Cold and Warm Rain Simulated Using a Global Nonhydrostatic Model without Cumulus Parameterization, and their Responses to Global Warming

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(Manuscript received 27 February 2014, in final form 20 November 2014)

Abstract

A global nonhydrostatic model was used to evaluate the reproduction skill of cold and warm rain over the ocean at low latitudes and investigate their responses to global warming. In response to global warming, surface precipitation at low latitudes (30°S–30°N) in the simulations using mesh sizes of 7 and 14 km (R7 and R14, respectively) increased by 1.9 % and 2.6 %, respectively. It was found that the increase in precipitation in the higher horizontal resolution model R7 was caused by the increase in cold and warm rain and that in R14 was due to the increase in cold rain. In R7, the net increase in cold rain occurred due to the increase in stronger precipitation (> 40 mm hr⁻¹), most of which compensated for the decrease in weaker precipitation (< 40 mm hr⁻¹). In contrast, warm rain increased in almost all ranges of precipitation intensity. The fractional coverage of warm (cold) rain increased (decreased) robustly for both mesh sizes in the simulations. Analysis of the contribution of dynamic and thermodynamic environmental changes to the changes in cold and warm rain revealed a strong dependency on dynamic regimes in their effects.

The lifespans of cold and warm clouds at low latitudes (defined by the ratio of the sum of cloud water and cloud ice paths to the precipitation flux) and possible changes related to global warming were also evaluated. On an average, in all precipitation intensities, there was no significant change in the longevity of cold clouds in response to global warming. In contrast, the lifespan of warm clouds was reduced in most of the sea surface temperature anomaly regimes.

Keywords tropical and subtropical atmospheres; precipitation; global warming; cloud longevity; global nonhydrostatic model; high-resolution simulation

1. Introduction

Warm rain, the precipitation that develops in clouds without accompanying ice processes (hereafter, warm clouds), is mainly generated by boundary-layer and cumulus congestus clouds with cloud top heights below the freezing level. On the other hand, cold rain is precipitation, which forms clouds in association with ice processes (hereafter, cold clouds). In contrast to the processes that generate cold rain, the major processes contributing to the development of warm clouds and associated warm rain are the collision and coalescence of droplets.

The importance of cold rain processes is relatively straightforward as they contribute to the direct driving forces associated with planetary-scale circulations, such as the Hadley circulation, through deep convection and sometimes bring about severe local weather.
Following extensive efforts, our understanding of the importance and roles of warm rain and associated cloud processes has improved considerably. For example, warm clouds are the primary contributor of heat and moisture to the lower atmosphere (Johnson et al. 1996, 1999). Vertical heat travels by boundary-layer cumulus clouds and their subsequent migration toward the equator modulate tropical thermodynamic structures, and can thereby modulate the width and strength of the Hadley circulation (Neggers et al. 2007). These processes then feedback to the behavior of deep convection. Another important role of warm clouds is preconditioning, which facilitates the development of deep convection by moistening the lower atmosphere (Brown and Zhang 1997; Takemi et al. 2004; Waite and Khouider 2010; Rapp et al. 2011). This process then affects larger-scale circulations, including monsoon onset and intraseasonal variability (Kiladis et al. 2005).

Warm rain frequently develops over the regions of warmer sea surface temperature (SST), and together with tropical warm rain (which occurs as persistent background rain in the tropics), it generally contributes to the modulation of atmospheric fields (Short and Nakamura 2000; Takayabu and Masunaga 2009; Liu and Zipser 2009). Observations by the Tropical Rainfall Measuring Mission (TRMM) satellite show that in addition to the contributions of deep clouds, remarkable latent heating also occurs on account of warm clouds over low latitudes (Kodama et al. 2009).

Lau and Wu (2003, hereafter LW03) show the stronger dependence of weaker warm rain that developed over a warm pool region of the subtropical and tropical Pacific Ocean on SST. Consequently, they speculate on two possible and contrasting changes in response to global warming—either increased warm rain will lead to a weakening tropical deep convection due to reduced vapor supply to the mid-to-high troposphere or it may act alternatively to intensify tropical deep convection due to the moistening of the lower atmosphere and strengthening of the low-level convergence of horizontal flow. Lau et al. (2005) then used a global climate model (GCM) to show that warm rain modulates atmospheric stability and subsequently, deep clouds. The combination of radiative forcing by the modulated upper part of deep clouds with changes in the parameter associated with warm rain processes can significantly alter the pattern of intraseasonal variability. This raises the interesting issue of how the activity of low-latitude warm rain as well as cold rain will change in the future, including under a warmer global atmosphere.

Most GCM studies still suffer from ambiguity associated with the behavior of simulated deep convection arising from cumulus parameterization (Randall 2003). However, using a high-resolution nonhydrostatic GCM without cumulus parameterization, which computes tropical convection explicitly, we can expect to reduce this ambiguity. For example, previous studies show realistic simulations of atmospheric disturbances over the scales of several thousand kilometers as well as diurnal variations (Miura et al. 2007; Nasuno et al. 2009; Sato et al. 2009; Noda et al. 2012, 2014).

Conventionally, time-integration of such high-resolution GCMs with, for example, meshes of 7 and 14 km has been limited to timescales shorter than intraseasonal due to large computational demand (e.g., Noda et al. 2010; Yamada et al. 2010; Yamada and Sato 2013). At higher resolutions, such as a 3.5-km mesh model, only 10-day integration has been possible until very recently. The recent development of a much faster supercomputer (the K computer; http://www.ais.c.u-cen.jp/en/kcomputer/) has enabled longer-term integration; e.g., year-long simulations using a 7-km mesh model. Here, we use data from the first long-term simulation by a nonhydrostatic GCM with mesh sizes of 7 and 14 km in order to assess the characteristics of simulated cold and warm rain, and their possible changes under a warmer global atmosphere.

2. Data

2.1 Model

The configuration of the nonhydrostatic icosahedral atmospheric model, (NICAM) (Tomita and Sato 2004; Satoh et al. 2008, 2014) used in this study was the same as that described by Noda et al. (2012). Cloud microphysics processes were computed using a one-moment scheme that predicts the mixing ratios of rainwater, graupel, and snow and also diagnoses cloud water and cloud ice (Tomita 2008). The mesh intervals used were approximately 7 and 14 km (R7 and R14, respectively). Although Inoue et al. (2008) showed that a mesh interval of 3.5 km generates better simulations of the cloud size distribution, R7 and R14 reproduce realistic behaviors of tropical cloud systems (Tomita et al. 2005; Miura et al. 2007; Sato et al. 2009). Using these two datasets of differing resolutions, R7 and R14, we were able to consider the precipitation statistics at least qualitatively. In addition, because of the relatively smaller computational demand of R7 and R14, it is possible to perform one-year experiments, which enable statistical analysis to include the seasonal cycle, although
these simulations are much heavier in terms of the computing resources required than those of conventional GCMs.

The experiment followed a time-slice approach (Bengtsson et al. 1996), which consisted of a control (CTL) experiment run from June 2004 to May 2005, and a global warmer-atmosphere experiment (GW) that began in May, and was time-integrated for one year at the end of the 21st century. The year 2004 was originally selected to evaluate the model performance of typhoon simulation because Japan experienced typhoon landfalls during that year, and validation studies of typhoons have been reported in several papers (e.g., Yamada et al. 2010; Yamada and Satoh 2013).

For CTL, the SST boundary conditions and sea ice concentrations (SICs) were the same as those used in the boreal summer 2004 experiment of Noda et al. (2012). The conditions for GW were developed using the dataset of the World Climate Research Program Coupled Model Inter-comparison Project phase 3 (CMIP3) and the method of Mizuta et al. (2008). The climate forcing for GW was created by adding the SST and SIC differences between the present day (1979–2003) and GW (2075–2099) periods to CTL. The carbon dioxide concentration in GW was uniformly twice that in CTL. The initial conditions for GW were taken from a present-day National Centers for Environmental Prediction reanalysis dataset for 00:00 UTC May 01, 2004, and spun up for one month. After confirming the similarity of the R14 result to that from R7, we will focus here on the results of the R7 run.

2.2 Satellite-based data

We used Global Precipitation Climatology Project (GPCP) data (Adler et al. 2003) to evaluate the spatial characteristics of the modeled surface precipitation. TRMM orbit data were also used to verify the vertical evolution of modeled clouds; namely, the storm top heights and SSTs from the 2A23 and 2A25 products were obtained for the period between June 01, 2004 and 30 May, 2005, which covers the same period as the simulation.

2.3 Definitions

For TRMM data, we defined a warm rain grid as the one that contained surface precipitation and had a storm top height of less than 5 km (LW03). For the model, based on detailed microphysical data, we used a more physically based definition. A cold rain grid was defined as the grid where the following three conditions were satisfied: 1) surface precipitation > 0 mm hr\(^{-1}\); 2) graupel > 0 at one or more levels of the column; and 3) the sum of all condensates > 0 at all levels below the layer in which condition 2 was satisfied. The second condition was stipulated because precipitation accompanying ice processes will be generated via graupel. The third condition ensures the selection of only vertically continuous deep cloud. We defined a warm rain grid as the one in which the first condition was satisfied and was not categorized as a cold rain grid. An additional explanation regarding the differences from the TRMM data is provided in the Appendix. Generally, some rain evaporates out entirely before reaching the surface; however, we refer to “rain” as simply that reaching the surface (i.e., surface precipitation).

Hereafter, the term “low latitudes” refers to the region between 30°S and 30°N, and any results that do not refer specifically to another period relate to the year simulated.

3. Results

3.1 Responses of surface precipitation to global warming

Figure 1 compares surface precipitation with GPCP data. The model simulates their spatial characteristics, although it overestimates the amplitude, especially along the oceanic intertropical convergence zone (ITCZ). In response to global warming, surface precipitation decreases over Central America, the Maritime Continent, and central America, but, in contrast, increases over the western Indian Ocean and the middle and eastern parts of the Pacific Ocean. The change in surface precipitation is positive, on an average, over low latitudes (1.9 % for R7 and 2.6 % for R14). As the gross behavior of the R7 and R14 simulations was similar, our discussion (below) will focus mainly on the results of the higher-resolution run (R7).

Recent GCM simulations appear to be consistent; e.g., modeled surface precipitation tends to have a less zonal contrast, such as an underestimation of surface precipitation near the ITCZ and overestimation around it (Solomon et al. 2007). Recent, CMIP5 results overestimate the surface precipitation in the tropical convergence zones, south of the equator in the Atlantic and the eastern Pacific, but underestimate it along the equator in the western Pacific (Flat and Marotzke 2013). In contrast, our modeling results tend to overestimate the precipitation along the ITCZ, except for the Amazon basin and underestimate the precipitation in the central–western Pacific.
Considering the responses to global warming, Sun et al. (2007) indicate the decrease in surface precipitation over most subtropical regions and support the so-called dry-get-drier hypothesis. Our results agree with theirs in that the increase in low-latitude precipitation occurs mainly over the central and eastern Pacific. In contrast, our results showed a decrease in the precipitation over most of the maritime continent, which differs from the CMIP5 result.

Figure 2 compares the frequency of the occurrence and intensity of the modeled warm and cold rain with net precipitation. Warm rain shows larger precipitation amounts relative to cold rain only off the west coasts of continents, and the ratio (warm/total) is often greater than 0.8 (Figs. 2a, b). In contrast, the contribution of warm rain to total precipitation intensity is much smaller than that of cold rain, and areas with relatively high amounts of warm rain are restricted to the subtropics and the northern and eastern parts of the ITCZ (Figs. 2c, d). These modeled results show a distribution of warm rain similar to previous TRMM observations (e.g., Short and Nakamura 2000; Takayabu and Masunaga 2009).

To illustrate these contributions to the total values more clearly, precipitation amounts, their frequencies of occurrence, and precipitation-weighted frequencies of occurrence are shown in Fig. 3. Previous studies (e.g., Johnson and Xie 2010; Ma et al. 2012) also emphasize that convective activity is strongly controlled not only by SST itself but also by an SST anomaly from the zonal mean (ΔSST). Thus, we present such a view in Fig. 3. The contribution from warm rain is distributed over a wide range of ΔSST (ca. −10°C to +5°C), but that of the cold rain region is mostly restricted to regions of −1°C ≤ ΔSST ≤ 4°C (Fig. 3a). As also shown in Fig. 2, the contribution of
warm rain to total precipitation is much smaller than that of cold rain over almost all ΔSST regimes (Fig. 3b). Consequently, the low-latitudinal mean of surface precipitation is mostly associated with cold clouds.

In response to global warming (Figs. 3c, d), the frequency of the occurrence of warm (cold) rain increases (decreases) around its maximum occurrence of ΔSST regimes of 0°C (2.7°C; Figs. 3a, c). Another noteworthy change may be the pronounced increase in warm rain occurrence around ΔSST = −8°C. The absolute amounts of warm and cold rain changes positively and negatively across 0°C almost symmetrically (Fig. 3d). Cold rain shows a relative increase (decrease) in weak (strong) and positive ΔSST regimes around 1.5°C (3.5°C), and, in contrast, the increase in warm rain occurs almost over the ΔSST regimes smaller than 4.0°C. Less cold rain in the higher ΔSST regions is mainly the consequence of the lower frequency of the occurrence of such higher ΔSST regions in GW (Figs. 3c, d). The fractional coverage of warm rain increases in both R7 and R14, but this change is more pronounced in R14 (Table 1). The changes of each rain type in the warmer atmosphere seem to be somewhat resolution dependent, at least qualitatively. That is, in the higher-resolution simulation (R7), both cold and warm rain increases. In particular, the increase in warm rain contributes primarily to the increase in total precipitation. In contrast, in the lower-resolution simulation (R14), the change in warm rain is much less than that of cold rain. These results suggest that the responses of warm rain processes are more sensitive to model resolution, and this is partly because of the better representation of small-scale convection in higher resolution models (Sato et al. 2009; Noda et al. 2012).
3.2 Thermodynamic and dynamic fields

To examine the response of tropical convection to global warming, we show vertical velocity at the 500 hPa altitude level ($\omega_{500}$ hPa day$^{-1}$) and low-level convergence ($\text{Conv}_{\text{Low}}$ s$^{-1}$) in both atmospheres in Fig 4. Low-level convergence is defined as follows:

$$\text{Conv}_{\text{Low}} = -\frac{1}{H} \int_{\text{surface}} \hat{\nabla}_h \cdot \vec{V} \, dz,$$

where $H = 2$ km, $\vec{V}$ is a velocity vector, and $\hat{\nabla}_h = \left( \frac{\partial}{\partial x}, \frac{\partial}{\partial y}, 0 \right)$ is a differential operator.

In response to global warming, the frequency of occurrence of intense (weak) magnitude $\omega_{500}$ decreases (increases) (Figs. 4a, b). The relatively strong convergence region ($> 2 \times 10^{-3}$ s$^{-1}$) decreases,
indicating that the strength of convergence weakens (Fig. 4d). The changes of $\omega_{500}$ along with weaker amplitudes of $\text{Conv}_{\text{Low}}$ indicate weakened activity of mean deep convection (accompanied by weaker and moderately stronger precipitation) in the warmer atmosphere at low latitudes. However, as will be seen later, particularly, very low frequency and much stronger precipitation increases, and this is the primary contributor to the increase in total precipitation due to global warming.

3.3 Increase in surface precipitation

The increase in low-latitude precipitation in a warmer atmosphere agrees with the traditional theory regarding the energy budget of tropical and subtropical atmospheres (Manabe and Strickler 1964; Manabe and Wetherald 1967). However, the greater increase in warm rain over cold rain in R7 is not straightforward, although the contribution of the former to total precipitation is one order of magnitude smaller than the latter (Table 1). To better understand this response, we estimated the potential for the generation of condensed water paths (WP) in warm and cold clouds in given atmosphere ($\text{WP}_{\text{cold}}$ and $\text{WP}_{\text{warm}}$, respectively; e.g., Rosenfeld and Woodkey 2000; Khain et al. 2001). We computed the lifting of an air parcel adiabatically from the lower atmosphere up to the level at which the air parcel loses buoyancy—the neutral buoyancy level (NBL)—after the air parcel drains its vapor to condensation. We used a level of 950 hPa as a representative of the lower atmospheric air to be lifted. We hereby define $\text{WP}_{\text{cold}}$ ($\text{WP}_{\text{warm}}$) as a WP for which the accompanying surface precipitation and the temperature of its NBL is $< 273.15$ K ($\geq 273.15$ K).

Figure 5 compares $\text{WP}_{\text{cold}}$ and $\text{WP}_{\text{warm}}$ along with their responses to global warming. In the present climate, $\text{WP}_{\text{warm}}$ is distributed across almost all low latitudes, except for the western Pacific and continents. In contrast, $\text{WP}_{\text{cold}}$ is restricted to regions in, and near to the ITCZ. In a warmer atmosphere, $\text{WP}_{\text{warm}}$ increases over almost the entire globe,
except the polar regions and, for example, small parts of the western Pacific, and $W_{\text{diag}}^\text{warm}$ is enhanced along the ITCZ. In total, the increase in $W_{\text{diag}}^\text{warm}$ outweighs that of $W_{\text{diag}}^\text{cold}$.

Realistically, actual liquid water and ice water contents differ from those estimated by adiabatically-lifted air parcels for several reasons. For example, during ascent, an air parcel should also lose buoyancy due to dilution. Nevertheless, if we assume that these quantities represent, in some part, the potential to generate condensates in each column, the pronounced increase in warm rain can be attributed partly to the local changes in the vertical thermodynamic structure of the warmer atmosphere.

Figure 6 shows the PDF and precipitation-weighted PDF as a function of its intensity. The occurrence frequency becomes noticeably larger in weaker precipitating regions, and these weaker precipitating regions (Fig. 6a) contribute more to total precipitation in both atmospheres (Fig. 6b). For example, the precipitating regions of $< 10 \text{ mm hr}^{-1}$ cover 99.0 % of the occurrence frequency, and they explain 33.0 % of the total precipitation in the CTL run of R7 (Figs. 6a, b). In response to global warming, the occurrence frequency of, in particular, weaker precipitation ($< 1 \text{ mm hr}^{-1}$) increases, and precipitation heavier than 1 mm hr$^{-1}$ becomes less frequent (Fig. 6d). Weaker and moderate precipitation ($< 40 \text{ mm hr}^{-1}$) decreases in frequency, and stronger precipitation ($> 40 \text{ mm hr}^{-1}$) becomes more frequent (Fig. 6e).

The increase in stronger cold rain ($> 40 \text{ mm hr}^{-1}$) almost compensates for the decrease in weaker cold rain ($< 40 \text{ mm hr}^{-1}$; Fig. 6f). Consequently, the slight positive of the residual contributes to the net increase in cold rain. In contrast, warm rain increases in almost all bins (thin line in Fig. 6f). In particular, the increase in weaker precipitation bins ($< 5 \text{ mm hr}^{-1}$) explains more than half the overall increase in warm rain. We showed in Table 1 that the net increase in warm rain is larger than that of cold rain in R7. More precisely, the present analysis reveals that the increase in warm rain associated with weak precipitation and that of cold rain associated with intense precipitation, are the main cause of the increase in total low-latitude precipitation due to global warming.

Similar to the present results, conventional GCMs also show an increase in the occurrence frequency of intense (weak) precipitation (Sun et al. 2007; Vecchi and Soden 2007; Chou et al. 2012). These precipitation responses were robustly observed in previous GCM studies (Knutti and Sedláček 2012; Liu et al. 2012; Huang et al. 2013). These similar precipitation
changes may imply the strong constraint of energy and hydrological balance in the low-latitude atmosphere, as also suggested by Liu et al. (2012).

3.4 Dynamic and thermodynamic variations

It is of interest to consider a possible link between warm and cold rain processes. To this end, we investigated the dynamic and thermodynamic relationships in the atmosphere in cold and warm rain regions (e.g., Emori and Brown 2005; Bony and Defresne 2005). Moreover, the probability density function of $\omega_{500}$, $F_\omega$, shows that the peak of $F_\omega$ is located at approximately 20 hPa day$^{-1}$ in a subsidence regime (solid line in Fig. 4a). Mean precipitation in each $\omega$ bin increases as the amplitude of $\omega_{500}$ increases (Fig. 7). In this presentation, the precipitation at each grid point and each time step can be further decomposed into either a warm or cold rain column:

\[
\delta P = \int_{-\infty}^{\infty} (F_\omega,\text{cold} + F_\omega,\text{warm} + \delta F_\omega,\text{cold} + \delta F_\omega,\text{warm})(P_\omega,\text{cold} + P_\omega,\text{warm} + \delta P_\omega,\text{cold} + \delta P_\omega,\text{warm}) \, d\omega
\]

Responses to the global warming of $P$, $\delta P$ can then be written as

\[
\delta P = \int_{-\infty}^{\infty} \frac{\delta F_\omega,\text{cold} P_\omega,\text{cold} + P_\omega,\text{warm}}{\partial \omega} \, d\omega + \int_{-\infty}^{\infty} \frac{\delta F_\omega,\text{warm} (P_\omega,\text{cold} + P_\omega,\text{warm})}{\partial \omega} \, d\omega
\]
shows the greatest amplitudes in general and acts with the change in precipitation. Among the terms, interaction between warm and cold rain processes, covariance terms, because they include the effects of thermodynamic components of the warm and cold represent covariance between changes in the dynamic thermodynamic element. Finally, the other four terms are those not directly related to the circulation or the spheric circulation, and the third and fourth terms of precipitation is related to the changes in atmospheric vapor. This section examines such a relation analysis regarding the hydrological cycle and atmospheric vapor. LW03, we define the residence time (RT in min) of in most low latitude areas. The major impact of an interaction between the changes to cold and warm rain (\( \tilde{C}_{cc} \) and \( \tilde{C}_{cw} \)) is weakening precipitation. Overall, the role of the covariance terms is minor, except for \( \tilde{C}_{cc} \), as the net positive effect of \( \tilde{C}_{cc} \) is the second largest after \( D_w \) and acts to increase low-latitude precipitation.

Figure 9 compares the RHS terms of Eq. (2) as a function of \( \omega_{500} \). The sum of the RHS terms (black line, almost overlaps with the blue line) acts to weaken precipitation in strongly ascending regimes, but strengthen it in moderately ascending and descending regimes. Particularly, the contribution of \( T_c \) differs depending on the dynamic regimes,

\[
\begin{align*}
\dot{D}_c + \dot{D}_w + \dot{T}_c + \dot{T}_w + \dot{C}_{cw} + \dot{C}_{cw} + \dot{C}_{cw} = \dot{C}_{cw},
\end{align*}
\]

where a hat, \( \hat{\cdot} \), indicates \( \int_{-\infty}^{\infty} d\omega \). The sum of the first and second terms on the right hand side (RHS) represents a dynamic component for which the variation of precipitation is related to the changes in atmospheric circulation, and the third and fourth terms are those not directly related to the circulation or the thermodynamic element. Finally, the other four terms represent covariance between changes in the dynamic and thermodynamic components of the warm and cold rain columns. Of particular interest are the roles of the covariance terms, because they include the effects of interaction between warm and cold rain processes.

Figure 8 compares the contributions of these terms with the change in precipitation. Among the terms, \( T_c \) shows the greatest amplitudes in general and acts to intensify precipitation over the central and eastern Pacific. However, the net \( T_c \) contribution is weakly positive due to strong negatives in other low-latitude regions. Furthermore, its positives in many regions compensate strong negatives in \( D_c \), except in some tropical regions such as the central and eastern Pacific. In contrast, those terms associated with warm rain (\( \dot{D}_w \) and \( \dot{T}_w \)) help to intensify precipitation across wider regions of the low latitudes, albeit with smaller amplitudes. For the covariance terms, \( \dot{C}_{cc} \) acts to intensify precipitation mostly over the low latitudes. Similarly, \( \dot{C}_{cw} \) is positive and the distribution of \( T_c \) mostly explains that of the total. Dependency of \( D_w \) and \( \dot{C}_{cc} \) on dynamic regimes are similar to each other. \( D_c \) plays a major role in reducing the precipitation in all dynamic regimes. Focusing on the covariance terms, \( \dot{C}_{cw} \) is the largest term to control the changes of precipitation, as also shown in Fig. 8. The magnitude of \( \dot{C}_{cw} \) is comparable to that of \( T_w \), but acts to decrease precipitation. The effects of \( \dot{C}_{cw} \) are \( \dot{C}_{cw} \) increasing and decreasing precipitation, respectively, but their impacts are minor.

3.5 Residence time

LW03 indicated the possibility of the increasing amounts of warm rain in a warmer atmosphere and argued for a likely link between the changes to warm and cold rain. Li et al. (2011) also conducted a similar analysis regarding the hydrological cycle and atmospheric vapor. This section examines such a relationship in the modeled hydrological cycle. Following LW03, we define the residence time (RT in min) of the water cycle associated with clouds as follows:

\[
\begin{align*}
RT_{cold} &= a WP / P_{cold}, \\
RT_{warm} &= a WP / P_{warm},
\end{align*}
\]
Fig. 8. Contributions of the terms (mm day$^{-1}$) on the right side of Eq. (2).

Fig. 9. Budget of Eq. (2), showing the sum of the RHS terms, and the contributions of $D_c$, $D_w$, $T_c$, $T_w$, $C_{cc}$, $C_{ww}$, $C_{cw}$, and $C_{wc}$. Values are binned every 10 hPa day$^{-1}$.
where WP (kg m$^{-2}$) is the sum of the liquid and ice water paths (LWP and IWP, respectively), $P_{\text{cold}}$ and $P_{\text{warm}}$ are surface precipitation from cold and warm clouds, respectively, and $a$ ($=60.0$) is a unit conversion coefficient. Increased residence times mean that the clouds remain in the atmosphere for longer, and thus have a greater effect on radiative fields. First, we calculated the RT at each time step and grid point and then arranged them as a function of SST or ΔSST in order to reflect the relationship of the cloud water/ice path, and their precipitation within the same precipitation systems. Note that, in contrast to our approach, LW03 estimated RT by first calculating the cloud water path and precipitation as a function of SST and then divided the former by the latter for each SST bin.

To confirm the relationship between RTs and SST under present climatic conditions, we compared the dependence of RT$_{\text{cold}}$ and RT$_{\text{warm}}$ on precipitation intensity as a function of SST using TRMM data in Fig. 10. RT increases as SST increases (contour lines) and becomes significantly larger, but discontinuously (in the warm pool regions). While the pattern of RT$_{\text{warm}}$ is similar to that of RT$_{\text{cold}}$ (color shades), the peak of RT$_{\text{warm}}$ is distributed 3°C cooler than RT$_{\text{cold}}$. The modeled result is similar to the TRMM result, except that the model tends to overestimate RTs compared with the observations.

Figure 11 shows the results of RT$_{\text{cold}}$ and RT$_{\text{warm}}$ for CTL in R7, in turn, as a function of ΔSST (e.g., Fig. 3) and their responses to global warming. In contrast to RT$_{\text{warm}}$, RT$_{\text{cold}}$ evidently increases in the warm pool regions (contour lines of Figs. 11a, b). The responses to global warming of RT$_{\text{cold}}$ and RT$_{\text{warm}}$ differ, i.e., the frequency of RT$_{\text{cold}}$ is larger in moderate ΔSST regimes (0°C ≤ ΔSST ≤ 4°C), and decreases in other areas (color shades). On the other hand, there is no noticeable response to global warming (i.e., signal of a positive and negative pair in the vertical direction).
In contrast, RT$_{\text{warm}}$ is reduced, especially for $\Delta$SST $> 4°C$ or $\leq 3°C$.

We also investigated the changes in RT with respect to its dependence on precipitation intensity (Fig. 12). Obviously, changes in RT depend not only on $\Delta$SST but also on precipitation intensity. Both RT$_{\text{cold}}$ and RT$_{\text{warm}}$ become longer as precipitation intensity weakens (contour lines). In response to global warming, the RT$_{\text{cold}}$ of moderate and strong precipitation becomes longer in negative $\Delta$SST regimes ($\Delta$SST between $-6°C$ and $0°C$; Figs. 12b, c). In contrast to RT$_{\text{cold}}$, for cases with weaker precipitation, RT$_{\text{warm}}$ becomes shorter (Fig. 12d). Such an increase in longevity is not apparent in cases of moderate and intense precipitation (Figs. 12e, f).

4. Summary

Recent progress in computing has made possible the time integration of the 7-km mesh global nonhydrostatic simulations over periods much longer than seasonal timescales. Based on data from the first of these longer-term simulations, we examined the possible changes to warm and cold clouds and associated precipitation at low latitudes in the R7 and R14 runs in response to global warming.

Surface precipitation at low latitudes ($30°S$–$30°N$) in the R7 and R14 model runs increase by 1.9 % and 2.6 %, respectively, in the warmer global atmosphere. Under the definitions of cold and warm rain used here, we confirmed that warm rain becomes more important in higher-resolution simulation (R7) in order to explain the simulated increase in surface precipitation. We also confirmed that the contributions from the increase in warm rain for almost all precipitation intensities, and cold rain for very intense precipitation ($>40$ mm hr$^{-1}$) act to increase net precipitation over the tropics and subtropics.

To investigate the possible link between the changes in cold and warm rain, we decomposed the changes of the two types of precipitation into the contributions from the dynamic and thermodynamic components and their covariance. Our results showed that the sum of the covariance terms related to the interaction between the changes in cold and warm rain ($\hat{C}_{\text{cw}}$ and $\hat{C}_{\text{wc}}$) was weakly negative across most of the low latitudes, and thus acts to weaken precipitation. However, the effects of the covariance terms, except for $\hat{C}_{\text{cc}}$, are minor.
We also examined the lifespans of warm and cold clouds as well as the changes related to global warming. Here, we defined cloud longevity using the ratio of cloud water path to precipitation flux. The lifespans of cold and warm clouds at low latitudes defined by the ratio of the sum of cloud water and cloud ice paths to the precipitation flux, and possible changes related to global warming were also evaluated. On an average, over all precipitation intensities, no significant change in longevity was found in cold rain due to global warming. In contrast, the lifespan of warm rain decreased in most SST anomaly regimes.

Our results also suggest the importance of changes in the relationship between clouds and precipitation, which would vary depending on the associated precipitation intensity, because these changes would significantly affect the hydrological cycle in a warmer atmosphere through, for example, cloud radiative processes. The present research focused on low-latitude oceanic regions; consequently, further investigations of cold and warm rain over land or in mid-latitudes are important areas for future research.

Acknowledgments

We thank Dr. Hirohiko Masunaga, Hydrospheric Atmospheric Research Center, Nagoya University, and two anonymous reviewers for useful comments. This study was partly supported by the Program for Risk Information on Climate Change and Strategic Programs for Innovative Research Field 3 from MEXT, Japan. The simulations in this study were performed using the K computer and the Earth Simulator.

Appendix: Definition of cold and warm rain

LW03 define warm rain, using the storm height obtained from the TRMM precipitation radar data, as being either below 0°C level or less than 5 km above the ground level (AGL). Their method contains a degree of uncertainty because the storm height alone does not guarantee the presence of liquid and ice phases, as was noted by LW03. To improve the classification of warm and cold rain columns, we selected a method based on the modeled microphysical scheme (Tomita 2008), in which modeled cold rain must be associated with the generation of graupel.

Figure A1 shows a typical vertical profile of a warm rain column that has been misclassified as a cold rain column. In this case, although very small graupel exists in the upper troposphere (ca. 14 km AGL), it does not relate to surface precipitation because there is no condensate in underlying air by 2 km AGL. The surface precipitation is obviously generated by a liquid cloud that develops below 2 km. For such cases, it is also necessary to consider condition 3; otherwise, such columns would all be categorized as cold rain columns. The generation of minor graupel seems to be not only a consequence of the advection of an anvil cloud that is cut off from an ambient convective region but also generated by the freezing of supersaturated vapor that occurs locally and intermittently. The ice content of these clouds is

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Fig. A1. Example of a warm rain column wrongly classified as a cold rain column, showing (a) vertical profiles of graupel (mg kg$^{-1}$) and (b) the sum of liquid water contents (mg kg$^{-1}$).
Fig. A2. As for Fig. 2, but applying the vertical continuity of surface precipitation and liquid condensates of the atmosphere.

Fig. A3. As for Fig. 2, but using a threshold of 18 dBz for the classification of warm and cold rain columns.
very small, but is commonly present. In addition, we also tested a condition that ensures the vertical continuity of warm rain, but the result was not noticeably different (Fig. A2).

To allow a closer comparison with the results in LW03, we conducted another sensitivity study. In this analysis, we estimated a radar echo based on the empirical diagnosis of Rutledge and Hobbs (1984) and defined a warm rain column as having radar reflectivity of >18 dBz (Short and Nakamura 2000; LW03). In this case, we see more warm rain (Fig. A3). This result shows that about 20% of the total tropical precipitation is categorized as warm rain; about 10% greater than in the body text (Table 1).

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