Chapter 7

SENSITIVITY OF TROPICAL CYCLONES TO LARGE-SCALE ENVIRONMENTS IN A GLOBAL NON-HYDROSTATIC MODEL WITH EXPLICIT CLOUD MICROPHYSICS

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ABSTRACT

Using a global non-hydrostatic model with a horizontal 14 km mesh, we investigated intensities and spatial distributions of tropical cyclones (TCs) and their relationships to large-scale environments. We conducted three cases of boreal summer experiments (June-October) using explicit cloud microphysics schemes without cumulus parameterization. The differences among the three experiments are the implemented physics schemes (i.e., a radiation scheme, a cloud microphysics scheme, a land model, and treatment of the sea surface temperature). One of the experiments is regarded as a control case. Only the radiation scheme used in another case differs from that in the control case in terms of the number of absorption bands. The other case uses a more comprehensive cloud microphysics scheme compared to the control case. The land and ocean processes are also changed in the last case. The radiation scheme in the last case is the same as that in the second case.

We examined the sensitivity of the simulated TC frequencies, tracks, and intensities. The relationship between the maximum wind speed (MWS) and minimum sea level pressure (MSLP) is quite similar among the cases. However, TC frequencies and tracks depend on cases; in terms of the frequency, the last case best reproduces the global observed cyclogenesis number, while the second case reproduces the most realistic intensity histogram of TCs. Comparisons of the relationship between the tropical cyclogenesis and the large-scale environment using a genesis potential index (GPI) show
that the spatial distribution of cyclogenesis is generally consistent with that of the GPI. Among the physical factors that contribute to the GPI, the absolute vorticity, relative humidity, and vertical wind shear are primarily relevant to the difference of the GPI. The change of these variables seems to be associated with convergence of the zonal wind at 850 hPa, and the zonal wind is affected by atmospheric circulation, such as the Walker circulation. Therefore, it is speculated that the changes in atmospheric circulation with the physical schemes play an important role for determining the spatial distribution of tropical cyclogenesis.

1. INTRODUCTION

Model projection of the change of tropical cyclone (TC) activity due to global warming is vigorously discussed [1] and remains a challenging problem and an interesting subject. A major problem in the approach with general circulation models (GCMs) comes from uncertainty in the cumulus parameterizations used in the hydrostatic framework [2]. GCMs were incapable of simulating TCs of realistic intensity under present-day climate conditions [3]. In addition, little consensus among models has been found in the projection of regional TC frequency under greenhouse-warmed conditions.

A reliable projection of regional TC change is sensitive to the pattern of relative sea surface temperature change which dominates tropical circulation and convective activity [4]. For a reliable model projection, it is absolutely imperative to improve the reproducibility of TC in present-day simulations. The sea surface temperature and associated atmospheric circulation are also important to understand TCs. Bengtsson et al. [5] emphasized that it would be misleading to focus only on the sea surface temperature. Vecchi and Soden [6] discussed a relationship between vertical wind shear and TCs over the tropical Atlantic and eastern Pacific under a global warming climate. They show that substantial increases in the tropical vertical wind shear are robust features of experiments that are operated with a suite of coupled ocean-atmosphere models forced by emissions Scenario A1B, and the substantial increases are connected to the model-projected decrease in the Pacific Walker circulation. Regarding the Western Pacific, Yokoi et al. [7] examined a relationship between TC frequency and simulated behavior of monsoon troughs using atmospheric-ocean coupled GCMs from the third phase of the World Climate Research Programme Coupled Model Intercomparison Project. They showed that TC frequency is attributable to the trough migration. In this chapter, we examine whether or not atmospheric circulation affects TCs by replacing the physical scheme. For a more realistic representation of the climate system and more relevant predictions, much higher resolution is a fundamental prerequisite [8].

In this chapter, to avoid uncertainty in the cumulus parameterizations used in the hydrostatic framework [2], we use the Nonhydrostatic ICosahedral Atmospheric Model (NICAM) [9, 10] without any cumulus parameterization. NICAM, a prototype global cloud-resolving model, has exhibited great promise in reproducing cloud-associated disturbances in a series of case studies on Madden-Julian oscillation (MJO) [11] and TCs [12-16]. In addition, the model has successfully captured monsoon-related intraseasonal disturbances [17, 18]. The success of NICAM in simulating various tropical cloud systems forms the basis of this type of TC projection study.
Although computational resources have recently increased, high-resolution global experiments, such as those with a resolution higher than 20 km, are difficult to run frequently. Murakami et al. [19] examined the resolution dependency for the projected intensities, genesis number, and spatial distributions of the genesis frequency of TCs using the Japan Meteorological Agency/Meteorological Research Institute Atmospheric General Circulation Model, which is among the highest resolution models available. They suggest that the highest resolution model performs best at representing the observed intense TC. In addition, their model indicates that differences in the basin-scale annual mean TC genesis number relative to the observation and the spatial distribution of TC genesis frequency were not critically dependent on the resolution. Yamada et al. [16] used NICAM to show that the experiment with a horizontal 14 km mesh reproduced realistic TC tracks and intensities. In this chapter, following [16], we use NICAM with a horizontal 14 km mesh for examining the sensitivity of the spatial distribution, frequency of tropical cyclogenesis, and intensity of TCs to the atmospheric circulation depending on physical schemes.

The structure of this chapter is as follows: in “2. Methodology”, the models and experimental setup and TC tracking methodology are described; in “3. TC genesis, track, and intensity”, simulated TC features with observed TCs are shown. In addition, a comparison of TCs and the large-scale environment is presented in “4. Relation between TC activity and environment section”; “5 Summary” addresses material covered herein, and our future plans are outlined.

2. METHODOLOGY

Using NICAM, we conducted three cases of experiments that include different types of physical schemes. In the present experiments, the horizontal grid interval is approximately 14 km for all simulations. There are 40 vertical levels from the surface up to 38 km, and their vertical intervals increase from 160 m to 2.9 km in height. Hereafter, we refer to the three experiments as CASE1, CASE2, and CASE3. The major model physical schemes are listed in Table 1. CASE1 and CASE2 use an explicit cloud microphysics scheme [20], which includes 3 categories of prognostic variables for water substances (water vapor, air-borne hydrometer, and precipitating hydrometer), and liquid and ice phases are partitioned by temperature-dependent function for both air-borne and precipitating hydrometeors. CASE3 uses a more comprehensive cloud microphysics scheme [21] with 6 categories (water vapor, snow, cloud water, ice, rain, and graupel) as prognostic variables. The difference between CASE1 and CASE2 is only the radiation scheme [22]. The number of absorption bands in CASE1 is 22, and that of CASE2 is 26. CASE3 employs a more sophisticated land model, called MATSIRO [23], while a simpler BUKETS model was implemented in the model configurations of CASE1 and CASE2. In addition, we implement a single layer slab ocean model only in CASE3. In the model, the sea surface temperature is nudged toward the observation. The physical scheme of CASE1 is detailed in Oouchi et al. [13, 18] and Noda et al. [24]. CASE2 is the same as the present-day experiment in Yamada et al. [16]. The physical scheme in CASE3 is the same as that documented in Satoh et al. [17].
Table 1. List of physical schemes in each experiment

<table>
<thead>
<tr>
<th>Physics / Case</th>
<th>CASE1</th>
<th>CASE2</th>
<th>CASE3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radiation</td>
<td>Sekiguchi and Nakajima [22]</td>
<td>Sekiguchi and Nakajima [22]</td>
<td>Sekiguchi and Nakajima [22]</td>
</tr>
<tr>
<td></td>
<td>(number of absorption bands: 22)</td>
<td>(number of absorption bands: 26)</td>
<td>(number of absorption bands: 26)</td>
</tr>
<tr>
<td>Land</td>
<td>BUCKET</td>
<td>BUCKET</td>
<td>Takata et al. [23]</td>
</tr>
<tr>
<td>Ocean</td>
<td>Fixed (weekly updating)</td>
<td>Fixed (weekly updating)</td>
<td>Slab ocean</td>
</tr>
</tbody>
</table>

The three experiments are performed for the same time period. All the simulations started on 1 June 2004 and ran for about 5 months. The initial atmospheric conditions for the simulations are the National Centers for Environmental Prediction (NCEP) Global Tropospheric Analyses. The sea surface temperature was provided by a weekly interpolated National Organization of Atmospheric Agency (NOAA)-Objectively Interpolated (OI) sea surface temperature.

The tracking method follows that of Oouchi et al. [2]. The maximum wind speed threshold (traced at 10-m height) is 17.5 m s\(^{-1}\) for NICAM. Caution is required in determining the threshold, to which TC frequency is quite sensitive. When the horizontal resolution is finer than about 10 km, a threshold of 17.5 m s\(^{-1}\) for the 10-m wind speed is appropriate [25]. TCs in the NICAM simulations are compared with observation (OBS) from the International Best Track Archive for Climate Stewardship (IBTrACS v03r01) provided by the NOAA. To examine the climatology in the simulations, we use NCEP Global Tropospheric Analyses for June to October as OBS.

### 3. TC GENESIS, TRACK AND INTENSITY

Figure 1 shows the spatial distributions of precipitation averaged for five months in each case and an observation which is calculated from the Global Satellite Mapping of Precipitation project (GSMaP) [26-29]. Here, we regard the GSMaP as OBS. The distribution in CASE2 resembles that of CASE1. Elaborating only the radiation scheme does not significantly influence the horizontal distribution of precipitation. CASE1 and CASE2 roughly capture the features of the observed horizontal distribution. However, CASE3 differs from the others. With CASE3 in particular, the precipitation is small near the Philippine Islands. It is noteworthy that the differences in the physical schemes have a large influence on the simulations.

Table 2 lists the frequencies of cyclogenesis observed in 2004, CASE1, CASE2, and CASE3. We follow Oouchi et al. [2] regarding the definition of ocean basins. Based on OBS, cyclogenesis from June to October 2004 is active over the Western and Eastern Pacific and the Atlantic basins. In contrast, cyclogenesis is inactive over the Indian Ocean basin. NICAM simulations reproduce the active cyclogenesis over the Western and Eastern Pacific and the Atlantic basin. Although CASE1 and CASE2 overestimate cyclogenesis over the Indian Ocean basin, CASE3 yields an improvement of the overestimation. Although CASE1 and
CASE3 overestimate cyclogenesis over the Atlantic basin, CASE2 suppresses the overestimation.

![Figure 1](image_url)

Figure 1. Averaged monthly precipitation \(\text{[mm day}^{-1}\text{]}\) from June to October 2004. The OBS is from GSMaP, and CASE1, CASE2, and CASE3 are from each simulation.

<table>
<thead>
<tr>
<th>Case / Basin</th>
<th>IO</th>
<th>WP</th>
<th>EP</th>
<th>AT</th>
<th>Globe</th>
</tr>
</thead>
<tbody>
<tr>
<td>OBS</td>
<td>2</td>
<td>20</td>
<td>10</td>
<td>13</td>
<td>45</td>
</tr>
<tr>
<td>CASE1</td>
<td>8</td>
<td>21</td>
<td>10</td>
<td>21</td>
<td>60</td>
</tr>
<tr>
<td>CASE2</td>
<td>9</td>
<td>22</td>
<td>12</td>
<td>13</td>
<td>56</td>
</tr>
<tr>
<td>CASE3</td>
<td>3</td>
<td>14</td>
<td>12</td>
<td>20</td>
<td>49</td>
</tr>
</tbody>
</table>

Regarding the cyclogenesis distributions in OBS (Figure 2, upper left panel) over the Western Pacific, the cyclogenesis extends northward and southward from 15°N, and, over the Eastern Pacific, it concentrates at some distance from the western coast of Mexico. For the Atlantic, TCs occur over the western Gulf of Mexico, near the east coast of the Florida Peninsula, and from the Caribbean ocean to the western coast of Africa. The NICAM simulations generally reproduce these distributions. However, there are some discrepancies. For CASE1, a TC is generated over a more widespread area than that of the OBS over the Atlantic basin. In contrast, for CASE2, cyclogenesis concentrates in the sea around the western coast of Africa. In addition, more TCs than OBS are generated over the area southeast of the Hawaiian Islands and are not generated at all over the East China Sea in CASE2. Regarding CASE3, the distribution shifts eastward over the Western Pacific basin.
Figure 2. Cyclogenesis density (number per 5° lat.-long. area per 5 months) distribution from June to October. Long dashed line means partitions of ocean basin [2]. The OBS is from IBTrACS v03r01, and CASE1, CASE2, and CASE3 are from each simulation.

To compare the observed tracks of TCs with those in the NICAM simulations, the track densities in the OBS and each case are shown in Figure 3. Over the Western Pacific, the TC moves northwesterly toward the East China Sea and then veers away to the Japan Islands in the OBS. CASE3 reproduces the change in direction to some extent, while many TCs move to Eurasia without the veering that takes place in CASE1 and CASE2. When focusing on the Eastern Pacific, we find that TCs in the simulations, except in CASE1, originate over the central and eastern Pacific Ocean and travel across the dateline; however, no such TC is noted in the OBS. In addition, in CASE3, most TCs originate near the west coast of Mexico (Figure 2, bottom right panel). Nevertheless, TCs travel across the dateline, as shown in Figure 3. This means that the TCs last longer than those in the observations and other simulations. Figure 3 also indicates that, although many TCs are generated near the west coast of Africa and traverse longitudinally over the Atlantic Ocean in CASE2 and CASE3, CASE1 reproduces the track density observed.

Figure 3. The same as Figure 2. However, these panels show Track density (number per 5° lat.-long. area per 5 months).
To illustrate the intensity of TCs, the histograms of the maximum 10-m wind speed (MWS) and the minimum sea level pressure (MSLP) in the life cycle of TCs are shown in Figures 4 and 5, respectively. In the OBS, the most intense TC develops to greater than 70 ms\(^{-1}\), while about 35\% of the TCs develop only to 30 ms\(^{-1}\) and less. In the NICAM simulations, however, the most intense TCs in each case do not develop to greater than 70 ms\(^{-1}\). CASE1 and CASE3 reveal a local maximum of 40 to 50 ms\(^{-1}\), while the OBS and CASE2 have a local maximum of 20 to 30 ms\(^{-1}\). Weak TCs in CASE3, which do not develop to greater than 30 ms\(^{-1}\), are smaller than those in the others.

Figure 4. Rank histograms about the maximum attained 10-m wind speed [ms\(^{-1}\)]. The OBS is from IBTrACS v03r01, and CASE1, CASE2, and CASE3 are from each simulation.

Figure 5. The same as Figure 4, except in the minimum sea level pressure [hPa].
In terms of MSLP, TCs in CASE1 and CASE2, which develop to less than 920 hPa, are more numerous than those in the OBS. In CASE3, there are no intense TCs that develop to less than 920 hPa. In contrast, there are also no weak TCs that maintain MSLP of greater than 1,000 hPa. Moreover, although the OBS show two peaks in 1,000 to 984 hPa and 952 to 936 hPa, the same peaks do not appear in these simulations.

A relationship between MSLP and MWS provides a way to assess the intensity and associated horizontal structure of the simulated TC. The scatter plots of both variables for all the TCs simulated in each case and observations are shown in Figure 6. The fitting line to the scatter plot in each case and the empirical relationship [30] are also shown in Figure 6. A comparison between the fitting lines in each simulation does not yield systematic differences. However, the scatter plots indicate that degrees of attainable TC intensity differ among the simulations. In a comparison of the simulations and the OBS, the maximum 10-m wind speed is less than that in the OBS when the MSLP is under 960 hPa.

Figure 6. Relationships between the minimum sea level pressure (MSLP) [hPa] and the maximum 10-m wind speed (MWS) [ms$^{-1}$] when a tropical cyclone has MSLP in the observations and NICAM simulations; the lines indicate the fitting line to the scatter plot in each case and the empirical relationship between MSLP and MWS [30]. The OBS is from IBTrACS v03r01, and CASE1, CASE2, and CASE3 are from each simulation.

4. RELATION BETWEEN TC ACTIVITY AND ENVIRONMENT

As discussed in the previous section, the intensity and genesis of a TC are sensitive to the model including physical schemes such as a cloud microphysics, a radiation scheme, a land model, and treatment of the sea surface temperature. Hereafter, we examine the factors that cause differences in intensity and genesis with the aid of a Genesis Potential Index (GPI) proposed by Emanuel and Nolan [31] and examined by Camargo et al. [32]. The normalized
GPI distributions in the NCEP Global Tropospheric Analyses and NICAM simulations are presented in Figure 7. Here, we regard NCEP Global Tropospheric Analyses as OBS. Comparing Figure 7 with Figure 2, the distributions of cyclogenesis are generally consistent with high GPI distributions in each simulation and the OBS. Therefore, comparing the change of the GPI and its components among cases is relevant. As CASE1 is a better simulation than the others to the extent of reproducing the distributions of cyclogenesis and the TC track, we regard CASE1 as a control case. The GPI is defined as:

$$GPI = 10^5 \eta^3 \left( \frac{H}{50} \right)^3 \left( \frac{\text{V}_{\text{pot}}}{70} \right)^3 (1 + 0.1V_{\text{shear}})^2,$$

where $\eta$ is the absolute vorticity [s$^{-1}$] at 850 hPa, H is the relative humidity [%] at 700 hPa, $V_{\text{pot}}$ is the potential intensity [ms$^{-1}$] [32, 33], and $V_{\text{shear}}$ is the vertical wind shear [ms$^{-1}$]. Here, to examine the GPI component that contributes most to the GPI change, we divide the GPI into four components, as shown below:

$$Vor = 10^5 \eta^3, \quad RH = \left( \frac{H}{50} \right)^3, \quad PI = \left( \frac{\text{V}_{\text{pot}}}{70} \right)^3, \quad Vsh = (1 + 0.1V_{\text{shear}})^2.$$
Figure 8. The difference in the GPI and its components by replacing the physics scheme, (a) CASE2 - CASE1 in the GPI, (b), (c), (d), and (e) show the contribution of factors in the GPI to the GPI change, which is calculated by taking the difference (CASE2 or CASE3 – CASE1) for only one of the contributing factors with the other factors set to CASE1 (i.e., $\Delta \text{GPI} = (\text{Vor}_{\text{CASE2}} - \text{Vor}_{\text{CASE1}}) \times \text{Vsh}_{\text{CASE1}} \times \text{RH}_{\text{CASE1}} \times \text{PI}_{\text{CASE1}}$, where Vor indicates the term associated with the absolute vorticity at 850 hPa, Vsh indicates the term associated with the vertical wind shear (difference between 200 and 850 hPa), RH indicates the term associated with the relative humidity at 700 hPa, PI indicates the term associated with the Potential Intensity, and subscript indicates a case of experiment), (f) – (j) are the same as (a) – (e) but for CASE3 instead of CASE2. The blue shade is unfavorable for tropical cyclogenesis, and the red shade is favorable.

Over the Indian Ocean, the GPI components in CASE3 are less favorable than those in CASE1. Although it is difficult to determine what factor makes the largest contribution to the change of the GPI between CASE3 and CASE1, the reduction of the GPI is consistent with the reduction of cyclogenesis.

Figures 8a and 8f also show that the GPI in CASE2 and CASE3 is smaller than that of CASE1 over west of the Western Pacific. In CASE3, the cyclogenesis area shifts eastward (Figure 2), and the high-GPI area also shifts eastward over the Western Pacific. The translation in the cyclogenesis area is consistent with the high-GPI area shift. Focusing on the
components of the GPI, we find that Vor and RH are unfavorable for cyclogensis. Precipitation decreases near the Philippine Islands (Figure 1), so it seems that the convection weakens there. Comparing CASE2 with CASE1, although the GPI reduced west of the Western Pacific, the frequency of cyclogensis does not largely change. Focusing on the difference in distribution of GPI components, although Vor (Figure 8b) and RH (Figure 8c) are increased, Vsh (Figure 7, 8d) and PI (Figure 8e) are decreased. We find that the local distribution of reduction of the GPI west of the Western Pacific resembles that of PI in form. Therefore, we suppose that PI is not a significant contribution to the change of cyclogensis over west of the Western Pacific, but it contributes to the change of the GPI in these simulations.

The distribution of cyclogensis in CASE2 and CASE3 is reduced and increased, respectively, from that in CASE1 (Figure 2) near the west coast of Mexico. Focusing on the differences in the GPI (Figures 8a and 7.8b), a contrasting structure is shown; the distribution of the warm and cold colors lies north and south, respectively, in a contrasting manner. When we examine Figure 8, it seems that Vor plays a key role. In addition, Vsh in CASE3 is favorable for TCs over the central Pacific (Figure 8i). Therefore, TCs last longer in CASE3 than in other simulations.

In CASE2 and CASE3, many TCs are generated near the west coast of Africa (Figure 2) over the Atlantic basin. In the OBS, however, cyclogeneses occur not only over the ocean area but also over the western Gulf of Mexico, near the east coast of the Florida Peninsula and the Caribbean Sea. Regarding the change of the GPI components between CASE2 and CASE1 and between CASE3 and CASE1, in particular Vor and RH are elevated near the west coast of Africa (Figures 8b, c and Figures 8g, h). In addition, over the Atlantic Ocean, Vsh in CASE2 and CASE3 is relatively favorable, comparing with that in CASE1 (Figures 8d and 7.8i). Therefore, TCs are not suppressed in CASE2 or CASE3 over the Atlantic Ocean by the vertical wind shear.

With regard to the large-scale environments among the three experiments, the horizontal distribution of the absolute vorticity, the relative humidity, and the vertical wind shear differ from each other. These variables are associated with the vertical and horizontal circulation in the atmosphere. We estimate that the difference of the spatial distributions of the GPI and its components results from change of simulated atmospheric circulation depending on the physical scheme.

Focusing on horizontal circulation, Figure 9 shows the zonal wind at 850 hPa averaged for 5 months in the NCEP Global Tropospheric Analysis and simulations. In CASE1, CASE2, and OBS, we find a westerly wind region from the Indian Ocean to the Philippine Islands. However, the westerly wind does not exist over the Philippine Islands in CASE3. CASE3 differs from CASE1 and CASE2 in cloud microphysics, land model, and treatment of the sea surface temperature; this is such that we cannot determine which scheme is most responsible. Alternatively, it is possible for the change in the westerly wind to result from interaction among replaced physics schemes. At a minimum, replacing physics schemes changes the horizontal circulation, and then the convergence region moves away from the Philippine Islands. Consequently, an updraft region disappears from the area near the Philippine Islands. It is also identifiable in Figure 1. Compared with CASE1 and CASE3, a reduction of precipitation in CASE3 is observed in the area over the Philippine Islands (Figure 1). The differences in physics schemes weaken the convection which produces the precipitation. Basically, updraft disappears or weakens so that water vapor is not supplied
from the lower troposphere. Hence, we suppose that RH in CASE3 becomes lower than that in CASE1 in the area over the Philippine Islands (Figure 8h). In addition, the distribution of the zonal wind changes (Figure 9) so that the distribution of absolute vorticity (Figures 8b and 7.8g) and vertical wind shear (Figures 8d and 7.8i) changes from CASE1 to CASE3. Therefore, we think that the cyclogenesis frequency distribution differs between CASE1 and CASE3 (Figure 2).

Figure 9. Averaged zonal wind speed at 850 hPa from June to October 2004. The blue shade is the easterly wind, and the red shade is the westerly wind. The OBS is from NCEP Global Tropospheric Analyses, and CASE1, CASE2, and CASE3 are from each simulation.

In terms of the intensity, Figures 8e and 7.8j show that PI in CASE2 and CASE3 are globally smaller than that in CASE1. In fact, the intensities of TCs simulated in CASE2 and CASE3 are weaker than those of the TCs simulated in CASE1 (Figures 4 and 7.5). In addition, the reduction in CASE3 is greater than that in CASE2. According to a comparison of the most intense TC in each simulation (Figure 4), that of CASE1 has the highest wind speed among all simulations, and that of CASE2 is stronger than that of CASE3. Therefore, we find that the PI reproduces actual intensity of the simulated TCs.

**SUMMARY**

In this chapter, we compare the frequencies, spatial distributions of the simulated cyclogenoses, and intensities of the simulated TCs in three experiments with those of OBS. The three experiments include different physical schemes such as a cloud microphysics, a radiation scheme, a land model, and treatment of the sea surface temperature. We thereby focus on differences in the physical schemes contributing to the spatial distribution of cyclogenesis and assess which large-scale environmental factors play an important role in those characteristics of TCs. The relationships between a TC and large-scale environments are discussed based on the GPI and its components. In the present experiments, the spatial distribution and frequency of tropical cyclogenesis are particularly sensitive to the physical scheme. In addition, the change of cyclogenesis is generally consistent with the change of the...
GPI when CASE1 is regarded as a control case according to the comparisons of the spatial distribution of cyclogenesis and track densities.

A radiation scheme used in CASE2 is different from that in CASE1. Over the Atlantic basin, its effect is to cause tropical cyclogenesis in CASE2 to be smaller than that in CASE1, and its frequency of cyclogenesis to be closer to the OBS than that in CASE1. In contrast, the spatial distribution of CASE1 is more realistic than that in CASE2 over the Atlantic basin. Cyclogenesis in CASE3 is smaller than that in CASE1 and CASE2 over the Indian Ocean, and its area shifts eastward over the Western Pacific as compared to that in other cases and the OBS.

By comparing spatial distributions of the GPI and its components in CASE2 and CASE3 with those in CASE1 (Figure 8), those spatial distributions are different from those in CASE1. These components of the GPI are associated with the atmospheric circulation, such as convergence of zonal wind at the lower troposphere (Figure 9) and the location of updraft or subsidence in the atmosphere. Therefore, we assume that these changes would result from differences in the atmospheric circulation depending on the physical scheme used. Bengtsson et al. [5] shows that considering the distribution of the relative sea surface temperature and the associated atmospheric circulation is important for understanding a TC. The comparison of the present simulations also indicates that the atmospheric circulation depends on not only the sea surface temperature but also the physical schemes. Therefore, we ought to carefully select a physical scheme in order to reproduce realistic atmospheric circulation.

In terms of the intensity of TCs, MSLP and MWS differ among the cases. For MSLP, NICAM basically reproduces the intensity of a TC compared with the OBSs, while MWS for an intense TC is underestimated. The previous study shows that their experiment also underestimates MWS for intense TCs [5]. The differences in TC intensity are consistent with changes of calculated PI from each simulation. The most intense TC is generated in CASE1. The TC intensity in CASE2 is weaker than that in CASE1 as the radiation scheme is replaced. Moreover, the TC has the weakest maximum intensity in CASE3, while there are a few TCs with MSLP or MWS of more than 1,000 hPa or less than 30 ms$^{-1}$, respectively. The simulated relationships between MWS and MSLP are compared to those obtained from the observations [30]. Figure 6 indicates that MWS for a given MSLP is weaker than that in the OBS for intense TCs. In addition, we examined the relationship in the three simulations. There is a slight difference in the relationship (Fig. 7). The 7 km mesh case [17] is much closer to the empirical relationship than the 14 km mesh cases. However, 14 km-mesh experiment used in this chapter reproduces more realistic intensity than the previous studies [2]. This is because GCRM using 14km has impressive strength at representing tropical convection and the convectively coupled disturbance compared with the conventional GCM [13]. Moreover such a propagating tropical wave influences on not only cyclogenesis but also rapid intensification of TC [34]. Thus, we assume that cyclogenesises and TCs intensity in the present simulations are more realistic.

In a future work, we will clarify how different physical schemes change the atmospheric circulation, such as the Walker circulation, that can influence characteristics of tropical cyclogenesis. The relationship between MWS and MSLP is associated with distance between the center of a TC and the location of MWS. Although we have a limitation on observed data of maximum wind radius, we hope that we can examine the horizontal structure and the maximum wind radius of the simulated TCs to verify whether the simulated TCs are realistic...
or not. We anticipate that findings in the future work will contribute to a reliable projection of TCs in a future global warming climate.

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