a warmer state after 30 days. The overestimation of the high cloud fraction causes a greenhouse effect that induces a warm bias in the troposphere. The efficient removal of cloud ice by aggregation and sedimentation processes are key processes for reducing the bias. We suggest the use of theoretical formulations as used in the NDW6 scheme to represent the detailed dependences of growth rates on microphysical parameters (see the appendix).

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APPENDIX

Autoconversion from Cloud Ice to Snow

In the NSW6 scheme, the conversion from cloud ice to snow is parameterized as follows:

$$\frac{\partial q_i}{\partial t} = -E_c(T) \frac{\rho(q_i - q_{\text{crit}})}{\tau_{\text{auto}}},$$

where $E_c$ is the sticking efficiency as a function of atmospheric temperature, $q_{\text{crit}}$ is a critical value to initiate the conversion process, and $\tau_{\text{auto}}$ is the relaxation time scale of the conversion process. Kodama et al. (2012) assumed that the values of $q_{\text{crit}}$ and $\tau_{\text{auto}}$ are global constants.
In the NDW6 scheme, based on Seifert and Beheng (2006), the stochastic collection equation is solved by assuming a binary collision to derive the conversion rate from cloud ice to snow via aggregation of cloud ice as follows:

$$\frac{\partial p_i}{\partial t} = -E_i(T) \int_0^\infty \int_0^\infty f_i(x)f_j(y)\alpha_i(x,y)\delta v_j(x,y) \, dx \, dy$$

(A2)

where $f_i(x)$ is the mass distribution function of cloud ice with a particle mass of $x$, $\alpha_i$ is the collisional cross section, and $|\delta v_j|$ is the relative terminal velocity between two colliding particles. In the NDW6 scheme, this equation is solved with several approximations. Seifert and Beheng (2006) should be referred to for the detailed derivation. Here, we attempt to show a simplified dependence of the growth rate on the bulk cloud parameters with further approximations. Assuming a monodisperse mass distribution function for $f_i(x)$, the mean projected area of cloud ice $A_i$ represents $\alpha_i$, and the mean terminal velocity of cloud ice $v_i$ represents $|\delta v_j|$, Eq. (A2) can be solved as follows:

$$\frac{\partial p_i}{\partial t} \sim -E_i(T)p_iN_iA_i v_i.$$  

(A3)

Furthermore, assuming that cloud ice consists of spherical particles that are sufficiently small to satisfy Stokes' law for evaluating their terminal velocities, the dependence of the growth rate is approximated as follows:

$$\frac{\partial p_i}{\partial t} \sim -E_i(T)p_iN_i\overline{D_i} v_i,$$  

(A4)

where $\overline{D_i}$ is the mean mass diameter. Comparing Eq. (A4) with Eq. (A1), the critical value speculated to initiate the conversion process ($q_{ictr}$) results from the strong dependence of the conversion rate on the mean mass diameter. In the theoretical formulation, the relaxation time scale and critical value are not global constants; instead, these values are dependent on the mixing ratio, number concentration, and mean mass diameter.

REFERENCES


