Response of Tropical Cyclone Activity and Structure to Global Warming in a High-Resolution Global Nonhydrostatic Model

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ABSTRACT

Future changes in tropical cyclone (TC) activity and structure are investigated using the outputs of a 14-km mesh climate simulation. A set of 30-yr simulations was performed under present-day and warmer climate conditions using a nonhydrostatic icosahedral atmospheric model with explicitly calculated convection. The model projected that the global frequency of TCs is reduced by 22.7%, the ratio of intense TCs is increased by 6.6%, and the precipitation rate within 100 km of the TC center increased by 11.8% under warmer climate conditions. These tendencies are consistent with previous studies using a hydrostatic global model with cumulus parameterization.

The responses of vertical and horizontal structures to global warming are investigated for TCs with the same intensity categories. For TCs whose minimum sea level pressure (SLP) reaches less than 980 hPa, the model predicted that tangential wind increases in the outside region of the eyewall. Increases in the tangential wind are related to the elevation of the tropopause caused by global warming. The tropopause rise induces an upward extension of the eyewall, resulting in an increase in latent heating in the upper layers of the inclined eyewall. Thus, SLP is reduced underneath the warmed eyewall regions through hydrostatic adjustment. The altered distribution of SLP enhances tangential winds in the outward region of the eyewall cloud. Hence, this study shows that the horizontal scale of TCs defined by a radius of 12 m s⁻¹ surface wind is projected to increase compared with the same intensity categories for SLP less than 980 hPa.

1. Introduction

Tropical cyclones (TCs) cause severe natural disasters, and the magnitude of damage is related to their frequency, intensity, and horizontal scales. TCs form over warm tropical oceans, and their primary energy source is latent heat from the ocean (Emanuel 2003). It is important that we understand how TCs are influenced by rising sea surface temperature (SST) due to global warming. Broccoli and Manabe (1990) applied a general circulation model (GCM) with a grid spacing of a few hundred kilometers to estimate global warming’s effects on TC formation frequency for the first time. They suggested that GCMs with finer resolutions are more...
appropriate for understanding the relationship between global warming and TC activity. Since their pioneering study, GCMs have been widely used to project future changes in global TC activities under global warming conditions (e.g., Bengtsson et al. 1996; Sugi et al. 2002; McDonald et al. 2005; Oouchi et al. 2006; Sugi et al. 2009; Zhao et al. 2009; Murakami et al. 2012a,b; Camargo 2013; Roberts et al. 2015; Wehner et al. 2015; Knutson et al. 2015). Knutson et al. (2010), Christensen et al. (2013), and Walsh et al. (2016) noted a reduction in global TC frequency but an increase in the ratio of intense TCs to total TCs.

The previous studies suggested that the decrease in TC formation is linked to either a reduction in upward mass flux associated with the weakening of tropical circulation (Sugi et al. 2002, 2012; Sugi and Yoshimura 2012; Walsh et al. 2015) or an increase in the saturation deficit of the middle troposphere (Emanuel et al. 2008). Emanuel (1987) projected that the maximum potential intensity of TCs is enhanced under warmer climate conditions using a theoretical model (Emanuel 1986). Satoh et al. (2015) proposed a mechanism for the relation between these apparently conflicting results. Using outputs of a 14-km mesh global nonhydrostatic model, they analyzed the changes in convective mass flux associated with TCs between experiments under the present-day and warmer climate conditions. They found that the ratio (here called $R_M$) of the convective mass flux associated with TCs to convective mass flux in the tropics remains almost constant between the present-day and warmer climate conditions. They also found that TC-associated convective mass flux increases as a TC intensifies. Therefore, as the total convective mass flux in the tropics decreases associated with the stabilization of the atmosphere (Sugi et al. 2002) as SST becomes warmer, the number of TCs decreases under the constraint that $R_M$ remains unchanged.

Regional changes in TC genesis are more uncertain than global change. Previous studies have predicted different responses of TC genesis frequency to global warming (e.g., Knutson et al. 2010; Christensen et al. 2013; Walsh et al. 2016). This discrepancy appears to arise from the relative SST change distribution in models (Sugi et al. 2009). In addition, TC formation is sensitive to the choice of convective parameterization and dynamical cores in the model (Murakami et al. 2012b; Walsh et al. 2013; Reed et al. 2015). To more reliably predict future TC activities, Murakami et al. (2012a) proposed the use of multiple models, emission scenarios, and SST changes. A global nonhydrostatic model is a new approach to assessing future changes in TC genesis.

Previous studies using observational datasets have indicated that horizontal sizes, such as the radius of gale force winds, or maximum wind speeds differ from TC to TC; they vary during their lifetime and are affected by environmental conditions (Merrill 1984; Weatherford and Gray 1988; Kimball and Mulekar 2004; Chan and Chan 2012; Chavas and Emanuel 2010; Chavas et al. 2015). Because of these observed variabilities, observational trends of TC size are not robust (e.g., there is a nonlinear relationship between size and intensity; Wu et al. 2015) and, to our knowledge thus far, no mechanism has been proposed for future changes in TC size (Kim et al. 2014).

The future changes in the structure of TCs, such as horizontal size, have not been well documented. This is because TCs simulated by GCMs are much weaker and much larger than observed TCs and do not have clear eyewall structures (McDonald et al. 2005; Camargo 2013). Computing advancements to date have enabled us to conduct long-term simulations using finer-resolution GCMs. Such higher-resolution GCMs have reproduced realistic primary circulation and warm-core structures (Murakami et al. 2012b; Manganello et al. 2012, 2014; Roberts et al. 2015).

Previous studies suggest that the change in TC intensity affects a projected change of TC horizontal size. For instance, a contraction of the maximum wind radius was reported by Kanada et al. (2013) using a downscaling technique. They indicated that the contraction is linked to the enhancement of inflow in a boundary layer due to intensification of TCs under warmer climate conditions. In their simulation, the composite mean of minimum sea level pressure (SLP) in warmer climate conditions was 20 hPa lower than that in present-day conditions. Knutson et al. (2015) documented a climatological change in the radius of $12\text{ m s}^{-1}$ surface wind (R12) using their downscaling framework. Their analysis included TCs not only at peak intensity but also when TCs are asymmetrical. They found that the median value of R12 in global TCs was almost unchanged, as the decrease in R12 of TCs in the western Pacific cancels out increases in other basins.

Improved understanding of the influence of TC intensity on horizontal size will be achieved by focusing on lifetime maximum intensity of TCs and categorizing TCs by their intensity. In addition, TC structures differ among individual TCs. Statistical comparisons are required to examine average behaviors of TCs in each intensity category. Recently, multidecadal climate simulations were conducted using a global nonhydrostatic model that resolves TC eyewall structure (Kodama et al. 2015; Satoh et al. 2015, 2017). We use the output of that model to investigate future changes in TC activities and structures, and consider a mechanism for structural change at the same intensity.
The remainder of this paper is organized as follows. Section 2 describes model settings, experimental design, methods for detecting TCs in model output, and observations used in model validation. Section 3 shows future changes in TC activities, precipitation associated with TCs, and horizontal and vertical structures of TCs. Section 4 discusses differences in TC activities between previous and this studies, and a mechanism for changes in TC structure. Section 5 presents summaries and concluding remarks.

2. Methodology and data

a. Experimental design and model setting

In this study we analyzed the outputs of two climate simulations (Kodama et al. 2015; Satoh et al. 2015, 2017). These simulations were conducted for a 30-yr period using the Nonhydrostatic Icosahedral Atmospheric Model (NICAM; Tomita and Satoh 2004; Satoh et al. 2008, 2014). Hereafter, we will refer to the present-day simulation as the control (CTL) run and to the warmer climate simulation as the global warming (GW) run.

Horizontal grid spacing was approximately 14 km, and convection was explicitly calculated using a single-moment bulk cloud microphysics scheme (Tomita 2008) without a convective scheme. Although the 14-km mesh resolution is not fine enough to resolve details of deep convection structures (Miyamoto et al. 2014), the simulated climatology is comparable to other GCMs (Kodama et al. 2015). In addition, a clear eyewall structure of relatively intense TCs is captured using the 14-km mesh resolution (Yamada et al. 2010; Yamada and Satoh 2013; Satoh et al. 2015). Moreover, Manganello et al. (2014) suggested that the lack of downward flow at the center of TCs might be caused by an incompleteness of cumulus parameterization or insufficient horizontal resolution. This suggestion is another motivation to use an explicit calculation of convectons without cumulus parameterization in this study.

The microphysics scheme classified water substances into six categories (water vapor, liquid cloud, ice cloud, rain, snow, and graupel). We used a broadband radiative transfer code called MSTRNX (Sekiguchi and Nakajima 2008) to calculate atmospheric radiation processes. We calculated the boundary layer processes using a second-order turbulence closure scheme (Nakanishi and Niino 2006; Noda et al. 2010). We solved the energy and water exchanges between atmosphere and land using a land surface model called Minimal Advanced Treatments of Surface Interaction and Runoff (MATSIRO; Takata et al. 2003). The atmospheric model was coupled to a slab ocean model. Calculated SSTs were nudged toward an observed SST in the CTL run, and a future SST in the GW run. We used the Hadley Centre SST dataset (Rayner et al. 2003) as the observed SST, and future SST was determined by adding future changes in monthly-averaged SSTs to observed SSTs based on the method of Mizuta et al. (2008). Future changes were estimated from the differences between the averaged SSTs from the periods 1979–2003 and 2075–99 in the 18-model ensemble mean by using outputs of the Special Report on Emissions Scenarios (SRES) A1B of the World Climate Research Program’s (WCRP) phase 3 of the Coupled Model Intercomparison Project (CMIP3; Mehl et al. 2007). The SST difference revealed an El Niño–like SST pattern (Mizuta et al. 2008; Fig. 3b). The emissions scenario of CO₂ concentration followed the SRES A1B scenario. Initial atmospheric conditions in both simulations were taken from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Dee et al. 2011). Both simulations were run for 30 years and seven months. The first seven months in each simulation were regarded as the spinup period and the final 30 years were the analyzed periods of 1979–2008 for the CTL run and 2075–2104 for the GW run.

b. Detection method of tropical cyclone in the simulation

To detect TCs in the simulation, we employed a detection method based on Sugi et al. (2002) and Nakano et al. (2015), which consists of two steps. In the first step, the algorithm searches grid points on which the SLP is 0.5 hPa less than the mean SLP over the surrounding 7° × 7° grid box. These grid points were candidates for the center of the TC. The next step combined these grid points into a TC track, which must satisfy the following criteria:

1) The maximum wind speed at a height of 10 m exceeds 17.5 m s⁻¹, which is searched over the 3° × 3° grid box surrounding the candidate grid for the center of the TC.
2) The maximum relative vorticity at 850 hPa exceeds 1.1 × 10⁻³ s⁻¹.
3) The temperature structure aloft has a marked warm core. That is, the sum of temperature deviations at 300, 500, and 700 hPa exceeds 2 K. Temperature deviation is defined as the difference between the temperature at the candidate grid and the mean temperature over the 7° × 7° grid box surrounding the candidate grid.
4) The wind speed at 850 hPa is higher than at 300 hPa.
5) The duration of each detected storm exceeds 36 h.
6) The TC formed within a limited range of latitudes (45°S–45°N).
These thresholds are independent of ocean basin. The wind speed threshold is objectively determined from the horizontal grid spacing (Walsh et al. 2007). The relative vorticity in criterion 2 was larger than in previous studies and was optimized to match the number of annual global mean TCs in CTL run with the observed number. The genesis position of a TC was defined as the grid point where the abovementioned thresholds were first satisfied.

c. Data

Simulated TCs in the CTL run were compared with best-track datasets compiled by the International Best Track Archive for Climate Stewardship (IBTrACS; Knapp et al. 2010), which comprise data from the World Meteorological Organization Region Specialized Meteorological Centers, Tropical Cyclone Warning Centers, and other national agencies. IBTrACS contains the position and maximum wind speed of cyclones over all ocean basins. Duplicate records from various agencies were consolidated into one record. Thus, these data are useful to evaluate the abilities of models to simulate global TC activity.

The maximum wind speeds in IBTrACS are time-averaged values, with averaging periods dependent on the agencies that furnished their best-track datasets (Knapp and Kruk 2010). We selected the National Hurricane Center (North Atlantic and eastern North Pacific; Jarvinen et al. 1984) and the Joint Typhoon Warning Center (western North Pacific, north Indian Ocean, and Southern Hemisphere; Chu et al. 2002) from the data archived in IBTrACS because both datasets use 1-min average wind speeds.

The best-track datasets include a parameter that characterizes TC nature during its life cycle: tropical depression, TC, and extratropical cyclone. However, this parameter was not completed in all of the historical records. As we are interested in TCs, we defined a tropical depression as a system in which the maximum sustained wind intensity did not exceed 17.5 m s\(^{-1}\) and excluded them from our analyses. Genesis positions of TCs were defined as the grid point where the maximum sustained wind speed first equaled or exceeded 17.5 m s\(^{-1}\).

3. Results

a. Genesis and track

Table 1 lists the number of TC geneses globally (GL), in the Northern and Southern hemispheres (NH and SH), and in six ocean basins: the north Indian Ocean (NI), western North Pacific (WP), eastern North Pacific (EP), North Atlantic (NA), south Indian Ocean (SI), and South Pacific (SP). These six ocean basins are separated by dashed lines in Fig. 1. The annual number of TCs in the CTL run is less in observations (OBS) in the NH and more than OBS in the SH. In the SH, the number of TCs in the CTL run is overestimated in the SI and SP. In the NH, the number of TCs in the CTL run is comparable to that of OBS in the NI and WP, whereas it is overestimated in the EP and underestimated in the NA compared to the OBS.

Figure 1 shows geographical distributions of TC genesis density. The density is defined as the number of TCs generated per year in each 5° x 5° grid box. This shows that the simulated geographical distribution of density (Fig. 1b) is similar to the OBS (Fig. 1a). In each ocean basin, however, positions of local maxima are not completely consistent with the OBS. For instance, the position in the WP shifts eastward compared to the OBS, and the genesis is considerably underestimated over the NA (Fig. 1c).

Figure 2 depicts geographical distributions of TC track density. The model simulates more TCs passing across the date line than in the OBS (Figs. 2a and 2b). This overestimation is associated with more TC geneses over the central Pacific in the model (Fig. 1b). In addition, fewer TCs are seen along the east coast of North
America and in the Gulf of Mexico and Caribbean Sea than in the OBS (Fig. 2c). This underestimation is associated with fewer TC geneses in these regions.

As for the impacts of global warming, globally averaged annual TC genesis is reduced by 22.7% (Table 1), which is generally agreement with previous studies (approximately 5%=30%; Christensen et al. 2013; Walsh et al. 2016). The reduction in the number of TCs is seen in all the ocean basins except for the SP. Figures 1d and 1e illustrate spatial distributions of the genesis density in the GW run and its change relative to the CTL run (GW-CTL), respectively. The genesis density increases near the date line in both hemispheres, and decreases prominently over the EP, especially near the west coast of Central America, although the future change in genesis density has a somewhat noisy distribution. These changes are more clearly seen in Fig. 2e. Figures 2d and 2e show TC track density in the GW run and its future change, respectively, indicating more TCs passing across the date line than in the CTL run near the northeast of Taiwan and islands in the SP: Vanuatu, Fiji, Samoa, and New Caledonia.

To interpret the basin-to-basin variability of TC genesis, Fig. 3 shows geographical distributions of the SST (Fig. 3a), vertical wind shear (Fig. 3c), vertical velocity at 500 hPa (Fig. 3e), and precipitation (Fig. 3g), together with their future change (GW run minus CTL run). The vertical wind shear is defined as the magnitude of
difference between zonal winds at 200 and 850 hPa. These four variables are averaged during boreal summer (July–October) for the NH and austral summer (December–March) for the SH, respectively. Figure 3b shows the future change in SST. In the equatorial region, the SST is increased throughout the central Pacific and the eastern Pacific, which is similar to El Niño pattern over the equatorial Pacific. The aforementioned future changes in TC genesis appear to be analogous to their responses to El Niño. For instance, Krishnamurthy et al. (2016) investigated the relationship between the inhibition of TC activity in the EP and NA and El Niño, revealing that the inhibition during El Niño events is attributable to the increase in the number of days with strong vertical wind shear. The NICAM projected increases in the vertical wind shear over the EP and NA under the warmer climate conditions (Fig. 3d), which might cause decreases in TC genesis. In the SP, Walsh (2015) suggested that TC geneses are increased east of 170°E and are associated with an eastward shift in tropical convection during an El Niño event. The model projected that TC genesis is decreased over the Coral Sea and increased over its east region (Fig. 1e) where the upward motion was increased (Fig. 3f). As for the WP, it is known that the location of TC genesis shifts southeast during El Niño conditions (Wang and Chan 2002; Chia and Ropelewski 2002; Chen et al. 2006). In the GW run, TC genesis and track are increased near the date line (Figs. 1e and 2e) where the vertical wind shear is decreased (Fig. 3d), and the convection is increased (Fig. 3h). In general, regional changes in TC genesis appear to be influenced by the response of large-scale circulation to El Niño–like increases in SST.

The global nonhydrostatic model has projected a significant decrease in the global annual TC frequency, which is qualitatively consistent with findings of previous
studies using conventional GCMs. In the subsequent subsection, we examine a response of TC intensity to global warming.

b. Intensity

Figure 4 shows a scatterplot of central SLPs and maximum sustained wind speeds of TCs (pressure–wind relationship), with an empirical relationship proposed by Atkinson and Holliday (1977). The central SLP for some TCs was not recorded in the IBTrACS dataset, depending on the date and ocean basin. Here, we use the dataset of TC intensity observations from 2001 to 2013 because SLP is available over all ocean basins since 2001. The model reproduces intense TCs with SLP less than 920 hPa as seen in the observation data. However, the maximum wind speed of intense TCs is underestimated compared to the observation data. A similar bias in the pressure–wind relationship has been reported in previous studies (Murakami and Sugi 2010; Manganello et al. 2012; Murakami et al. 2012b; Roberts et al. 2015). This bias seems to be related to an incompleteness of convective schemes or insufficient horizontal grid spacing to resolve pressure gradients. For instance, Satoh et al. (2012) and Ohno et al. (2016) showed that bias in the pressure–wind relationship was reduced by using the output of 7-km NICAM simulations; maximum wind speeds become...
stronger for the same SLP. The pressure–wind relationship is almost unchanged between the CTL and GW runs, which is similar to previous studies (Murakami et al. 2012b; Knutson et al. 2015).

Figure 5 illustrates normalized TC frequencies with respect to the total TC frequencies in each ocean basin (Table 2). TCs are categorized with respect to the minimum central SLP during their life cycles (i.e., lifetime maximum intensity) following Roberts et al. (2015). Intensity categories are defined as follows: TS (SLP > 994 hPa), Cat1 (980 < SLP ≤ 994 hPa), Cat2 (965 < SLP ≤ 980 hPa), Cat3 (945 < SLP ≤ 965 hPa), Cat4 (920 < SLP ≤ 945 hPa), and Cat5 (≤920 hPa). The ocean basin is defined as the location where the TC reaches its lifetime maximum intensity as opposed to its generated location. Although different sample numbers between observation data and the CTL run (Table 2) are used, biases are still seen. Compared to observations in all ocean basins, the model overestimates TCs in Cat1–Cat3 and underestimates TCs in TS, Cat4, and Cat5.

In response to global warming, the genesis rate of Cat4 and Cat5 TCs (i.e., intense TCs) increases over the WP, SI, and SP globally, reflecting changes over the WP, SI, and SP. A clear change in the number of intense TCs is seen in the EP; it seems that intense TCs do not occur there at all in the simulated future climate. TC genesis decreases over the eastern half of the EP and increases over the western half of the ocean basin in the GW run (Fig. 1d). A greater fraction of TCs generated in the EP intrudes into the WP (Fig. 2d), reaching their lifetime maximum intensity over the WP rather than the EP. Consequently, intense TCs are not counted in the EP.

Table 4 lists the mean lifetime minimum SLPs in each ocean basin. The mean SLP deepens over ocean basins (WP, SI, and SP) where the number of Cat4 and Cat5 TCs increased under warmer climate conditions (Figs. 5c,f,g). The results of the global nonhydrostatic model show increases in the ratio of global intense TCs to total TCs and in global TC mean intensity, supporting the findings of previous studies (Christensen et al. 2013).

Christensen et al. (2013) reported an increase in TC associated precipitation. Future precipitation change is an interesting topic in climate research. In the next subsection, we analyze a future change in precipitation using the outputs of the global nonhydrostatic model.

c. Precipitation

Figure 6 shows the radial profiles of azimuthally averaged rainfall rate for TCs, both globally and in six
ocean basins. These profiles are composites of TCs at the time when TCs reach their lifetime maximum intensity. In the global TC composite (Fig. 6a), increases in rainfall rate increases are seen at all radii under warmer climate conditions, although the radius of the peak is unchanged between the CTL and GW runs. The area-mean rainfall rate within a 100-km radius from the TC center increases by 11.8%.

The responses of rainfall profiles to global warming were found to differ among ocean basins (Figs. 6b–g).

**TABLE 2.** The numbers of TCs generated over various ocean regions obtained from the IBTrACS dataset (OBS) and from the control and global warming runs. The ocean basin is defined as a location at which TC reaches its lifetime maximum intensity, instead of its generated location. The numbers in parentheses are the numbers of TC in the South Atlantic, which is included in the global (GL) and Southern Hemisphere (SH) regions. Abbreviations are as in Table 1.

| Basin  | Period    | GL  | NH | SH  | NI  | WP  | EP  | NA  | SI  | SP  |
|--------|-----------|-----|----|-----|-----|-----|-----|-----|-----|-----|
| OBS    | 2001–2013 | 910 (2) | 669 | 241 (2) | 47 | 267 | 173 | 182 | 170 | 69  |
| CTL    | 1979–2008 | 2560 (4) | 1661 | 899 (4) | 174 | 836 | 517 | 134 | 539 | 356 |
| GW     | 2075–2104 | 1979 (7) | 1265 | 714 (7) | 155 | 768 | 263 | 79  | 339 | 368 |
Clear increases in rainfall rates outside of peaks in the CTL run are seen in TC composites in the WP, SI, and SP (Figs. 6c.f.g). These increases are statistically significant. In these three ocean basins, the model projected increases in the ratio of intense TCs to total TCs (Figs. 5c.f.g) and their mean intensities (Table 4). However, in the case of composite TCs in the NI and EP (Figs. 6b,d), rainfall increases are clear around the center of the TCs. These two ocean basins show decreases in the rate of intense TCs (Figs. 5d.e) and their mean intensities (Table 4). The radius of the rainfall rate peak shifts toward the center of the TC only in the EP (Fig. 6d). In the NA, the rainfall profile response is different from those in other ocean basins, as rainfall rate clearly decreases within 100 km in the NA (Fig. 6e).

TC intensity decreased over the NI, EP, and NA. However, the responses of precipitation around the TCs differ between these three ocean basins. This inconsistency seems to be related to TC size. In the subsequent subsection, we consider a change in horizontal TC size due to global warming.

d. Horizontal size

We compared the horizontal size using R12, which is defined as the radius of a 12 m s⁻¹ surface wind. For the simulations, we used the radial profile of azimuthal averaged tangential wind velocity at a height of 10 m ($V_{10}$) to estimate R12. Figure 7 shows variations of R12, both globally and in each ocean basin when TCs reach their lifetime maximum intensity. For TCs in the mature stages, Chavas and Emanuel (2010) estimated the global median value of R12 at 197 km using a satellite dataset. Knutson et al. (2015) showed that the median value is approximately 260 km when sampling TCs axisymmetrically (see their Fig. 5). Although previous studies and the present study used different definitions of R12, the model underestimates the median value (140 km) relative to the observational estimates. However, a clear contrast is captured; median values of R12 in the NI and EP are smaller than those in other ocean basins. This contrast between basins is consistent with observational estimates (Chavas and Emanuel 2010; Knutson et al. 2015).

The mean values of R12 widen over the WP, SI, and SP at a statistically significant level as a result of global warming. This means that TCs with a larger R12 are more frequently generated in warmer climate conditions. Over the NI and EP, the mean values of R12 in the GW run are reduced relative to those in the CTL run at a statistically significant level. Figure 7 shows a smaller increase in the mean value of R12 over the NA than those in the WP, SI, and SP. Although the model projected decreases in TC intensity over the NI, EP, and NA (Fig. 5 and Table 4), the mean values of R12 decreased over the NI and EP, and increased over the NA. This opposite response of horizontal size might contribute to the different responses of precipitation due to global warming.
global warming: rainfall rates increase inside its peak in the NI and EP and decrease around TCs in the NA as shown in Fig. 6.

Increases in R12 over the WP, SI, and SP are concurrent with increases in the proportion of intense TCs to total TCs. In addition, decreases in R12 over the NI and EP are also concurrent with decreases in the proportion of intense TCs to all TCs. Therefore, it is speculated that the simulated change in the horizontal size of TCs is associated with a change in the proportion of intense TCs (except for in the NA).

**e. Radial profile of tangential wind velocity**

Because the horizontal size of a TC varies among individual TCs (Fig. 7) (Merrill 1984; Weatherford and Gray 1988; Kimball and Mulekar 2004; Chan and Chan 2012; Chavas and Emanuel 2010; Chavas et al. 2015; Wu et al. 2015), a careful analysis is required to assess future changes in TC structure. As described in the previous subsection, the change in the mean values of R12 is associated with the change in the proportion of intense TCs to total TCs in the model. As the number...
of intense TCs increase, R12 widens. To elucidate structural changes due to global warming, we categorize the cyclones according to their lifetime maximum intensity (the minimum SLP) in order to examine the change in TC structures of the same categories. In addition, we will focus on the structure of TCs at their lifetime maximum intensities to avoid effects of growth or decay.

Figure 8 shows radial profiles of composite azimuthal mean $V_{T10}$ for each TC intensity category. Table 5 lists the numbers of TCs for each intensity category in the CTL and GW runs. In general, as categories become more intense, the maximum $V_{T10}$ radii become larger. As for global warming response, maximum $V_{T10}$ radii are almost the same in the CTL and GW runs. For almost all categories (Cat1–Cat5 except for TS), $V_{T10}$ increases outside the radius of maximum $V_{T10}$ and decreases inside under warmer climate conditions. The increase is more prominent with stronger intensity (SLP ≥ 980 hPa).

The radial profile of $V_{T10}$ is associated with that of the SLP through the gradient wind balance. The response of Cat4 TCs is larger than that in other categories and is statistically significant. Figure 10 shows radius–height cross sections of vertical, radial, and tangential velocities for composite Cat4 TCs. Maximum vertical velocity radii tilt outward with height (Figs. 10a,b), indicating that their eyewalls slope outward. The maximum vertical velocity of the GW run (about 70 cm s$^{-1}$) is smaller than that of the CTL run (about 80 cm s$^{-1}$). There is a region of slight increase in the vertical velocities in the upper layers near the tropopause and near the outer edge of the updraft (Fig. 10b). Although the maximum vertical velocity decreases in the GW run, convective mass flux associated with TCs increases as shown by Satoh et al. (2015) because the area of upward motion widens. The clear increase in vertical velocity in the upper troposphere is associated with the increase of tropopause height in the GW run. TC eyewalls become taller under warmer climate conditions as reported in the previous study (Yamada et al. 2010). The changes in the upper troposphere are also seen in other variables; radial and tangential velocities ($V_T$) (Figs. 10d,f), temperature anomalies (Fig. 11b), water content (Fig. 11d), and latent heating (Fig. 11f) are all increased. A clear
change in radial velocity is seen in the boundary layer (Figs. 10c,d). Inflow (negative value) significantly weakens within the radius around 100 km. It is speculated that this weakening is associated with a decrease in the pressure gradient force within RSLP.

In general, $V_T$ increases outside the radius of its maximum at all heights under warmer climate conditions (Fig. 10f). The radial profiles of the change in $V_T$ below 8 km are the same as those in $V_{T10}$ (Figs. 8b and 9c). Above 8 km, $V_T$ becomes larger irrespective of the radius. This intensification of $V_T$ is interpreted based on the thermal wind relationship described below. Figures 11a and 11b show composite radius–height cross sections of azimuthal averaged temperature anomalies for the CTL and GW runs, respectively.

### Table 5. The numbers of tropical cyclones for each intensity category.

| Unit | TS | Cat1 | Cat2 | Cat3 | Cat4 | Cat5 |
|------|----|------|------|------|------|------|
| SLP  | hPa| 994 < SLP | 980 < SLP ≤ 994 | 965 < SLP ≤ 980 | 945 < SLP ≤ 965 | 920 < SLP ≤ 945 | SLP ≤ 920 |
| CTL  | No.| 250  | 740  | 696  | 572  | 267  | 35   |
| GW   | No.| 185  | 585  | 474  | 413  | 271  | 51   |
and GW runs, respectively. The temperature anomaly is defined as the deviation from the mean temperature outside a 1500-km radius from the TC center. Radial intervals between isotherms in the GW run are larger than those in the CTL run [e.g., the 4-K isotherm reaches a radius of 200 km at a height of 12 km in the GW run (Fig. 11b) but not in the CTL run (Fig. 11a)]. The radial temperature gradient decreases under warmer climate conditions. As required by the thermal wind relationship, the vertical $V_T$ gradient of the GW run must be smaller than that of the CTL run.

The temperature anomaly near the center of TCs represents a warm core. The intensity of the warm core (i.e., the maximum value of the temperature anomaly at the center of TCs) is almost unchanged (less than 1 K) between the CTL and GW runs. The height of the warm core maximum is elevated from around 9 km (CTL) to around 10 km (GW). Clear warming is seen in the upper troposphere, which is mainly associated with the elevation of tropopause.

Figures 11c and 11d show the radius–height cross sections of water content for the CTL and GW runs, respectively. Below the melting level (designated by the $0^\circ$C isotherm with thin dotted lines in Figs. 10 and 11), water content increases near the outer edge of the eyewall (20–150 km) and decreases near the inner edge. This dipole structure indicates that the eyewall shifts outward under warmer climate conditions. Both the melting level and tropopause height (thin dashed lines in Figs. 10 and 11) are clearly shifted upward. Associated with changes in the melting level and the tropopause, water content increases in the upper troposphere and decreases between altitudes of 12 and 15 km (Fig. 11d). Latent heating is mainly related to condensation heating due to convection. To focus on convective clouds, radius–height cross sections
of latent heating are illustrated in Figs. 10e and 10f. A $1.0 \times 10^{-3}$ K s$^{-1}$ isoline reaches around 13 km in the CTL run and around 14 km in the GW run. An increase in latent heating is significant at radii exceeding 50–150 km, and its maximum is observed at a radius of 84 km (Fig. 11f). This radius almost corresponds to the RSLP (Fig. 9a).

g. Response of surface pressure to an increase in latent heating in the eyewall region

The increase in latent heating warms the surrounding atmosphere when the effect of its heating is larger than that of adiabatic cooling due to updrafts. The latent heating in the GW run was larger than that in the...
CTL run around the outer and upper edges of the eyewall (Fig. 11f). Associated with the increase in latent heating, the temperature anomaly increases near the outer edge of the eyewall (Fig. 11b). This warmer air affects SLP underneath via hydrostatic adjustment (Wang 2009). To elucidate the impact of warmer air on SLP, we estimate changes in SLP associated with warming aloft (Holland 1997), which is formulated as follows:

$$\Delta P_S = \frac{P_S}{T_v(P_S)} \int_{P_T}^{P_S} \Delta T_v d \ln p,$$

where $p$ is the pressure (Pa) and $P_S$ and $P_T$ indicate the SLP (Pa) and pressure (Pa) at the tropopause,
respectively; $T_V$ is the virtual temperature (K) written as

$$T_V = T(1 + 0.61q),$$

where $q$ is the water vapor mixing ratio (kg kg$^{-1}$).

Figure 12 shows the radial profiles of the difference in the hydrostatic change in SLP between composite Cat4 TCs in the CTL and GW runs, together with the difference in simulated SLP, which is the same as the gray line in Fig. 9a. Both lines depict similar profiles, and their radii of maximum pressure decrease at a radius of 84 km. Therefore, it is considered that the vertical extent of the eyewall affects the radial SLP profile. Although one suspects that radiative heating contributes to the SLP change, its contribution is small relative to that of latent heating (figure not shown).

4. Discussion

In this study, we have examined future changes in TC activity and structure using the output from an AMIP-type simulation with a 14-km mesh global nonhydrostatic model, NICAM.

a. TC activity

Previous studies on future projections of TC activity documented decreases in annual global TC number and increases in the ratio of intense TCs to total TCs, mean intensity of TCs, and rainfall rate associated with TCs (Knutson et al. 2010; Christensen et al. 2013; Walsh et al. 2016). Our results are qualitatively consistent with previous studies. However, our results show some discrepancies with previous studies in terms of projected future changes in each ocean basin. For instance, in most GCMs, a decrease in TC frequency in the SH is larger than that in the NH (Walsh et al. 2015). However, in the present study TCs are slightly increased in the SP (although not statistically significant) and their reduction rate in the NH is larger than that in the SH. This opposite response might be attributed to the models used and experimental configurations. Moreover, uncertainty in the tracking scheme complicates the interpretation of differences in responses of TCs to global warming among models (Tory et al. 2013; Walsh et al. 2015, 2016). Coordinated intercomparison experiments with a common setting will be required to decompose the reasons for the discrepancy and improve our understanding (e.g., Walsh et al. 2015; Haarsma et al. 2016).

Future changes in intense TCs (Cat4 and Cat5) should be viewed with caution because the nonhydrostatic model used in this study tends to overestimate the number of TCs for Cat1–Cat3 but underestimate intense TCs (Fig. 5). A similar discrepancy is seen in other models (e.g., Zhao et al. 2009; Knutson et al. 2015). Statistical refinement of TC intensity might be useful for understanding future changes in intense TCs as examined in Zhao and Held (2010) and Sugi et al. (2017). Further study is required to understand why certain models show this common discrepancy.

b. Structural change

Kim et al. (2014) demonstrated that the horizontal size of TCs increases under a scenario with a doubling of carbon dioxide, which is consistent with our results (Fig. 7). Knutson et al. (2015) estimated future changes in the horizontal size of TC (R12). The global-averaged R12 is almost unchanged under warmer climate conditions, whereas the regional-averaged R12 changes and their signs are different among the ocean basins. In this study, the nonhydrostatic model also projected a different response of R12 among the ocean basins and an increase in the mean value of R12 in the WP. This response in the WP is different from that given by Knutson et al. (2015), in which R12 in the WP decreased. This difference might be associated with future changes in the horizontal distribution of the TC track in the WP. Figure 8 of Knutson et al. (2015) showed a large decrease between 20° and 30°N (compared with the equator side), although a zonal shift in track density is seen in our results (Fig. 2d). At high latitudes, the Coriolis force should become large relative to the
equator side; therefore, based on the gradient wind relationship, R12 will be enlarged. It is speculated that this difference yields the opposite responses given by Knutson et al. (2015) and the present study. Because of the large variability of R12 (Fig. 7), the projected change in R12 might also depend on other factors such as changes in environmental conditions and the details of the model settings.

We found a relationship between changes in the vertical and radial structures of TCs. The vertical extent of the eyewall becomes deeper under warmer climate conditions. The deepening of the eyewall alters the SLP radial profile through hydrostatic adjustment, resulting in an increase in tangential wind outside of the RSLP (Fig. 9c) and a widening of the R12.

Wang (2009) discussed the relationship between diabatic heating due to convection and SLP change. He related the heating of an atmospheric column to the change in surface pressure underneath the column using hydrostatic adjustment, and evaluated the effects of total diabatic heating on TC structure and intensity by artificially modifying the heating and cooling rate due to cloud processes in the model. His results revealed that the heating of outer spiral rainbands decreases the surface pressure underneath those rainbands (cooling increases surface pressure), which affects TC structure and intensity. The rising eyewall produces more condensation heat (Fig. 11f), playing a similar role to the outer rainbands in Wang (2009). The concept of hydrostatic adjustment supports the validity of the relationship between changes in vertical and radial structures in this study.

Changes in tangential velocity $V_T$ were not homogeneous in the radial direction, increasing outside RSLP and decreasing inside RSLP (Fig. 9c). Figure 12 shows that RSLP coincides with the estimated hydrostatic change in SLP, which is associated with warming aloft. We have indicated that the radial profile of SLP is altered by an increase in diabatic heating in the upper edge of the eyewall. As the eyewall slopes outward, RSLP is larger than the radius of maximum wind (Figs. 9a,c). However, previous studies using a regional model indicated that the eyewall slope varies even though the model employs a horizontal grid spacing smaller than a few kilometers (Fierro et al. 2009; Gentry and Lackmann 2010; Sun et al. 2013). Finer-resolution models simulate smaller radii of maximum wind than coarser ones, and thus the eyewall slope becomes more upright as the horizontal grid spacing becomes smaller. The magnitude of eyewall slope modulates RSLP position. When the eyewall slope becomes larger, RSLP becomes distant from the radius of maximum wind. In the present study, the nonhydrostatic model employed a horizontal 14-km mesh, which is more than that in higher-resolution models used in previous studies (Fierro et al. 2009; Gentry and Lackmann 2010; Sun et al. 2013). The eyewall slope in this study declines more than the slopes in models with finer resolution. As for observations, Hazelton and Hart (2013) calculated the eyewall slope from radar reflectivity between heights of 2 and 11 km, obtaining a mean slope of 58.5° from the vertical (with standard deviation 5.5°). The simulated mean and standard deviation of the Cat4 TCs were calculated from the maximum radii of water contents between heights of 2 and 11 km; these were 68.5° and 25.4° for the CTL run and 63.7° and 35.5° for the GW run, respectively.

Relative to observations, the simulated eyewall slopes outward to a greater extent (i.e., the location of latent heating becomes far from the center of TCs). This bias might undermine the reliability of our results relating to the relationship between the vertical extension of the eyewall and increases in tangential wind. However, Shapiro and Willoughby (1982) examined not only the sensitivity of the heating location on secondary circulation but also the responses of tangential wind and pressure using the Sawyer–Eliassen model (Eliassen 1951). They showed a decrease in pressure near the radius of maximum wind and an increase in tangential wind outside the radius of maximum wind, despite the radial location of the heating source. Encouragingly, their results show general agreement with our results.

5. Summary and concluding remarks

In this study, we investigated future changes in TC activity and structure using output from long-term and high-resolution simulations (14-km mesh) using a nonhydrostatic model called NICAM. In general, the model simulated the distribution of TC geneses and tracks reasonably well, although TC activity is underestimated over the NA and overestimated over the central Pacific (Figs. 1 and 2). Murakami et al. (2014) evaluated contribution of model biases in simulating the present-day climate to future projection, and suggested that reducing the model biases helps to derive significant signals of future changes. Despite the biases in the present-day climate, the trends in TC activity are consistent with previous studies (Knutson et al. 2010; Christensen et al. 2013; Walsh et al. 2016). The global frequency of TCs is reduced by 22.7%; however, intense TCs categorized as Cat4 and Cat5 increased by 6.6%, and the precipitation rate within 100 km of the TC center increased by 11.8%. Future changes in the frequency of intense TCs should be viewed with caution because, according their intensity, the model has a bias in their frequency distribution (i.e., overestimation of Cat1–Cat3 TCs and underestimation of TS, Cat4, and Cat5 TCs) (Figs. 5 and 6).
Regarding changes in TC structure, the model predicts an increase in $V_r$ outside the outer edge of the eyewall for intense TCs (less than 980 hPa; Fig. 8). The reason why the increase is limited outside the outer edge of the eyewall is linked to a decrease in SLP under the eyewall. The decrease in SLP is related to the elevated tropopause height under a global warming scenario. The top of the eyewall reached the height of the elevated tropopause and increased latent heating at the upper edge of the eyewall (Fig. 11). Increased latent heating warms the atmosphere and decreases the SLP underneath the heating area primarily by hydrostatic adjustment (Wang 2009), which alters the radial pressure gradient. Therefore, tangential wind increases outside the outer edge of the eyewall. As the eyewall slopes outward with height, the decrease in SLP maximizes under the outer edge of the eyewall. Eyewall slope determines the radius of the maximum decrease in SLP and plays a primary role in this mechanism. Compared with observations and higher-resolution regional models, this model tends to simulate a more inclined eyewall slope because of its coarser grid interval. This might undermine the reliability of projected change in tangential wind under warmer climate conditions. However, based on diagnostics of a previous study using the Sawyer–Eliassen model, despite the radial location of heating, tangential wind is increased outside the radius of maximum wind and pressure is decreased. This supports our result of an increase in tangential wind outside the eyewall due to the vertical extent of the eyewall under warmer climate conditions.

The 14-km mesh is insufficient to well represent TC structure. In particular, the simulated eyewall tends to slope more than in observations due to the coarse resolution. Moreover, the structure simulated by the model might be subject to artifacts due to the model resolution. For reducing uncertainties, and understanding limitations, further studies are required by using finer resolution.

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