Impacts of Ice Phase Processes on Tropical Cyclone Development

Masahiro SAWADA and Toshiki IWASAKI

Department of Geophysics, Graduate School of Science, Tohoku University, Sendai, Japan

(Manuscript received 16 June 2006, in final form 6 May 2007)

Abstract

Idealized cloud-resolving simulations with the horizontal resolution of 2 km are carried out to investigate effects of ice-phase processes on the development and structures of tropical cyclones (TCs). A comparison between cold-rain and warm-rain simulations shows that ice-phase processes delay the TC organization, and decrease the area-averaged kinetic energy. The ice-phase processes also shrink the TC size; for example, the radius of the storm-force wind area (over 25 m s\(^{-1}\)) in the cold-rain simulation is two-thirds of that in the warm-rain one.

The TC evolution depends greatly on strong cooling due to melting and sublimation of snow and graupel around the melting layer outside of the TC eyewall. The cooling reduces the pressure gradient below the melting level, and weakens the inflow toward the TC center. It suppresses the inward transport of high absolute angular momentum (AAM), and decreases the energy conversion rate from the available potential energy to kinetic energy of axisymmetric flows. As a result, the reduction of AAM around the eyewall shrinks the TC eyewall size, and the reduction of energy conversion rate delays the TC organization.

The influence of the terminal fall velocities of snow and graupel is also examined by performing sensitivity experiments, with the horizontal resolution of 5 km. The results show that the increased terminal fall velocity lowers the actual melting altitude, and enhances the TC development through the change in the vertical profile of diabatic heating.

1. Introduction

Cumulus convection plays an important role in tropical cyclone (TC) formation and development (Riehl and Malkus 1961; Yanai 1961). In the course of TC development, individual cumulus convective cells are organized into a cyclonic circulation, and some of the released latent heat is converted into kinetic energy. Such processes are known as the conditional instability of the second kind (CISK), and they have been successfully explained with appropriate parameterization schemes for deep cumulus ensembles (Charney and Eliassen 1964; Yamasaki 1968; Ooyama 1969). In general, cumulus parameterization schemes have several empirical parameters that greatly influence the development, structure, and track of TCs (Iwasaki et al. 1987; Baik et al. 1991). Iwasaki et al. (1987) noted that the TC track is affected by the TC size through the interaction between the sustained wind and the environmental flow.

Ambiguities in empirical parameters of parameterization schemes prevent us from clarifying TC development and improving its numerical prediction. Such arbitrariness in cumulus parameterization can be eliminated by using cloud-resolving models, with proper cloud microphysical parameterization schemes. However, cloud-resolving models require large computing resources. Thus, early attempts of cloud-resolving simulation were made using two-dimensional axisymmetric models. In those attempts, it was shown that ice-phase pro-
cesses delays TC development, but deepens central sea-level pressure (Lord et al. 1984; Willoughby et al. 1984; Lord and Lord 1988). Those studies also indicated that the downdrafts enhanced by melting cooling of snow reduce low-level inflow, and limit vortex-scale ascent. Moreover, numerical simulations with ice-phase processes (cold-rain experiments) tended to have multiple convective rings, and to shrink the eyewall (Fig. 6 and Fig. 8 in Willoughby et al. 1984; Fig. 4 in Lord and Lord 1988). Hausman et al. (2006) remarked that ice-phase processes assist the formation of convective rings through extensive stratiform precipitation with slow terminal fall velocity. The stratiform precipitation delays the development of the TC, weakens the intensity, and enlarges the eyewall in comparison with the cases without ice-phase processes (warm-rain experiment). The differences between cold-rain and warm-rain simulations have been primarily attributed to effects of terminal fall velocity, and secondarily to the latent heat due to ice-phase processes. These results are significantly different from the results of Willoughby et al. (1984) and Lord et al. (1984), in which ice-phase processes cause smaller eyewalls and deeper central sea-level pressure. In addition, the axisymmetric models can not capture three-dimensional structures, including rainbands and vortex Rossby waves, which may affect TC development and structures.

Recently, advancements in computer technology have enabled us to perform three-dimensional cloud-resolving simulations of TC development, in which cloud microphysical processes are explicitly expressed without a cumulus parameterization scheme. Wang (2002), Franklin et al. (2005), Zhu and Zhang (2006), and McFarquhar et al. (2006) investigated effects of ice-phase processes on TC. Wang (2002), based on 5-km mesh numerical simulations under an idealized environment, indicated that melting and evaporative cooling enhances downdrafts and dries the subcloud layer, and it might suppress the TC development. In a simulation of Hurricane Bonnie (1998) using a 4-km mesh model, Zhu and Zhang (2006) showed that latent heating of fusion due to the Bergeron process intensifies the TC, and it formed a symmetric and compact eyewall. Their impacts on TC intensity are different from those found by Wang (2002). They suggested that the differences arise from physical parameterization schemes or environments of TCs. However, the real reason for these differences has not been clarified. Eastin et al. (2005) indicated that 90% of convective vertical velocity cores (defined as $|u| \geq 1 \text{ m s}^{-1}$ for a horizontal extent of at least 0.5 km) are smaller than 4 km in diameter.

Thus, the purpose of this study is to investigate the impacts of ice-phase processes on TC development, intensity, size, and structure, based on a three-dimensional cloud-resolving model with a 2-km mesh. For this purpose, we conduct idealized experiments under a horizontally uniform basic state with a constant Coriolis parameter ($f$-plane), in order to remove various disturbances associated with a realistic environment. Several sensitivity experiments on the change in diabatic heating are also made to clarify the effects of ice-phase processes on TC development, TC size, and structure.

2. Model description and experimental design

A high resolution nonhydrostatic model (JMA-NHM, Saito et al. 2006) is used as a cloud-resolving model in this study. The computational domain is set to an area of 1200 km × 1200 km covered by 2-km mesh or 5-km mesh, depending on the sensitivity experiments. Rayleigh damping is applied within 100 km from lateral boundaries. The model has 42 variable vertical layers from the surface to 25.02 km, with the lowermost layer being 40 m thick. The top and bottom boundaries are assumed to be rigid. The region above 19.26 km is set to a sponge layer that damps vertically propagating gravity waves. Horizontal and vertical advection terms are formulated in fourth-order and second-order flux forms, respectively. To suppress unrealistic numerical noises, flux correction is applied to the advection term (Kato 1998).

The model describes full physics for subgrid scale phenomena and/or diabatic phenomena. The control experiment adopts the double-moment bulk parameterization as cloud microphysics, based on that by Murakami (1990) and Murakami et al. (1994). The mixing ratios of six species (water vapor, cloud water, rainwater,
cloud ice, snow, and graupel) and the number concentrations of three species (cloud ice, snow, and graupel) are treated as prognostic variables in the model. Turbulent closure is formulated by Klemp and Wilhelmson (1978) and Deardorff (1980), with nonlocal mixing effects (Sun and Chang 1986). Surface fluxes over the sea are estimated with the bulk formula by Kondo (1975). Long-wave and short-wave radiation, and cloud-radiation interactions (Sugi et al. 1990) are included.

We examine TC development under an idealized motionless condition, in which the Coriolis parameter $f$ is assumed to be constant at $10^{-5}$N ($f$-plane). The horizontally uniform sea surface temperature is fixed at 303 K. The basic state of the horizontally uniform temperature and humidity profiles is given by averaging the reanalysis data, ERA40, over the subtropical region of the western North Pacific (EQ-25°N, 120°E–160°E) in August for 5 years (1998–2002).

An initial cyclonic vortex is given as follows (Kurihara and Tuleya 1974; Rosenthal 1978; Nasuno and Yamasaki 1997):

$$v = \frac{2\hat{v}(r/r)}{1 + (r/r)^3},$$

where $v$ is the tangential wind, $r$ is the distance from the center of the TC, $\hat{v}$ is 20 m s$^{-1}$, and $r$ is 150 km. The maximum tangential wind reaches 21.17 m s$^{-1}$ at a radius of 119.1 km from the center. The tangential wind decreases linearly from the surface to 10 km. The horizontal and vertical wind speed are assumed to be zero above a height of 10 km.

The control experiment incorporates a realistic cloud microphysical parameterization scheme described above, while in the warm-rain experiment the Kessler-type cloud microphysical parameterization scheme (Kessler 1969) is used. By comparing the control experiment with the warm-rain one, the roles of ice-phase processes in TC development are demonstrated.

In order to clarify the effects of ice-phase processes on TC development through the change in diabatic heating, three sensitivity experiments (no_iceh, no_icec, and adjust_tv) are conducted, as indicated in Table 1. The no_iceh experiment excludes the freezing and deposition heating from the control simulation, and

the no_icec experiment excludes the melting and sublimation cooling. In the adjust_tv experiment, the simulation is identical to the control simulation in which the terminal velocities of snow and graupel are as fast as that of rainwater. These results are described in Section 4.

Table 1. Description of the outline in the sensitivity experiments.

<table>
<thead>
<tr>
<th>name</th>
<th>outline</th>
</tr>
</thead>
<tbody>
<tr>
<td>control</td>
<td>include full ice physics. 2-km mesh or 5-km mesh are used.</td>
</tr>
<tr>
<td>warm</td>
<td>warm-rain process only. 2-km mesh or 5-km mesh are used.</td>
</tr>
<tr>
<td>no_iceh</td>
<td>cold-rain process except that the freezing and deposition heating are excluded with 5-km mesh.</td>
</tr>
<tr>
<td>no_icec</td>
<td>cold-rain process except that the melting and sublimation cooling are excluded with 5-km mesh.</td>
</tr>
<tr>
<td>adjust_tv</td>
<td>cold-rain process in which the terminal fall velocities of snow and graupel are equal to that of rainwater with 5-km mesh.</td>
</tr>
</tbody>
</table>

3. Comparison between the control and warm-rain experiments

3.1 Time evolution and structure of TC

Figure 1 shows impacts of ice-phase processes on the time evolution of simulated TC from the comparison between the control and warm-rain experiments. In the control experiment, the TC starts to deepen at $T = 33$ hours, and it has the maximum tangential wind of 85 m s$^{-1}$ and the minimum sea-level pressure (SLP) of 930 hPa around $T = 60$ hours. An area-averaged precipitation rate within a radius of 200 km from the center of TC (Fig. 1c) gradually increases till $T = 75$ hours, and then it reaches a quasi-steady state. From Fig. 1d, an area-averaged kinetic energy within a radius of 300 km from the center of TC becomes four times stronger at $T = 120$ hours than at the initial state. In the warm-rain experiment without ice-phase processes, the deepening starts about 15 hours earlier than that in the control experiment. The maximum tangential wind exceeds 85 m s$^{-1}$, and the minimum SLP drops to 935 hPa in the warm-rain experiment.
The area-averaged precipitation rate reaches a quasi-steady state rapidly at $T = 27$ hours. After then, the precipitation rate begins to increase gradually at $T = 75$ hours, and it becomes twice higher than that in the control experiment at $T = 102$ hours. The difference in the precipitation rate indicates that convective activity near the eyewall is more vigorous in the warm-rain experiment than in the control one. The area-averaged kinetic energy becomes twice stronger in the warm-rain experiment than in the control one, although the maximum tangential wind is not largely different. The cause for the larger kinetic energy in the warm-rain experiment is examined from the energy conversion rate from the available potential energy (APE) to the kinetic energy of axisymmetric motion (KM). The energy conversion rate $C(\text{APE}, \text{KM})$ (cf., Tuleya and Kurihara 1975) is given by

Fig. 1. Impacts of ice-phase processes on the time evolution of an idealized tropical cyclone. The panels depict (a) minimum sea-level pressure, (b) maximum azimuthally averaged tangential wind, (c) area-averaged precipitation rate within a radius of 200 km from the center of the TC, and (d) area-averaged kinetic energy within a radius of 300 km from the center of the TC; the solid and dashed lines indicate the control experiment including ice-phase processes and the warm-rain experiment, respectively.
\[ C(APE, KM) = \frac{1}{S} \int_{z_0}^{Z_t} \int_{0}^{2\pi} u \cdot \frac{\partial p}{\partial r} r \, dr \, d\lambda \, dz, \]

where \( u, p, \lambda, z, R, Z_t, \) and \( S \) are the radial wind, the pressure, the azimuth, the height, the radius of integrated area (300 km), the top height of calculation domain (25 km), and the integrated area \( (\pi R^2) \), respectively. In the warm-rain experiment, the conversion term \( C(APE, KM) \) averaged from 24 to 120 hours is 50.7 J s\(^{-1}\) m\(^{-2}\), which is about twice larger than that in the control experiment (23.6 J s\(^{-1}\) m\(^{-2}\)). This results in a much larger time tendency of kinetic energy (8.50 J s\(^{-1}\) m\(^{-2}\)), in comparison with 3.54 J s\(^{-1}\) m\(^{-2}\) in the control experiment. The residual [i.e., \( C(APE, KM) \) minus the tendency of kinetic energy] is considered to mainly dissipate through the surface friction (Tuleya and Kurihara 1975). The cause for large difference in the energy conversion rate is examined \( C(APE, KM) \), between the control and warm-rain experiments in Subsection 3.2. The above-mentioned results indicate that ice-phase processes suppress the production of kinetic energy, and delay the timing of TC deepening.

Figure 2 displays the horizontal structures of vertically integrated total water condensates of the control and warm-rain experiments at \( T = 120 \) hours, when the TC almost matures in both experiments. The eyewall clouds are clearly presented in both experiments as the maxima of total water condensates surrounding the TC eye. The radius of the eyewall is about 30 km and 60 km in the control and warm-rain experiments, respectively. This indicates that ice-phase processes shrink the TC eyewall size. The impacts on TC eyewall size in our simulations are much larger than those in the simulations by Wang (2002, see Figs. 7 and 8 of his paper). The difference could be caused by setting of experiments, such as model resolution, physics, and initial environmental conditions. More detailed intercomparisons are desired in our future issues.

Next, we examine how ice-phase processes affect TC eyewall size and structures. Figure 3 exhibits the radial-height cross sections of azimuthally averaged structure at the mature stage between \( T = 108 \) hours and 120 hours. The tangential winds show that the horizontal extent of the strong wind region (\( \geq 50 \) m s\(^{-1}\)) is about 40 km smaller in the control experiment than in the warm-rain one (Figs. 3a and 3b). The control experiment has a larger radial pressure gradient inside of a radius of 60 km. The warm-rain experiment has a larger radial pressure gradient in the outside. Figures 3c and 3d show the azimuthally averaged verti-
cal velocity and absolute angular momentum (AAM). A larger eyewall size, with stronger updrafts, is simulated in the warm-rain experiment than it is in the control one. The AAM, as seen from the center of the TC on the $f$-plane, is defined as

$$AAM = r\bar{v} + \frac{\bar{r}^2}{2},$$

where $\bar{v}$ is the azimuthally averaged tangential wind. In both experiments, the AAM is almost conserved along the axis with eyewall updrafts. The AAM along the eyewall is smaller in the control experiment than it is in the warm-rain one. The control experiment has a weak centrifugal force associated with the small AAM, and the steep pressure gradient near the eyewall. Both can reduce the eyewall size.

The mixing ratio of total water condensates in the control and warm-rain experiments has a peak value at a height of about 9 km and 5 km, respectively (Figs. 3e and 3f). In the control experiment, the terminal fall velocities of snow and graupel are much slower than that of rainwater, so that the ice condensates exist above the melting layer for a long time. From the dynamical point of view, the ice condensates enhance the water loading effect in the control experiment.

As described above, the weak pressure gradi-
ent and large AAM in the warm-rain experiment are favorable for the eyewall size to enlarge. To quantitatively examine the main contributor to determine the eyewall size, we assume a gradient wind balance in the vicinity of the eyewall,

$$\frac{v^2}{r_e} + f v = p_r,$$

where $r_e$ and $p_r$ are the radius of the eyewall, and the radial pressure gradient per unit mass, respectively. In the left hand side of Eq. (4), the Coriolis terms are negligibly smaller than centrifugal force near the eyewall, then we approximate it as,

$$\frac{v^2}{r_e} \approx p_r.$$

The relative angular momentum at the eyewall is substitute, $m_e = r_e v$, for Eq. (5):

$$r_e^3 \approx \frac{m_e^2}{p_r}.$$

The vertically averaged values between 2–8.5 km are applied to Eq. (6), since the radial pressure gradient depends on the height. The averaged AAM value increases by 60% in the warm-rain experiment, in comparison with the control one (Figs. 3c and 3d). However, the averaged pressure gradient force decreases by 15% in the warm-rain experiment, in comparison with the control one (Figs. 3a and 3b). In Eq. (6), the warm-rain experiment increases $(m_e^2)^{1/3}$ by 36%, and $(p_r)^{-1/3}$ by 5%, respectively. These estimations indicate that the difference in eyewall size is mainly attributed to the centrifugal force.

### 3.2 Secondary circulation and diabatic heating

The AAM at the eyewall is almost half in the control experiment, in comparison with that in the warm-rain one. To understand the difference in AAM, the axisymmetric secondary circulation is examined. Figures 4a and 4b show the radial-height cross-sections of azimuthally averaged mass streamfunctions between $T = 108$ hours and 120 hours, in the control and warm-rain experiments. The mass streamfunction $\psi$ (cf., Schubert and Hack 1983) is given by

$$\psi(r,z) = -r \int_0^z \overline{\rho u}(r,z) \, dz,$$

where $\rho$ and $\overline{\rho u}$ are the density and the azimuthally averaged radial component of momentum, respectively. The streamlines indicate the axisymmetric secondary circulation of the TC. The secondary circulation is much weaker in the control experiment than in the warm-rain one. In particular, the inflows found between the top of the boundary layer (about a 1-km height), to the melting level (about a 5.6-km height), are slower in the control experiment.

The influences of the secondary circulation on the AAM transport and its distribution are examined. The azimuthally averaged AAM fluxes, shown in Figs. 4c and 4d, demonstrate that the inward AAM flux, below a 6-km height, is considerably smaller in the control experiment than in the warm-rain one. This is because ice-phase processes make the secondary circulation weaker. Since the surface friction dissipates the AAM in the boundary layer within a short time, the mid-low level inflows (from the top of the boundary layer to the melting level) in the warm-rain experiment transport a large AAM inward, and sustain a high AAM. At a radius of 100 km, the mid-low level inward of the AAM transport is 75% as large as that in the boundary layer. In contrast, the control experiment simulates a weak mid-low level inflows, which leads to the reduction in AAM at the eyewall. In the control experiment, the mid-low level inward AAM transport is 20% as large as that in the boundary layer. The above analysis indicates that the mid-low level inflows play a significant role in the generation of kinetic energy. Therefore, the larger energy conversion rate $C(\text{APE, KM})$ in the warm-rain experiment, described in Subsection 3.1, is brought from three times stronger inflows between 1–5.6 km through the larger pressure gradient than those in the control experiment.

To clarify the impacts of ice-phase processes on mid-low level inflows, the relationship between the secondary circulation and diabatic heating is examined using the thermodynamic equation in cylindrical coordinates without diffusion.

$$\frac{\partial \theta}{\partial t} = -u \frac{\partial \theta}{\partial r} - v \frac{\partial \theta}{\partial \lambda} - \frac{\partial \theta}{\partial z} + Q,$$
where $\theta$ is the potential temperature, $w$ is the vertical velocity, and $Q$ is the diabatic heating. Concentrate on major terms, we have

$$w_q \approx \frac{Q}{\frac{\partial \theta}{\partial z}}. \tag{9}$$

Here, $w_q$ is a vertical velocity derived from the diabatic heating. Substituting (9) into the definition of mass streamfunction $\rho w \equiv \frac{\partial \psi}{\partial r}$ gives

$$\psi_q = \int_0^r \rho w_q dr \approx \int_0^r \rho \frac{Q}{\frac{\partial \theta}{\partial z}} dr. \tag{10}$$

Fig. 4. Secondary circulation of simulated tropical cyclones at their mature stages, azimuthally averaged between $T = 108$ hours and 120 hours, in the (a) control and (b) warm-rain experiments. The contour shows the radial-height cross section of mass streamfunctions, with intervals of $3.0 \times 10^8$ kg s$^{-1}$. The thick dashed line denotes the axis of the eyewall updrafts, defined as the maximum vertical velocity at each height. The bottom two panels display the absolute angular momentum (AAM) with contour intervals of $4.0 \times 10^5$ m$^2$ s$^{-1}$, and AAM flux with vectors in the (c) control and (d) warm-rain experiments.
where $\psi_q$ is a mass streamfunction induced by diabatic heating. Comparing $\psi_q$ derived from the diabatic heating with $\psi$ derived from the vertical velocity, quite similar structures are found except below the boundary layer and at the upper outflow layer. This result indicates that the difference in the secondary circulation is substantially brought from that in diabatic heating.

To explain the impacts of diabatic cooling on the secondary circulation, based on Eq. (10), the azimuthally averaged mass streamfunctions are derived from the diabatic cooling ($\psi_{\text{cooling}}$) between $T = 108$ hours and 120 hours as shown in Fig. 5. In the control experiment, the negative values of $\psi_{\text{cooling}}$ offset $\psi$ by about 50% around the melting layer (see Figs. 5a and 4a). Therefore, the ice-phase processes suppress mid-low level inflows below the melting layer, which reduce the inward AAM transport. The suppressed mid-low level inflows reduce the generation of kinetic energy. In contrast, $\psi_{\text{cooling}}$ around the melting layer is weaker in the warm-rain experiment than in the control one, and hardly offsets $\psi$ (see Figs. 5b and 4b).

Next, influences of ice-phase processes on diabatic heating are examined. Figure 6 shows the vertical profile of azimuthally averaged diabatic heating at a mature stage of the TC between $T = 108$ hours and 120 hours. In the control experiment, the total diabatic heating rates are decomposed into three categories. Figures 6a, 6b and 6c show the sums of condensational heating and evaporative cooling rates ($\text{cond} + \text{evp}$), freezing heating and melting cooling rates ($\text{frz} + \text{mlt}$), and deposition heating and sublimation cooling rates ($\text{dep} + \text{slm}$), respectively. Updrafts mainly cause the large condensational and depositional heating below the melting layer, and in the upper troposphere, respectively (see Figs. 6a and 6c). They also cause the small freezing heating above the melting layer.

Fig. 5. Azimuthally averaged secondary circulation derived from the diabatic cooling due to melting, sublimation and evaporation, between $T = 108$ hours and 120 hours, in the (a) control, and (b) warm-rain experiments. The contour shows the radial-height cross section of mass streamfunction, estimated by Eq. (10) with intervals of $3.0 \times 10^6$ kg s$^{-1}$. See text for details.
layer as well (Fig. 6b). Melting and sublimation cooling spreads in the layer with a depth of a few km below and above the melting layer, respectively (Figs. 6b and 6c). The cooling rate of graupel is four times larger than that of snow around the melting layer. Total heating shown in Fig. 6d has the structure with double peaks, and the maximum value exceeding 80 K h\(^{-1}\) is found in the eyewall updrafts. Another specific feature is the existence of cooling on the outside of the eyewall around the melting layer.

In the warm-rain experiment, the maximum value of condensational heating is about 70 K h\(^{-1}\). An area-averaged total diabatic heating within a radius of 150 km from the TC center is two times larger than that in the control experiment. The larger latent heating is consistent with the more intense precipitation, and sustains the larger kinetic energy in the warm-rain experiment. Below the boundary layer, the accumulated evaporative cooling within a radius of 150 km is about 160% stronger in the warm-rain experiment than that in the control one. The melting and sublimation of snow and graupel largely change the vertical profile of cooling between 4–7 km from that in the warm-rain experiment. The features of each heating rate of the ice-phase processes are similar to those obtained in the numerical simulation of Hurricane Andrew (Zhang et al. 2002).

Since melting and sublimation cooling may enhance convective downdrafts, and form cold pools, as indicated by Wang (2002), the occurrence frequency of strong downdrafts within a radius of 400 km from the TC center and the simulated cold pool are examined for \(T = 108-120\) hours. In the control experiment, the occurrence of downdrafts stronger than \(-1\) m s\(^{-1}\) at a height of 0.93 km decreases by about 50% in comparison with the warm-rain experiment. The cold-pool area, defined as the potential temperature anomaly less than \(-1\) K from the azimuthally averaged potential temperature, is also reduced by about 45% in the control experiment, because evaporative cooling is much

The apparent coincidence of the supersaturation for water and subsaturation for ice (Figs. 6a and 6c) is due to the azimuthal average.
weaker than that in the warm-rain experiment. The above-mentioned results show that the melting and sublimation cooling does not significantly enhance convective downdrafts and cold-pool formation.

4. Sensitivity experiments

Major differences between the cloud physics in the control and the warm-rain experiments are the diabatic heating due to phase change and the sedimentation of condensates. To examine these two processes on TC development, three sensitivity experiments are conducted: In the no_iceh and no_icec experiment, the freezing and deposition heating, and the melting and sublimation cooling are removed from the control experiment, respectively. The adjust_tv experiment includes the same ice-phase processes as those in the control experiment, while terminal fall velocities of snow and graupel are adjusted to that of rainwater. To save the computational resource, a 5-km mesh model is used in place of a 2-km mesh model in these sensitivity experiments. The influence of horizontal resolution is discussed in the next Subsection 4.1.

4.1 Influence of horizontal resolution

Since the cumulus convection simulated in the model is sensitive to the horizontal resolution (Weisman et al. 1997), the reproducibility of the low resolution experiments is examined through a comparison between a 5-km mesh and the 2-km mesh experiments. Figure 7 shows the time evolution of the minimum SLP and area-averaged kinetic energy in the no_iceh (thin dashed), no_icec (thin solid), and adjust_tv experiments (dashed-dotted); the thick solid, and thick dashed lines are the same as those in Figs. 1a and 1d. In the control and warm-rain experiments, the evolution and intensity of minimum SLP, and area-averaged kinetic energy have little difference between the 5-km mesh (see Figs. 7a and 7b) and the 2-km mesh experiment (see Figs. 1a and 1d). The 5-km mesh experiments are able to capture the TC evolution, such as the timing of TC deepening and maximum intensity. The eyewall sizes
in the 5-km mesh experiments are also equal to those in the 2-km mesh experiments.

4.2 Effects of diabatic heating/cooling

The no_iceh experiment does not change the timing of TC deepening from the results of the control experiment (Fig. 7). The minimum SLP reaches 940 hPa, which is about 10 hPa larger than that in the control experiment. The area-averaged kinetic energy is 10% smaller in the no_iceh experiment than in the control one.

To illustrate the effects of ice-phase processes on the axisymmetric secondary circulation of the TC, Figs. 8 and 9 show radial-height cross sections of azimuthally averaged mass streamfunction, and diabatic heating between $T = 108$ hours and 120 hours in the control, warm-rain, no_iceh, no_icec, and adjust_tv experiments. The no_iceh experiment slightly weakens the secondary circulation in the upper troposphere, in comparison with the control experiment, but this weakening does not make any significant influence on the AAM transport, and the eyewall size. Therefore, the freezing and deposition heating has little impacts on the cumulus organization and TC structures.

In the no_iceh experiment, the azimuthally averaged diabatic heating has a similar structure to that in the control experiment (see Fig. 9). The maximum heating rate at a height of 2 km is about 10 K smaller in the no_iceh experiment than in the control one, and the area-averaged diabatic heating within a radius of 150 km decreases by 10–70% above a height of 9 km. This leads to the reduction of accumulated upward mass flux within a radius of 150 km by about 20%; hence, the TC development is suppressed by weakened eyewall updrafts associated with the decrease of diabatic heating.

The no_iceh experiment shows that the TC starts to deepen 18 hours earlier, it develops slower and the minimum SLP is 20 hPa weaker than the control experiment. The area-averaged kinetic energy reaches almost a steady state at $T = 84$ hours, and it becomes about 80% larger than that in the control experiment. Therefore, the TC development is greatly sensitive to the melting and sublimation cooling.

The cooling has also large impacts on the axisymmetric secondary circulation and the TC size (Fig. 8d). The secondary circulation is much stronger in the no_icec experiment, than in the control one, especially below the melting level. This is consistent with the discussions in Subsection 3.2. Without the cooling, the condensation heating increases the mass streamfunction. The mid-low level inflow brings a high AAM air into the eyewall, which increases AAM by 60% at the eyewall in the no_icec experiment, in comparison with the control one. The large AAM causes such large eyewall size, as found in the warm-rain experiment. The intensified inflow below the melting level is sustained by the diabatic heating. The area-averaged diabatic heating, within a radius of 150 km is 30% larger in the no_icec experiment than in the control one. In the no_icec experiment, the diabatic cooling in the melting layer almost disappears because of the exclusion of melting and sublimation (Fig. 9d), and the mass streamfunction (Fig. 8d) has similar characteristics to that in the warm-rain experiment (Fig. 8b). Therefore, the melting and sublimation cooling mainly causes the difference of TC organization, between the control and warm-rain experiments. This cooling, however, is insufficient to perfectly explain the impacts of ice-phase processes on the TC. In the no_icec experiment, the TC deepening also starts earlier, but the maximum intensity and the area-averaged kinetic energy are weaker in comparison with the results in the warm-rain experiment.

4.3 Effects of the terminal fall velocity

Previous works have shown that the terminal fall velocity of condensates influences the TC evolution and structures (Lord and Lord 1988; Wang 2002; Hausman et al. 2006; McFarquhar et al. 2006). Hausman et al. (2006) remarked, from the simulation results with an axisymmetric model, that the slow fall-speed enhances convective downdrafts, which suppresses the TC development. On the contrary, McFarquhar et al. (2006), by using a three-dimensional model, showed that different terminal fall velocities of graupel do not clearly affect the occurrence frequency of strong downdrafts. To understand the impacts of terminal fall velocity on the formation of convectively strong downdrafts, a sensitivity experiment is performed regarding terminal fall velocities of
Fig. 8. Same as Figs. 4a and 4b but for (a) control (thick solid), (b) warm (thick dashed), (c) no_iceh (thin dashed), (d) no_icec (thin solid), and (e) adjust_tv (dashed-dotted). It should be noted that the figures show the results of 5-km mesh experiments.
condensates. The terminal fall velocity is assumed to be \( V(D) = a_0 D^{b_0} \), where \( a_0 \) and \( b_0 \) are determined by the condensate property, such as the shape and density; \( D \) is the particle diameter of the condensate that is diagnosed in the scheme. \( a_0 \) and \( b_0 \) for snow and graupel is replaced with that for rainwater, so that for condensates with the same \( D \) are the terminal fall velocities of snow and graupel equal to that of rainwater. In general, the terminal fall velocity of rainwater is several times faster than those of snow and graupel.

When the terminal fall velocities of snow and graupel are made faster, the diabatic heating profile is modified through the cooling processes. In the adjust_tv experiment, the TC deepening starts about 12 hours earlier, and the kinetic energy becomes 50% larger than those in the control one. The fast terminal fall velocities have little influence on the frequency of convectively strong downdrafts, while they intensify the secondary circulation (Fig. 8e). The intensified secondary circulation increases the eyewall size and storm-force wind area through the high AAM transport. In order to clarify the effects of terminal fall velocity on the heating and cooling profiles, the diabatic heating and cooling profiles are examined from Fig. 9. The fast terminal fall velocities of snow and graupel reduce the cooling region (\( < -1 \text{ K h}^{-1} \)) around the melting layer. The area-averaged diabatic cooling rate, within a radius of 150 km, decreases by about 70% between 4–7 km in the adjust_tv experiment (Fig. 9e), in comparison with that in the control one. In contrast, below a height of 2 km evaporative cooling in the adjust_tv experiment is distributed very similar to that in the control experiment. In the adjust_tv experiment, the diabatic cooling decreases around the melting layer, because of considerable amounts of snow and graupel particles fall into the sea surface without melting or sublimation. This is the reason why the impacts of fast terminal fall velocity are similar to those of melting and sublimation cooling that is found from the no_icec experiment.

5. Conclusion

In this study, three-dimensional cloud-resolving simulations of tropical cyclones (TCs) under the idealized environments are carried out to clarify the effects of ice-phase processes. Comparisons between the cold-rain (control) and warm-rain experiments show that ice-phase processes have a significant influence on the development, intensity, and structure of TCs. Ice-phase processes considerably delay the timing of TC deepening. Moreover, ice-
phase processes reduce by 50% the production of the kinetic energy from the available potential energy, in comparison with the warm-rain experiment, although they hardly change the maximum tangential wind.

In ice-phase processes, the TC development is very sensitive to the melting and sublimation cooling generated around the melting layer. This is ascertained from the secondary circulation that is much weaker in the control experiment, than in the warm-rain one. According to the relationship between radial mass stream-function and diabatic heating that is derived from the thermodynamic equation, the melting and sublimation cooling could suppress the secondary circulation, especially the inflow into the layer between the top of the boundary layer (about a 1-km height) and the melting level (about a 5.6-km height). This leads to the weaker inward transport of high absolute angular momentum (AAM) below the melting layer, and to the lower energy conversion rate from the available potential energy to the kinetic energy of axisymmetric flows. Thus, the melting and sublimation cooling around the melting layer significantly decreases the generation of kinetic energy, and consequently delays the start time of TC deepening. From the viewpoint of CISK, the present experiments show that ice-phase processes have a negative impact on TC organization.

Ice-phase processes also reduce the TC eye-wall size by about a half, and narrow the radius of the storm-force wind area (≥25 m s⁻¹) by 65%, in comparison with the warm-rain experiment. The shrunken TC size is closely related to the smaller AAM. The horizontal scale of TC is an important characteristic for the TC movement, because the TC track prediction is sensitive to its size (Iwasaki et al. 1987).

To examine the diabatic heating due to phase change, and the sedimentation of condensates, these three sensitivity experiments are performed. When the melting and sublimation cooling is neglected, the TC starts to deepen earlier, it forms a larger eyewall size accompanied with a stronger secondary circulation. When the freezing and deposition heating is neglected, the secondary circulation is slightly suppressed in the upper layer, so that the kinetic energy of the TC is reduced. However, it hardly changes the TC size and its eyewall size. When the terminal fall velocities of snow and graupel are adjusted to that of rainwater, the TC development is delayed, and the TC size is reduced. This is because some of the ice particles reach the sea surface without melting and sublimation because of the increased fall speed. In other words, slow terminal fall velocities of ice condensates suppress the TC development, this is similar to the melting and sublimation cooling.

Concerning the effects of ice-phase processes, it is still not clear to what degree the present interpretation is generalized. Our results, using 2-km mesh experiments, are somewhat different from those obtained by Wang (2002), using 5-km mesh experiments. Many uncertainties are in cloud microphysics. For realistic TC simulations, the cloud microphysics parameters must be optimized through intensive validation with observations. This will be more important in the future, when cloud-resolving models are utilized for both predictions of the TC intensity and track. There are many processes that affect the actual TC development. The relative importance of ice-phase processes should be studied on realistic situations.

Acknowledgments

The authers are grateful to two anonymous reviewers and the editor, Dr. Kato, for their critical and constructive comments, which help to improve the manuscript. The simulations were conducted using the Japan Meteorological Agency Nonhydrostatic Model (JMA-NHM) developed by the Meteorological Research Institute and the Numerical Prediction of the Japan Meteorological Agency. Part of the numerical experimental results in this work were obtained using the supercomputing resources (SX-7) at the Information Synergy Center, Tohoku University.

References


