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Comparison of Convective and Stratiform Precipitation Properties in Developing and Nondeveloping Tropical Disturbances Observed by the Global Precipitation Measurement over the Western North Pacific

Ke WANG

*Center for Monsoon System Research, Institute of Atmospheric Physics, Chinese Academy of Sciences, China
University of Chinese Academy of Sciences, China*

Guanghua CHEN

Center for Monsoon System Research, Institute of Atmospheric Physics, Chinese Academy of Sciences, China

Xinxin BI

*Key Laboratory of Meteorological Disaster, Ministry of Education/
Joint International Research Laboratory of Climate and Environment Change/
Collaborative Innovation Center on Forecast and Evaluation of Meteorological Disasters,
Nanjing University of Information Science and Technology, China*

Donglei SHI and Kexin CHEN

*Center for Monsoon System Research, Institute of Atmospheric Physics, Chinese Academy of Sciences, China
University of Chinese Academy of Sciences, China*

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Abstract

The tropical oceans spawn hundreds of tropical disturbances during the tropical cyclone (TC) peak season every year, but only a small fraction eventually develop into TCs. In this study, using observations from the Global Precipitation Measurement (GPM) satellite, tropical disturbances over the western North Pacific (WNP) from July to October during 2014–2016 are categorized into developing and nondeveloping groups to investigate the differences between satellite-retrieved convective and stratiform precipitation properties in both the inner-core (within 200 km of the disturbance center) and outer-core (within 200–400 km of the disturbance center) regions. The developing disturbances experience a remarkably more oscillatory process in the inner-core region than in the outer-core region. The large areal coverage of strong rainfall in the inner-core region of the disturbance breaks into scattered remnants and then reorganizes and strengthens near the disturbance center again. Contrarily, the precipitation characteristics in the nondeveloping group evolve more smoothly. It can be summarized that disturbances prone to developing into a TC over the WNP satisfy two essential preconditions in terms of precipitation characteristics. First, a large fraction of stratiform precipitation covers the region that is within 400 km from the disturbance center. The mean vertically integrated unconditional latent heating rate of stratiform and convective precipitation in the developing group above 5.5 km is 6.6 K h^{-1} and 2.4 K h^{-1} , respectively; thus, the

Corresponding author: Guanghua Chen, Center for Monsoon System Research, Institute of Atmospheric Physics, Chinese Academy of Science, P.O.Box 2718, Beijing 100029, China
E-mail: cgh@mail.iap.ac.cn
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stratiform rainfall makes a major contribution to the warming of the upper troposphere. Second, strong convective precipitation occurs within the inner-core region. Compared with stratiform precipitation, which plays a critical role in warming the mid-to-upper levels, the most striking feature of convective precipitation is that it heats the mid-to-lower troposphere. Overall, the formation of TCs evolving from parent disturbances can be regarded as an outcome of the joint contribution from the two distinct types (convective and stratiform) of precipitation clouds.

Keywords convective and stratiform precipitation; Global Precipitation Measurement; developing and nondeveloping disturbances

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1. Introduction

The formation of organized tropical cyclones (TCs) from unorganized cumulus convection is often accompanied by multiple-scale air–sea interactions. Favorable large-scale environmental fields are regarded as necessary conditions for TC formation: warm sea-surface temperature (SST), weak vertical wind shear, high mid-level moisture, and moist instability (Gray 1968). Over the Atlantic, within the critical layer where the zonal flow equals the easterly wave propagation speed, a nearly closed Lagrangian circulation can provide a favorable dynamic and thermodynamic sweet point for vorticity convergence and protect convection from unfavorable environment and dry air intrusion, which plays a significant role in TC formation (e.g., Dunkerton et al. 2009; Wang et al. 2010a; Wang 2012).

Over the Pacific, the tropical upper-tropospheric trough (Wang and Wu 2016, 2018), the monsoon trough (Wu and Duan 2015; Wu et al. 2013), and tropical waves (Chen and Huang 2009; Chen and Chou 2014) can jointly affect TC genesis. Although the synoptic circulation patterns that contribute to the formation of TCs vary from ocean to ocean, it is accepted that TC formation experiences indispensable mesoscale evolution. There are two distinct theories for the TC formation: top-down development and bottom-up development. For the top-down development, stratiform rainfall from a mesoscale convective system (MCS) can spin up a mid-level vortex, followed by the downward extension of the vortex, which can build up a near-surface cyclonic circulation (e.g., Bister and Emanuel 1997; Ritchie and Holland 1997). Contrarily, the bottom-up development emphasizes the effect of deep convection and associated vorticity anomaly on TC formation. Specifically, the convective

bursts (Tory et al. 2006) or vortical hot towers (VHTs) form, leading to the formation of a mesoscale vortex, which develops near the low-level cyclonic circulation center and eventually intensifies into a self-sustaining entity (Montgomery et al. 2006). Both the top-down and bottom-up theories are closely related to the behaviors of the two major types of precipitation cloud in the tropics: stratiform and convective. These two types of precipitation clouds have dynamically and thermodynamically different vertical profiles. The dynamic characteristics of stratiform precipitation cloud are featured by divergence at the lower and higher troposphere and mid-troposphere convergence, whereas convective precipitation cloud is divergent at the higher troposphere and convergent at the lower troposphere. Correspondingly, in stratiform precipitation, the diabatic cooling is associated with downdrafts, and evaporation is prominent below the melting layer, whereas the updrafts and diabatic heating occur in the upper troposphere. In convective precipitation, the strong updrafts and condensation heating are distributed throughout the whole depth of the troposphere (Houze 1997). When strong convection becomes weak, it transforms into an intermediary stage, in which the vertical profile of divergence appears more like stratiform than convective precipitation, except that the convergence layer is slightly higher than that in stratiform precipitation (Mapes and Houze 1993). The intermediary precipitation clouds are usually generated from older and less-active convection.

Numerous modeling and observation studies have been devoted to identifying the roles of the different cloud types in TC formation, structure, and intensity changes based on their distinct dynamic and thermodynamic features. For instance, using a high-resolution numerical simulation, Wang (2014) distinguished the simulated clouds into deep convection and cumulus

congestus and found that the cumulus congestus clouds humidify the lower-to-middle layers and stimulate a near-surface cyclonic circulation before TC genesis. Contrarily, the deep convective clouds play a prominent role in moistening the upper layers and strengthening the cyclonic circulation at the upper levels. Furthermore, Chen et al. (2018) examined the role of stratiform and convective diabatic processes in the changes in TC intensity and structural evolution by artificially modifying stratiform and convective heating and cooling in the near-core region.

On the basis of the observations, Fritz et al. (2016) studied the rainfall evolution during the TC genesis using the Tropical Rainfall Measuring Mission (TRMM) precipitation radar (PR) dataset. By tracing the precipitation types associated with the proto-vortex disturbances back to 3 days before the genesis of the named storms in the North Atlantic (NATL), Fritz et al. (2016) found that the areal coverage of stratiform precipitation increases despite the nearly constant rainfall rate from day -3 to day -1 before the TC formation. In addition, using the top height of the 20 dBZ PR, they partitioned the convective precipitation clouds into three categories—shallow cumulus, mid-level convection (namely, cumulus congestus), and deep convection—which jointly contribute to the TC formation. Deep convection has the highest rainfall rate, although this type of precipitation has a small areal coverage. Contrarily, mid-level convection with the largest areal coverage percentage among the three types of convective precipitation (shallow, mid-level, and deep convection) plays a major role in the contribution to the total precipitation. In addition, the variation in the areal coverage of stratiform and convective precipitation clouds has the potential to predict the rapid intensification (RI) of a TC. Using 16-yr TRMM PR observations, Tao et al. (2017) found that the percentage of stratiform precipitation significantly increases in all shear-relative quadrants when TCs start RI. Compared with RI storms, slowly intensifying storms have a relatively smaller fraction of stratiform coverage within the TC inner-core region. Furthermore, the presence of vigorous convective precipitation in the upshear-left quadrant also plays a critical role in TC intensification. Overall, using the TRMM database, most previous studies have focused on the differences between convective and precipitation properties during the changes in TC intensity (Jiang 2012; Kieper and Jiang 2012; Zagrodnik and Jiang 2014).

Over the tropical oceans, the large-scale environment can spawn several hundreds of disturbances

during the TC peak season every year, but only a small fraction develop into TCs. Using global daily reanalysis data, Peng et al. (2012) studied the large-scale environmental differences between disturbances that develop into TCs over the NATL and those that don't. For the disturbances that develop into TCs, they found that the thermodynamic fields (such as water vapor content, rainfall rate, and SST) are more instructive than the dynamic variables (such as maximum relative vorticity at 700 hPa, vertical shear, translational speed, and horizontal shear). Given the pronounced differences in the large-scale environmental background over the Atlantic and the western North Pacific (WNP), Fu et al. (2012) examined the developing and nondeveloping disturbances over the WNP and drew the opposite conclusion, that is, the dynamic fields play a more important role in TC genesis than the thermodynamic fields. In addition to the discrepancy in the large-scale environmental conditions, the properties of convection and precipitation are also anticipated to differ between developing and nondeveloping disturbances. However, this has not yet received much attention.

The Global Precipitation Measurement (GPM) satellite was launched in February 2014, with a more sensitive onboard Dual-Frequency Precipitation Radar (DPR) than TRMM PR and 13 Microwave Imager (GMI) channels. Using the GPM dataset, this study compares the differences in the evolution and three-dimensional structure of key convection and precipitation types for developing and nondeveloping disturbances over the WNP. The precipitation properties in the TC formation process may exhibit marked regional characteristics. The large-scale environmental circulation controlling most of the TC formations in the NATL is related to the easterly wave, whereas the WNP features various synoptic circulation patterns. Therefore, it would make more sense to study the distinctive precipitation characteristics in the WNP. This is unique compared with previous studies, which have paid more attention to the developing group in the NATL or large-scale environmental conditions during TC genesis. Comprehensive comparisons of the structure and evolution in the different cloud types between the developing and nondeveloping disturbances can help determine the convection and precipitation properties and guide predictions during the TC pregenesis stage.

The remainder of this paper is structured as follows. The next section describes the dataset and analysis methods. The differences in convection and precipitation characteristics between the inner- and outer-core

regions for the developing and nondeveloping disturbances are presented in Section 3, and a summary is given in Section 4.

2. Data and methodology

2.1 Data

In this study, the data analyzed to determine the convection and rainfall properties were obtained from the GPM DPR (available online <https://pmm.nasa.gov/gpm>). The DPR is formed from a Ka-band precipitation radar (KaPR) running at 35.5 GHz with a 125-km swath and a Ku-band precipitation radar (KuPR) operating at 13.6 GHz with a 245-km swath. The shorter wavelength of the KaPR makes it easier to measure light rain and solid precipitation (Hamada and Takayabu 2016). Moreover, the differential attenuation for the DPR is more accurate than the Ku- or Ka-band attenuations derived separately (Meneghini et al. 2012). The GPM DPR standard level 2 (L2) product (2ADPR), version 05A (V05A), is used in this study (Iguchi et al. 2012; Seto and Iguchi 2015). The 2ADPR product records three scanning modes: normal scan (NS), matched scan, and high sensitivity scan. In this study, we used the NS mode, in which the inner swath of the KuPR and KaPR swaths are measured synchronously to match the KuPR and KaPR beams. The NS consists of the three-dimensional radar reflectivity and precipitation rate and two-dimensional variables, including the precipitation type and the near-surface precipitation rate, which are analyzed in this study. The precipitation pixels are classified into three main categories: stratiform, convective, and other. The horizontal resolution of all variables in 2ADPR is 5 km, and the vertical resolution is 125 m for the vertical profiles of radar reflectivity and rainfall rate.

The 2ASLH_V05A product, as well as the 2ADPR product, is employed to obtain latent heating. The composite latent heating profiles for the precipitation pixels are used to examine the thermodynamic properties of the different precipitation types. The GPM Spectral Latent Heating (SLH) algorithm in the tropics is identical to that in the TRMM PR data (Shige et al. 2004, 2007, 2008). Shige et al. (2004) proposed the SLH algorithm and applied it to TRMM PR products. The specific methods are as follows: (i) using the variables (precipitation-top height, rain rates at the surface and melting level, and rain type) of the cloud-resolving model (CRM) simulation to construct the heating profile lookup tables of three precipitation types (convective, shallow stratiform, and anvil) and then verifying the reliability of lookup tables; (ii)

according to the PR information observed by TRMM (precipitation-top height, rain rates at the surface and melting level, and rain type) to identify the heating profile from lookup tables. Shige et al. (2007) improved this algorithm. According to the ice and liquid water process, the convective heating profile was divided into upper-level heating and lower-level heating, respectively. Contrarily, the heating profile of stratiform precipitation is shifted up or down according to the melting layer. In this study, the three-dimensional profiles of latent heating were also composited based on the classification of rainfall types defined in the 2ADPR product.

The disturbances of interest are confined spatially from the equator to 30°N and 100°E–150°W and temporally from July to October. Owing to the GPM operation starting from spring 2014, the year period in this study spans from 2014 to 2016.

2.2 Categorization of developing and nondeveloping disturbances

The 6-hourly data from the European Centre for Medium-Range Weather Forecasts interim reanalysis (ERA-Interim, Berrisford et al. 2011) with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ are used to identify two groups of disturbances: developing and nondeveloping disturbances. In the first group, tropical disturbances develop into at least tropical depressions (TDs), with a maximum sustained wind speed over 25 kt (12.9 m s^{-1}). Using the best track data from the Joint Typhoon Warning Center (JTWC) that specify the intensity of cyclone by using 1-min mean sustained wind speed based on the Dvorak technique (Song et al. 2010; Barcikowska et al. 2012), the time at which a tropical disturbance first upgrades to a TD is indicated as day 0. Then, the cyclonic circulation center is traced back 3 days according to the minimum of the 3–8-day filtered 850-hPa stream function, which are designated as days -1 , -2 , and -3 , respectively. The TC information in the JTWC dataset includes the central location at the disturbance stage before upgrading to TDs. For example, for typhoon Nangka (2015), the center location during its disturbance stage is 8.8°N and 172.2°E in the JTWC dataset, and our positioning is 9°N and 172.5°E. The comparison confirms that the TC centers identified by the minimum of the 3–8-day filtered stream function coincide well with those recorded in the JTWC dataset, justifying the method used in this study to identify the disturbance centers. In some developing cases, the disturbances intensify rapidly, such that only 1 days or 2 days before genesis can be traced. In such cases, the characteristic vari-

Table 1. The number of developing and nondeveloping disturbances over the WNP from 2014 to 2016.

	Developing					Nondeveloping				
	July	August	September	October	Total	July	August	September	October	Total
2014	5	0	7	1	13	8	15	3	6	32
2015	5	4	5	4	18	6	13	8	8	35
2016	5	6	6	3	20	10	5	11	9	35
Total	15	10	18	8	51	24	33	22	23	102

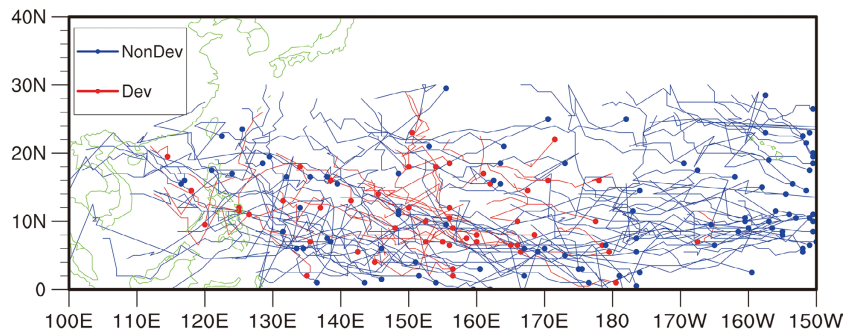


Fig. 1. The tracks (from day -3 to day -1) of all developing and nondeveloping disturbances over the WNP. Blue (red) solid lines refer to the nondeveloping (developing) disturbance tracks. The filled dots indicate the disturbance center locations on day -3 .

ables are composited together with those in other developing cases on a corresponding day. Table 1 presents the distribution of the number of developing and nondeveloping disturbances from July to October during 2014–2016. There are 51 developing cases, among which the total number in October is the lowest and no TDs formed in August 2014 due to the unfavorable large-scale conditions (Bian et al. 2018).

The second group consists of the nondeveloping disturbances that fail to develop into TDs. Similarly, the nondeveloping disturbances are manually traced back 3 days before they dissipate. Five criteria are used to select the nondeveloping cases: (1) the mean radius of cyclonic circulation must be larger than 300 km; (2) there must be a closed center in the stream function; (3) if the disturbance merges into other larger and stronger circulation or is too obscure to identify its center, its life is considered to terminate; (4) the stream function center should be located in the domain of interest (0° – 30° N and 100° E– 150° W); and (5) its lifetime is required to last for at least 3 days. Day 0 for the nondeveloping disturbances refers to the time of life termination. As a result, a total of 102 nondeveloping cases are selected. As can be seen in Table 1, the number of nondeveloping cases is highest in August, suggesting that the synoptic disturbances

over the WNP are the most active in August, although the proportion of disturbances upgrading to TDs is the lowest due to the relatively small number of developing cases in August.

As it takes about 92.5 min for one GPM orbit to scan the earth once, the original 6-hourly TC center locations are linearly interpolated into 1-hourly intervals to obtain a sufficient number of samples. Figure 1 presents the tracks of all developing and nondeveloping disturbances selected in the 3 years. It is evident that almost all developing disturbances emerge to the west of the dateline. Some developing disturbances can be found on the equatorward side of 5° N, though the planetary vorticity is small, indicating that the developing disturbances in the equatorial region could be linked with the equatorial waves in collaboration with the favorable environmental conditions.

2.3 Orbit selection

Fritz et al. (2016) used the TRMM satellite data to study the different precipitation characteristics of TCs that originated within the wave pouches in the Atlantic. They defined the inner-core region as the area within 200 km of the disturbance center. The precipitation characteristics in this inner-core region are mainly related to the meso- β -scale process, whereas

the entire circulation of the disturbance represents the meso- α -scale process (Wang 2018). Therefore, to study the precipitation evolution and difference between the inner- and outer-core regions during the TC formation, the GPM-retrieved variables within the radii of 200 km (inner-core region) and 200–400 km (outer-core region) from the disturbance center are analyzed. This comparison confirms that the conclusions are not sensitive to the selection of the inner- and outer-core region, because the different choices of radii for both the inner-core region (100, 150, or 250 km) and outer-core region (200–500 or 200–600 km) obtain qualitatively consistent results (not shown). To avoid the influence of small-coverage overpass, the other imposed criterion is that an effective GPM DPR swath has at least 4000 pixels (including non-precipitation pixels) within 400 km from the centers of the developing and nondeveloping disturbances. As a result, the total numbers of effective orbits in the developing and nondeveloping groups from day -3 to day 0 are 71 and 179, respectively. These orbits swept through 44 developing disturbances and 104 nondeveloping disturbances, respectively. The composite on a given day is the average over 1 day later. For example, the composite on day -1 in the developing group refers to the mean overpass samples from day -1 to day 0.

3. Results

3.1 Spatiotemporal evolution of precipitation

The disturbance development and dissipation in the developing and nondeveloping groups are accompanied by precipitation variation, and the spatiotemporal evolution of the near-surface precipitation reflects the characteristics of the two types of disturbance. Figure 2 presents the composite near-surface rainfall rate over the 3 days before TC generation or disturbance dissipation within 400 km from the disturbance center. We define the threshold of strong precipitation as 1 mm h^{-1} . Table 2 presents the percentages of strong precipitation pixels to the total precipitating pixels in both the inner- and outer-core regions for the developing and nondeveloping groups. The statistical significance determined by bootstrap test (Efron and Gong 1983) at the 90 % level is denoted by black dot, indicating that there are sufficient strong precipitation samples at the given pixel. For the developing group, the synthetic near-surface rainfall rate on day -3 (Fig. 2a) indicates that in the inner-core region, the number of strong precipitation pixels accounts for 10.3 % of all precipitation pixels. In addition, the strong precipitation occurs near the disturbance center and is well-organized, as represented by the contiguous

coverage of strong precipitation. In the outer-core region, such a strong precipitation with large contiguous coverage is mainly distributed on the east side of the disturbance center, and the number of precipitation pixels only accounts for 4.8 % of all precipitation pixels. However, the outer-core characteristics of the strong precipitation for the developing group on day -3 aren't statistically significant at a 90 % confidence level, probably due to the limited sample size. On day -2 (Fig. 2b), the strong precipitation areas in both the inner- and outer-core regions become looser and are no longer organized as day -3 . The number of strong precipitation pixels in the inner- and outer-core regions accounts for 7.7 % and 5.6 % of all precipitation pixels, respectively. On day -1 (Fig. 2c), the strong precipitation within the inner-core region begins to revive, reorganizing around the disturbance center. The percentage of strong rainfall in the inner- and outer-core regions increases to 9.7 % and 6.5 %, respectively. From the spatiotemporal distribution characteristics of the mean surface rainfall rate in the developing group, it can be found that the strong precipitation in the inner-core region experiences an oscillatory development, with the well-organized structure breaking down into scattered reservoirs with time (Figs. 2a, b) and then rebuilding in the vicinity of the disturbance center (Fig. 2c). Contrarily, the percentage of strong precipitation in the outer-core region increases gradually with time. Overall, the proportion of strong precipitation in the inner-core region is higher than that in the outer-core region.

Compared with the developing group, the occurrence of strong precipitation in the nondeveloping group is scattered and intermittent only away from the center of the disturbance (Figs. 2d–f). From day -3 to day -1 , the ratios of strong precipitation pixels to the total precipitation pixels in the inner-/outer-core regions are 2.6 %/1.7 %, 1.4 %/1.2 %, and 0.4 %/0.7 %, respectively, which are significantly less than those in the developing group, indicating that the percentages of strong rainfall in both the inner- and outer-core regions gradually decrease with time during the disturbance dissipation in the nondeveloping group.

3.2 Statistics of convective and stratiform precipitation properties

The precipitation pixels in the 2ADPR product are classified into three rainfall types: stratiform, convective, and other. As the “other” rain type mostly comprises anvil rainfall, the mean near-surface precipitation rate and the proportion of anvil rainfall are small enough to be neglected in this study. In this sub-

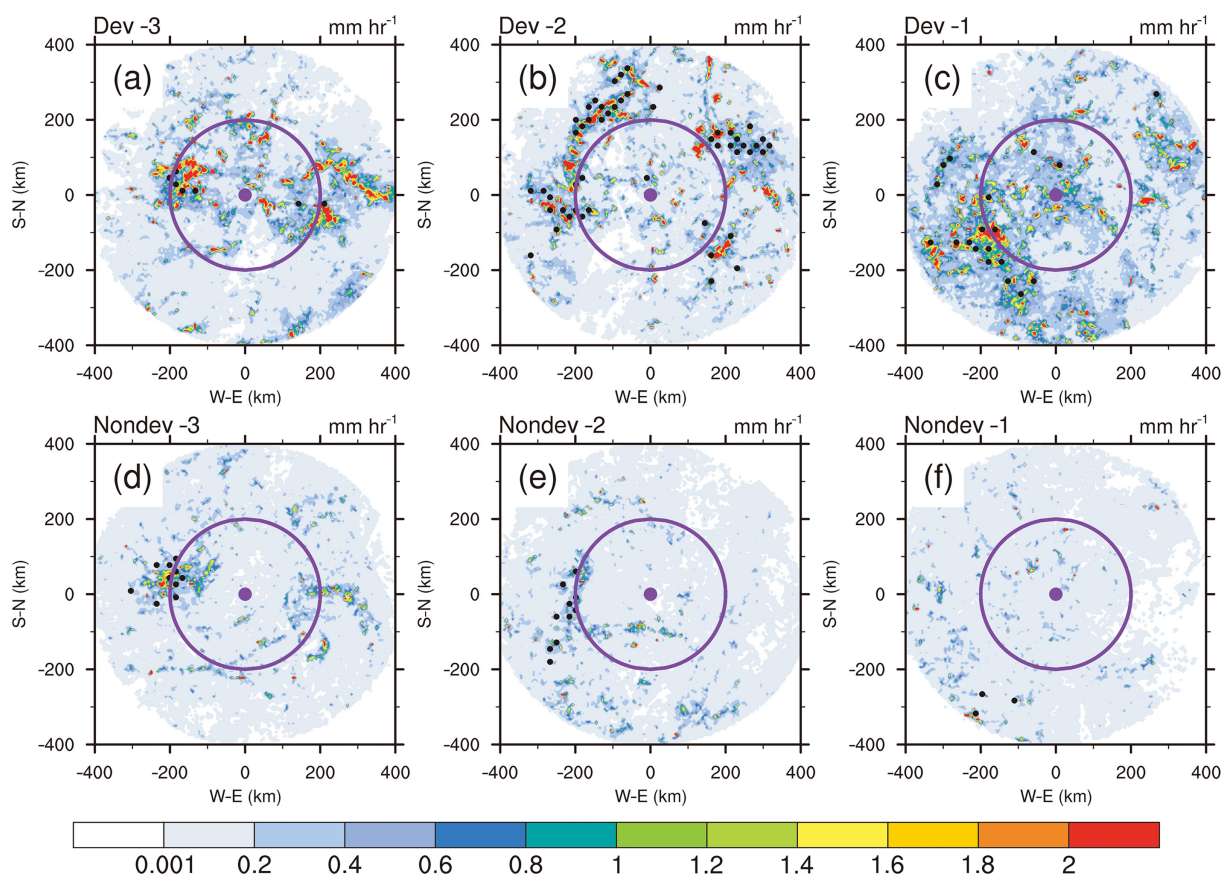


Fig. 2. The composite mean near-surface rainfall rates (mm h^{-1}) at different times for the developing and nondeveloping groups. The dark violet point indicates the center of the disturbance, and the dark violet circle indicates the radius of 200 km from the disturbance center. Statistical significance at the 90 % level is indicated by black dot.

section, four indicators are used to quantitatively study the characteristics of convective and stratiform types in the inner- and outer-core regions in the two groups. Taking the inner-core region as an example (Fig. 3), the first indicator is the percentage of the number of precipitation pixels of a certain precipitation type to the total number of precipitation pixels (Fig. 3a, $N_{\text{str/con}}/N_{\text{precipitating}}$), which reflects the areal fraction of a specific precipitation type in the precipitation region. The second indicator is the ratio of the specific precipitation pixels to the total pixels (including non-precipitation grid points), which represents the coverage area of the different precipitation types (Fig. 3b, $N_{\text{str/con}}/N_{\text{total}}$). The sum of stratiform and convective areal coverage percentages occupies the large majority of the precipitation coverage, and the residual percentage nearly accounts for the non-precipitation coverage percentage. The third indicator is referred to as the conditional mean rainfall rate (Fig. 3c,

Table 2. The percentages of strong precipitation pixels to the total precipitating pixels in both the inner- and outer-core regions at different times for the developing and nondeveloping groups.

	Developing		Nondeveloping	
	inner-core	outer-core	inner-core	outer-core
Day -3	10.3 %	4.8 %	2.6 %	1.7 %
Day -2	7.7 %	5.6 %	1.4 %	1.2 %
Day -1	9.7 %	6.5 %	0.4 %	0.7 %

$\Sigma R_{\text{str/con}}/N_{\text{str/con}}$), which is calculated as the total precipitation rate divided by the total number of precipitation pixels for a certain precipitation type that signifies the precipitation intensity. The fourth indicator is the unconditional mean rainfall rate (Fig. 3d, $\Sigma R_{\text{str/con}}/N_{\text{total}}$), which is calculated as the total precipitation rate of a certain precipitation type divided by the total number

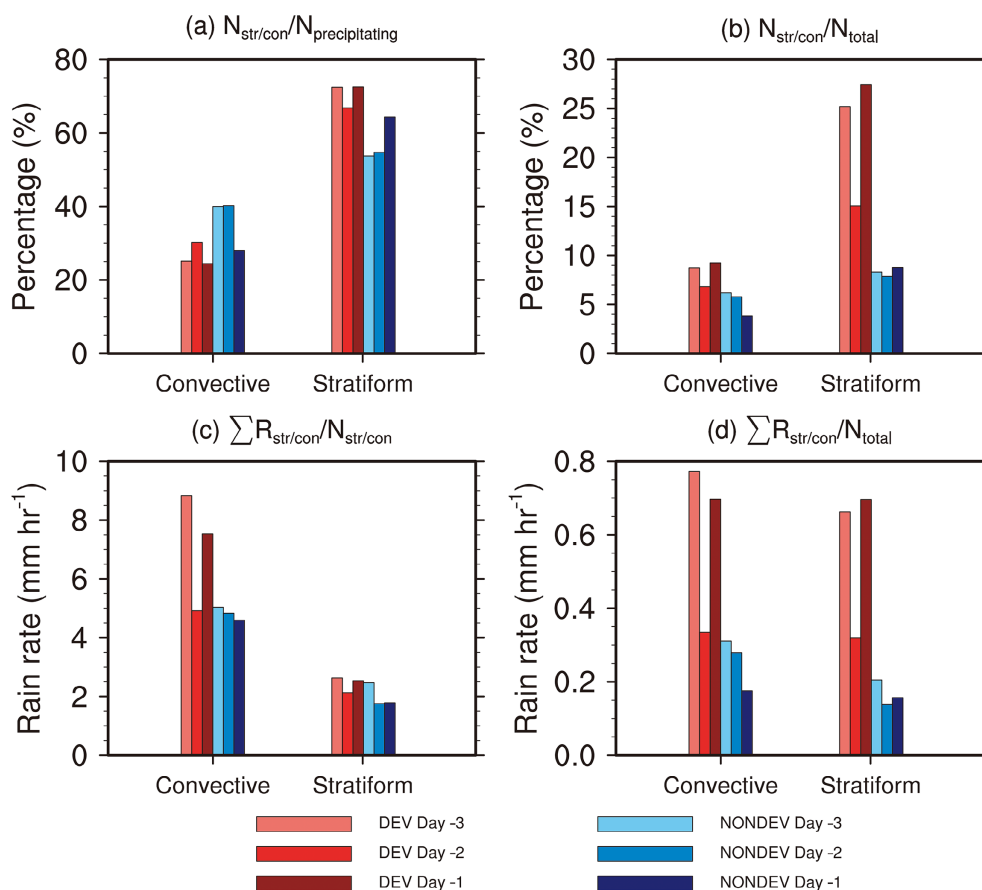


Fig. 3. (a) The ratio of the specific precipitation pixels to the total precipitation pixels ($N_{str/con}/N_{precipitating}$), (b) the ratio of the specific precipitation pixels to the total pixels ($N_{str/con}/N_{total}$), (c) the conditional mean near-surface rainfall rate ($\sum R_{str/con}/N_{str/con}$), and (d) the unconditional mean near-surface rainfall rate ($\sum R_{str/con}/N_{total}$) within a 200-km radius of the circulation center. Red (blue) bars with light-to-dark colors represent the different days in the developing (nondeveloping) group.

of pixels (including non-precipitation pixels). The unconditional mean rainfall rate is equivalent to the product of the second and third indicators, which can be regarded as the standardized total precipitation rate of this precipitation type.

We first compare the differences between the convective and stratiform precipitation properties in the developing and nondeveloping groups in the inner-core region. As presented in Fig. 3a, the stratiform precipitation pixels in the developing group account for about 71 % of the total precipitation pixels, which is similar to the result of Schumacher and Houze (2003), who found that 73 % of the tropical precipitation area was covered by stratiform precipitation using the TRMM PR data from 1998 to 2000. The percentage of stratiform precipitation in the developing group

is higher than that in the nondeveloping group, which implies that more well-organized precipitation systems occur in the developing group. Contrary to the counterpart in the developing group, the percentage of convective precipitation in the nondeveloping group is more dominant. Specifically, the percentages of convective and stratiform precipitation averaged over the 3 days in the developing (nondeveloping) group are 27 % (36 %) and 71 % (58 %), respectively, indicating the potential importance of stratiform precipitation in TC formation. In terms of the temporal evolution, the percentage of convective (stratiform) precipitation in the developing cases varies, with 25 % (72 %) on day -3, 30 % (67 %) on day -2, and 24 % (73 %) on day -1. In the nondeveloping group, the convective precipitation percentage decreases gradually from day

–3 to day –1, whereas the proportion of stratiform precipitation exhibits a progressive increase.

To further examine the areal coverages of the different precipitation types, Fig. 3b presents the occurrence frequencies of convective and stratiform precipitation pixels normalized by the total pixels (including non-precipitation pixels). The average total precipitation areal coverage of both convective and stratiform from day –3 to day –1 in the developing group makes up 30.8 %, which is much larger than the 13.6 % in the nondeveloping group, indicating that the nondeveloping group has more non-precipitation pixels in the inner-core region. In both the developing and nondeveloping groups, the areal coverages of stratiform precipitation are higher than those of convective precipitation, but the contrast in the areal coverage between stratiform and convective precipitations is more substantial in the developing group than in the nondeveloping group. The mean areal coverage of stratiform precipitation from day –3 to day –1 in the developing group is 14.3 % higher than that of convective precipitation, whereas the corresponding difference in the nondeveloping group is only 3.1 %. The moistening exerted by stratiform precipitation in the RI events was studied by Tao et al. (2017), who found that RI storms are accompanied by higher occurrence and larger areal coverage of stratiform rainfall around the storm center compared with the non-RI storms. In this study, TC formation is also collocated with the larger areal coverage of stratiform precipitation. Furthermore, the areal coverage of stratiform precipitation decreases from day –3 to day –2 and increases from day –2 to day –1 in the nondeveloping group, whereas convective precipitation decreases from day –3 to day –1. Contrarily, the areal coverages of both convective and stratiform precipitations decrease from day –3 to day –2 and increase from day –2 to day –1 in the developing group, indicating an oscillatory development in the developing group, as described in Figs. 2a–c. Moreover, compared with convective precipitation, the reduction in stratiform precipitation coverage is more pronounced from day –3 to day –2.

This oscillatory development is also observed in other cases of TC genesis. For instance, during the formation of typhoon Manyi (2001), Xu et al. (2014) found that stratiform precipitation started developing with downdrafts after the active convection triggered convective precipitation, and then, both the categories of precipitation became active until a few hours later. Li et al. (2006) used the numerical model TCM3 to study the causes of the oscillatory development of low-level vorticity during the TC formation. Their

results revealed that the consumption and recovery of moisture in the planetary boundary layer (PBL) and the interaction between the PBL and the circulation led to oscillatory development. Specifically, the strong updrafts related to deep convection will transport the water vapor from the PBL to the upper-level troposphere, leading to the consumption of PBL moisture. Subsequently, the evaporation of convective rainfall caused the mid-level cooling and the downdrafts. Meanwhile, the moisture in PBL will be supplied by the sea-surface evaporation, which will increase the instability of the atmosphere and initiate new convection, thus showing the oscillatory development of vorticity.

In addition to the areal coverage of stratiform and convective precipitations, the conditional mean rainfall rate that reflects the precipitation intensity is presented in Fig. 3c. It is clear that the conditional mean near-surface precipitation rate of convective precipitation is much stronger than that of stratiform precipitation. Moreover, the near-surface convective precipitation rates in the developing group are greater than those in the nondeveloping group. The evolutions of both stratiform and convective precipitation intensity in the developing group have the same trends as the variation of areal coverage percentage (Figs. 3b, c), with a decrease from day –3 to day –2 and an increase from day –2 to day –1. In particular, the intensity of convective precipitation indicates a significant decrease of 44.0 % from day –3 to day –2. As a result, stratiform and convective precipitations in the developing group on day –2 exhibit distinct characteristics in the inner-core region during the oscillatory development, that is, stratiform precipitation features a large reduction in areal coverage (Fig. 3b), whereas convective precipitation is characterized by a significant reduction in precipitation intensity (Fig. 3c). The t-test shows that the differences in convective rainfall rates between day –3 and day –2 and day –2 and day –1 are all statistically significant with a 99.9 % confidence level, indicating the existence of oscillatory development of convective rain rate in the inner-core region for the developing group. In the nondeveloping group, the convective precipitation intensity indicates a slight decrease from day –3 to day –1, whereas the stratiform precipitation intensity decreases from day –3 to day –2 and then remains almost unchanged.

Contrary to the conditional mean rainfall rate in Fig. 3c, Figure 3d presents the unconditional mean near-surface rainfall rates of the different precipitation types. Recall that the unconditional mean near-surface precipitation rate ($\Sigma R_{\text{str/con}}/N_{\text{total}}$) is equal to the product

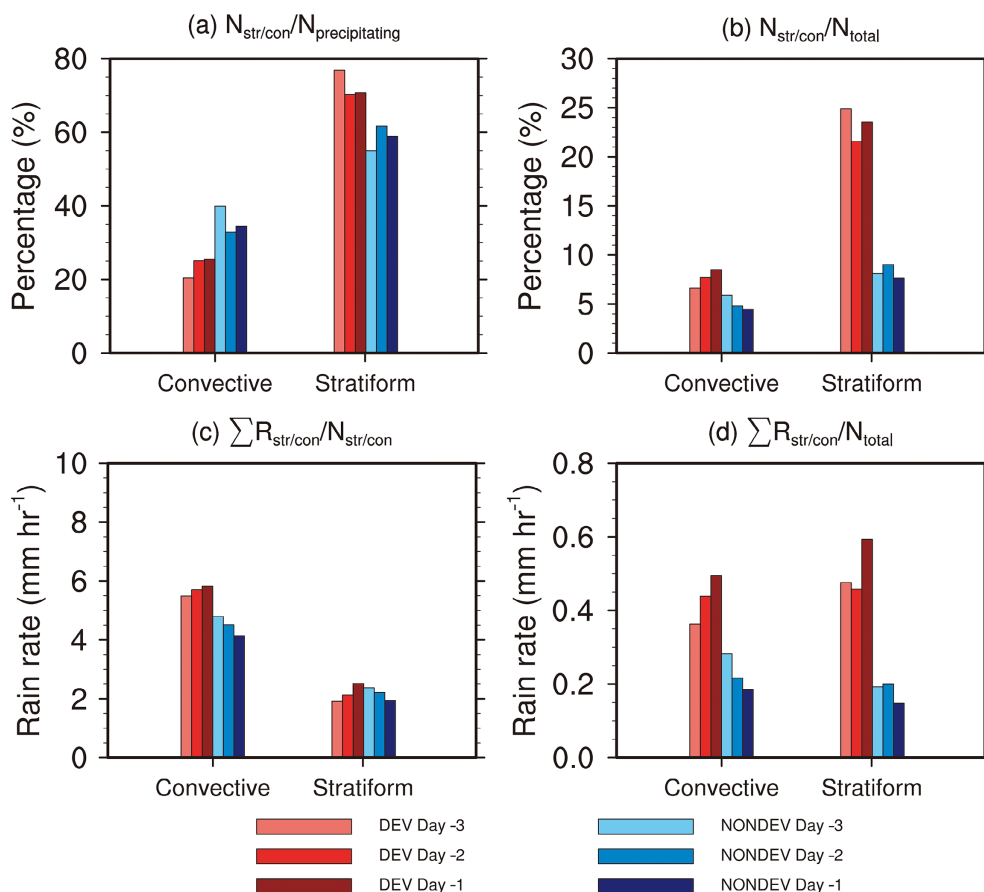


Fig. 4. As in Fig. 3, but within the radii of 200 km and 400 km of the circulation center.

of precipitation areal coverage ($N_{\text{str/con}}/N_{\text{total}}$) and precipitation intensity ($\sum R_{\text{str/con}}/N_{\text{str/con}}$). Despite a much lower areal coverage of convective rainfall than stratiform rainfall in the developing group, as presented in Fig. 3b, the unconditional mean convective precipitation rate is stronger than the stratiform counterpart due to the higher intensity of convective precipitation (Figs. 3c, d). In the nondeveloping group, given the distinct difference in precipitation intensity and the moderate contrast in areal coverage between convective and stratiform precipitations, the unconditional mean near-surface convective precipitation rate is dominant over the stratiform one.

The similarities and dissimilarities in the precipitation characteristics between the outer-core (200–400 km) and inner-core (0–200 km) regions are presented in Fig. 4. The main qualitative similarities in the precipitation characteristics within these two regions are as follows: (1) The percentage of stratiform precipitation in the developing group is higher than that in the

nondeveloping group (Fig. 4a), with the 3-day mean stratiform precipitation pixels accounting for 72.6 % (58.5 %) of the total precipitation pixels in the developing (nondeveloping) group. (2) The nondeveloping group in the outer-core region also comprises more non-precipitation pixels, that is, the 3-day mean percentage of non-precipitation pixels in the nondeveloping group is 17.6 % higher than that in the developing group (Fig. 4b). (3) The mean intensity of convective precipitation in the developing group is higher than the counterpart in the nondeveloping group (Fig. 4c), with the 3-day mean convective rainfall rate in the developing group being 1.2 mm h⁻¹ higher than that in the nondeveloping group. (4) The unconditional convective rainfall rate of convective precipitation is comparable to that of stratiform precipitation in the developing group but is stronger in the nondeveloping group (Fig. 4d).

The dissimilarities are depicted by the quantitative differences in the precipitation characteristics in the

different groups. One striking feature is that no notable oscillatory development occurs in the outer-core region of the developing group, especially in terms of the areal coverage of stratiform precipitation, and the conditional and unconditional mean rainfall rate of convective precipitation. This indicates that, during the development stage, the precipitation systems in the inner-core region apparently wax and wane, whereas the precipitation systems in the outer-core region experience relatively stable evolution. In addition, the 3-day mean conditional rainfall rate of convective precipitation in the inner- and outer-core regions is 7.1 mm h^{-1} and 5.7 mm h^{-1} , respectively, whereas for stratiform precipitation, it is 2.4 mm h^{-1} and 2.2 mm h^{-1} , respectively. This indicates that the difference in precipitation intensity between the inner- and outer-core regions is more remarkable for convective precipitation (Figs. 3c, 4c), resulting in a total unconditional convective precipitation rate that is less than the stratiform precipitation rate in the outer-core region in the developing group (Fig. 4d).

On the basis of the above analyses and comparisons, the preconditions for the development of tropical disturbances into TCs over the WNP are as follows. First, vast stratiform clouds covering the disturbance core regions are required. In particular, the areal coverage of stratiform precipitation can act as a proxy for the system organization degree in the developing group. The widely spreading stratiform precipitation can continuously humidify the lower atmosphere over a large area and spin up mid-level vortices from MCSs, followed by downward extension that can promote a near-surface vortex (Bister and Emanuel 1997). Using observations, Didlake and Houze (2013) also found that secondary circulation caused by stratiform precipitation leads to the convergence of angular momentum and increases the tangential wind.

Second, strong convective precipitation occurs within the inner-core region of developing disturbances. Hot towers and convective bursts near the vortex center are related to the intensification of TC (Steranka et al. 1986). The intense convection in the developing group can transport more water vapor upward to the higher levels, along with the collection of surrounding larger raindrops or ice particles in the fall path to produce a more substantial concentration of liquid water. Furthermore, although the finite energy of the atmosphere and ocean can only make the strong convections maintain for a short time, the vorticity anomalies after the convection outbreak can maintain for a long time, which is conducive to the merging of MCV and stimulation of the secondary circulation (Montgomery

et al. 2006). For a balanced vortex with a heating effect released by the VHT in the inner-core region, Vigh and Schubert (2009) studied the responses of the vortex to the diabatic heating that plays a critical role in the formation of the warm core in TCs. The responses are formulated by three factors: baroclinity, static stability, and inertial stability, which is higher near the vortex center. Diabatic heating more effectively intensifies the vortex when it occurs within the radius of maximum wind (Schubert and Hack 1982; Hack and Schubert 1986; Nolan et al. 2007; Rogers et al. 2016). Thus, the deep-tropospheric convective episodes in the inner-core region can effectively warm the atmospheric column to facilitate TC formation.

3.3 *The vertical properties of stratiform and convective precipitation*

To compare the differences in the vertical properties of stratiform and convective precipitation in the developing and nondeveloping groups, the vertical profiles of three-dimensional conditional rainfall rate, reflectivity, and unconditional latent heating in the inner-core region are presented in Fig. 5. For both the rainfall rate and reflectivity, we adopt the range bins above “binClutterFreeBottom” to avoid the contaminated data (Hamada and Takayabu 2016). As presented in Figs. 5a, b, the maximum conditional rainfall intensity of stratiform and convective precipitation in the two groups of disturbances is located near the surface. The convective rainfall rate decreases rapidly with height, and the stratiform rainfall rate decreases sharply with height above 4 km. The comparison of the vertical precipitation profiles in the two groups reveals that the most striking difference is that the intensity of convective precipitation in the developing group is higher than that in the nondeveloping group, but the intensity in the developing group oscillates over the 3 days (Fig. 5a). Moreover, the mean stratiform rainfall rate in the developing group is slightly stronger than that in the nondeveloping group.

The observations show that stratiform precipitation exists in a layer of pronounced high reflectivity around 4–5 km, referred to as the “bright band”, in which the downward-settling ice particles melt (Stewart et al. 1984). As presented in Figs. 5c, d, the stratiform rainfall features a bright band located near the 4.5-km height, consistent with previous observational findings (Fritz et al. 2016). Contrarily, the maximum reflectivity of convective precipitation is around 3.5 km, reflecting more precipitation particles concentrating in the lower troposphere. Moreover, the maximum 30-dBZ echo-top height that is treated as a proxy for

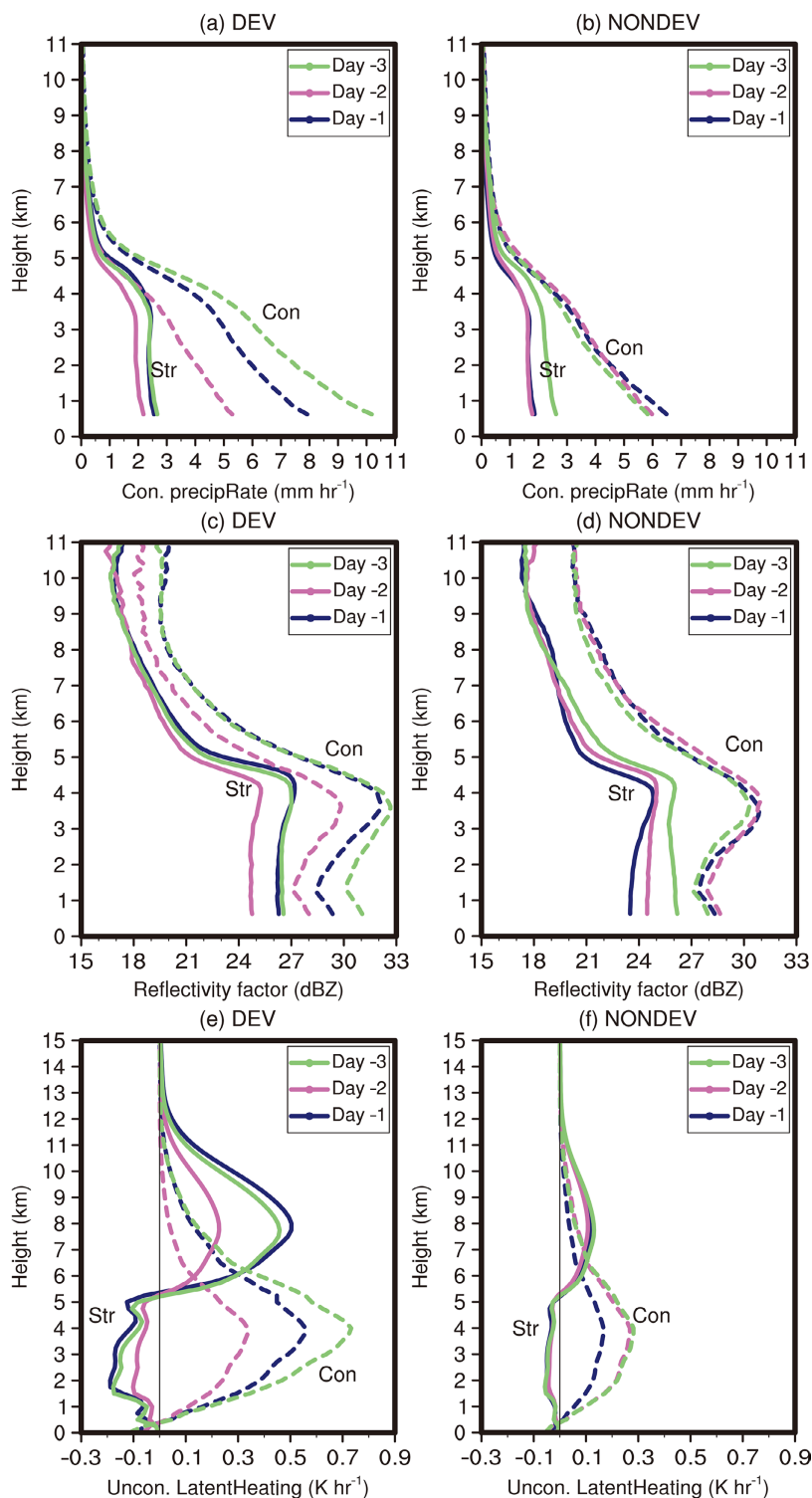


Fig. 5. The composited vertical profiles of (a, b) conditional precipitation rate (mm h^{-1}), (c, d) reflectivity (dBZ), and (e, f) unconditional latent heating rate (K h^{-1}) of the different precipitation types within a 200-km radius of the circulation center: Str (stratiform, solid line) and Con (convective, dashed line) precipitation in the (a, c, and e) developing and (b, d, and f) nondeveloping categories.

the convective intensity (Liu et al. 2015) also indicates that the convective intensities of the developing group on day -3 and day -1 are higher than those in the nondeveloping group (not shown). Although there is not an obvious difference in the vertical profile of the stratiform precipitation rate between day -2 and day -1 in the nondeveloping group, the reflectivity on day -1 is weaker than that on day -2, and the height of the bright band also decreases on day -1.

In addition to the contribution of the different precipitation types to rainfall, the role of vertical heating played by latent heating is also taken into account in Figs. 5e, f, which presents the composited vertical profiles of the unconditional latent heating rate. Stratiform precipitation shows a distinct latent heating vertical profile that is characterized by diabatic heating peaking around 8 km and diabatic cooling below the melting layer due to the melting cooling of solid precipitation particles and evaporative cooling of raindrops. As for the levels above 5.5 km, stratiform and convective precipitations jointly contribute to the warming of the upper troposphere in both the developing and nondeveloping groups, among which stratiform precipitation has a predominant role in the upper-tropospheric warming in the developing group due to the larger areal coverage (Figs. 3a, b, 5e, f). Specifically, the 3-day mean unconditional stratiform latent heating rate at 8-km height in the developing/nondeveloping group is 0.39/0.11 K h⁻¹. And the 3-day mean unconditional stratiform latent heating from 5.5- to 15-km heights contributes to 74 % of the total warming at the mid-to-upper levels in the developing group. In addition, the unconditional latent cooling effects of stratiform precipitation below the melting layer in the developing group are also stronger than those in the nondeveloping group. Given the larger percentage of stratiform precipitation in the developing cases, as described in Figs. 3a, b, the total amount of latent heating in the upper troposphere and upward flux of moisture induced by the stratiform precipitation clouds are dominant in the developing cases. As a result, besides the moistening effect in the lower troposphere, the stratiform clouds in the developing group also make a major contribution to the warming of the upper troposphere. Contrary to the lower-level warming, the upper-level warming can efficiently lower the pressure and height at the lower levels to facilitate TD development, which is in good agreement with previous studies (Hirschberg and Fritsch 1993; Jusem and Atlas 1991; Chen and Zhang 2013). Using a five-level quasi-geostrophic model, Hirschberg and Fritsch (1993) prescribed the diabatic heating at dif-

ferent altitudes to examine the atmospheric responses and found that the upper-tropospheric heating produces a more negative lower-level height tendency than the equivalent lower-tropospheric heating. Therefore, the strong accumulated upper-tropospheric heating mainly associated with stratiform precipitation in the developing group can play a critical role in the subsequent TD development.

As for the convective precipitation system, strong updrafts in convective precipitation can transport a large amount of vapor aloft to release substantial latent heat by condensation and warming of the mid-to-lower troposphere (Figs. 5a, e). It can also be seen in Fig. 5 that the maximum heating height coincides with the maximum reflectivity height, and the greater the convective precipitation intensity, the higher the latent heating rate. Therefore, the strong convective rainfall intensity in the developing group corresponds to the warm latent heating, whereas the almost constant small convective precipitation rate in the nondeveloping group corresponds to the weak latent heating without obvious daily variation. Although the precipitation rate of stratiform precipitation is lower than that of convective precipitation in the developing group (Fig. 5a), its proportion is much larger than that of convective precipitation (Fig. 3b). The mean vertically integrated unconditional latent heating rate of stratiform and convective precipitation in the whole circulation of the developing group above 5.5 km is 6.6 K h⁻¹ and 2.4 K h⁻¹ (Fig. 5e), respectively; thus, the stratiform rainfall has a predominant role in the upper-tropospheric warming. Compared with stratiform precipitation, the most striking feature of convective precipitation is characterized by latent heating located in the lower troposphere, such that the lower-level convective heating can offset the stratiform cooling. The thermodynamic conditions near the disturbance center are crucial to the formation of TC. For instance, Wang (2012) demonstrated that convective heating peaked around 4-km height and concentrated near the pouch center before the formation of hurricane Felix (2007) based on a high-resolution simulation, in which the larger latent heating rate in the deep convection can stimulate a secondary circulation and a cyclonic circulation near the surface, whereas stratiform precipitation with a higher heating height dedicates to spinning up near the middle troposphere. Thus, the dynamic and thermodynamic conditions reinforce each other and jointly promote the organization of convective structures and TC formation.

4. Summary and discussion

Hundreds of tropical disturbances emerge over the WNP during the TC peak season each year, among which a small fraction upgrade into TDs and most of them dissipate after lasting for several days. Using the 3-year observational data from the GPM DPR and the SLH dataset, the characteristics of convective and stratiform precipitations in the inner- and outer-core regions in developing and nondeveloping disturbances are examined. The focus is on the evolution and comparison of precipitation cloud characteristics during the 3 days before the TC genesis and disturbance dissipation.

We firstly talk about the source of uncertainty about our statistics. Over the tropical ocean, the most reliable and easily available high-resolution data should be the detections from polar orbit satellites and geostationary satellites. There are some limitations in the identification of the disturbance center only with the geostationary satellite products as the disturbance circulation is still relatively loose. Therefore, this study tracks the disturbance center using ERA-Interim data. The JTWC records the center position after the disturbance upgrades to TD intensity. Using the tracing method in this paper to identify the position when the disturbance first upgrades to a TD and then compare it with the JTWC data, the reliability of the tracing method in this study can be verified. Compared with the 3-year JTWC data, the TD position identified in this paper has an average absolute bias of 0.64° in the meridional and 0.59° in the zonal direction, respectively. The error is within an acceptable range for this study, and the source of the biases is probably related to the model error of the reanalysis data.

The composite results indicate that the developing disturbances experience an oscillatory process. The large areal coverage of strong rainfall in the disturbance inner-core region breaks into the scattered remnants on day -2 and then reorganizes and strengthens near the disturbance center again. Contrarily, the occurrence of vigorous precipitation in the nondeveloping group is farther away from the disturbance center. Within a 400-km radius from the disturbance center, the key features before the disturbance develops into a TC is the increasing proportion and the widely spreading distribution of strong rainfall. For the nondeveloping disturbances, the proportion of strong precipitation within the 400-km circulation region gradually decreases with time.

Regarding the different precipitation types, the statistics reveal that the stratiform and convective

precipitation coverages and convective intensity in the developing group in both the inner- and outer-core regions are larger than those in the nondeveloping group. Contrarily, the coverage and intensity of convective precipitation in the developing group in the inner-core region are larger and stronger than the counterparts in the outer-core region. Moreover, the precipitation behaviors in the inner-core region, such as stratiform precipitation coverage and convective precipitation intensity, experience a more remarkably oscillatory process than those in the outer-core region. Contrarily, the precipitation characteristics in the nondeveloping group evolve more smoothly.

It can be summarized that the disturbances prone to developing into a TC over the WNP satisfy the following two essential preconditions in terms of precipitation characteristics. First, a large fraction of stratiform precipitation covers the core regions of the disturbances. Second, strong convective precipitation occurs within the inner-core region. Convective and stratiform precipitations have a similar contribution to the total rainfall in the lower troposphere. Compared with stratiform precipitation that plays a critical role in the warming of the mid-to-upper levels, the most striking feature of convective precipitation is that it heats the mid-to-lower troposphere.

Although the vertical heating profiles of different precipitation types are synthesized by using the GPM 2ASLH products, it is necessary to explain the limitations of this algorithm. Park and Elsberry (2013) used Doppler radar to retrieve the heating and cooling rates of two developing disturbances, two nondeveloping disturbances, and two mature TC outer rainbands in the WNP. These heating/cooling profiles represented the thermodynamic structure of MCS in individual samples. They compared the retrieved results with the SLH products of the TRMM PR and found that the SLH algorithm missed the cooling effect of convective-scale downdrafts that form under the strong tilted updraft in MCS, which can be attributed to the fact that the heating profile lookup tables are constructed by the long-term and large-scale average of CRM simulation. Consequently, it is more reasonable to apply the SLH products to the thermal budget of large-scale or multi-sample statistics (Shige et al. 2009; Takayabu et al. 2010). In this paper, the thermodynamic profiles of the chosen precipitating pixels are used to composite the heating profile. The 3-day total numbers of precipitating pixels for the developing and nondeveloping groups in the inner-core regions are 38905 and 43138, respectively. Therefore, the composite of the large-sample precipitation profiles

in this study can reflect the primary thermodynamic characteristics of precipitation in the disturbance.

Overall, TC formations evolving from parent disturbances can be regarded as an outcome of the joint contribution from the two distinct types of precipitation cloud. It is imperative to quantitatively evaluate the individual importance of the different precipitation clouds in TC formation by numerical modeling, which is an ongoing research topic and will be reported elsewhere.

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