Response of Ice and Liquid Water Paths of Tropical Cyclones to Global Warming Simulated by a Global Nonhydrostatic Model with Explicit Cloud Microphysics

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(Manuscript received 15 March 2013, in final form 7 August 2013)

ABSTRACT

Cloud feedback plays a key role in the future climate projection. Using global nonhydrostatic model (GNHM) simulation data for a present-day [control (CTL)] and a warmer [global warming (GW)] experiment, the authors estimate the contribution of tropical cyclones (TCs) to ice water paths (IWP) and liquid water paths (LWP) associated with TCs and their changes between CTL and GW experiments. They use GNHM with a 14-km horizontal mesh for explicitly calculating cloud microphysics without cumulus parameterization. This dataset shows that the cyclogenesis under GW conditions reduces to approximately 70% of that under CTL conditions, as shown in a previous study, and the tropical averaged IWP (LWP) is reduced by approximately 2.76% (0.86%). Horizontal distributions of IWP and LWP changes seem to be closely related to TC track changes. To isolate the contributions of IWP/LWP associated with TCs, the authors first examine the radial distributions of IWP/LWP from the TC center at their mature stages and find that they generally increase for more intense TCs. As the intense TC in GW increases, the IWP and LWP around the TC center in GW becomes larger than that in CTL. The authors next define the TC area as the region within 500 km from the TC center at its mature stages. They find that the TC’s contribution to the total tropical IWP (LWP) is 4.93% (3.00%) in CTL and 5.84% (3.69%) in GW. Although this indicates that the TC’s contributions to the tropical IWP/LWP are small, IWP/LWP changes in each basin behave in a manner similar to those of the cyclogenesis and track changes under GW.

1. Introduction

Cloud feedback is the main cause of uncertainties in climate sensitivity (Soden and Held 2006). For climate-projection simulations, understanding the response mechanisms of clouds due to global warming is important. Bony and Dufresne (2005) showed that the albedo of low clouds is extremely important for the uncertainty of cloud feedback. Zelinka et al. (2012a) reported that changes in high clouds induce wider ranges of longwave and shortwave cloud feedback across models than those in low clouds. Because the spread in high-cloud-induced longwave and shortwave components is partially compensatory, however, the change in upper clouds brought about by climate change does not contribute much to the uncertainty compared with that in low clouds, although the upper cloud feedback, such as the iris effect (Lindzen et al. 2001) and fixed anvil temperature hypothesis (Hartmann and Larson 2002), remains uncertain. Future changes in the upper cloud fraction because of global warming do not exhibit a robust response. Zelinka and Hartmann (2010) showed that the combined mean of the 15 climate models for the A2 scenario of International Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) caused a decrease in the amount of upper
clouds in the tropics (see also Zelinka et al. 2012b). However, Collins and Satoh (2009) and Satoh et al. (2012a) recently argued that the amount of upper clouds might be increased under global warming conditions by using a global nonhydrostatic model with explicit cloud microphysics processes; such a global atmospheric model is different from climate models (Satoh et al. 2008). Miura et al. (2005) and Wyant et al. (2006) also showed cloud responses different from those obtained by relatively coarse-resolution climate models.

Upper clouds are related to the ice phase of water. Satoh et al. (2012a) showed that the ice water path (IWP) generally decreases under a global warming condition for a different set of model parameters, in spite of the upper cloud spread in the tropics. They argued by using a simple mechanistic model that the reduction of the convective mass flux resulting from global warming decreases the IWP. Such a decrease in IWP is also different from the response analyzed for the relatively coarse-resolution climate models (Zelinka et al. 2012b). It is unknown whether these differences result from the different treatments of convective clouds (i.e., explicit versus parameterized).

The response of tropical cyclone (TC) activity to global warming has been discussed in several previous studies (Knutson et al. 2010; Emanuel et al. 2008; Randall et al. 2007). TCs transport the water substance upward and contribute to the upper cloud covers. With respect to the abovementioned cloud changes, the extent of the contribution of the TCs to the cloud feedback remains a question. Before answering this question, we need to know the extent of the contribution of the clouds associated with TCs to the global ice and liquid water contents. As for precipitation, Rodgers et al. (2000) estimated the contribution of TCs to precipitation over the North Pacific in their active period as 7% by using satellite observation data. However, Kubota and Wang (2009) argued that the contribution of TCs to precipitation over the north of the Philippines ranges from 50% to 60% using stationary data. Kamahori (2012) also estimated to the TC rain contribution using the Tropical Rainfall Measuring Mission (TRMM) satellite data. Such a spread of the contribution of TC comes from the different definitions of the TC domain. These studies indicate that the contribution of TC to the total precipitation in the tropics and hence to
the cloud feedback might not be negligible, although thus far no detailed analysis on how the TCs contribute to the global cloud amount instead of precipitation has been conducted.

Yamada et al. (2010) analyzed the TC changes simulated by the Nonhydrostatic Icosahedral Atmospheric Model (NICAM; Tomita and Satoh 2004; Satoh et al. 2008) and showed that the total TC number decreases and that the strongest TCs intensify under the global warming condition; this is consistent with the findings of several other numerical studies (e.g., Oouchi et al. 2006). Yamada et al. (2010) further showed that the eyewall around TC center becomes deeper under the relatively warm condition. Their simulations resolve the TC eye structure, and the cloud properties of a TC can be analyzed because deep convective clouds are explicitly calculated using a cloud microphysics scheme. In this study, using the global nonhydrostatic model simulation dataset of Yamada et al. (2010), we examine the responses of TC clouds to global warming by evaluating the IWP and liquid water path (LWP) associated with TCs, and the contributions of TC clouds to the tropical clouds.

2. Methodology

We adopt the simulation dataset produced by Yamada et al. (2010) using NICAM (Tomita and Satoh 2004; Satoh et al. 2008). The horizontal mesh interval is 14 km, and a cloud microphysics scheme (Grabowski 1998) is employed without using convective parameterization. Although the resolution is relatively coarser than that
generally used for nonhydrostatic models, the simulated results obtained using the 14-km-mesh NICAM are quantitatively similar to those obtained using the 3.5- or 7-km-mesh NICAM as far as the large-scale organized clouds are concerned (Tomita et al. 2005; Oouchi et al. 2009; Sato et al. 2009; Noda et al. 2010). Yamada et al. (2010) showed that the NICAM simulation under the present condition reproduced TC tracks and intensities comparable to observations. Thus, the dataset is useful for a statistical analysis of TCs. The model settings are the same as those of the CS4MYNN case in Iga et al. (2011) and Satoh et al. (2012a). Iga et al. (2011) reported that the IWP simulated by NICAM with this setting is larger than that estimated from CloudSat, although it is within the range of values in the general circulation model (GCM) simulations that contributed to the IPCC AR4 (Waliser et al. 2009).

The TC detection method used in this study is almost the same as those described in Yamada et al. (2010). The detection method is improved to capture weak TCs and exclude weak disturbances having a short lifetime of less than 36 h and the extratropical cyclone; thus, the number of TCs detected in this study is slightly different from that detected in Yamada et al. (2010). The fifth criterion of Oouchi et al. (2006) (i.e., the magnitude relationship between the maximum wind speed at 850 and 300 hPa) is replaced with the regional mean difference between wind velocity components at 850 and 300 hPa for each grid point. This results in the exclusion of extratropical cyclones from the detected tracks. Revised and additional analyses are shown in the appendix of this paper.

The analysis period is 5 months from June to October for both the present and the future global warming climate conditions; the two experiments are referred to as control (CTL) and global warming (GW), respectively. The number of TCs simulated for the CTL experiment is 49, while that for the GW experiment is 36. Although the integration period is too short to examine the interannual variability, we can extract the statistical properties of TCs and their relations to cloud changes.

3. Results

a. IWP and LWP associated with TC

We first analyze the radial distributions of the IWP and LWP of all simulated TCs. Figures 1 and 2 show the radial distributions of tangentially averaged IWP and LWP, respectively, and the minimum central sea level pressure (SLP) of the individual TCs at their lifetime.
maximum intensity, where the vertical axes represent separate TCs and are ordered by the minimum central SLP. The deepest SLP of the CTL experiment is 902 hPa and that of the GW experiment is 871 hPa, showing the intensification of the deepest TC as described by Yamada et al. (2010). The maximum IWP (or LWP) is located almost at the eyewall in the radial direction. Precisely, the maximum IWP has a larger radius than the maximum LWP, since IWP mainly resides in the upper levels and the eyewalls incline outward as the altitude increases, as shown later in the case of the vertical structure. Figures 1 and 2 indicate that the radius of the eyewall generally increases as TC becomes stronger for a moderately strong TC having a central SLP of more than approximately 920 hPa. For TCs having deeper TC, it seems that the radius of the eyewall again slightly shrinks, while these cases might be only fluctuations. Weatherford and Gray (1988) indicated that, using aircraft reconnaissance data, the radius of the TC eyewall cloud widens outward in proportion to its intensity and shrinks again for intense TC. The relationship between the TC intensities and radius of the eyewall cloud in the NICAM simulation is qualitatively consistent with Weatherford and Gray’s (1988) finding for TCs having more than approximately 920 hPa. However, it may be a major issue that the smaller and intense TCs are more poorly resolved at the 14-km grid spacing. There is a need to be careful when quantitatively discussing the eyewall structure. A comparison of the CTL and GW experiments shows that the radius of the eyewall in the GW experiment increases relative to the radius of the maximum tangential wind. It seems that the increases in the maximum radius are caused from the increase in the top of ice and liquid water clouds, rather than the expansion of the radius of the eyewall cloud. This is confirmed by the composite of the radial distributions of IWP and LWP shown in Fig. 3. The radius is normalized by the radius of the maximum tangential wind. The normalized radius of the maximum IWP is approximately 1.5, which is broader than that of the maximum LWP. The difference in IWP (LWP) of the two experiments shows that IWP (LWP) in the GW experiment is generally larger than that in the CTL experiment, particularly outside the eyewalls. The radius of the maximum value of IWP (LWP) clearly increases: for IWP the radius increases from 0.9 (CTL) to 1.0 (GW), while for LWP it increases from 1.4 (CTL) to 1.5 (GW). According to the Welch t test, increases are statistically significant at the 90% and 85% confidence levels for IWP and LWP, respectively.

Figure 4 shows the vertical structure of the composites of ice water contents (IWC) and liquid water contents (LWC) for the CTL and GW experiments. Embedded are the contours of temperature. As the temperature increases, the vertical extent of the LWC becomes deeper. The maximum value of LWC also increases.
These two effects explain the increase in LWP; the maximum value of LWP changes from 7.0 to 9.6 kg m\(^{-2}\), as shown in Fig. 3. As for IWC, the maximum value is also intensified. As the temperature increases, both the melting level designated by the 0°C contour and the tropopause height (or the top level of the smallest contour of IWC, 0.2 g m\(^{-3}\)) rise, and the vertical extent of IWC does not increase appreciably. Moreover, the air density decreases at relatively high altitudes. These combined effects indicate that the increase in IWP is not as large as that in LWP; the maximum value of IWP changes from 6.0 to 7.1 kg m\(^{-2}\), as shown in Fig. 3.

From Figs. 1–4 we can define the TC areas in which the TC-associated IWP (TC-IWP) and LWP (TC-LWP) are isolated from the environment. We define the TC-IWP (or TC-LWP) as the horizontally averaged IWP (LWP) within 500 km from the TC center at its lifetime maximum intensity. The color shades in Figs. 1 and 2 almost vanish at approximately 500 km; this implies that both IWP and LWP become less than 1 kg m\(^{-2}\) at this radius. Figure 3 shows the values of IWP and LWP at the normalized radius of 5, which is considerably smaller than 500 km, are approximately 10% of the peak values of IWP and LWP, respectively. Figure 5 shows the relation between the minimum SLP and TC-IWP (TC-LWP) for the CTL experiment (black) and the GW experiment (red). Although large scatters are seen in these panels, both TC-IWP and TC-LWP generally increase as the minimum SLP decreases. In particular, for the TCs whose SLP is lower than 970 hPa, the values of IWP clearly increase as the minimum SLP decreases. Such dependence becomes obscure for the weaker TCs with the minimum SLP of more than 1000 hPa; these TCs no longer have clear eyewalls and both IWP and

![Fig. 6. Horizontal distributions of 5-month average total IWP amount (kg m\(^{-2}\)) in (a) the CTL and (b) the GW experiments. (c) Latitudinal distribution. Black and red lines denote the CTL and GW experiments, respectively. (d)–(f) As in (a)–(c), but for the IWP excluding the contribution of TCs. (g) Contribution of the IWP associated with TCs to the total IWP in the CTL experiment (%). (h) As in (g), but for the GW experiment.](image-url)
LWP are the largest near the center of the TCs (Fig. 1). As for the comparison between the CTL and GW experiments, Fig. 5 shows that both TC-IWP and TC-LWP become more abundant for the same intensity of TC under the warmer condition, as inferred from Figs. 1–4.

b. The contributions of TC on the environmental IWP and LWP

The above analysis shows that both TC-IWP and TC-LWP generally increase under the GW condition. However, as Yamada et al. (2010) showed and Figs. 1 and 2 indicate, the total number of TCs decreases under the GW condition. Thus, the next question arises: How are the total IWP and LWP associated with all the TC changes and how do they contribute to the global distributions of IWP and LWP? Figures 6 and 7 show the horizontal distributions of the IWP and the LWP, respectively. In these figures, the total IWP and LWP distributions (Figs. 6a,b and 7a,b), the contributions of TC-IWP and TC-LWP (Figs. 6g,h and 7g,h) to the total IWP/LWP, and the contributions of others cloud systems (Figs. 6d,e and 7d,e) are shown. The latter part (Figs. 6d,e and 7d,e) is obtained by subtracting the TC contributions (Figs. 6g,h and 7g,h) from the total IWP/LWP (Figs. 6a,b and 7a,b).

The values of IWP and LWP are relatively higher over the intertropical convergence zone (ITCZ) and the South Pacific convergence zone (SPCZ) (Figs. 6a and 7a). The IWP and the LWP over the ITCZ further increase in the GW experiment (Figs. 6b and 7b). The horizontal distributions of the differences in IWP and LWP are shown in Figs. 8a,c, respectively. The IWP and the LWP significantly increase over the western North Pacific in the GW experiment. The differences of the zonal mean IWP and LWP between the CTL and GW experiments are shown in Figs. 8b,d, respectively. The tropical average IWP (30°S–30°N) is 87.46 g m⁻² for CTL and 85.05 g m⁻² for GW. IWP decreases by approximately 2.76% from CTL to GW. Such a reduction in IWP is consistent with that found in a previous study.
using NICAM [e.g., CS4MYNN case in Table 2 of Satoh et al. (2012a)]. Figure 8b shows that the IWP over the SPCZ decreases from CTL to GW. Over the ITCZ, the peak of the IWP for GW moves toward the equator. It seems that the increase in the IWP from CTL to GW in the tropics in the Northern Hemisphere (10°–30°N) results from the increase in the IWP over the North Pacific.

The area of large IWP over the North Pacific for GW is closely related to the area of the high TC track density, as shown in Fig. 9, which is reproduced from Yamada et al. (2010). There are differences between Fig. 9 and Fig. 1 of Yamada et al. (2010), which is caused by the above-mentioned improvement of the tracking method. Tables 1 and 2 show the averages of the IWP and the LWP in different domains, respectively; from the left column to the right, the values are averaged over the global domain, the tropical region (30°S–30°N), and each basin (the Indian Ocean, the western Pacific, the eastern Pacific, and the Atlantic) in the tropics in the Northern Hemisphere (0°–30°N). The globally averaged IWP decreases, whereas the globally averaged LWP increases from CTL to GW, irrespective of the inclusion of the TC contributions (the first column on the left). With respect to the tropical average, both the IWP and the LWP decrease (the second column from the left). Both the IWP and the LWP decrease in the Indian Ocean and the Atlantic, while they increase in the eastern and western Pacific. These changes almost correspond to the changes in the number of TCs in each basin (Table 3). In particular, in the Indian Ocean and the Atlantic, the simulation shows a drastic decrease in the number of TCs, corresponding to the noticeable decrease in the IWP and the LWP in these basins.

The abovementioned tables indicate that the changes in the IWP and the LWP are closely related to the changes in the TC tracks. However, even when the contribution of TCs (i.e., TC-IWP or TC-LWP) is subtracted from the total IWP (LWP), the distributions of IWP (LWP) are very similar to those of the total IWP.
and is consistent with the change in the TC tracks shown in Fig. 9. Table 1 shows that the tropical averaged IWP except for the contribution of TCs is 83.15 g m$^{-2}$ in the case of CTL and 80.08 g m$^{-2}$ in the case of GW. It also shows that the contribution of TC-IWP to the total IWP over the tropics is merely 4.93% and 5.84% in the case of CTL and GW, respectively. When the contribution of TC-IWP is excluded, the reduction rate of IWP from CTL to GW becomes 3.70%, which is slightly larger than the total reduction of IWP including the contributions of TC (2.76%). The TC's contribution to LWP and its change because of global warming are similar to the TC's contribution to IWP and its change.

The above results suggest that the TC's contribution to the total IWP/LWP is small and may not be important. However, the previous studies report that a contribution of TCs to the precipitation depends on the area and that it is more than 50% over certain specific islands (Rodgers et al. 2000; Kubota and Wang 2009). Near the TC tracks, the contribution ratio of TC to the total IWP and LWP are more than 50% in this study. Kubota and Wang (2009) report the contribution of the TCs to the precipitation is large over the north of the Philippines in the present climate. In this study, the contribution of TC-IWP and TC-LWP echo those of CTL but for GW. GW/CTL (%) refers to the change rate of the total IWP between CTL and GW. GW/CTL-noTC (%) denotes the same as GW/CTL but excludes the TC contribution to the total IWP. The negative values mean reduction.

### Table 1. Summary of the averaged IWP (g m$^{-2}$) for CTL and GW experiments carried out across the globe, tropics (30°S-30°N), and each tropical ocean basin (the Indian Ocean: 30°–100°E; the western Pacific: 100°E–180°; the eastern Pacific: 180°–90°W; and the Atlantic: 90°W–0°) in the Northern Hemisphere (0°–30°N). CTL-total (g m$^{-2}$), CTL-noTC (g m$^{-2}$), and TC/CTL (%) denote the total IWP, IWP without TC, and the rate of the TC contribution to the total IWP, respectively. GW-total (g m$^{-2}$), GW-noTC (g m$^{-2}$), and TC/GW (%) echo those of CTL but for GW. GW/CTL (%) refers to the change rate of the total IWP between CTL and GW. GW/CTL-noTC (%) denotes the same as GW/CTL but excludes the TC contribution to the total IWP. The negative values mean reduction.

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<th>Globe</th>
<th>Tropics</th>
<th>North Indian Ocean</th>
<th>Northwestern Pacific</th>
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<th>North Atlantic</th>
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<tr>
<td>CTL-total (g m$^{-2}$)</td>
<td>78.70</td>
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<td>127.88</td>
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<td>100.01</td>
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<td>76.44</td>
<td>83.15</td>
<td>121.33</td>
<td>166.78</td>
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<td>TC/CTL (%)</td>
<td>2.88</td>
<td>4.93</td>
<td>5.13</td>
<td>12.11</td>
<td>1.24</td>
<td>4.48</td>
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<td>GW-total (g m$^{-2}$)</td>
<td>77.39</td>
<td>85.05</td>
<td>107.16</td>
<td>230.54</td>
<td>114.67</td>
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<td>GW-noTC (g m$^{-2}$)</td>
<td>74.83</td>
<td>80.08</td>
<td>104.67</td>
<td>191.49</td>
<td>112.49</td>
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<td>TC/GW (%)</td>
<td>3.31</td>
<td>5.84</td>
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<td>16.94</td>
<td>1.90</td>
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<td>GW/CTL (%)</td>
<td>-1.67</td>
<td>-2.76</td>
<td>-16.21</td>
<td>21.48</td>
<td>14.66</td>
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<td>GW/CTL-noTC (%)</td>
<td>-2.10</td>
<td>-3.70</td>
<td>-13.73</td>
<td>14.81</td>
<td>13.89</td>
<td>-11.76</td>
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of each TC generally increase under the global warming condition. The contribution of TC-IWP and TC-LWP is determined not only by the number of TCs but also by the intensity, size, and duration of the TCs.

4. Summary and discussion

A future change in clouds is most ambiguous because of climate sensitivity of the climate model projection studies. Cloud radiative forcing and its change are closely related to cloud properties such as cloud fraction, thickness, cloud sizes, and the amount of ice and liquid water contents. In particular, one needs to understand how IWP and LWP change because of global warming in order to identify the factors responsible for the changes in cloud forcing (Zelinka et al. 2012b).

Satoh et al. (2012a) have argued, using a global non-hydrostatic model (NICAM) with explicit cloud microphysics, that the IWP decreases under a global warming condition and the reduction of convective mass flux under the warming condition mainly causes the decrease in the IWP. Yamada et al. (2010) have shown by analyzing the NICAM experimental data of the CTL and GW experiments that the number of TCs decreases and the stronger TCs intensify because of global warming. By examining the responses of TC clouds, we evaluated the IWP and the LWP associated with TCs (TC-IWP and TC-LWP) and contributions of TC clouds to the global and tropical cloud distributions. The TC-IWP and the TC-LWP increase as the TC intensifies in both CTL and GW experiments. In the case of GW, the radius of the eyewall relative to the radius of the maximum tangential wind, the IWP, and the LWP increase. Consequently, the TC-IWP and the TC-LWP in the case of GW become larger than those in the case of CTL.

The contributions of TC-IWP (TC-LWP) to the tropical averaged IWP (LWP) are 4.93% (3.00%) in the case of CTL and 5.84% (3.69%) in the case of GW (Tables 1, 2). Although these contributions might not be significant, the contribution rate is large over areas with frequent and dense TCs in the cases of both CTL and GW (i.e., north of the Philippines; as shown in Figs. 6g,h and 7g,h). The distribution of the TC contributions of the present-climate experiment (CTL) is consistent with the findings of previous observational studies using precipitation (Rodgers et al. 2000; Kubota and Wang 2009).

In spite of an approximately 30% reduction of cyclogenesis over the globe because of global warming, the contributions of TC-IWP (TC-LWP) to the total IWP (LWP) do not change appreciably. It seems that the increase in TC-IWP (TC-LWP) under the global warming condition compensates for the reduction of the contribution of TCs to total IWP (LWP) resulting from the reduction of cyclogenesis. The contributions of TC-IWP and TC-LWP are determined not only by the number of TCs but also by the intensity, size, and the duration of TCs.

Satoh et al. (2012a) showed that the upward mass flux weakens under the global warming condition; this is consistent with the finding of several other studies (Held and Soden 2006; Vecchi and Soden 2007). Satoh et al. (2012a) also analyzed that the reduction of mass flux is almost explained by the reduction of the area of the upward convective core. The reduction of the area of the upward convective core is consistent with the decrease in the number of TCs under the global warming condition. The present study, however, reveals

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<td>GW/CTL-noTC (%)</td>
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<th>Table 3. TC genesis numbers in just one season from June to October across the globe and each ocean basin. Observations are obtained from IBTrACS WMO (Knapp et al. 2010) version v03r04 for June–October 2004, except for tropical depressions. These data are published on the web (<a href="http://www.ncdc.noaa.gov/oa/ibtracs/">http://www.ncdc.noaa.gov/oa/ibtracs/</a>). The numbers in parentheses are the genesis numbers over the Southern Hemisphere.</th>
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that the contribution of TCs to the reduction of the total IWP and LWP is not large. Thus, only the decrease in the number of TCs does not explain the reduction of the area of the upward convective core and the mass flux. However, Figs. 8 and 9 show that the changes in the TC track almost correspond to the changes in the IWP and LWP distributions. This suggests that the distributions of the IWP and the LWP associated with synoptically organized convective systems including tropical depressions and cloud clusters change similarly to the distributions of the TC-IWP and the TC-LWP; that is, the area of the high TC frequency is the active region of deep convection. Although we speculate that the other types of convective systems than TCs have similar characteristics of and changes in the IWP and the LWP under the global warming condition, it requires further analysis. Now, the next step is to clarify the contributions of the IWP, LWP, and radiative forcing of different types of convective systems with respect to the dependence on cloud characteristics such as cloud size. This requires a statistical analysis carried out using satellite observation data to evaluate numerical data, following the procedure developed by Inoue et al. (2008).

Acknowledgments. The numerical experiments were performed using the Earth Simulator of JAMSTEC under the framework of the KAKUSHIN project funded by the Ministry of Education, Culture, Sports, Science and Technology (MEXT), Japan. The present study was partially supported by Program for Generation of Climate Change Risk Information and Strategic Programs for Innovative Research “Field 3” funded by MEXT. The figures were obtained using the Grid Analysis and Display System or GNUPLOT. We would like to express our sincere gratitude to Drs. K. Oouchi, C. Kodama, and A. T. Noda and the NICAM development members for their helpful discussions and to the anonymous reviewers for their valuable comments. The authors would also like to thank Enago (http://www.enago.jp) for the English language review.

APPENDIX

Verification of Statistical Structures of TCs and Their Changes

The overall behaviors of TCs simulated using the present NICAM setting have been reported by Yamada et al. (2010). However, they have not directly mentioned the lifetime of TCs, which appears to be important for discussing contributions of TC to the total IWP/LWP. Since the study of Yamada et al. (2010), the tracking method has improved. For comparison with observational data, we use the International Best Track Archive for Climate Stewardship (IBTrACS) World Meteorological Organization (WMO) dataset (Knapp et al. 2010) instead of the Unisys Corporation website (Unisys 2013), because IBTrACS includes the dataset of the Unisys Corporation website and the SLP data are available for all ocean basins in IBTrACS. It is important to note that the duration of the outputs of the present NICAM simulations average 90 min for two-dimensional variables and comprise snapshot data for three-dimensional variables.

a. Comparison of the CTL simulation with observations (IBTrACS WMO)

The CTL simulation reproduces a realistic number of the global cyclogenesis frequencies (Table 3). However, there are discrepancies in the regional cyclogenesis between the CTL simulation and observation; for example, an underestimation of the cyclogenesis over the eastern Pacific and an overestimation of the cyclogenesis over the Indian Ocean. The overestimation may be associated with a precipitation bias in the Indian Ocean in this version of NICAM (Oouchi et al. 2009). With respect to the underestimation over the northeastern Pacific, NICAM with the horizontal 14-km mesh might not resolve the smaller-sized TCs that tend to be generated more over the northeastern Pacific than over the other ocean basins (Chavas and Emanuel 2010; Strachan et al. 2013).

Figure A1 shows the probability density functions (PDFs) of the lifetime maximum 10-m wind speed and the lifetime minimum SLP for all TCs in each simulation and the observation. It is important to note that IBTrACS includes information from various agencies that use different averaging periods for the 10-m wind speed (Knapp et al. 2010). In this study, the observed 10-m wind speed averaged for 10 min is converted to the wind speed for an average of 1 min by assuming that the ratio of the 10-m wind speed averaged for 10 min to that averaged for 1 min is 0.88, although this assumption is a source of uncertainty (Knapp et al. 2010).

Figure A1 shows that the CTL simulation reproduces the PDF of the minimum SLP in the observation well but that the PDF of the 10-m wind speed is not similar to that in IBTrACS. It appears that TCs having the strongest wind speed are not simulated, and the TCs with intensities around 50 ms \(^{-1}\) are overrepresented. Differences in the averaging time periods of the wind speed may cause this discrepancy. However, even though an average wind speed for 1 min is used, TCs having the strongest wind speed are not simulated (Satoh et al. 2012b). They reported that the maximum wind speed is underestimated relative to the minimum SLP for intense TC in the NICAM simulation.
compared with the empirical relationship between the SLP and maximum wind speed (Atkinson and Holliday 1977). This bias may be because the horizontal 14-km mesh is coarser to resolve TCs that are strong and small in size. The TC frequency over the eastern Pacific is sensitive to the model settings (e.g., cloud microphysics scheme) in the NICAM simulations (Yamada et al. 2012). The bias of the relationship between the maximum wind speed and SLP is also observed in GCMs (e.g., Satoh et al. 2012b, their Fig. 13; Manganello et al. 2012, their Fig. 6; Murakami and Sugi 2010, their Fig. 1). The improvement of the horizontal grid spacing or cumulus parameterization plays a key role in reducing the bias (Manganello et al. 2012; Murakami et al. 2012b).

The contributions of TC to the total IWP/LWP are thought to depend on time, when TCs are present. Figure A2 shows the relationship between the TC intensity (SLP) and its lifetime. The TC lifetime in the observation is defined as any period in which the maximum 10-m wind speed is larger than 17.5 m s$^{-1}$, while the TC lifetime in the simulation is defined by the duration up to the disappearance of the TC from cyclogenesis. The disappearance of the TC is defined as the first step when one of the thresholds for the tracking method has not been satisfied throughout 36 h. As the central SLP of TCs becomes deeper, its lifetime becomes longer. This tendency is captured well by the NICAM simulation. Figure A2 shows that the simulation reproduces the lifetime of TCs almost comparable to the observations.

b. Change of TC structure from the CTL to the GW experiment

With respect to the effect of global warming, Table 3 clearly shows reductions of cyclogenesis over the Indian Ocean and Atlantic basins. In this simulation, the reduction of the TC frequency over the Indian Ocean is consistent with that of the genesis potential index (GPI; Emanuel and Nolan 2004), and the vertical wind shear contributes to the reduction of the cyclogenesis over the

Fig. A1. PDFs of the (a) maximum 10-m wind speed and (b) minimum SLP at the lifetime maximum intensity of all TCs. The transverse values mean the borderlines of each bin. The blue, black, and red boxes indicate observations and the CTL and GW simulations, respectively. The values labeled on the horizontal axes indicate the upper and lower limits of each bin and are equal to the Saffir–Simpson hurricane scales. The observations are obtained from IBTrACS WMO for June to October 2004, with the exception of tropical depressions.

Fig. A2. Relationship between TC intensity (SLP) and its lifetime. The blue, black, and red marks indicate IBTrACS WMO data and the CTL and GW simulations, respectively. The TC lifetime is defined as the period for which the maximum 1-min averaged 10-m wind speed is more than 17.5 m s$^{-1}$ for IBTrACS WMO and from the cyclogenesis to the passing of the TC for the simulation.
Fig. A3. Composited structures of tropical cyclone for all TCs at their lifetime minimum SLP in CTL and GW simulations. Shown are the radial distributions of the (a) SLP (hPa), (b) precipitation (mm day$^{-1}$), and (c) 10-m tangential and (d) radial wind speed (m s$^{-1}$). The black and red lines denote CTL and GW simulations, respectively. Also shown are the radius–height cross sections of the (e1),(e2) tangential wind speed (m s$^{-1}$); (f1),(f2) radial wind speed (m s$^{-1}$); (g1),(g2) vertical wind speed (cm s$^{-1}$); and (h1),(h2) temperature anomalies (K) for (e1)–(h1) CTL and (e2)–(h2) GW simulations.
TABLE A1. Lists of the mean of the maximum 10-m wind speed (MWS), central SLP, lifetime, and rainfall rate (within 500 km of the TC center) of all individual TCs at their lifetime minimum intensity in each experiment and observation. Observations are obtained from IBTrACS WMO, with the exception of tropical depressions.

<table>
<thead>
<tr>
<th></th>
<th>MWS (m s(^{-1}))</th>
<th>SLP (hPa)</th>
<th>Lifetime (h)</th>
<th>Rainfall rate (mm day(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTL</td>
<td>40.07</td>
<td>959.20</td>
<td>146.08</td>
<td>90.42</td>
</tr>
<tr>
<td>GW</td>
<td>42.56</td>
<td>945.46</td>
<td>135.17</td>
<td>128.83</td>
</tr>
<tr>
<td>Obs</td>
<td>41.93</td>
<td>965.88</td>
<td>128.81</td>
<td>—</td>
</tr>
</tbody>
</table>

North Atlantic (Yamada et al. 2010). In terms of the change of the cyclogenesis frequency over the northwestern Pacific, a slight difference exists between Table 3 and Table 1 of Yamada et al. (2010), because the tracking method is improved, as mentioned above (see section 2).

It should be noted that the revised result shows that the cyclogenesis frequency increases over the northwestern Pacific, which is consistent with the change of GPI in the present experiments. This increase may be considered to be contrary to the results of other GCM studies. However, the regional TC frequency change varies among GCMs (Knutson et al. 2010) and depends on the change of SST because of global warming (Sugi et al. 2009; Murakami et al. 2012a). In addition, our experiments comprise only one boreal summer for each climate condition. Therefore, the TC frequency change over the northwestern Pacific may be variable and may be contrary to that in other GCM studies. In this study, we do not intend to discuss the robust response of the TC genesis frequency but the relationship between the changes of TC frequency and IWP/LWP.

Figure A1 shows that the TC intensity in GW becomes stronger than that in CTL. Figure A2 shows that the lifetime of TC in GW becomes shorter than that in CTL. This suggests that the lifetime shortening because of global warming plays a role in reducing the contribution of TC to the total IWP and LWP.

Figure A3 indicates composited structures of the TC for all TCs at their minimum SLP in the CTL and GW simulations, respectively. Each panel shows the radial distributions of the SLP (Fig. A3a), precipitation (Fig. A3b), 10-m tangential wind speed (Fig. A3c), and radial wind speed (Fig. A3d) and the radius–height cross section of the tangential wind speed (Fig. A3e1,e2), radial wind speed (Figs. A3f1,f2), vertical wind speed (Figs. A3g1,g2), and temperature anomalies (Figs. A3h1,h2). The mean intensities in GW are stronger than those in CTL (Table A1). Consequently, all variables in GW may have larger amplitudes than those in CTL. However, the altitude of the outflow at the upper troposphere (Figs. A3e1,e2,f1,f2) and warm core (Figs. A3h1,h2) increases because of the rises in tropopause height resulting from global warming.

In the present experiments, the TC intensity in the GW simulation increases compared with that in the CTL simulation. However, the TC frequency is reduced and the lifetime of TC is shortened in the GW simulation. Consequently, in the present results, the contributions of the TC to the tropical IWP and LWP do not almost change because of global warming.

REFERENCES


