Sensitivity of Tropical Cyclone Feedback on the Intensity of the Western Pacific Subtropical High to Microphysics Schemes

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ABSTRACT

The Advanced Research version of Weather Research and Forecasting (WRF-ARW) Model is used to examine the sensitivity of a simulated tropical cyclone (TC) track and the associated intensity of the western Pacific subtropical high (WPSH) to microphysical parameterization (MP) schemes. It is found that the simulated WPSH is sensitive to MP schemes only when TCs are active over the western North Pacific. WRF fails to capture TC tracks because of errors in the simulation of the WPSH intensity. The failed simulation of WPSH intensity and TC track can be attributed to the overestimated convection in the TC eyewall region, which is caused by inappropriate MP schemes. In other words, the MP affects the simulation of the TC activity, which influences the simulation of WPSH intensity and, thus, TC track. The feedback of the TC to WPSH plays a critical role in the model behavior of the simulation. Further analysis suggests that the overestimated convection in the TC eyewall results in excessive anvil clouds and showers in the middle and upper troposphere. As the simulated TC approaches the WPSH, the excessive anvil clouds extend far away from the TC center and reach the area of the WPSH. Because of the condensation of the anvil clouds’ outflows and showers, a huge amount of latent heat is released into the atmosphere and warms the air above the freezing level at about 500 hPa. Meanwhile, the evaporative (melting) process of hydrometers in the descending flow takes place below the freezing level and cools the air in the lower and middle troposphere. As a result, the simulated WPSH intensity is weakened, and the TC turns northward earlier than in observations.

1. Introduction

As a dynamical downscaling tool of global general circulation models (GCMs), regional climate models (RCMs) have been increasingly used in climate research for more than 20 years (Dickinson et al. 1989; Giorgi 1990; Giorgi et al. 1993a,b; McGregor 1997; Leung et al. 2003; Wang et al. 2003; Pal et al. 2007; Liang et al. 2012). The principle behind the RCM technique is that, given a large-scale atmospheric forcing, a limited-area model with appropriate high-resolution surface information (e.g., complex topography, land–sea contrast, land use) and a detailed description of physical processes can generate realistic regional climate simulations that are consistent with the driving fields of either reanalysis data or GCM output but with high-resolution (both spatial and temporal) information (Wang et al. 2003). The RCMs have been proven to be able to improve the simulation of regional climate (e.g., Europe, New Zealand, Yangtze–Huai River valley, and East Asia) with great detail (e.g., Giorgi 1990; Jones et al. 1995; Renwick et al. 1998; Wang et al. 2000; Zhong 2006).

Despite the significant progress in regional climate modeling studies during the past 20 years or so, RCMs still exhibit relatively low skill in simulating the regional
climate in the tropics, especially in simulations of the East Asian monsoon system (McGregor 1997; Wang and Wang 2001; Zhong 2006; Zhou et al. 2008, 2009). One of the primary reasons for this failure is that convection in the tropics cannot be well represented in current RCMs. Previous studies have suggested that the RCM simulation of monsoon circulation is very sensitive to different cumulus parameterization (CP) schemes (Zhang 1994; Leung et al. 1999; Lee and Suh 2000; Sun et al. 2014). In fact, various microphysics parameterization (MP) schemes also have a significant impact on convection through microphysical heating/cooling. While great efforts have been taken to investigate the impact of CP schemes, the sensitivity of the simulated monsoon circulation to MP schemes, which are equally important, is neglected.

Because of the limited skills of RCMs in simulating convection in the tropics, large biases often appear in the East Asian monsoon simulation when a tropical cyclone (TC) is active over the western North Pacific (WNP) (e.g., Giorgi et al. 1999; Lee and Suh 2000; Zhong 2006; Zhong and Hu 2007; Fudeyasu et al. 2010). One of the major synoptic systems affecting the Asian monsoon is the western Pacific subtropical high (WPSH), which controls various mesoscale systems, such as tropical storms (Flatau et al. 1994; Carr and Elsberry 2000a,b). Zhong (2006) suggested that RCMs can realistically simulate the WPSH system under most conditions but they often fail when TCs are active over the western North Pacific (see his Fig. 1). This is further confirmed by Cha et al. (2011), who showed that typhoons can cause significant systematic errors in long-term regional climate simulations over East Asia because of their impact on large-scale environmental flow (see their Fig. 6). Recent studies indicate that the erratic departure of the simulated TC track from its realistic position is possibly a primary reason for RCMs’ failures in simulating the East Asian summer monsoon (Zhong 2006; Kubota and Wang 2009; Cha et al. 2011). The unrealistic position of the simulated TC over the WNP can lead to errors in simulation of the extent and intensity of the WPSH and a subsequent failure in East Asian summer monsoon simulation.

Despite the consistent and substantial improvements in TC track forecasts, large errors still exist and are of great concern given the overall improvement and the societal expectations accompanying such improvement. Numerous studies have examined various factors, including environment wind errors and storm structure errors, that may lead to inaccurate TC position forecasts in numerical models. Carr and Elsberry (2000b) found that errors in the simulated TC track could be attributed to an unrealistic description of the interaction between tropical cyclones and midlatitude systems. Their distance and spatial scales may significantly affect their interaction. Wang and Holland (1996) suggested that the diabatic heating could affect the TC motion through downward-penetrating flows that are associated with the anticyclonic potential vorticity (PV) anomalies aloft, which are continuously generated by diabatic heating. Asymmetric divergent flows associated with convective asymmetries within the vortex core region also contribute to the downward-penetrating flows.

While the TC position forecast “busts” can be related to errors in the structure and intensity of the TC vortex (e.g., McTaggart-Cowan et al. 2006), errors in the environmental wind appear to be dominant on errors in the TC track forecast (Galarneau and Davis 2013). For example, in the forecasts of TC Ike (2008) based on three operational global models, the environmental wind field in all three models steered TC Ike into southern Texas instead of recurving it over the Gulf of Mexico (Brennan and Majumdar 2011). This error in the forecasted TC position was attributed to an excessive zonal elongation of the subtropical anticyclone over the southern United States, which induced a more easterly steering flow over the Gulf of Mexico. This error in the structure of the subtropical ridge could be traced back to errors in the environmental initial condition of the model (Komaromi et al. 2011). As suggested by McGregor (1997), however, the midlatitude systems are generally reproduced well by regional models because of the strong large-scale forcing and few cumulus convections. Thus, the relationship between the TC motion and midlatitude systems (e.g., subtropical high) is an interesting issue that needs to be addressed.

While numerous studies have demonstrated the impact of microphysical schemes on hurricane intensity forecasts (e.g., Lord et al. 1984; Wang 2002; McFarquhar et al. 2006; Zhu and Zhang 2006), little is known about its impact on TC track forecasts. Fovell and Su (2007) and Fovell et al. (2009) argued that, without imposed large-scale flow, various cloud microphysics and cumulus schemes in a single model could produce large hurricane motion deviations that could finally lead to significantly different track forecasts. The large hurricane motion deviations are caused by the direct or indirect impact of microphysical assumptions on storm structure.

Our overall goal in this study is to explore the physical mechanism behind the current model’s failure in simulating the WPSH when TCs are active over the WNP. The objective of this paper is twofold. We will first investigate the sensitivity of TC motion and WPSH to MP schemes and then try to reveal the possible reasons and the involved physical processes. This paper is organized as follows. Section 2 describes the numerical model used in this study and experiment’s design. Section 3 shows
the simulation results using different MP schemes. The relationship between the TC motion and WPSH is discussed in section 4. Conclusions and a discussion are given in the final section.

2. Model configuration and experimental design

Typhoon Megi (2010) was the most intense TC with the longest lifespan over the western North Pacific and South China Sea in 2010. Megi originated over an area of disturbed weather about 600 km to the east of the Philippine Archipelago around 0000 UTC 12 October 2010 and developed quickly through the day. The Joint Typhoon Warning Center (JTWC) classified the system as a tropical depression at 0900 UTC 13 October. Possibly because of the strong influence of the WPSH, the system moved slowly in a west-northwest direction toward the Philippines. Meanwhile, the depression intensified to a tropical storm around 1200 UTC 13 October. Later, on 14 October, an eye-like structure of the storm could be seen clearly from satellite images. The Japan Meteorological Agency (JMA) upgraded Megi to a severe tropical storm, and JTWC upgraded it to a category 1 typhoon. JMA upgraded Megi to a typhoon on 15 October (Kieu et al. 2012; Wang et al. 2013).

The storm initially moved northwestward and then turned west-southwestward along the southern periphery of the WPSH. During this process, it underwent significant intensification as a result of the highly favorable conditions for tropical storm development. By the time it made its first landfall on the Philippines on 18 October, Megi had become one of the strongest tropical cyclones recorded to make landfall. After crossing the Luzon Island, Megi moved slowly, as a trough over central China deepened and extended over the South China Sea, leading to a break in the subtropical ridge. Because of the strong influence of the trough and the weakening subtropical high over the South China Sea, Megi experienced a sharp northward turning around 0000 UTC 19 October, then turned north-northeastward. It weakened into a tropical storm and finally a tropical depression when it made its second landfall at Zhangpu in Fujian Province, China, on 23 October.

The model used in this study is the Advanced Research version of Weather Research and Forecasting Model, version 3.3.1 (WRF V3.3.1) (Skamarock et al. 2008). The initial and lateral boundary conditions are obtained from the 1° × 1° National Centers for Environmental Prediction (NCEP) Final Analysis data (FNL) at 6-h intervals (http://rda.ucar.edu/datasets/ds083.2). A 20-km-resolution domain is set up for the simulation of Megi and WPSH. There are 36 uneven σ levels extending from the surface to the model top at 50 hPa. The model domain is centered at 22°N, 122°E, with 160 (north–south) × 180 (east–west) grid points, including complex topography and land–sea contrast. It extends far enough south to allow capture of the withdrawal of WPSH and the turning of Megi. The simulation starts at 0000 UTC 14 October and ends at 0000 UTC 24 October 2010, with a total of 240 h of integration that covers the entire life span of Megi. The time interval for the output of model results is 1 h.

The physical parameterizations used in this study include (i) the Grell–Dévényi (GD) CP scheme (Grell and Dévényi 2002), (ii) the Mellor–Yamada–Janjić boundary layer scheme (Mellor and Yamada 1982; Janjić 2001) coupled with the Monin–Obukhov surface layer scheme (Monin and Obukhov 1954; Janjić 1996, 2001), (iii) the Rapid Radiative Transfer Model (RRTM) (Mlawer et al. 1997) for longwave radiation calculation and Goddard scheme (Chou and Suarez 1994) for shortwave radiation calculation, and (iv) the five-layer thermal diffusion scheme (Skamarock et al. 2008) for land surface processes. Four experiments with different MP schemes are conducted in this study. All other physical schemes and model settings are the same in the four experiments. Results using these four different MP schemes are compared to investigate the sensitivity of TC and WPSH simulations to different MP schemes. Moreover, to verify the performance of different MPs in simulating the WPSH when TCs are absent over the WNP, we have also conducted an additional suite of experiments using the four MP schemes. These experiments start at 0000 UTC 1 November and end at 0000 UTC 30 November 2010 with a total of 1 month of integration. TCs are absent over the WNP during this period. Except for the different simulation period, the model domain and configuration in the four additional experiments are all consistent with that in the former experiments.

The MP schemes used in this study include 1) the WRF single-moment 3-class (WSM3) MP scheme (Hong et al. 2004), 2) the Lin MP scheme (Lin et al. 1983; Chen and Sun 2002), 3) the WRF single-moment 6-class (WSM6) MP scheme (Hong and Lim 2006), and 4) the Thompson MP scheme (Thompson et al. 2004, 2008). More specifically, the WSM3 scheme predicts three categories of hydrometers: vapor, cloud water/ice, and rain/snow. This is a so-called simple ice scheme. A major difference between the WSM3 and other approaches is that the diagnostic equation used for calculation of ice number concentration is based on ice mass content rather than on temperature. It follows Dudhia (1989) to assume cloud water and rain for temperatures above freezing, and cloud ice and snow for temperatures below freezing. The Lin scheme includes six classes of hydrometeors: water vapor, cloud water, rain, cloud ice, snow, and graupel. All parameterization production terms are based on Lin et al.
Corrections and modifications, including a saturation adjustment following Tao et al. (1989) and ice sedimentation. The scheme is taken from the Purdue cloud model, and the details can be found in Chen and Sun (2002). The WSM6 scheme is similar to the WSM3 simple ice scheme. However, vapor, rain, snow, graupel, cloud ice, and cloud water are held in six different arrays. Thus, it allows the existence of supercooled water and a gradual melting of snow falling below the melting layer. Some of the graupel-related terms follow Lin et al. (1983), but its ice-phase behavior is much different because of the changes of Hong et al. (2004). The Thompson scheme is a new bulk microphysical parameterization (BMP) that has been developed for use with WRF or other mesoscale models. Unlike any other BMP, the assumed snow size distribution depends on both ice water content and temperature and is represented by a sum of exponential and gamma distributions. Furthermore, snow assumes a nonspherical shape with a bulk density that varies inversely with diameter, as found in observations. This is different to almost all other BMPs that assume a spherical snow with constant density. More details of the four MP schemes can be found in Skamarock et al. (2008).

3. Simulation results with different MP schemes

Zhong (2006) suggested that the RCMs can realistically reproduce the regional circulation systems (e.g., WPSH). However, they cannot perform equivalently well when TCs are active over the WNP (see his Fig. 1). Cha et al. (2011) proposed that typhoons are responsible for the significant systematic errors in long-term regional climate simulations over East Asia (see their Fig. 6) because of their impact on the large-scale environment. For the purpose of this study, we discuss the performance of RCM under two conditions: when TCs are absent and when TCs are active over the WNP. Figure 1 shows the geopotential height in NCEP reanalysis data and the simulated geopotential height at 500 hPa at 0000 UTC 30 November in the four additional experiments, which are initialized at 0000 UTC 1 November. The results are verified against NCEP reanalysis data. It is found that the simulated WPSH is not sensitive to MP schemes, since all the sensitivity experiments perform well in simulating the WPSH in terms of the geopotential height contour of 5880 m after 29 days of integration. Results of this suite of experiments clearly indicate that the simulated WPSH is not sensitive to MP schemes when TCs are absent over the WNP.

Next, we will investigate the sensitivity of WPSH intensity and TC track to the MP schemes when TCs are active over the WNP. Previous studies suggested that the large-scale environmental flow played a critical role in determining TC motion. Without imposed large-scale flow, however, a self-propagating vortex motion can be modulated distinctly by microphysics (Fovell et al. 2009). Fovell and Su (2007) showed that various cloud microphysics assumptions, together with CPs, can have a great impact on hurricane tracks simulated in a regional-scale model at 30-km horizontal resolution, and the impact is quite significant even in relatively short-range (2 days) forecasts. Sun et al. (2014) demonstrated that, even with different CP schemes in the sensitivity experiments, the change of microphysical latent heating is still largely responsible for the discrepancies in the WPSH and TC motion simulations.

Figure 2 compares the storm tracks simulated in the sensitivity experiments with the JTWC best track. The model with the WSM3 MP scheme can well reproduce the track of Megi before and after its turning, with an average track error of about 50 km at 6-hour intervals, but it performs not so well with the other three MP schemes. We use control run (CR) to refer to the WSM3 experiment and parallel runs (PRs) to refer to the Lin, WSM6, and Thompson experiments. All experiments realistically simulate the northwestward movement of Megi before 0000 UTC 17 October and the west-southwestward movement along the southern periphery of the WPSH until the storm crossed the Luzon Island. Large differences...
between results of the four experiments occur after 1800 UTC 18 October. The simulated storm in PRs turns northward earlier than observation, whereas in CR, it continues to move westward and turn northward over the South China Sea at about 1800 UTC 19 October. Apparently, the simulated storm track is sensitive to the choice of MP scheme.

It has been pointed out that the TC track over the WNP is mainly determined by two factors. One is the environmental flow, especially the steering flow in the southern part of the WPSH at 500 hPa (e.g., Chan and Gray 1982; Wu et al. 2005; Zhong 2006), and the other is the dynamic and thermodynamic structure of TC itself [e.g., Holland (1983); Fiorino and Elsberry (1989); Wu and Wang (2000, hereafter WW00)]. Therefore, theoretically the MP has two ways to affect the TC motion. First, the MP can modulate the TC motion by influencing the dynamic and thermodynamic structure of TC. Second, the MP can affect the steering flow of a TC by influencing the extension and withdrawal of the WPSH. In the following section, we will discuss which one is responsible for the failed simulation of TC motion in PRs.

In this study, the potential vorticity tendency (PVT) diagnosis technique (Chan 1984; WW00) is utilized to estimate contributions of the TC structure and environmental flow to TC motion. Based on their (baroclinic) simulation of a TC on a $\beta$ plane without basic flow, WW00 suggested that the motion of the modeled TC follows the maximum local PVT with azimuthal wavenumber 1 (WN1) in the midtroposphere. The PVT is given by the sum of the contributions from the horizontal advection (HA), vertical transportation (VT), and diabatic heating (DH). The influence of physical processes on the vortex motion at each level can be identified by the PVT diagnosis approach. As suggested by WW00, the velocity of the TC motion is estimated based on PVTs in Eq. (1), and the individual contributions of various terms in Eq. (2) can be estimated in the pressure coordinates:

$$\frac{dP}{dt} = -V_{PV} \cdot \nabla P_S$$ and

$$\frac{dP}{dt} = \Lambda_1 \left[ -V \cdot \nabla V - \frac{\partial P}{\partial p} g V_3 \cdot \left( -\frac{\nabla q}{C_F \pi} + \mathbf{V} \times \mathbf{F} \right) \right],$$

where $P$ is PV, $P_S$ is the symmetric component of PV, $V$ is the horizontal air motion, $V_{PV}$ is the vortex motion speed estimated from the WN1 component of the PVT, $q$ is the three-dimensional absolute vorticity vector, $p$ is the pressure, $\Lambda_1$ denotes an operator to obtain the wavenumber-1 component, $V_3$ is the three-dimensional gradient, $\theta$ is potential temperature, $\omega$ is the vertical velocity in the pressure coordinates, and $Q$ and $F$ denote diabatic heating rate and friction, respectively. We apply Eq. (1) to each grid point (denoted by subscript $i$). This will result in a set of linear algebraic equations:

$$\frac{dP}{dt}_{1i} = -c_x \left( \frac{\partial P_S}{\partial x} \right)_{1i} - c_y \left( \frac{\partial P_S}{\partial y} \right)_{1i},$$

The zonal ($c_x$) and meridional ($c_y$) components of the velocity of the vortex motion at each level will be determined from Eq. (3). Considering a specified domain in the vicinity of the TC center, which is within a radius of 300 km from the TC center, we use the least squares method to estimate $c_x$ and $c_y$ by minimizing:

$$\sum_{i \in N} \left[ \left( \frac{dP}{dt}_{1i} \right) + c_x \left( \frac{\partial P_S}{\partial x} \right) + c_y \left( \frac{\partial P_S}{\partial y} \right) \right],$$

where $N$ is the number of total grid points in the specified domain. The steering by environmental flow is included in HA, which is the first term on the right-hand side of Eq. (2), while VT (the second term) and DH (the third term) mainly depend on the dynamic and thermodynamic structure of the simulated TC that is sensitive to the MP scheme.
Figure 3 shows the temporal variation of the vertically averaged TC speed calculated from the PVT equations, as well as its individual contributions of HA, VA, and DH in the WSM3 (CR) and WSM6 experiments (PR). The time period is from 17 to 21 October, which corresponds to the period before and after the time when the TC track forecast between the CR and PR becomes significantly different. Because of the various vertical extents of the positive PV anomalies in CR and PR (figure omitted), the mean speed is calculated from 800 to 500 hPa. The results indicate that the mean speed of TC estimated from $V_{PV}$ is almost identical to that calculated from the time-dependent variational TC center position $V_C$ in most cases. Thus, the PVT diagnosis approach is equally efficient in estimating the motion of TC in a real TC case as it is in the ideal case in WW00.

According to Eq. (2), the contribution of individual physical process on the difference in TC motion between the CR and PR can be determined. In the following analysis we focus on the period before 0000 UTC 19 October, when the difference in TC position is insignificant. This is because, once the difference in TC position becomes significant, the contribution of each single term in Eq. (2) may be affected by the difference in the surrounding environment and thus cannot be used to fully explain the difference in TC motion. Because of the zonal extension of WPSH and its strong large-scale forcing in the CR, the contribution of HA is larger than that in PR (i.e., $V_{PV-HA}$), leading to the difference in TC zonal motion (Figs. 3b,e). Although in this case the meridional wind is weaker than zonal wind, it still contributes greatly to the TC motion along the meridional direction (Figs. 3c,f). Moreover, the magnitude of zonal
$V_{PV-HA}$ in CR is much larger than that in PR by up to 4 m s$^{-1}$, but the magnitude of meridional $V_{PV-HA}$ in the PR is notably larger than that in the CR by up to 2 m s$^{-1}$, especially before 1800 UTC 18 October, when the forecast track difference between CR and PR is not significant. This indicates that $V_{PV-HA}$ slows down the westward motion of the TC but accelerates its northward motion in the PR. Compared to the contribution of $V_{PV-HA}$, $V_{PV-VT}$ contributes little to the TC motion before 0000 UTC 19 October. Hence it is not the direct and major reason for the difference in TC motion (i.e., $V_{PV}$) between the two experiments. As suggested by WW00 and Peng et al. (1999), because of a fast adjustment between the relatively asymmetric flow (HA) and the asymmetric DH, the temporal variations of $V_{PV-HA}$ and $V_{PV-DH}$ contributions are out of phase. Although $V_{PV-DH}$ contributes little to the difference in zonal TC motion under the condition of strong zonal steering flow (i.e., zonal $V_{PV-HA}$), it plays an important role in determining the difference in meridional TC moving, since the meridional steering flow is relatively weak. This is consistent with the findings of Wang and Holland (1996). The contribution of DH, however, is smaller than that of HA in the difference in meridional TC motion, except for a short period from approximately 1800 UTC 18 October to 0600 UTC 19 October, when the forecast TC track difference between CR and PR is significant (Fig. 3i). Therefore, it is the contribution of HA ($V_{PV-HA}$) that is responsible for the difference, especially the TC’s northward turning ahead of time in PR. The MP directly affects TC motion by changing the environmental flow near the TC (e.g., HA in PVT). The environment flow is closely related to the intensity and extent of WPSH. Moreover, by comparing the result using the WSM3 with that using the other two MP schemes (i.e., Lin and Thompson scheme), we can reach similar conclusions.

Figure 4 shows the geopotential height in NCEP reanalysis data and the simulated geopotential height (m) at 500 hPa in the four sensitivity experiments at 0000 UTC (a)–(e) 15 Oct and (f)–(j) 18 Oct.

![Figure 4](image-url)
height at 500 hPa in NCEP reanalysis data (Fig. 4f), the WPSH does not break near Taiwan in terms of the geopotential height contour of 5880 m, which is basically consistent with the results in CR (Fig. 4g) but substantially different from the results in PRs (Figs. 4h–j). Therefore, the break of WPSH near Taiwan in PRs at 0000 UTC 18 October is unrealistic. Compared with the better model performance in simulating the WPSH when TCs are absent over WNP, we find that the unrealistic break of WPSH in PRs is closely related to the unrealistic simulation of TC motion (or track) when the TC is active over the WNP (Figs. 1, 2, and 4).

Meanwhile, note that there is an additional disturbance at 500 hPa near 15°N, 110°E that is not present in the CR (Figs. 4f–h). It could be the result of the initial conditions and microphysics parameterization. However, the impact of the disturbance on the large-scale steering flow and thus the storm track in PRs is very limited for the following three reasons. First, as suggested by Brand (1970), the rotation of the binary system is sharply dependent upon separation distance when the distance is smaller than 12° latitude. At 0000 UTC 18 October, the distance between the two storms is larger than 12° latitude. Such a long distance makes it impossible for the two storms to interact with each other. Second, the difference between the intensity of the two storms is huge. Thus, the impact of the weaker storm on the stronger storm is very limited. Third, and most importantly, because of the long distance, the fictitious disturbance is unlikely to exert any impact on the WPSH in PRs, which determines the large-scale steering flow and thus TC track. Apparently the unrealistic break of WPSH in PRs cannot be attributed to the small fictitious disturbance at 0000 UTC 18 October prior to the significant departure of the simulated TC track.

As mentioned above, we find that the unrealistic positions of TC over the WNP are responsible for the failure in simulating the extent and intensity of WPSH. Note that this statement is not contradicting with the earlier argument that the TC track over the WNP is mainly determined by the steering flow of WPSH. In this study, as the TC track over the WNP is mainly determined by the steering flow of WPSH, the unrealistic break of WPSH causes errors in the simulation of TC motion and, thus, unrealistic positions of the simulated TC in PRs. Subsequently, as the simulated TC position deviates from the observation, errors are also found in the WPSH simulation in PRs. The feedback loop between the inaccurately simulated TC position and the WPSH eventually leads to large biases in simulations of WPSH and TC.

This study suggests that an inappropriate MP scheme is among those factors that are responsible for the failure in TC motion simulation. The MP scheme influences the environmental flow of TC through its impact on HA. The unrealistic change in the steering flow along the southern edge of the WPSH is the primary reason for the early turning of TC in PRs. Sensitivity tests can simulate the WPSH quite well when TCs are absent (Fig. 1), implying that the failure in WPSH simulations in PRs may be related to the unrealistic description of convectons in TCs (McGregor 1997). However, as shown in Fig. 4, the WPSH has broken in the sensitivity region (SR; shown in Fig. 2), when the center of SR is within a radius of about 500 km of the TC center at 0000 UTC 18 October. The question of how TC convections can lead to the break of WPSH in SR from such a distant place is not answered yet. To address this question, in the next section we will discuss the variation of WPSH with respect to the MP scheme and investigate how the MP scheme affects the DH in TC and leads to the break of WPSH subsequently.

4. Relationship between the WPSH and TC track
   a. Temperature profile

Since the failure in simulation of the Megi track in PRs is directly linked to the unrealistic break of the WPSH and associated errors in steering flow, we define an SR (shown in Fig. 2; 22°–27°N, 125°–130°E) where the variation of WPSH in PRs is significantly different from that in the CR. The difference in WPSH intensity over the SR simulated by the PRs and CR serves as an important index to evaluate model performance in simulation of the WPSH.

Detailed analysis reveals that the unrealistic break of the WPSH in PRs can be attributed to the bias in simulation of temperature profile, which affects the geopotential height and thus the intensity of the WPSH. The equation for temperature tendency is written as follows:

$$\frac{\partial T}{\partial t} = -\mathbf{V} \cdot \nabla T - \omega \left( \frac{\partial T}{\partial \rho} - \frac{1}{C_p \rho} \dot{Q} \right) + \frac{\dot{Q}}{C_p},$$

where $T$ is the temperature and $\rho$ is the air density. The three terms on the right-hand side of Eq. (5) represent contribution of HA, VT, and DH, respectively.

Figure 5 shows the time–height cross section of the differences in geopotential height and temperature averaged over the SR between results of the WSM6 run (PR) and WSM3 run (CR). Terms contributing to temperature tendency difference (i.e., HA, VT, and DH) are also shown in Fig. 5. The unrealistic split of the WPSH in PRs can be attributed to the large bias in temperature variation after 0000 UTC 17 October when the storm is approaching SR (Fig. 5b). Compared with
FIG. 5. Height–time cross sections of difference in (a) geopotential height (m), (b) temperature (°C), and the terms contributing to temperature tendency (°C day$^{-1}$)—(c) HA, (d) VT, (e) DH, and (f) HA + VT + DH—between PR and CR averaged over the SR.
that in CR, the simulated temperature in PRs is colder in the lower troposphere below 500 hPa by up to −2.7°C but warmer in the upper troposphere above 500 hPa by up to 2.2°C. As suggested by Pauley and Smith (1988), such a difference in temperature profile in the PR may not lead to notable changes in the surface pressure, but it causes a decrease of geopotential height at 500 hPa with a magnitude of −29 m under hydrostatic constraint conditions (Figs. 5a,b) and results in shrinking of the WPSH areal extent (Fig. 4i), which is eventually responsible for the unrealistic earlier turning of the TC in PR (Fig. 2).

The differences in terms contributing to the temperature tendency between the CR and PR are also shown in Fig. 5. In contrast to its great contribution to TC motion, the contribution of HA to temperature tendency is much smaller than that of VT and DH. Its contribution to differences in the temperature profile between CR and PR, however, cannot be neglected. Consistent with that of temperature, difference in HA between the two experiments is positive (negative) above (below) 500 hPa from 1200 UTC 16 October to 0000 UTC 18 October, corresponding to the time period when the TC is active in areas south of the SR (Fig. 5c). Interestingly, the effects of VT and DH on temperature tendency are always out of phase. As suggested by Mapes and Houze (1995) and Zhang et al. (2002), it indicates an approximate equilibrium relationship between the diabatic heating and adiabatic cooling (warming) caused by upward (downward) motion (Figs. 5d,e). However, because of the large magnitude of DH, the effect of VT cannot completely offset that of DH on temperature tendency. Thus, the profile of HA + VT + DH is still roughly similar to that of DH; the latter is sensitive to the choice of MP scheme and contributes greatly to differences in the temperature profile between CR and PR (Figs. 5e,f). Apparently the DH difference between the two experiments is a major reason that causes the different temperature profiles shown in Fig. 5b. Temperature is warmer (colder) in the PR than in the CR above (below) 500 hPa, which is a dividing line that separates upper-level temperature response to MP from that in the lower level. The MP scheme in WRF affects the temperature profile by influencing the profiles of VT and DH. Large biases in VT and DH eventually lead to the unrealistic break of WPSH and thus the earlier turning of TC.

To further investigate the main source of the DH difference between the PR and CR, we examine the diabatic heating rates caused by radiation parameterization $Q_{RA}$, boundary layer parameterization $Q_{BL}$, cumulus parameterization $Q_{CP}$, and microphysics parameterization $Q_{MP}$, which are ranked according to their calling sequence in WRF. Note that the calling sequence for various physical schemes has no impact on the results. This is because in WRF, the physical variables are not updated immediately after any specific physical scheme is called. Only the tendencies are saved. The physical variables are updated only after all the physical schemes are called and then move on to next time step. Total diabatic heating rate $Q$ can be expressed as

$$Q = Q_{MP} + Q_{CP} + Q_{RA} + Q_{BL}.$$  (6)

Differences in each component of the diabatic heating rate between PR and CR are shown in Fig. 6. Difference in the MP heating is the largest above 500 hPa and after 17 October. Its vertical distribution and temporal variability is quite similar to that of DH differences despite the slight difference in magnitude between them (Figs. 5e and 6a). Above (below) the 500-hPa geopotential height, the MP component can be considered as an important heat source (sink) for the difference in temperature between the PR and CR after 0000 UTC 17 October. This result clearly indicates that, compared to the other three components of diabatic heating rate (i.e., the diabatic heating rate caused by $Q_{CP}$, $Q_{BL}$, and $Q_{RA}$), $Q_{MP}$ plays the most important role in determining the difference in DH and temperature. Large positive (negative) values of difference in both DH and temperature between the CR and PR appear above (below) 500 hPa, which is a dividing line between different responses to MP schemes in the upper and lower atmosphere (Fig. 6a). Hence, the difference in the microphysics scheme in the CR and PR directly leads to a large discrepancy in the Megi and WPSH simulation. Note that the differences between PR and CR caused by cumulus, boundary layer, and radiation are a consequence of using different MP schemes and play a secondary role in determining the difference in DH, since all physical schemes are the same except for microphysics in our four experiments. Specifically, although the $Q_{CP}$ difference is notably weaker than the $Q_{MP}$ difference between PR and CR, it contributes to the warming in the upper and middle troposphere and cooling in the lower troposphere, especially around 0000 UTC 19 October (Fig. 6b). While it is true that the $Q_{RA}$ difference is much smaller than the $Q_{MP}$ difference, there is a difference of up to 2°C in the upper troposphere (around 200–300 hPa), which is consistent with the magnitude in temperature changes related to cloud–radiation interaction simulated by Jin et al. (2014). This relatively small change could have a substantial impact on the environment given its wider coverage (Fig. 6c). The $Q_{BL}$ difference is much weaker than the $Q_{MP}$, $Q_{CP}$, and $Q_{RA}$ differences and mainly concentrates in the boundary layer with little effects above (Fig. 6d).
However, it does not mean that the other three schemes (i.e., cumulus, boundary layer, and radiation schemes) contribute little to the vertical distributions of DH and, thus, temperature differences. According to the results in the CP sensitivity experiments in Sun et al. (2014), the CP scheme contributes to DH not only by changing the temperature profile in terms of $Q_{\text{CP}}$ but also by changing the moisture profile and thus the microphysical latent heating in terms of $Q_{\text{MP}}$.

Because of the importance of the diabatic heating caused by MP and by CP, we further investigate the role of MP and CP on temperature in the storm. Figure 7 provides a Hovmöller diagram of the azimuthal-averaged diabatic heating from MP and CP at 500 hPa for the WSM3 and WSM6 cases and the difference between them. It shows that the CP heating has a varying trend similar to that of the MP heating. In the WSM3 and WSM6 cases, although the magnitude of CP heating is much smaller than that of MP heating, the extension of CP heating is much larger than that of MP heating. Thus, similar to the effect of MP heating, the CP heating also has great influence on convection near the storm. More importantly, compared with that in the WSM3 case, both the MP heating and CP heating in the WSM6 are smaller in the inner region near the eyewall (i.e., the region within a radius of 200 km from TC center) but larger in the outer region covered by spiral rainbands (i.e., the region outside a radius of 200 km from TC center). As suggested by Sun et al. (2013a,b), less diabatic heating release around the eyewall area leads to less decrease in central pressure and a smaller pressure gradient near the eyewall in the WSM6 case (figure omitted). Subsequently, the eyewall in the WSM6 case will expand farther outward under the weaker pressure gradient, resulting in a much larger storm size (figure omitted). On the other hand, more MP and CP heating release is found in the outer spiral

![Figure 6](image-url)
rainbands in the WSM6 case than in the WSM3 case (Fig. 7). As suggested by Wang (2009), MP and CP heating in the outer spiral rainbands can make the convection more active, which is in favor of the increase in storm size. Thereby, the less heating release near the eyewall and more heating release in the outer spiral rainbands both contribute to the increase in the simulated TC size in the WSM6 case and, thus, result in the large difference in TC size between the WSM6 case and WSM3 case. The difference in TC size leads to a different extension of anvil clouds (Fig. 11), which finally determines the difference in WPSH intensity and TC track simulations between the WSM6 case and WSM3 case (Figs. 2 and 4).

Phase changes of hydrometeors are important physical processes described in MP. We will investigate the distribution of the hydrometeors and its impact on temperature in the next section.

**b. Distribution and source of the hydrometeors**

Because of the importance of the phase changes of hydrometeors for the WPSH and TC simulations, it is necessary to compare the hydrometeor distributions of the WSM3 and WSM6 experiments with the observations. As mentioned above, we focus on the period before the difference in TC position is insignificant between the WSM3 case and WSM6 case. This is because, once the difference in TC position becomes significant, the contribution of the MP scheme may be affected by the difference in the surrounding environment and thus cannot be used to fully explain the difference in hydrometeor distribution between the WSM3 case and WSM6 case.

Figure 8a displays the high-resolution (0.1° × 0.1°) blackbody temperature (TBB) observed by the FengYun-2E (FY-2E) meteorological satellite, while Figs. 8b and 8c show two top views of the total mixing ratio of model-simulated hydrometeors for the WSM3 and WSM6 experiments at 0000 UTC 18 October. TBB indicates cloud-top heights if clouds are present below the satellite and are used as an index of convective activity. When equivalent cloud amount and contained hydrometeors increases, TBB decreases, which means deeper convections were observed (Kubota et al. 2005). Despite the fact that it is meaningless to compare the values of simulated hydrometeors with the values of observed TBB...
because of their unit difference, we can still use the observed TBB to validate the spatial distributions of the simulated hydrometeors in the WSM3 and WSM6 cases. It is encouraging that both the WSM3 and WSM6 cases well reproduce the main characteristics of the inner and outer spiral rainbands. Although it is not possible to predict the detailed distribution of convective cells along the spiral bands, the model does simulate well the distributions of intense and organized (convective and stratiform) clouds in the spiral rainbands. Both the observed and simulated TC spiral rainbands present strong asymmetry. The inner spiral rainbands are mainly distributed to the east of the TC center, while parts of the outer spiral rainbands appear to the southeast, east, and northeast of the TC. On the other hand, compared with that of the WSM6 case, the hydrometeor distribution in the WSM3 case is more consistent with the TBB distribution in observations. Both the observed TBB distribution and the simulated hydrometeor distribution in the WSM3 case show that the spiral rainbands are collapsed and break over the aforementioned SR, which is to the northeast of the TC, while the simulated hydrometeor distribution in the WSM6 case displays intense rainbands over the SR that connect inner and outer spiral rainbands in the northeastern portion of the TC. As mentioned in section 4a, the difference in the simulated hydrometeors and the related phase change over the SR plays an important role in determining the differences in temperature profile and, thus, the differences in WPSH intensity between the two cases. We will further discuss this issue in the following.

The WSM3 scheme includes ice and snow processes suitable for mesoscale grid sizes (Hong et al. 2004). It is a simple ice scheme that predicts water vapor, cloud water/ice, and rain/snow (Skamarock et al. 2008). It follows Dudhia (1989) to assume cloud water and rain for temperatures below the freezing level, and cloud ice and snow for temperatures above the freezing level. This scheme is computationally efficient for the inclusion of ice processes and is thus widely used in regional climate modeling studies (e.g., Liang et al. 2012; Mooney et al. 2013). Note that not every hydrometeor is stored as individual array and output from the model in the WSM3 scheme. Cloud water and ice are stored as the cloud component of the hydrometeors in the output from the model, while rain and snow are stored as the precipitating component (Braun 2006). Figure 9 shows the height–time cross sections of the simulated hydrometeors averaged over the SR in the WSM3 experiment. The simulated hydrometeors are mostly distributed in the levels between 600 and 200 hPa and below 850 hPa. Consistent with the microphysical heating, the hydrometeors in the upper levels become notable after 0000 UTC 17 October, when the simulated storm is approaching SR (Fig. 9a). The distribution of the cloud component (i.e., cloud water and ice) is basically similar to that of total hydrometeors, especially in the levels below 850 hPa (Fig. 9b). As suggested by Dudhia (1989), the cloud component in the low (middle and upper) troposphere below (above) freezing level can be taken as cloud water (ice). The magnitude and vertical extent of the cloud component are all much larger than that of the precipitating component (Fig. 9c). Thus, the cloud component can be considered as the major component of the total hydrometeors, while the precipitating component is minor. In addition, the precipitating component is only distributed in the levels between 650 and 300 hPa in the form of snow. From 0000 to 1200 UTC 17 October, no precipitation occurs at the surface, implying that all the precipitating component has evaporated before arriving at the surface (Fig. 9c).

The 6-class scheme (i.e., the WSM6 scheme) is similar to the WSM3 simple ice scheme but includes descriptions
of graupel and its associated processes. Different from the WSM3 scheme, in the WSM6, water vapor, rain, snow, cloud ice, cloud water, and graupel are held in six different arrays. Hence, it allows supercooled water to exist and describes the gradual melting of snow and graupel during their falling below the melting layer. Some of the graupel-related terms in the WSM6 are described following Lin et al. (1983), but its ice-phase behavior is different because of the changes of Hong et al. (2004).

Figure 10 shows the height–time cross sections of the simulated hydrometeors and their individual contributions (g kg$^{-1}$) in the WSM3 experiment averaged in SR: (a) total hydrometeors, (b) cloud water/ice, and (c) rain/snow.
FIG. 10. As in Fig. 9, but for the simulated hydrometeors and their individual contributions (g kg$^{-1}$) in the WSM6 experiment: (a) total hydrometeors, (b) cloud water, (c) cloud ice, (d) rainwater, (e) snow, and (f) graupel.
extent and thus can be considered as the major part of the precipitating component. Consistent with that in WSM3, snow is concentrated in the levels between 600 and 200 hPa after 0000 UTC 17 October in the WSM6 case (Fig. 10e). The magnitude of snow in WSM6, however, is about 3 times larger than that in the WSM3 and contributes greatly to the difference in hydrometeors between the WSM6 case and the WSM3 case. Moreover, the WSM6 case with the higher concentration of snow produces more stratiform precipitation compared to the WSM3 case (Figs. 10d,e). This result is consistent with other modeling studies (Lord et al. 1984; McCumber et al. 1991). The simulated graupel only exists in a limited area near 550 hPa, and its value is much smaller than that of rain in the WSM6 case (Fig. 10f).

Overall, both cloud and precipitating components in the WSM6 show much larger values and vertical extents than those in the WSM3 case. This is an important reason for the difference in DH and temperature between the WSM6 and WSM3 cases. Since there is low water condensation in the WPSH region (Noone et al. 2011; Noone 2012), the large amount of the simulated hydrometeors in the WSM6 over the SR must be transported from somewhere else. To understand the source of the hydrometeors in the SR, we depict the simulated hydrometeors at different levels (e.g., 300, 500, and 700 hPa) for the WSM3 and WSM6 cases in Fig. 11. Compared to those in WSM3, the simulated hydrometeors in WSM6 extend to a much larger area far away from the storm center in the upper troposphere despite fewer hydrometeors near the eyewall. Note that the difference in hydrometeors between WSM6 and WSM3 also varies with height. WSM6 produces many more anvil clouds above 500 hPa in terms of hydrometeors in the SR. These anvil clouds are far away from the storm center and can reach the SR, which is consistent with the large amount of cloud ice and snow in Figs. 10c and 10e. At the 700-hPa level, the WSM6 produces more hydrometeors...
in terms of precipitation in the SR, which is consistent with the larger value of rainwater in Fig. 10d. However, there is no significant difference in hydrometeors between the two cases at 0000 UTC 15 October before the storm entering into the model domain (figure omitted). The difference in hydrometeors between the WSM6 and WSM3 cases can be attributed to the excessive anvil clouds that extend far away from the TC center and reach the area of the WPSH in the WSM6 case. In other words, it is the overestimated anvil clouds extending from the TC in the upper troposphere that are responsible for the excessive hydrometeors over the SR in the WSM6 case. We will further discuss this issue in the next section.

c. The role of phase change

As discussed above, the microphysical heating (i.e., DH) caused by phase changes of water substance and excessive anvil clouds in WSM6 are responsible for the temperature difference between the WSM6 and WSM3 cases. Thus, it is necessary to investigate the hydrometeors and related phase changes in the TC region as the storm is approaching the SR.

Figure 12 shows azimuthal-averaged cross sections of microphysical heating and temperature in the two experiments and their difference at 0000 UTC 18 October. The grid spacing of 20 km is too coarse to well reproduce the structure of the TC eyewall (Fierro et al. 2009; Sun et al. 2013a), and the radius of the simulated eyewall in the two cases is notably larger than the observed radius (figure omitted). The microphysical heating is mostly concentrated in the area of eyewall convections and extends outward with height following the eyewall slope. Stern and Nolan (2009) showed that the outward slope of the eyewall with height is directly proportional to the radius of the eyewall. As expected, the WSM6 case is characterized by a less upright (i.e., larger slope) eyewall in terms of the latent heat release because of the larger eyewall radius when compared with the WSM3 case. As a result, underneath the more tilted eyewall in the WSM6 case, the downdrafts induced by evaporation of rain and melting of snow and graupel make the subcloud layers inflow drier and cooler than that in the WSM3, resulting in lower temperature below 500 hPa and in the boundary layer (Figs. 12a,b). This result is consistent with the modeling studies of Yang et al. (2007). Comparing with results in the WSM3 case, the WSM6 case can produce more microphysical latent heating near the eyewall on one hand and exhibit a feature of cooling in the upper levels and heating in the lower levels, which can promote the development of convection by increasing the convective instability on the other hand. This leads to a stronger secondary circulation and supports a larger low-level inward mass flux and larger volumes of mixed-phase particles aloft near the eyewall in the WSM6 case than in the WSM3 case (figure omitted).

As suggested by Powell (1990), the stratiform precipitation of the storm extends outward as anvil rain, and some of the strong anvil showers in the WSM6 case can generate penetrative downdrafts below the anvil cloud. This leads to an extended heat source due to condensation in the anvil cloud and an extended heat sink due to precipitation evaporation below the anvil cloud (Willoughby 1988). Note that the anvil clouds here refer to cloud water and cloud ice, while the anvil showers refer to rain, snow, and graupel in this study. Because of the wider and thicker anvil cloud in the WSM6 case, both the warming above and the cooling below 500 hPa are all stronger than those in the WSM3 case. This mechanism explains the temperature difference in the outer region of the TC and SR (Figs. 12c and 5b). Consistent with
Stossmeister and Barnes (1992), the changed temperature profile would cause a decrease of geopotential height at 500 hPa in SR (Fig. 5a) and weakening of WPSH (Fig. 4i), leading to an earlier northward turn of the TC in the WSM6 case compared to that in the WSM3 case (Fig. 2).

To further investigate the impacts of the MP scheme on temperature distribution, we examine individual components (i.e., cloud and precipitating component) related to microphysical heating. Figure 13 shows the azimuthal-averaged cross sections of cloud and precipitating components and hydrometeors in the two cases and their differences at 0000 UTC 18 October. As suggested by Wang (2002), cloud water is formed by condensation of supersaturated water vapor; thus, a high concentration of cloud component is closely related to the outward-tilted updrafts in the eyewall (Figs. 13a,b). In the WSM3 case with a simple ice phase, the cloud component (i.e., cloud water and ice) is concentrated in the eyewall region with a maximum at about 600 hPa. It contributes greatly to the eyewall updraft because of the related condensation warming effects (Fig. 13a). The distribution of cloud component in the WSM6 case is similar to that in the WSM3 case, except for a larger value and broader scope (Fig. 13b). This induces an excessive heating not only in the storm eyewall region, but also in the outer region of the storm, especially in the levels above 500 hPa (Fig. 13c). This is the main reason for higher temperature above 500 hPa in the WSM6 case (Fig. 12c).

The precipitating component in both the WSM3 and WSM6 cases is mostly concentrated in the middle and upper troposphere (above freezing level) in the form of snow, which melts into rain in the saturated region or sublimates to water vapor in the unsaturated region as it falls through the freezing (or melting) level. Below the freezing level, rainwater forms through conversion from the cloud water and grows by collecting cloud water and by melting of both snow and graupel, resulting in a large concentration there. Note that, along with the melting of snow, the evaporation of rain during its falling process is responsible for the cooling in the low troposphere below the freezing level (Figs. 13d,e). Although the maximum value of precipitating component in the WSM6 case is not larger and sometimes smaller than that in the WSM3 case, its extent is larger in terms of hydrometeors because of the larger eyewall slope. This results in a much greater extent of cooling below the freezing level because...
of melting and evaporating effects (Fig. 13f), which contributes greatly to the lower temperature below 500 hPa in the outer region of the storm in the WSM6 case (Fig. 12c) when compared with that in the WSM3 case.

Overall, compared with the WSM3 case, it is the difference in the cloud (precipitating) component that is responsible for the higher (lower) temperature above (below) the freezing level in the outer region of the storm in the WSM6 case. As the storm approaches the SR, the outer region of the storm extends and reaches the SR at 0000 UTC 18 October. The temperature profile in the SR is basically consistent with the temperature profile in the outer region of the storm and leads to a decrease in the geopotential height at 500 hPa, which subsequently contributes to the unrealistic break of WPSH in the WSM6 case.

d. Possible mechanisms

As mentioned in the section of PVT diagnosis, the failure in the simulation of the TC motion (e.g., earlier turning) is attributed to the unrealistic break of the WPSH in PRs. Further analysis reveals that the large bias in DH over the area of TC convections is responsible for the break of WPSH in the PR when the center of SR is far away from the TC center. How the MP scheme affects the DH and results in the break of the WPSH and an earlier turning of the TC in the PRs is summarized as follows.

The physical mechanisms responsible for the sensitivity of the WPSH and TC track to MP schemes are identified through comprehensive diagnosis and are schematically summarized in Fig. 14. Compared with the CR, the simulations with different MP schemes in PRs overestimate hydrometeors in the storm eyewall region. The excess hydrometeors in PRs extend outward with height following the eyewall slope and reach the outer region of the TC in the upper troposphere and thus result in excessive anvil clouds and showers (i.e., cloud and precipitating components) in the upper and middle troposphere. As the simulated TC approaches the WPSH, the excessive anvil clouds in PRs extend far away from the TC center and reach the area of the WPSH. The overestimated cloud and precipitating components in PRs cause condensation warming above 500 hPa and evaporative (melting) cooling below 500 hPa (freezing level) in the outer region of the TC. Such a temperature pattern is quite clear over the SR when the TC is approaching, leading to a decrease of 500-hPa geopotential height and the unrealistic break of WPSH in SR. This, in turn, prompts the earlier turning of TC.

5. Conclusions and discussion

The sensitivity of TC motion and the WPSH to MP schemes has been studied using the nonhydrostatic mesoscale WRF. It is found that the model simulation of the WPSH is usually realistic under normal conditions with the absence of TC activities, but it is quite sensitive to the MP scheme when TCs are active over the WNP. Unrealistic representation of microphysical process in TCs is responsible for the failed simulation of the WPSH. In this study, the simulation using the WSM3 scheme presents a much better result in terms of TC track and WPSH intensity, compared to results using the other three MP schemes. Consistent with the previous studies, the TC motion is sensitive to MP schemes.

It is noteworthy that there is no MP scheme appropriate for all TC cases because of the specificity of each TC case. Although the simulation with the WSM3 scheme produces a better result than those with the other three MP schemes in this study, it does not necessarily mean that the WSM3 scheme is more suitable than other MPs for the simulation of the WPSH when TCs are active over the WNP. The focus of this paper is on the sensitivity of the simulated TC motion and WPSH to MP schemes, instead of on the advantage of the WSM3 scheme in simulating the WPSH. Our conclusions are based on the comparisons between sensitivity experiments and less dependent on the simulation results.
In this study, the simulated storm intensities are not presented, because the 20-km grid spacing is too coarse to simulate the storm intensity accurately. For this reason, numerous RCM studies with coarse resolution are conducted to investigate the TC track and frequency rather than the TC intensity. On the other hand, the TC intensity may not be important in our efforts to understand how the TC feedback affects the simulation of WPSH intensity. Although the simulated TC intensity in PRs is weaker than that in the CR during the track turning (figure omitted), it is the TC in PRs rather than that in CR that weakens the simulated WPSH intensity, resulting in the unrealistic break of WPSH and, subsequently, the early turning of the TC. Thereby, the TC intensity plays an insignificant role in determining the simulation of the WPSH intensity and the TC track. Hence, the TC intensity analysis does not make much sense in the present study.

The potential vorticity tendency (PVT) diagnosis technique is utilized to estimate the contributions of the horizontal advection (HA), vertical transportation (VT), and diabatic heating (DH) to TC motion. Their differences between the PR and CR simulations are analyzed. HA makes the largest contribution to the difference in TC motion between the PR and CR. It slows down the westward movement but accelerates the northward movement of the TC in the PR. VT and DH also affect the TC motion, but their impacts are less significant than that of HA. It is the difference in HA that is responsible for the different TC motion simulated by the PR and CR, especially Megi’s unrealistic early northward turning in PR. The MP directly contributes to the early turning of the TC in PRs by changing the environmental flow (i.e., steering flow) of the TC. Because of the unrealistic break of the WPSH, the steering flow to the south of the WPSH is changed and results in the early turning of the TC in PR.

Further analysis shows that the TC feedback on the intensity of the WPSH plays an important role in determining the break of the WPSH and thus the early turning of the TC in PRs. The physical mechanisms responsible for the sensitivity of the WPSH and TC track simulation to MP schemes are identified through comprehensive diagnostic analysis. The large bias in DH simulation over the area of TC convection is responsible for the break of the WPSH in PRs. Specifically, compared with the WSM3 in CR, the failure in the simulation of the WPSH intensity and TC track using the other MP schemes in PRs is attributed to the over-estimation of anvil clouds, which extend far away from the TC center and reach the area of the WPSH. On the other hand, compared with that in the CR, both less heating near the eyewall and more heating in the outer spiral rainbands in PRs contribute to the increase of the simulated TC size and thus result in the large difference in TC size between PRs and CR. This contributes to the difference in the scope of anvil clouds extending over the upper troposphere over the WPSH and thus the difference in WPSH intensity and TC track between PRs and CR. The MP schemes used in PRs produce excessive hydrometeors in the outer region of the TC with more cloud and precipitation, which are reflected in the excessive anvil clouds and showers. This leads to condensation warming above 500 hPa and evaporative cooling below 500 hPa (freezing level) in the TC outer region, which results in the large bias in the simulations of DH and thus temperature profile. As the TC is approaching the WPSH, such a pattern of vertical temperature distribution causes the decrease of 500-hPa geopotential height and subsequent break of WPSH. Errors in the WPSH simulation described above change the large-scale steering flow, leading to the early turning of the TC. Through a series of feedback loops, the model eventually fails in the simulation of both the TC and the WPSH. It is important to note that, although the simulated intensity of the TC is much weaker than the observation, it does not imply that the “feedback” of the simulated TC to the simulated WPSH is insignificant. MP schemes cannot directly impact the simulation of the WPSH without the feedback of the TC; however, it may indirectly impact the simulation by affecting the TC activity. In other words, the MP affects the simulation of the TC activity (e.g., eyewall convection, anvil cloud and showers, hydrometeors distributions), which subsequently influences the simulation of WPSH intensity. The TC feedback on the intensity of WPSH plays a critical role in the model behavior of simulation for both the WPSH and TC.

Note that some of the conclusions are based on the comparison between the WSM3 case (CR) and WSM6 case (PR) in section 4. Similar results are found if we compare the WSM3 case and the other two cases (i.e., the Lin and Thompson cases). This is because vapor, rain, snow, cloud ice, cloud water, and graupel are held in six different arrays in all three MP schemes (i.e., WSM6, Lin, and Thompson) used in PRs. Unlike in the WSM3 scheme, supercooled water is allowed to exist in the three schemes. Gradual melting of snow and graupel falling below the melting layer is also described in the three MP schemes used in PRs. As a result, all three PRs produce a similar volume of hydrometeors in the outer region of the TC, which is much larger than that in the WSM3 case, especially in terms of cloud ice and snow. Note that the three experiments in PRs have similar simulation results in terms of the TC motion, WPSH intensity, and vertical distribution of hydrometeors and temperature. In fact, compared to the WSM3 scheme, the increase in the category of hydrometeors in the other
three MP schemes may improve the realism of the results. However, just as Mass et al. (2002) suggested, it does not necessarily ensure the improvement of the forecasts. This is because the increase in the category of hydrometeors may introduce extra errors in the simulation. In our present study, all three MPs in the PR experiments overestimate the hydrometeors in the TC, resulting in large biases in simulating the WPSH and TC track. It is necessary to improve the available MPs or develop a new MP that is appropriate for the TC–WPSH simulations. This will be a topic for our future research. In addition, although the CR with the WSM3 scheme produces the best results in this study, it does not necessarily imply that simulations with the WSM3 scheme can always perform better than other MPs for different TCs and WPSH simulations, since the performance of a specific physics parameterization scheme may be case dependent.

The results of this study are helpful for understanding the impact of various MP schemes on the WPSH and TC track simulation. The physical mechanism revealed in this study not only helps us better understand the TC and WPSH dynamics but also provides insight for further improvement in the MPs. It is noteworthy that, in addition to the MP scheme, the model-simulated anvil clouds may also be sensitive to other model physics schemes (e.g., cumulus, boundary layer parameterization schemes). In future work, we will investigate how changes in these physics schemes affect the model simulation of the WPSH and TC track.

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