Impacts of Evaporation of Rainwater on Tropical Cyclone Structure and Intensity—A Revisit

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ABSTRACT

The impact of evaporation of rainwater on tropical cyclone (TC) intensity and structure is revisited in this study. Evaporative cooling can result in strong downdrafts and produce low–equivalent potential temperature air in the inflow boundary layer, particularly in the region outside the eyewall, significantly suppressing eyewall convection and reducing the final intensity of a TC. Different from earlier findings, results from this study show that outer rainbands still form but are short lived in the absence of evaporation. Evaporation of rainwater is shown to facilitate the formation of outer rainbands indirectly by reducing the cooling due to melting of ice particles outside the inner core, not by the cold-pool dynamics, as previously believed. Only exclusion of evaporation in the eyewall region or the rapid filamentation zone has a very weak effect on the inner-core size change of a TC, whereas how evaporation in the outer core affects the inner-core size depends on how active the inner rainbands are. More (less) active inner rainbands may lead to an increase (a decrease) in the inner-core size.

1. Introduction

In the presence of a variety of degrees of convection, complicated cloud microphysics are critical to the development and maintenance of a tropical cyclone (TC). One of the vital effects of cloud microphysical processes on TC structure and intensity is related to latent heating and cooling due to conversion between water categories. It has been long recognized that latent heating from condensation considerably governs the development and maintenance of a TC (Riehl and Malkus 1961; Yanai 1961; Shapiro and Willoughby 1982; Möller and Shapiro 2002; Bui et al. 2009; Wang 2009; Fudeyasu and Wang 2011; Li et al. 2014). Regardless of the causes of cooling, effects of diabatic cooling on changes in TC structure and intensity have recently received considerable attention. Wang (2009) evaluated how the evaporative cooling in outer spiral rainbands affects TC structure and intensity using the cloud-resolving TC model TCM4 and found that cooling in outer rainbands tends to maintain both the intensity and inner-core compactness of a TC from the hydrostatic adjustment perspective. Diabatic cooling in the rapid filamentation zone where inner rainbands are healthy also has considerable influences on TC structure and intensity (Wang 2008a; Li and Wang 2012b). More recently, Li et al. (2014) demonstrated that diabatic cooling in the rapid filamentation zone substantially limits both the storm intensity and the inner-core size and is unfavorable for the development of an annular-hurricane.
Two major sources of cooling associated with conversion of water categories are evaporation of rainwater and melting of cloud ice, snow, and graupel. Evaporation of raindrops (i.e., conversion from rainwater into water vapor) can occur in different regions of a TC. Many previous studies have evaluated the impacts of evaporation and the associated evaporative cooling on TC structure, intensity, and motion.

Since direct and accurate measurement of evaporation and the associated diabatic cooling in a TC are greatly arduous, various numerical simulations have been commonly used to investigate how evaporation affects TCs. Axisymmetric models were employed in several earlier studies. Yamasaki (1983) found that the boundary layer cold pools formed because evaporative cooling strengthens mesoscale convective activities and thereby the secondary circulation in the simulated TC. Willoughby et al. (1984) and Lord et al. (1984) showed that downdrafts driven by evaporative cooling favor the occurrence of convective rings outside of the eyewall, which slowed down the vortex intensification. Bister (2001) evaluated the effect of evaporation on intensification of TCs at different latitudes by artificially omitting the evaporative cooling in an axisymmetric model but allowing rainwater to convert into water vapor. She found that the onset of rapid intensification of the simulated TC at lower latitudes was as early as that at higher latitudes if evaporative cooling was neglected, whereas the onset at lower latitudes was much earlier than that at higher latitudes when evaporative cooling was included. Much smaller inertial stability in the outer core and greater Ekman pumping in lower-latitude TCs favor the occurrence of inner-core convection and hence the earlier onset of rapid intensification (Bister 2001). Based on the results of a cloud-resolving axisymmetric model, Frisius and Hasselbeck (2009) demonstrated that evaporative cooling decelerates the storm development, reduces the final intensity, and poses intensity fluctuations after the TC becomes matured.

Three-dimensional numerical models have also been employed to further evaluate the effect of evaporation on TC structure and intensity. Wang (2002) carried out several sensitivity experiments for idealized TCs with the hydrostatic TC model TCM3 and found that evaporation of raindrops together with melting of snow and graupel caused strong downdrafts, which were arguably responsible for the formation of spiral rainbands in the outer-core region. Such downdrafts limit the intensification rate and the final intensity of the simulated TC, as found in earlier axisymmetric models. Zhu and Zhang (2006) explored how varying cloud microphysical processes sway the intensity, precipitation, and inner-core structure of Hurricane Bonnie (1998). The experiment with evaporation turned off produced the deepest storm among their experiments with the smallest radius of maximum wind, a wider eyewall, and the strongest ascending motion in the eyewall. These features were also confirmed in Pattnaik and Krishnamurti (2007) and Frisius and Hasselbeck (2009). More recently, Sawada and Iwasaki (2010a,b) further studied the impacts of evaporation on TC evolution and structure. They found that convective subsidence forced by evaporative cooling can diminish the enthalpy in the boundary layer and slow down the storm development. They also showed that the formation of outer rainbands was closely related to cold-pool dynamics in the boundary layer. Namely, convergence between the upstream pre-existing cold pool due to evaporative cooling and low-level inflow contributed to nascent convective cells at the upstream end of the cold pool. Later, those convective cells successively organized into spiral-shaped rainbands. The development of outer rainbands expands the TC size through the enhanced secondary circulation as a result of condensational heating. They thus suggested that “evaporative cooling might greatly increase the kinetic energy of a TC and its size.”

Based on striking differences in dynamical and thermodynamic traits, a typical TC can be roughly divided into four portions in the radial direction: the eye, where precipitation is almost free; the eyewall, where deep convective clouds form an outwardly tilted wall with large diabatic heating; the rapid filamentation zone, where active inner spiral rainbands favorably exist (Wang 2008a; Li and Wang 2012a,b, Li et al. 2014); and the outer core, where outer spiral rainbands form and develop (Wang 2009; Li and Wang 2012a,b). The above-mentioned studies took into account the influence of evaporation in the entire circulation of a TC, whereas other studies have focused on the roles of evaporation in certain portions of a TC. For instance, the surface cold pools triggered by evaporative downdrafts in outer rainbands (Eastin et al. 2012) may visibly modulate TC evolution (Houze 2010). Subsidence in active outer rainbands brings low-equivalent potential temperature ($\theta_e$) air because of evaporative cooling into the inflow boundary layer. The air can be advected to the core region and subsequently entrained into the eyewall, thus suppressing eyewall convection and reducing the TC intensity (Barnes et al. 1983; Powell 1990a,b). Such a reduction becomes more notable if cloud condensation nuclei are increased in the outer-rainband region because of the enhanced evaporative cooling from the increased precipitation (Carrio and Cotton 2011; Cotton et al. 2012). Yang et al. (2007) deemed that the storm simulated with a three-dimensional TC model was more intense than that simulated with its axisymmetric version. They attributed this difference to the increased air–sea entropy.
deficit caused by the enhanced cooling in the inflow boundary layer from evaporation in addition to melting due to the more outwardly tilted eyewall simulated in the axisymmetric model.

Although previous studies have revealed that evaporation, even in certain regions of a TC, can have great impacts on TC structure and intensity, potential different impacts of evaporation in different regions of a TC have not been comparatively evaluated yet. Therefore, as an extension of earlier studies, we will revisit the effects of evaporation of raindrops on TC structure and intensity, with specific attention to the dynamical and thermodynamic roles of evaporation in different portions of a TC. A brief description of the model used and the experimental design are given in section 2. Results and discussion are presented in sections 3 and 4, respectively. Conclusions are drawn in section 5.

2. Model and experimental design

The model employed here is the fully compressible, nonhydrostatic atmospheric model TCM4 (Wang 2007). The model has been used for studying many issues associated with TCs, including the inner-core asymmetric structure (Wang 2007), rapid filamentation zone (Wang 2008a), annular hurricane structure (Wang 2008b), size change (Wang and Xu 2010; Xu and Wang 2010a,b), and spiral rainbands (Li and Wang 2012a,b).

The model equations are formulated in Cartesian coordinates in the horizontal and in mass coordinate in the vertical. The model assumes a flat lower boundary at the ocean surface with a uniform unperturbed surface pressure of 1010 hPa. The model top is set at about 40 km above the sea surface. A sponge upper-boundary condition similar to that used in Durran and Klemp (1983) is applied to absorb the upward-propagating sound and gravity waves. The physical parameterizations in TCM4 include a turbulence kinetic energy $E$ and dissipation $\varepsilon$ parameterization for subgrid-scale vertical turbulent mixing (Langland and Liou 1996), a modified Monin–Obukhov scheme for surface flux calculation (Fairall et al. 2003), an explicit treatment of mixed-phase cloud microphysics, a nonlinear fourth-order horizontal diffusion for all prognostic variables except for that related to the mass conservation equation, a simple Newtonian cooling term added to the perturbation potential temperature equation to mimic the longwave radiative cooling (Rotunno and Emanuel 1987), and the dissipative heating related to the turbulent kinetic energy dissipation rate from the $E-\varepsilon$ turbulence closure scheme.

The model domain is quadruply nested with two-way interactive nesting and with all the inner meshes automatically moving to follow the center of the storm. The model has 26 levels in the vertical and has mesh sizes of $201 \times 181$, $109 \times 109$, $127 \times 127$, and $163 \times 163$ grid points with horizontal grid increments of 67.5, 22.5, 7.5, and 2.5 km for the four meshes, respectively. A quiescent environment is considered in this study, and no cumulus parameterization is employed even in the two outermost meshes since convection occurs mainly within 200 km from the storm center and is covered by the innermost mesh. The model is initialized with an axisymmetric cyclonic vortex on an $f$ plane at 18°N over the ocean with a uniform sea surface temperature of 29°C. The initial thermodynamic profile of the unperturbed model atmosphere is defined as the western Pacific clear-sky environment given by Gray et al. (1975). Given the tangential wind field for the initial cyclonic vortex, which has a maximum wind speed of 25 m s$^{-1}$ at a radius of 80 km at the surface and decreases sinusoidally with height, the corresponding mass and thermodynamic fields are obtained by solving the nonlinear balance equation (Wang 2001).

In the control experiment (CTL), the above default model settings were used and the model was integrated for 96 h. To investigate the effects of evaporation in different regions of the TC on TC structure and intensity, we conducted five sensitivity experiments (see Table 1). All sensitivity experiments were initialized after the 42-h spinup in CTL. By 42-h spinup, the modeled storm had a radius of maximum wind (RMW) of about 20 km (not shown) and a rapid filamentation zone where the azimuthal-mean filamentation time is less than 45 min in the troposphere inside a radius of 60 km [see Fig. 1 in Li

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Description</th>
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<tbody>
<tr>
<td>CTL</td>
<td>Control simulation with all default model settings.</td>
</tr>
<tr>
<td>EW</td>
<td>As in CTL, except without evaporation of raindrops inside of the 40-km radius after 42 h of simulation.</td>
</tr>
<tr>
<td>RFZ</td>
<td>As in CTL, except without evaporation of raindrops between 40- and 60-km radii after 42 h of simulation and with a transition zone between 60- and 70-km radii where the evaporative effect was linearly diminished to zero with the decreasing radius.</td>
</tr>
<tr>
<td>OUT</td>
<td>As in CTL, except without evaporation of raindrops outside of the 70-km radius after 42 h of simulation and with a transition zone between 60- and 70-km radii where the evaporative effect was linearly diminished to zero with the increasing radius.</td>
</tr>
<tr>
<td>NO</td>
<td>As in CTL, except without evaporation of raindrops on all grids after 42 h of simulation.</td>
</tr>
<tr>
<td>A_NO</td>
<td>As in NO, except without evaporative cooling while allowing evaporation-related conversion from rainwater into water vapor.</td>
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In addition, no active outer spiral rainbands were present at this time because it was in the inactive phase of the quasi-periodic outer-rainband activity as documented in Li and Wang (2012a). In the first experiment, evaporation of rainwater was turned off inside a radius of 40 km (EW). Considering the eyewall width of about 20 km and the tilting of the eyewall, experiment EW meant ignorance of evaporation in the eyewall region. In the second experiment, evaporation was excluded in the rapid filamentation zone where inner rainbands often prevail between 40- and 60-km radii with a transition zone between 60- and 70-km radii where the evaporative effect was linearly diminished to zero at 60-km radius (RFZ). To evaluate the effects of evaporation in the outer core where outer rainbands usually occur, evaporation outside of the 70-km radius was excluded in the third experiment (OUT) with a transition zone between 60- and 70-km radii where the evaporative effect was linearly diminished to zero at 70-km radius. We found that the RMW of the storm was nearly constant (~20 km) with time and the rapid filamentation zone was kept inside the radius of about 60 km in experiments EW, RFZ, and OUT (not shown). Therefore, the regions where evaporation of rainwater was barred remained unchanged during the integration in these experiments.

In the fourth experiment, evaporation was turned off on all grids (NO) as was done in some of previous studies. To examine the possible differences in the effects of evaporation and evaporative cooling, an additional (the fifth) experiment (A_NO) was performed with the removal of evaporative cooling on all grids but allowing evaporation-related conversion from rainwater into water vapor as conducted in Sawada and Iwasaki (2010a).

3. Results

a. Intensity

A comparison of intensity changes of the simulated TCs is depicted in Fig. 1. The storms can be grouped into three families based on their distinct intensity changes. The TCs simulated in CTL, EW, and RFZ showed similar intensity changes, with a slow intensification through about 60 h of simulation. Afterward, the storms reached their mature stages but with considerable oscillations in
intensity, mainly resulting from the quasi-periodic behavior of outer rainbands as documented in Li and Wang (2012a). The minimum sea level pressure (MSLP) and the lowest model-level maximum wind speed of the storms in CTL, EW, and RFZ were approximately 923 hPa and 65 m s\(^{-1}\) at 96 h, respectively (Figs. 1a and 1b). The second family was the TCs in OUT and NO that tended to persistently intensify (Fig. 1). The storm in NO (OUT) became stronger than those in CTL, EW, and RFZ after 51 (64) h of simulation, with an MSLP of about 880 (899) hPa at 96 h (Fig. 1a). In contrast, the TC simulated in A_NO experienced explosive intensification from 42 to 60 h and obtained an MSLP of 847 hPa and a maximum wind speed of 92 m s\(^{-1}\) at 96 h (Figs. 1a and 1b), which was much stronger than the TCs in all other experiments.

The results shown in Fig. 1 demonstrate that evaporation of raindrops plays a role in suppressing the final intensity of a TC as noted in earlier studies (Wang 2002; Zhu and Zhang 2006). In particular, evaporation in the outer-core region largely limits the TC intensity. The differences in intensity were likely subject to the differences in entropy in the inflow boundary layer (Fig. 2). At 42 h, the azimuthally averaged value of \(\theta_e\) at \(z = 0.3\) km dramatically decreased with the increasing radius in all experiments (Fig. 2)—a result of reduction due to pre-existing outer rainbands (Li and Wang 2012a). As \(\theta_e\) in the boundary layer gradually recovered by extracting energy from the underlying ocean, outer rainbands re-formed and became active in about 8 h in CTL, EW, and RFZ (Figs. 3a–c), whereas the outer rainbands in OUT and NO, which will be discussed in the next subsection, were much less organized (Figs. 3d and 3e). Following the growth of outer rainbands, \(\theta_e\) in the boundary layer progressively lessened again in CTL, EW, and RFZ (Figs. 2a–c). The \(\theta_e\) decrease resulted likely from low-entropy air related to downdrafts in outer rainbands. Observations have indicated that downdrafts in an outer rainband tend to form in two distinct sectors—namely, the upwind convective sector (Barnes et al. 1983; Powell 1990a,b; Hence and Houze 2008; Didlake and Houze 2009) and the downwind stratiform sector (Didlake and Houze 2013b). This was also identified in high-resolution simulations (Li and Wang 2012b; Moon and Nolan 2015). In the convective sector of an outer rainband, two types of convective-scale downdrafts referred to as the low-level downdraft and the inner-edge downdraft correlate spatially with convective-scale updrafts (Barnes et al. 1983; Hence and Houze 2008; Didlake and Houze 2009). Both the downdraft features could spur downward owing to evaporative cooling (Didlake and Houze 2009) at low levels and even infiltrate the boundary layer, producing low-entropy air therein. As a result, a prominent decrease in boundary layer \(\theta_e\) occurred relatively far away from the TC center in CTL, EW, and RFZ (Figs. 2a–c) as outer rainbands were active (Figs. 3a–c). The low-\(\theta_e\) air could be advected radially inward into the eyewall region, lessening the \(\theta_e\) values therein. In the downwind stratiform sector of an outer rainband, mesoscale subsidence due to sublimation and evaporation exists (Hence and Houze 2008; Li and Wang 2012b; Didlake and Houze 2013b). Observations showed that some of the descending air with low entropy could continue downward into the boundary layer (Didlake and Houze 2013b), thereby reducing local \(\theta_e\). Figure 4 depicts the azimuthal-mean \(\theta_e\) values averaged during the period 78–84 h when outer rainbands were active in experiments CTL, EW, and RFZ but absent in experiments OUT, NO, and A_NO (Fig. 3). Immediately outside of the inner core (a radius of about 60 km) a low-entropy tongue clearly extended downward into the boundary layer in CTL, EW, and RFZ (Figs. 4a–c), which was not seen in OUT, NO, and A_NO (Figs. 4d–f). In summary, the downdrafts from both the convective and stratiform sectors of healthy outer rainbands simulated in experiments CTL, EW, and RFZ flushed the boundary layer with low-entropy air, which was advected radially inward and mixed into the eyewall (Figs. 2a–c). This implied a frustration of the thermodynamic cycle of the TC heat engine. Less kinetic energy could be generated and the final intensity of the TCs simulated in CTL, EW, and RFZ was thus weaker than that in OUT and NO (Fig. 1).

The limiting effect of evaporation on TC intensity can also be explained by comparing the intensity of TCs simulated in OUT and NO. The ultimate intensity of the storm in OUT was lower than that in NO (Fig. 1), suggesting that evaporation in the inner core reduced the TC intensity. Figure 2d shows that the \(\theta_e\) value (<357 K) at \(z = 0.3\) km between 20- and 50-km radii was smaller than that in the surrounding areas after 54 h of simulation in OUT. The evaporative cooling in inner rainbands was responsible for such a reduction in \(\theta_e\). In contrast, \(\theta_e\) in NO kept larger than 357 K at 0.3-km height after the recovery of the boundary layer (Fig. 2e). Resultantly, the final intensity of the TC in OUT was weaker than that in NO. Note that much smaller values of \(\theta_e\) (<355 K) between 20- and 50-km radii in OUT were found during the period 54–64 h (Fig. 2d), which was related to the enhanced evaporative cooling associated with more active inner rainbands. The storm in OUT was even weaker than those simulated in other experiments at that time (Fig. 1). This is seemingly inconsistent with the result in RFZ in which evaporation in inner rainbands was turned off but the storm intensity was still similar to that in CTL.
On one hand, convection in the inner rainbands was more active in OUT, as will be discussed in the next subsection, and the corresponding evaporative cooling related to the inner rainbands was relatively large. On the other hand, low-$\theta_e$ air in RFZ originated mainly from outer rainbands (Fig. 2c). The explosive intensification and the strongest final intensity of the storm simulated in A_NO were
attributed to the absence of the evaporative cooling effect (Fig. 2f) and an unphysical feedback loop. The evaporation of rainwater without evaporative cooling in A_NO would produce more condensation of the previously evaporated rainwater in convection and thus more latent heating later on, leading to an unphysical increase in TC intensity (Frisius and Hasselbeck 2009).
b. Structure

1) Precipitation

Figure 3 depicts the radius–time Hovmöller diagrams of the azimuthal-mean surface rain rate in all experiments. The surface rain rate shows very similar features in CTL, EW, and RFZ. The heaviest rainfall occurred in the eyewall, and the annular region with the azimuthal-mean surface rain rate greater than 70 mm h\(^{-1}\) was about 10 km wide (Figs. 3a–c). The quasi-periodic nature of rainfall outside of the 70-km radius in the three experiments was...
related to the quasi-periodic behavior of radially outward-propagating outer spiral rainbands as already discussed in Li and Wang (2012a).

The largest rainfall rate also occurred in the eyewall in OUT and NO, but the strong rainfall in the eyewall was obviously wider than that in CTL, EW, and RFZ (Figs. 3d and 3e). The annular regions with the azimuthal-mean surface rain rate greater than 70 mm h\(^{-1}\) in OUT and NO were nearly doubled, up to about 20 km at 96 h of simulation (Figs. 3d and 3e), indicating that evaporation, particularly in the outer core, considerably reduced the radial range of strong rainfall in the eyewall. In addition, precipitation in the inner core in OUT and NO extended farther outward from the storm center owing to the outward expansion of stratiform clouds in the absence of evaporation.

Although there was no quasi-periodic activity of outer rainbands in OUT and NO (Figs. 3d and 3e), outer spiral rainbands indeed formed during the period 50–70 h of simulation (Figs. 5 and 6). As seen in Fig. 5, an outer rainband with embedded active convective cells spiraled cyclonically from the southwest to the east of the storm center at 55 h in OUT. Similar banded structure was found in NO (e.g., at 55 h in Fig. 6) as well. Sawada and Iwasaki (2010a,b) related the formation of outer rainbands to cold-pool dynamics, which suggests that evaporation was key to outer-rainband initiation. However, evidence in Figs. 5 and 6 suggests that outer rainbands can still form even without evaporation of rainwater. Even in A_NO where evaporation was allowed but evaporative cooling was excluded, as in Sawada and Iwasaki (2010a,b), well-organized outer rainbands still appeared (e.g., between 52 h 30 min and 62 h 30 min of simulation shown in Fig. 7). This suggests that evaporation is not a necessity for outer-rainband formation of a TC. Li and Wang (2012a) found that outer spiral rainbands are often triggered by reinvigoration of inner-rainband remnants immediately outside the rapid filamentation zone and local inertial instability in the upper troposphere. Indeed, outer rainbands formed in the early integration of all sensitivity experiments (Figs. 3d–f), because subsidence was predominant in the outer-core region. Figure 8 compares the azimuthal-mean vertical velocity averaged from \(z = 3\) to \(z = 10\) km in CTL, OUT, NO, and A_NO. We can see that subsidence prevailed outside the 70-km radius after 65 h of simulation in OUT, NO, and A_NO (Figs. 8b–d), which is in sharp contrast to pronounced ascending motion in the inner core. Furthermore, downdrafts in the outer-core region were related closely to the strengthened diabatic cooling from the melting of cloud ice, snow, and graupel (Lord et al. 1984; Wang 2002). Actually, the stronger and wider eyewall convection (namely, the wider range of high surface rain rates in the eyewall in Figs. 3d–f) after 65 h of simulation enhanced the outward transport of ice particles in the upper troposphere due to the stronger outflow. As a result, the melting of a large amount of cloud ice, snow, and graupel produced large diabatic cooling and subsidence in the outer-core region (Figs. 8b–d), suppressing the formation and development of outer rainbands. Overturning flow associated with the robust eyewall and inner-rainband convection (diabatic heating) could also redound to the sinking motion in the outer-core region. However, its contribution was expected to be secondary because the overturning flow was in general one order smaller in magnitude than the subsidence shown in Fig. 8 (Moon and Nolan 2010; Fudeyasu and Wang 2011; Zhu and Zhu 2014).

2) INNER-CORE SIZE

Recently, more attention has been given to changes in the size of TCs and their possible impact on TC intensification (Wang 2009; Xu and Wang 2010a; Chan and Chan 2012, 2013; Knaff et al. 2014; Carrasco et al. 2014; Li et al. 2014). In this subsection we will examine how evaporation may affect the inner-core size defined as the radius of azimuthal-mean near-surface damaging-force wind (25.7 m s\(^{-1}\)) outside of the eyewall (Xu and Wang 2010a,b).

As shown in Fig. 9, the inner-core size of the TCs in EW and RFZ shared very similar evolution with that in CTL, with the size increasing through 63 h, decreasing between 63 and 70 h, regrowing from 70 to about 84 h, and remaining approximately at a 77-km radius afterward. It seems that only exclusion of evaporation in the eyewall region or the rapid filamentation zone did not markedly affect the inner-core size of a TC. By contrast, the inner-core size of the TCs in OUT and NO was considerably larger than that in CTL, EW, and RFZ (Fig. 9), with the maximum values of 87 km in OUT and 85 km in NO. This implies that evaporation in the outer-core region limited the inner-core size of TCs.

Note that the inner-core size of the storm in OUT was always larger than that in NO after 47 h of simulation.
FIG. 5. Plan view of the simulated reflectivity at $z = 3$ km from 50 h to 73 h 45 min of simulation in experiment OUT.
Fig. 6. As in Fig. 5, but for experiment NO.
Fig. 7. As in Fig. 5, but for experiment A_NO.
FIG. 8. Radius–time Hovmöller diagram of the azimuthal-mean vertical velocity (shading) and latent cooling rate (contours; K h$^{-1}$) averaged from $z = 3$ to $z = 10$ km in experiments (a) CTL, (b) OUT, (c) NO, and (d) A_NO. The latent cooling rates are contoured at −0.1 (thin), −0.4 (intermediate), and −0.8 (thick) K h$^{-1}$. 
Figure 9. Time evolution of the radius of the azimuthal-mean damaging-force wind (25.7 m s\(^{-1}\)) in all six experiments.

(Fig. 9). The more rapid increase in the inner-core size in OUT was mainly due to more active inner rainbands than in NO. Figures 10a and 10c compare the radius–time cross sections of the azimuthal-mean divergence at \(Z = 0.3\) km in OUT and NO. A contrasting feature between Figs. 10a and 10c is the occurrence of a locally enhanced convergence zone in the vicinity of the radius of 65 km in OUT. Sporadic downdrafts and corresponding divergence were found in the boundary layer between the radii of 40 and 60 km in OUT (Figs. 10a and 10b), which were likely associated with the evaporative cooling in the more active convection. The divergent flow could contribute to the locally enhanced convergence immediately outside of the inner core. As a result, additional updrafts (convection) arose in the convergence zone (Fig. 10b), fostering the development of inner rainbands likely through axisymmetrization. Figure 10b also implies that the forced inner rainbands propagated radially inward from the radius of about 65 km to the outer edge of the eyewall (e.g., from 47 to 53 h and from 62 to 67 h). In the experiment NO, no locally enhanced convection was found near the edge of the inner core (Fig. 10d).

To further examine the effects of inner rainbands on the inner-core size change, we conducted absolute angular momentum \([\text{AAM}; \quad M = (1/2)f r^2 + \nu v, \text{where } r \text{ is radius, } f \text{ is the Coriolis parameter, and } \nu \text{ is the tangential wind}]\) budgets. The azimuthal-mean AAM budget equation in cylindrical coordinates centered at the TC center can be written as

\[
\frac{\partial M}{\partial t} = -ru\bar{\nabla}_u - w\frac{\partial \bar{M}}{\partial z} - rw\bar{\nabla}_w - w\frac{\partial \bar{M}}{\partial z} + \bar{r}F + \bar{D},
\]

where \(u \) and \(w \) are radial and vertical components of wind; \(z \) is height; and \(t, F, \) and \(D \) are time, vertical turbulent mixing (including surface friction) of tangential winds, and horizontal diffusion of tangential winds, respectively. The overbar and prime respectively denote the azimuthal-mean and the eddy components of a given variable. Note that \(\zeta_t = \bar{\nabla}_u + \partial (\nu v)/\partial r \) is the azimuthal-mean vertical component of absolute vorticity, and the eddy pressure gradient term is negligible compared with other terms and thus is excluded in (1). The main contributions to the local tendency of the azimuthal-mean AAM on the right-hand side of (1) are the mean radial flux of absolute vorticity, the mean vertical advection of \(M \), the corresponding eddy fluxes, and vertical mixing terms.

Figures 11 and 12 show results for the AAM budgets averaged from 48 to 54 h in OUT and NO, respectively. Radial advection of AAM by strong inflow contributed positively to the AAM tendency in the boundary layer (Figs. 11a and 12a), with additional positive contributions by the inflow associated with inner rainbands near the 65-km radius in the lower troposphere in OUT (Fig. 11a; Fudeyasu and Wang 2011). The contribution to the AAM tendency by the vertical AAM advection (Figs. 11b and 12b) largely offset the contribution by the radial advection, with larger positive contribution by vertical advection between 30 and 60 km, which was associated with active inward-propagating inner rainbands in OUT (Fig. 11b). As a result, the mean-flow advection contributed positively to the AAM budget in the boundary layer (Figs. 11c and 12c), and the corresponding contribution particularly outside of the radius of 60 km in OUT (Fig. 11c) was larger than that in NO (Fig. 12c). Note that a maximum positive contribution by the total eddy flux appeared near the 60-km radius in the midtroposphere in OUT (Fig. 11f), indicative of the role of convective activity in the convergence zone in accelerating the local tangential winds. Note also that the net positive contribution to the AAM budget outside of 35 km in the boundary layer in OUT was much larger than that in NO (Figs. 11h and 12h)—mainly a result of the contribution by the azimuthal-mean secondary circulation associated with more active inner rainbands in OUT. This is consistent with the faster increase in the inner-core size in OUT than in NO during the period from 48 to 54 h of simulation (Fig. 9). The result further confirms that convection (latent heating) outside the eyewall, particularly in the rapid filamentation zone, is essential for the inner-core size increase as noted in previous numerical (Pendergrass and Willoughby 2009; Fudeyasu and Wang 2011; Li et al. 2014) and observational (Didlake and Houze 2013a) studies. During most of the period, convection in the rapid filamentation zone in OUT, NO, and A_NO was more active than that in
FIG. 10. Radius–time Hovmöller diagram of (top) azimuthal-mean divergence in experiments (a) OUT and (c) NO and (bottom) vertical velocity at $z = 300$ m in experiments (b) OUT and (d) NO.
CTL, EW, and RFZ because of higher $\theta_v$ in the boundary layer (Fig. 2). The inner-core size of the TCs in OUT and NO was thus larger than that in CTL, EW, and RFZ (Fig. 9).

The importance of convection in inner rainbands to the inner-core size increase can be further visualized from the different inner-core size in OUT and NO between 42 and 47 h of simulation. The inner-core size of the storm in OUT was subtly smaller than that in NO during that period (Fig. 9). The evaporation-induced sinking motion in OUT between 45- and 60-km radii (Fig. 10b) was in sharp contrast to the predominantly

Fig. 11. (a) The azimuthal-mean radial AAM advection, (b) the azimuthal-mean vertical AAM advection, (c) mean secondary circulation, (d) radial eddy AAM advection, (e) vertical eddy AAM advection, (f) total eddy advection, and (g) surface friction, vertical mixing, and horizontal diffusion contributions to (h) the net AAM budget averaged from 48 to 54 h in experiment OUT.
upward motion therein in NO (Fig. 10d). Therefore, the inactivity of inner rainbands and relatively weaker inward advection of AAM in the boundary layer made the inner-core size in OUT smaller than that in NO between 42 and 47 h (Fig. 9), although updrafts also appeared near the 65-km radius in OUT (Fig. 10b).

Note that the inner-core size of the TC simulated in NO tended to decrease after 84 h of simulation (Fig. 9). Figures 13a and 13c show the azimuthal-mean relative humidity and downward vertical motion for the storms in CTL and NO, respectively. Visible subsidence occupied the troposphere outside of the 60-km radius in
experiment NO (Fig. 13c), resulting mainly from diabatic cooling from melting of ice-phase hydrometeors (Figs. 13d). Moreover, Fig. 13c clearly shows downdrafts in the rapid filamentation zone in NO during the period from 90 to 96 h with two downdraft cores at $z = 3$ km near the radius of 40 km and at $z = 7.5$ km near the radius of 45 km, respectively, whereas updrafts were dominant in the inner core in CTL (Figs. 13a and 13b).

Hence and Houze (2012) examined the vertical structure of TC rainbands by analyzing data from the Tropical Rainfall Measuring Mission Precipitation Radar (TRMM PR) and revealed significant vertical suppression of inner-rainband activity possibly by locally enhanced vertical shear due to the outflow layer from the eyewall. Meanwhile, the layer immediately above the melting level was filled with slowly falling ice particles, which were ejected from the eyewall and seeded clouds below. This scene seems to be testified in NO. Figure 13d indicates that much stronger outflow strengthened the local shear in the upper layers during 90–96 h of simulation in NO, thus limiting the vertical development of inner-rainband convection. At the same time, compared with that in CTL (Fig. 13b), more ice particles spread farther from the eyewall and more ice fallouts from the outflow occurred in the rapid filamentation zone in NO (Fig. 13d). Resultantly, sublimation of ice particles in association likely with the upper-level intrusion of relatively dry air (relative humidity less than 90%; Fig. 13c) and low-level melting of ice particles, together with the constraint on the depth of the inner-rainband convection, led to net latent cooling and downdrafts therein (Fig. 13d). The downdrafts further diluted convection in inner rainbands in NO during this period (also see Fig. 8b). Hence and Houze (2012) also found a stronger convective suppression at the upper levels of inner rainbands in more intense TCs, which tend to have larger vertical shear from outflow. This is supported in experiment NO where the storm tended to become more intense after 90 h as defined by the maximum wind speed (Fig. 1b), so that the inner rainbands appeared to be much less active.
Fig. 8b. Note that the most intense storm in experiment A_NO (Fig. 1) had the largest inner-core size (Fig. 9), which seems inconsistent with the abovementioned stronger vertical suppression of inner-rainband convection and thus inner-core size in more intense TCs. In fact, Figs. 3f and 7 show that the strong eyewall convection of the TC in A_NO was much wider, and the more radially outward extension of enhanced convection and diabatic heating hence favored the noticeable growth of inner-core size.

The AAM budget averaged between 90 and 96 h for NO indicated that contribution by vertical advection (Fig. 14b) between the radii of 40 and 60 km in the troposphere was even smaller than that averaged between 48 and 54 h (Fig. 12b) because of much less active inner rainbands. The boundary layer inflow associated with
inner rainbands (not shown) and the corresponding radial advection of AAM tended to weaken. As a result, large surface friction (Fig. 14g) led to a net negative AAM tendency outside of the 50-km radius in the boundary layer and the lower troposphere (Fig. 14h), thereby reducing the inner-core size after 84 h of simulation (Fig. 9).

4. Discussion

Both Sawada and Iwasaki (2010b) and Li and Wang (2012a) proposed distinct mechanisms to explain the formation of outer rainbands. Sawada and Iwasaki (2010b) related the formation of outer rainbands to evaporative cooling. New convective cells can be initialized upstream of the preexisting surface cold pool triggered by evaporative cooling because of the convergence between the asymmetric outflow associated with the cold pool and the boundary layer inflow. These cells are finally organized into new spirally shaped outer rainbands. Since no active outer rainbands formed in the experiment without evaporative cooling in Sawada and Iwasaki (2010a,b), they further stated that evaporation is key to the formation of outer rainbands. However, Didlake and Houze (2009) hypothesized that the cold pool dynamics associated with convective-scale downdrafts make for the growth and sustenance rather than the formation of outer rainbands in terms of the development of new convective cells on the upwind end of an outer rainband (principal rainband). The numerical results in this study serve as a demonstration of Didlake and Houze’s (2009) hypothesis.

Despite being not as well organized as those simulated in CTL, EW, and RFZ, outer rainbands still formed in experiments OUT, NO, and even in A_NO (Fig. 3f) in which outer rainbands were inactive. During the period from 50 to 70 h of simulation in OUT, NO, and A_NO, high \( \theta_e \) in the lower troposphere (Figs. 2d–f) in the absence of strong downdrafts outside of the radius of 70 km (Fig. 8) facilitated the reinvigoration of inner-rainband remnants and thus the formation of outer rainbands (Figs. 3d–f), although the surface cold pools associated with evaporative cooling were not present. When considerable subsidence related to cooling due to melting of enhanced ice particles dominated the outer-core region, convection and outer rainbands was greatly suppressed therein. A careful examination of the result in Sawada and Iwasaki (2010a) shows evidence of such latent cooling as well. Their Fig. 6b shows that diabatic cooling overwhelmed diabatic heating outside the radius of 60 km in the troposphere, thereby producing net latent cooling. Corresponding downdrafts were expected to take place (their Fig. 8d) and convection was considerably suppressed on the mature stage of the simulated TC. Therefore, our numerical simulation results strongly suggest that the cold-pool dynamics associated with evaporation proposed by Sawada and Iwasaki (2010a,b) lead to the growth and/or sustenance of convective cells (or outer rainbands) rather than the initiation of outer rainbands, which is consistent with the observational results in Didlake and Houze (2009).

Evaporation is propitious to the formation of outer rainbands by indirectly abating cooling due to melting of ice particles ejected from the eyewall, which causes mesoscale downdrafts outside of the inner core. Furthermore, in the light of the outer-rainband formation mechanism proposed by Li and Wang (2012a), outer rainbands would hardly form as inner-rainband activity is considerably suppressed as seen after 84 h of simulation in NO. Sawada and Iwasaki (2010a) also pointed out that evaporation contributes to the enlargement of TC size, which is roughly referred to as the azimuthal-mean tangential wind vertically averaged from the surface to the midtroposphere (see their Fig. 3). Large diabatic heating in evaporation-induced spiral rainbands (viz., outer rainbands) drives the secondary circulation and enhances the inward transport of AAM outside the eyewall, leading to the increase in TC size, as previously documented in Wang (2009) and Fudeyasu and Wang (2011). However, the inner-core sizes of the storms simulated in OUT and NO in this study were larger than that in CTL (Fig. 9), although outer rainbands in OUT and NO were less active compared with those in CTL. In particular, the most rapid increase and the maximum of the inner-core size occurred in experiment A_NO (Fig. 9) in which evaporation was allowed while the evaporative cooling was removed as done in Sawada and Iwasaki (2010a,b).

Even though we took into account the azimuthal-mean tangential wind vertically averaged from the surface to the midtroposphere, the TC size was still significantly grew in A_NO (Fig. 3f) in which outer rainbands were inactive. The TC size increase was also found in other evaporation-free simulations using different numerical models [e.g., Fig. 15c in Frisisius and Hasselbeck (2009)].

It should be noted that in this study we evaluated the response of the inner-core size, which was defined as the radius of azimuthal-mean near-surface damaging force wind, to evaporation of rainwater. Therefore, more attention was given to the wind change near the surface. The results demonstrate that diabatic heating in inner
rainbands (or rapid filamentation zone) can drive strong inflow not only in the midtroposphere but also in the boundary layer outside the eyewall [see Fig. 8e in Fudeyasu and Wang (2011)], predominantly contributing to the tangential wind increase in the boundary layer outside the eyewall and thus the inner-core size increase [Fig. 11e in Fudeyasu and Wang (2011)]. This result is reminiscent of the observational finding in Didlake and Houze (2013a). They diagnosed the radar-retrieved wind field of rainbands in Hurricane Rita (2005) and deemed that stronger inflow and weaker, shallower updrafts associated with convective cells in inner rainbands constrain tangential wind acceleration to a low altitude. By contrast, the inflow in response to diabatic heating in outer rainbands was relatively weaker in the boundary layer [e.g., Fig. 8e in Fudeyasu and Wang (2011) and Fig. 7a in Didlake and Houze (2013a)]. In addition, Didlake and Houze (2013b) documented that the mesoscale stratiform portion of an outer band enables the strengthening of local tangential winds—hence, conducive to the inner-core size increase. However, the storm inner-core sizes were larger in OUT and NO in which outer rainbands were less active. Therefore, diabatic heating in inner rainbands may contribute equally or even more to the inner-core size increase than that in outer rainbands.

5. Conclusions

In this study, we have revisited the effects of evaporation of raindrops in different regions on TC structure and intensity by performing several sensitivity numerical experiments under idealized conditions, with special attention to the dynamical and thermodynamic roles of evaporation in the eyewall, the rapid filamentation zone, and the outer core of a TC.

Consistent with previous findings, evaporation of raindrops, particularly in the outer-core region, is shown to suppress the final intensity of a TC, whereas the effects of evaporation in the eyewall or in the rapid filamentation zone on TC intensity are very limited. The low-$\theta_e$ air in the inflow boundary layer resulting from evaporative cooling is transported radially inward and is mixed into the eyewall region, reducing $\theta_e$ under the eyewall and hence suppressing eyewall convection and reducing the TC intensity.

Evaporation in the eyewall and in the rapid filamentation zone hardly affects the activity of outer rainbands. The experiments without evaporation in the eyewall region or in the rapid filamentation zone shared the quasi-periodic appearance of outer rainbands similar to that in the control experiment. However, evaporation in the outer core of a TC can significantly influence the activity of outer rainbands. Unlike in Sawada and Iwasaki (2010a,b), in this study outer rainbands could still form with evaporation in the outer-core region switched off if the inner-rainband remnants can redevelop outside the rapid filamentation zone (Li and Wang 2012a). These outer rainbands are usually short lived and rarely move farther outward because of the lack of surface cold pools. Our results suggest that the cold-pool dynamics associated with evaporation lead to the growth and/or sustenance of outer rainbands rather than to their initiation. With evaporation removed, convection in the eyewall became stronger and wider, leading to more cloud ice, snow, and graupel transported radially outward in the upper-troposphere outflow layer. Melting of a larger amount of cloud ice, snow, and graupel produces larger diabatic cooling, together with the convective overturning flow whose role is secondary though, causing considerable subsidence in the outer core and suppressing the formation and development of outer rainbands.

Sensitivity experiments also show that exclusion of evaporation in either the eyewall or the rapid filamentation zone alone has a very weak effect on changes in the inner-core size of a storm. However, evaporation in the outer-core region can affect the inner-core size considerably. Without evaporation outside the inner core, inner rainbands became more active with the larger inner-core size. This suggests that evaporation in the outer-core region generally limits the inner-core size. Diabatic heating outside the eyewall, particularly in the rapid filamentation zone, is essential for the increase in the inner-core size by enhancing the boundary layer inflow and transporting AAM radially inward. In the experiment with evaporation turned off in all regions, the inner-core size might decrease when inner-rainband activity was suppressed because of the vertical suppression of convection and enhanced net diabatic cooling associated with sublimation and melting of ice hydrometeors therein.

In summary, evaporation could affect not only the intensity of a TC but also the activity of outer rainbands, leading to change in the inner-core size. A new finding in this study is that change in evaporation of rainwater can modulate the amount and distribution of other water substances (such as cloud ice, snow, and graupel) and their interactions, thereby further affecting the distribution of diabatic heating and thus the TC structure and intensity changes. The current in situ measurements could not allow accurate evaluation on the sophisticated interactions among different hydrometeors and their phase changes. This makes it difficult to determine how reasonably the microphysical processes are parameterized in high-resolution atmospheric models. The pronounced sensitivity of the simulated TC structure and intensity to evaporation, particularly outside the inner core, strongly
suggests that further effort must be made to validate and improve parameterization of cloud microphysics in numerical models for TC forecasting. In addition, as other cooling sources, the effects of melting and sublimation of ice hydrometeors, as well as the nonlinear interactions among these processes, on TC structure and intensity and their changes are also worth studying further in the future.

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