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Abstract: During the movement of Typhoon Hato (2017) over land, heavy rainfall occurred when the spiral rainband which was about 100 km distance away from the center of the typhoon passed the Dayao Mountain (with an elevation of 1.2 km). In this study, the structures and forming mechanism of the heavy rainband along the mountain range are investigated by using high-resolution model simulations. The results show the importance of topography in causing the heavy rainband. Upslope of the steep terrain lifts the cyclonic flow to produce strong upward motion when the rainband passes across with high wind speed. At the same time, the warm and humid air is lifted to the steep slope, causing unstable energy to accumulate over the windward slope, which is conducive to the occurrence of rainfall. In particular, the convective cells generated upstream of rainband will further strengthen and develop due to the uplift when they move close to the mountain foot. Some precipitation particles in the convective cells fall to the ground while others move downstream with the intense updrafts, forming heavy rainfall near the summit. As a result, the largest accumulative rainfall coincides well with the orientation of the mountain ridge.

Key words: typhoon; spiral rainband; orographic influence
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1 INTRODUCTION

Typhoon is one of the most destructive weather systems in the southeast coastal and inland areas of China, and there is an average of 7-8 typhoons landing in China each year (Chen and Meng [1]). Landfalling typhoons not only bring flooding and wind damage to coastal areas, but also bring heavy rainfall to inland areas, sometimes even causing urban flooding and landslides (Xu et al. [2]). Rainfall in typhoons exhibits elongated and organized band structure, called “rainbands” or “spiral rainbands”. These rainbands may be dozens to hundreds of kilometers from the typhoon center, cyclonically spiral into the eye wall of typhoon, and have an open spiral geometry in contrast to the quasi-circular eyewall geometry. The rainbands are the most prominent feature of typhoons and contribute significantly to the rainfall (Willoughby et al. [3]; Houze [4]).

In recent decades, with the application of radar and other advanced observational methods and the improvement of numerical simulation, more attention has been paid to the smaller-scale convective system of the spiral rainband. Many studies have pointed out that rainband is generally characterized by extensive regions of stratiform precipitation with embedded convective cells with various organized degrees (Willoughby et al. [3]; May [5]). Henze et al. [6] further observed and studied the spiral rainband of typhoons over the Atlantic Ocean, which has deepened the understanding of the convective activities and precipitation of mesoscale spiral rainband. However, most previous studies used the idealized conditions with no environmental flow, such as the typhoons over the ocean. The rainband under complex environmental influences is less discussed, especially the change of the underlying surface, multi-scale system interactions, etc. (Meng and Zhang [7]; Moon and NoLan [8]). In recent years, rainfall research after typhoon landing has attracted much attention, which is inseparable from the influence of topographic changes (Lin et al. [9]; Wu et al. [10]; Yang et al. [11]; Dehart and Houze [12]; Li Y et al. [13]). But little is known about the structural change of rainband and the related physical mechanisms.
The typhoon severe rainfall over land is often related to the effect of topography (Yu and Tsai [14-15]; Fan et al. [16]). Typhoon circulation can be influenced by topography, and if the mountain along the flow direction has sufficient height and width, the rainfall can be maximum on the windward slope due to the uplift (Lin et al. [9]; Lee et al. [17]). As Typhoon Longwang (2000) approached the large topography in northern Taiwan, Yu and Cheng [18] recorded the structural characteristics of the outer rainbow by radar, and suggested that specific parts of the rainbow moving over the topography could determine the rainfall process, with the maximum precipitation often corresponding to the strongest wind and the precipitation enhancement by topography is proportional to the speed of upward motion. Dehart and Houze [12] studied the change of the surrounding vertical rainfall structure when a hurricane made landfall in the complex terrain of eastern Mexico. The results showed that in the region with upward motion, the radar reflectivity increased, and a small area of severe convection occurred.

However, if the slope is not long enough (small hills) but can produce cloud and rainfall, the rainfall can be described in a seeder-feeder process (Hill et al. [19]). In a previous study of Hurricane Dean (2007) over the Dominica Island (Smith et al. [20]), the results showed that the Dominica Island is too narrow, the wind is too fast for convective cells to have time to form, and the enhanced topographic precipitation is caused by the accumulated growth of raindrops and their hydromorphic contents. These studies indicate that the specific shape of the terrain (height, scale, slope, geometry, etc.) has a profound influence on the typhoon precipitation characteristics.

Compared with previous studies on topographic precipitation from middle-latitude weather systems (e.g., front and cyclone), the flow patterns in typhoon environment are significantly different. In general, the areas visited by typhoon rainbow are usually characterized by abundant moisture and high winds at low levels (Yu and Cheng [18]; Yu and Tsai [14]). Precipitation systems are easily triggered on the windward slope of mountains as they approach/ pass through, involving complex land surface processes including diverse topographic dynamics, heat exchange and cloud microphysics effects (Hill et al. [19]; Smith et al. [20]; Lin et al. [21]). Tang et al. [22] focused on the effects of topography-induced gravity waves and the characteristics of topographic convection on rainfall in their study of Typhoon Nari (2001). Convective cells were frequently triggered on steep windward slopes in the distant rainbow, and gravity waves can change the vertical motion and precipitation in the leeward side. Jtl et al. [23] also studied the interactions between the locally unstable air flow on the leeward slope and the precipitation particles on the windward side under a high Froude number (Fr > 1) with high wind speed and humidity.

The above studies mainly focused on the impact of island topography or coastal topography on typhoon rainfall, while this study focused on the impact of inland mountains on rainbow after typhoon landing. When Typhoon Hato (2017) reached the central-eastern region of Guangxi, the northern rainbow passed through the Dayao Mountain, and this individual case provided an opportunity to investigate the evolution of the rainbow under the effects of terrain. The Dayao Mountain is oriented in a northeast-southwest direction, about 110 kilometers long and 45 kilometers wide, with a general elevation of 1200 meters above the sea level. It is the highest peak in the central-eastern region of Guangxi, and its southeast slope is one of the rainfall centers in Guangxi, with the annual precipitation of more than 2000 mm. In this study, in order to deepen the understanding on the typhoon-related orographic rainfall, we use the high-resolution numerical model to analyze the structural characteristics of the precipitation system when the rainbow passed through the Dayao Mountain. We hope the results could provide some objective bases for the forecast of local short-term heavy rainfall brought by the typhoon.

The remainder of this paper is organized as follows. The overview of Typhoon Hato is introduced in Section 2. In order to understand the orographic influence on typhoon rainbow, the model set-up and the design of sensitivity experiment is described in Section 3. The simulated results are discussed in detail in Section 4. Section 5 analyzes the influence mechanism of the terrain on the rainfall increase and the structure characteristics of the rainbow in the Dayao Mountain, and Section 6 summarizes the main conclusions.

2 OVERVIEW OF TYPHOO HATO

Typhoon Hato was one of the most severe typhoons that affected the coastal and inland areas of southern China in 2017. It was generated at 0600 UTC on August 20 in the northwest Pacific Ocean, southeast of Taiwan. Afterwards, it moved westward into the South China Sea being a severe typhoon, and made landfall at 0435 UTC on August 23 in the southern coast of Zuhai City, Guangdong Province, with a maximum wind of 45 m s⁻¹ and a minimum pressure of 950 hPa. At 1200 UTC on August 23, it entered Guangxi from the junction of Yulin and Wuzhou with the intensity of severe tropical storm and continued to move westward. Then, it weakened to a tropical depression at 0600 UTC on August 24. After that, it moved into Yunnan Province, and stopped being numbered at 0900 UTC on August 24.

During the landfall, it brought violent storms and significant damage to the Guangdong coastal cities, such
as Zhuhai and Shenzhen. Moreover, under the influence of Hato, the rainfall and local heavy rainstorm occurred in southwestern Guangdong, central and southern Guangxi, southwestern Guizhou, and Yunnan Province from August 23 to 25. The maximum hourly rainfall reached 60–80 mm, accompanied by strong winds, which caused mountain torrents, landslides, mud-rock flows and other secondary disasters in many places.

3 DATA AND EXPERIMENTAL DESIGN

The following data are used to evaluate the model simulation. The 6-h intensity and position data of Typhoon Hato are derived from the best-track data of tropical cyclone (TC) provided by the Regional Specialized Meteorological Center-Tokyo Typhoon Center (RSMC-Tokyo) of the Japan Meteorological Agency. The ground-based weather radar data from the China Meteorological Administration are used to analyze the structure of Typhoon Hato. The rainfall data are from the hourly precipitation grid dataset (version 1.0) integrated on the basis of the surface observed precipitation in China and the Climate Prediction Center’s morphing technique (CMORPH) precipitation products, with a temporal resolution of 1 hour and a spatial resolution of 0.1° x 0.1°.

Numerical simulation is performed by using the Weather Research and Forecasting (WRF) model version 3.7.1 (Skamarock et al. [24]), which is a fully compressible and nonhydrostatic mesoscale model with a terrain-following vertical coordinate. The domain used in the simulation is two-way interactive and triply nested as shown in Fig. 2. The inner domain is moveable with the TC via tracking the 500 hPa TC center. From the outer domain to the inner domain, the horizontal grid spacing are respectively 9 km, 3 km and 1 km, and there are 319 x 268, 622 x 388 and 601 x 601 grid points, respectively. There are 36 vertical levels with the model top being 50 hPa, and the coordinate system is used for the vertical direction.

The integration period is 36 hours from 0000 UTC on August 23 to 1200 UTC on August 24, 2017. Physical parameterizations selected for the simulation include the revised Monin-Obukhov surface layer scheme, the unified Noah land surface model, the Yonsei University boundary layer parameterization scheme, the Dudhia shortwave radiation scheme and the Rapid Radiative Transfer Model for long-wave radiation. The simulation process uses the Lin microphysics and does not use the cumulus cloud parameterization scheme. The combination of these physical parameterization schemes has been extensively tested in simulating landfall typhoons in previous studies (Yang et al. [25], Yang et al. [26]). The integral time steps for the three domains are 45 seconds. The initial and boundary conditions for WRF simulations are provided by the National Centers for Environmental Prediction Global Forecast System analyses, with a spatial resolution of 0.25 x 0.25 and a temporal resolution of 6 hours. The model outputs from the innermost 1 km domain are produced every 15 minutes.

To analyze the orographic influence of the Dayao Mountain on the precipitation distribution and the rainband of Hato, the following two experiments are designed:

1) Control test (CTL): It adopts the actual terrain, and is used for comparison with the sensitivity test.

2) Sensitivity test (NoTer): The terrain is removed, and the height of terrain above 200 m in the Dayao Mountain area is set as 200 m.

The geographical distribution of the Dayao Mountain is shown in Fig. 1, and the blue box indicates the area where the sensitivity test is conducted. It should be noted that the other parameters are set identically for the two experiments. Based on the comprehensive consideration of model resolution and analysis area, the simulation results of the innermost domain are used for comparative analysis.

![Figure 1](image_url) (a) Geographical features around the Dayao Mountain, with terrain height indicated by light to dark colors, black markers indicating the center of Typhoon Hato during its passage, and the location of the Doppler radar site at Liuzhou is denoted by the black triangle. (b) Three-dimensional structure of Dayao Mountain in the blue box.

4 RESULTS

4.1 Track and intensity of Typhoon Hato

The typhoon track, maximum surface wind speed \( (V_{\text{MAX}}) \) and minimum sea level pressure \( (P_{\text{MIN}}) \) of the two simulated experiments are superimposed on the observed data from RSMC-Tokyo, and the results are shown in Fig. 2. Overall, the simulated paths are
basically consistent with the observations after about 6 h of integration, and moving directions are all northwesterly with less deviation (Fig. 2a). Fig. 2b shows that the intensity evolution of Typhoon Hato is simulated reasonably well, showing the characteristics of weakening after landfall and continuous weakening later. However, it should be noted that in the initial period of simulation, the simulated intensity has a deviation due to the physical instability of the model, the initial field data and other problems. Despite the simulated bias in the initial time, the track and intensity are similar to the observed when passing across Guangxi. Therefore, the effects of Dayao Mountain on typhoon rainband should be reasonably captured by the model in this case (see next section). Besides, the simulated typhoon path and intensity of the CTL and NoTer experiments are basically the same. It indicates that the Dayao Mountain has almost no influence on the track and intensity of Typhoon Hato.

Figure 2. (a) The initial nest configuration (d01, d02, and d03 denote the outermost, second, and innermost domains, respectively), initial (0000 UTC 23 August) surface wind (wind barbs) overlaid with the 6-hourly RSRC-Tokyo best track (Obs) and simulated TC tracks. (b) Temporal evolution of the maximum surface wind speed (upper panel, units: m s⁻¹) and minimum sea level pressure (lower panel, units: hPa).

4.2 Rainband structure

Figure 3 compares the radar reflectivity distribution in the model-simulated results and real-time radar observation from the Liuzhou radar site. The model results reflect the inner-core asymmetric eyewall and spiral rainband pattern. However, due to the limited coverage of radar, there are not enough observational data available to be compared with the eyewall in the east area of Typhoon Hato (Figs. 3a-3c). Nevertheless, both of the observation and simulations show obvious asymmetric structure, with strong convective rainband with the reflectivity of 40-50 dBZ in the eastern and southern sides within 100 km distance away from the typhoon’s center (Figs. 3a-3c). This can imply that the overall size and rainband structure of the simulated typhoon is relatively realistic compared with that of the real typhoon.

4.3 Accumulated rainfall

Figure 4 shows the comparisons of 5-hour accumulated rainfall near the Dayao Mountain between the observation and two experiments during 1500-2000 UTC on August 23. During this period, the rainband on the north of Typhoon Hato, 50-150 km away from the center, passing through the Dayao Mountain (in blue square of Fig. 1a), which belongs to the inner rainband of the typhoon (Figs. 3d-3i). Fig. 4a shows the rainfall from CMORPH precipitation products, and the rainfall is distributed in a northeast-southwest direction, which is consistent with the direction of the mountain range. In the middle of the mountain, there is a large-value center with the rainfall more than 70 mm, which is distributed in a ring shape at the mountain peak. The rainfall on the north side of the mountain gradually decreases to 30-50 mm (Fig. 4a). The rainfall area in the CTL is relatively consistent with the observation, though the model overestimates the precipitation in the middle of the mountain (Fig. 4b). Moreover, it can be seen more clearly that the rainfall is distributed in a northwest-southeast band on the southeast side of the mountain in simulated experiment with high resolution (Fig. 4b). In the NoTer experiment, the rainfall is 20-50 mm and it decreases significantly after removing the mountain terrain, and the weaker rainband is still distributed in a spiral band (Fig. 4c). Thus, the local topography changes the local precipitation of the typhoon, and the precipitation intensity is well correlated with the topographic height.

5 MECHANISM ANALYSIS OF OROGRAPHIC RAINFALL

5.1 Convective systems enhancement and upstream environment

Figure 5 shows the horizontal distributions of radar reflectivity and wind field at 2 km height when the rainband passed through the mountain. Figs. 5a-5c
OUYANG Ping, WANG Yong-qing, et al. No.4 indicate that in the CTL experiment the radar reflectivity in the mountain area (> 40 dBZ) is higher than the surroundings, which is consistent with the rainfall distribution. The southeasterly cyclonic flow is perpendicular to the direction of the mountain range, and the wind speed reaches 30 m s⁻¹ or more (Figs. 5a-

Figure 3. (a)-(c) Observed Doppler radar reflectivity (dBZ) from the lowest elevation scan of Liuzhou. (d)-(f) and (g)-(i) simulated radar reflectivity at 2 km height during 1530-1730 UTC on August 23 in the CTL and NoTer experiments. Black dashed circles indicate 100 km and 200 km from the typhoon center.

Figure 4. 5-hour accumulated rainfall (shading, mm) within the blue rectangle of Fig. 1a, when Typhoon Hato passed the Dayao Mountain during 1500-2000 UTC on August 23 from the (a) CMORPH precipitation products (observation), (b) CTL experiment and (c) NoTer experiment. The solid black lines with light to dark colors indicate the terrain heights of 400 m, 700 m, 1000 m and 1300m, respectively.
the high wind speed could play an important role for air passing a mountain. In the NoTer experiment, most areas of the rainband belongs to the stratiform precipitation with several convective cells embedded, and the horizontal wind field is similar to that in the CTL experiment (Figs. 5d-5f).

On studying this type of overhill wet airflow, we can introduce the wet Froude number \( F_w = U/N, H \) for analysis (Chu and Lin [27]). Where \( U \) is the speed of upstream wind that is perpendicular to the mountain range, \( H \) is the terrain height above the sea level, \( N_w (= g/θ, \partial θ/∂z) \) is the wet Brunt-Väisälä frequency or wet static stability, and \( F_w \) is one of the most important dimensionless parameters to determine whether the airflow tends to climb or bypass mountains. For a given terrain height, the weak low-level wind \( (F_w < 1) \) tends to be blocked by the terrain and thus induces a convergence on the windward side of the terrain. However, the strong low-level wind \( (F_w > 1) \) tends to climb over the terrain (Smolarkiewicz and Rotunno [28-29]; Li et al. [13]). The height of the Dayao Mountain is around 1200 meters and the average wind speed upstream is 26-37 m s\(^{-1}\). The calculated \( N_w \) is about 0.013 s\(^{-1}\), and this value is closely related to the typical magnitude of low-level static stability (about 10\(^{-2}\) s\(^{-1}\)) (Yu and Cheng [18]; Yu and Tsai [14]). After calculation the \( F_w \) is about 1.5-2.5, which is even larger for lower terrain. Thus, the airflow is more likely to climb over the hill, and produce uplift on the windward slope side, favoring the rainfall enhancement.

\[ w < \frac{∂θ}{∂z} \]

\[ w \text{ is one of the most important parameters for the propagation of orographic precipitation, it may not completely represent some flow characteristics, and the environmental conditions of rainband are also very important to the convection development (Chen and Lin [13]). We analyze the T-logP diagram at point Q, located on the upslope of the mountain in the CTL experiment, where the air is saturated below 500 hPa and clearly shows a humid and saturated environment with a lifting condensation level (LCL) at about 850 hPa and the convective available potential energy (CAPE) not exceeding 400 J kg\(^{-1}\) (Fig. 6a). This configuration with small CAPE and large wind speed is similar to the flow regime III.}

\[ \frac{∂θ}{∂z} < g/θ \]

\[ \text{vol} \text{. } 27 \]

\[ \text{Fig. 5. Horizontal distribution of simulated radar reflectivity (shading, dBZ) and wind vectors at 2 km height during 1530-1730 UTC on August 23 within the blue rectangle of Fig. 1a in (a)-(c) the CTL experiment and (d)-(f) the NoTer experiment. The lines A-A1, B-B1 and C-C1 are the locations of the corresponding vertical profiles in Fig. 8. The black dashed elongated box S1 perpendicular to the Dayao Mountain indicates the area for calculating mean radar reflectivity along the box shown in Fig. 10. The solid black lines with light to dark colors indicate the terrain heights of 400 m, 700 m, 1000 m and 1300 m, respectively.} \]

Although \( F_w \) may serve as an effective control parameter for the propagation of orographic precipitation, it may not completely represent some flow characteristics, and the environmental conditions of rainband are also very important to the convection development (Chen and Lin [13]). We analyze the T-logP diagram at point Q, located on the upslope of the mountain in the CTL experiment, where the air is saturated below 500 hPa and clearly shows a humid and saturated environment with a lifting condensation level (LCL) at about 850 hPa and the convective available potential energy (CAPE) not exceeding 400 J kg\(^{-1}\) (Fig. 6a). This configuration with small CAPE and large wind speed is similar to the flow regime III.
or IV proposed by Chen and Lin [30]. In the NoTer experiment, after removing the mountain, the CAPE is lower, indicating a weaker convection in the rainband (Fig. 6b).

5.2 Orographic lifting

The horizontal distribution of vertical velocity in the two experiments can be seen in Fig. 7. The distribution of vertical velocity in the CTL experiment is consistent with the terrain, with an updraft on the windward side and a downdraft on the leeward side (Figs. 7a-7c). This structure can be shown more clearly in the vertical cross-section. The vertical motion on the upslope gradually increases from bottom to summit, and peaks before the slope decreases, with a maximum of more than 4 m s\(^{-1}\) and a vertical thickness of about 8 km (Figs. 8d-8f). Similarly, convective cells develop vigorously and extend to a higher level along the windward upslope (Figs. 8a-8c). Besides, there is a weak divergence on the windward slope at the low levels, and a weak convergence on the leeward slope. Meanwhile, the horizontal wind speed of over-mountain airflow increases with height below 2 km level in the vertical direction. The maximum wind of 37 m s\(^{-1}\) is at the height between 5 and 7 km (Fig. 8a). It can be explained that as the velocity of overhill airflow increases with height on the windside, causing the contraction of the vertical air column and the divergence of the horizontal air, and the leeward side is the opposite (Sun [32]).

To further illustrate the relationship between the topography and upward motion, the forced vertical velocity \( W_f \) resulted from the lifting effect of the mountain is defined as follows (Yu and Cheng [37]):

\[
W_f(x, y, t) = u(h, t) \frac{\partial h(x, y)}{\partial x} + v(h, t) \frac{\partial h(x, y)}{\partial y},
\]

where \( h \) is the terrain height, and \( u \) and \( v \) represent the east-west and south-north flow components of upstream oncoming wind, respectively. According to Wu et al. [10], \( u \) and \( v \) are derived from the wind vectors of the lowest model layer. Eq. 1 has been used in many previous studies to evaluate the relative importance of orographic lifting and other convective forcings associated with synoptic and/or mesoscale systems (Lin et al. [21]; Dong et al. [33]; Lin and Wu [34]). The calculated results are close to the difference of the CTL minus NoTer (Figs. 7g-7l). Therefore, the low-level vertical velocity mainly comes from terrain forcing, rather than the vertical motion in the rainband.

On the lee side, vertical velocity is negative and is less than \(-3 \) m s\(^{-1}\) as air flows downhill, but this downdraft changes suddenly, with air moving upward (Figs. 8d-8f). At 1530 UTC, the updraft is 1-3 m s\(^{-1}\) and there is convergence zone below the upward motion in the profile (Fig. 8a). Similarly, there is increased upward motion on the leeward side at 1630 and 1730 UTC (Fig. 8b and Fig. 8e). Jtl et al. [23] stated that the upward motion at lee side should be considered as the results of lee side convergence in typhoon environment. It is also similar to the hydraulic jump effect at the lee side described by Durran et al. [14], that can be explained as a kind of energy conversion in which the potential energy is converted into kinematic energy along the downslope. Thus, the over-mountain flow forms a dense flow near the summit, which accelerates downhill along the slope to create an upward lee side motion. However, because there is too little rainwater on the leeward side, and the strengthening effect of updraft on precipitation is weak, so much less precipitation occurs on the leeward side.

Figure 9 shows the vertical cross section of vertical velocity, equivalent potential temperature, and specific humidity. The maximum specific humidity is
the black shading area represents the terrain height near the Dayao Mountain.

Figure 7. Horizontal distribution of simulated vertical velocity (shading, units: m s\(^{-1}\)) at 2 km height during 1530–1730 UTC on August 23 for (a)-(c) the CTL experiment, (d)-(f) the NoTer experiment, (g)-(i) the difference of the CTL experiment minus NoTer experiment, and (j)-(l) the terrain-forced vertical velocity calculated from Eq. 1. The solid black lines with light to dark colors indicate the terrain heights of 400 m, 700 m, 1000 m and 1300 m, respectively.

Figure 8. (a)-(c) Vertical cross-sections of the simulated radar reflectivity (shading, units: dBZ) and wind vectors crossing the lines AA1, BB1 and CC1 depicted in Figs. 5a-5c. (d)-(f) the same as (a)-(c), but for the vertical velocity (shading, units: m s\(^{-1}\)), horizontal wind speed (green solid lines, units: m s\(^{-1}\)) and divergence (the solid and dashed contours are the positive and negative values of divergence, units: 10\(^{-5}\) s\(^{-1}\)); the black shading area represents the terrain height near the Dayao Mountain.
at low levels and decreases with height (Figs. 9a–9c). Fundamentally, there is a large amount of low-level moist airflow in the typhoon rainband (Jt et al. [32]). The water vapor carried by the lower wind accumulates in front of the mountain, forcing the air to move upward and transport water vapor due to the orographic lifting, then the moisture and energy of the air column over the windward slope increase continuously, which is showed that the large value area of equivalent potential temperature extends upward, reaching the maximum at 4–6 km height (Figs. 9a–9c). This moist airflow is easy to condense and release latent heat when it is lifted, which further increases the instability of the area and the development of vertical movement. Tang et al. [21] reported that the steep topography greatly enhances the upward motion on the windward slope, leading to heavy rainfall and the release of huge latent heat.

5.3 Temporal evolution characteristics of the rainfall system

To further analyze the characteristics of rainfall system influenced by topography, we select the region S1 in Fig. 5c, which is perpendicular to the mountain direction and roughly parallel to the flow passing through. The convective system intensifies and the reflectivity reaches more than 40 dBZ most of the time on the mountain peak when the rainband passed through (Fig. 10). Thus, there is a question, is it possible that these convective cells are triggered on the windward slope? About the generation of convective cells, it can be estimated roughly by advection time and cloud growth time (Stein et al. [35]). The advection time is calculated by \( a / U \), where \( a \) is the mountain half-width and \( U \) is the velocity of the over-mountain flow. The timing of cloud growth is controlled by microphysical processes that are quite complex and difficult to estimate (Kirshbaum and Smith [37]). The advection time is about 9 minutes by calculation, it is difficult to determine whether the ascending time of air is long enough to form convective clouds. On the other hand, Chen and Lin regarded that it is more inclined to form stratiform clouds precipitation at the peaks when strong wind passes through small mountains with large \( F_a \) and small CAPE [30].

Therefore, we guess it may be related to the passing of the typhoon rainband. The typhoon rainband not only contains abundant water vapor and strong wind, but also is composed of many convective cells, which move cyclonically with the wind field. When passing through the terrain, these convective cells are bound to be affected. Moreover, though the rainband passing Dayao Mountain is not strong convection area, it also contains convective cells with different organizational degrees (Figs. 3d-3i). We can see that a lot of convective cells exist upstream before reaching

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**Figure 9.** Same vertical profiles as Figs. 8a-8c, but for the vertical velocity (shading, units: m s\(^{-1}\)), equivalent potential temperature (black solid lines, units: K) and specific humidity (green solid lines, units: g kg\(^{-1}\)).

**Figure 10.** (a) Temporal variation of simulated radar reflectivity (shading, units: dBZ) and vertical velocity (black solid and dashed lines indicate the positive and negative values, respectively, units: m s\(^{-1}\)) at 2 km height from 1300 UTC to 2000 UTC on August 23 when the rainband passed over the Dayao Mountain. The selected area is within the rectangular black box S1 (shown in Fig. 5c), and the data is averaged in the direction normal to the orientation of the box. The corresponding terrain height is denoted by the black shading.
the foot of the mountain in the rainband (Fig. 10). As the study of Smith et al. [30], it seems that the passage of typhoon rainband plays an important role in the enhancement of precipitation in island of Dominica, where with a very short up slope range and there are no orographic convections. In the next section, we will discuss in detail the influence of terrain on these convective cells in the rainband.

5.4 The effect of topography on convective cells in the rainband

To illustrate the influence of topography, we compare the structure characteristics of convective cells in the rainband around the Dayao Mountain in the CTL and NoTer experiment. Firstly, four movable convective cells are selected in the CTL experiment from 1745 UTC to 1815 UTC, marked as 1, 2, 3 and 4. They are 100-150 km away from the typhoon center, with spatial scales of 10-20 km, and are generated in the upstream of the rainband (Fig. 11a). When moving downstream with the wind, the convective cells which are farther away from the mountain first strengthen at 1800 UTC and then weaken at 1815 UTC (1 and 2). In addition, the convective cells that have reached the mountain foot become stronger and larger after climbing the mountain, no matter they strengthened or weakened before (3 and 4). However, they do not stay for a long time at the mountain peak, and then gradually weaken and die out on the leeward slope (Figs. 11a-11c). The convective cells are marked as a, b, c and d in NoTer experiment. These convective cells are also generated at the tail of the spiral rainband and weaken to below 30 dBZ as the flow moves downstream (Figs. 11d-11f). This situation is consistent with previous studies. When there is no influence of external circumstances such as topography, generally, initial convective cells are in the upstream of rainband, the matured convective cells are in the midstream, and the extensive stratiform precipitation is in the downstream (May [5], Akter and Tsuboki [31]).

Figure 11. Horizontal distributions of radar reflectivity (shading, units: dBZ) and wind vectors at 2 km height during 1745–1815 UTC on August 23 in (a) the CTL experiment and (b) the NoTer experiment. The grey solid line indicates the terrain height of 500 m.
Figure 12 shows the vertical cross-sections of the convective cells in Fig. 11. The vertical thickness of these convective cells is about 7 km, and there are obvious upward motion and downward motion in cells 1 and 2, which is the structure of mature convective cells. In the cells 3 and 4 on the windward slope there is strong upward motion to reach 7 km, they seem to be developing and strengthening further (Fig. 12a). In the NoTer experiment, the cells in the downstream are in the weakening stage with gradually increasing downdrafts and more stratiform cloud precipitation. This shows that when convective cells in the rainband passes through the mountain, the upward movement will be strengthened by the orographic lifting, resulting the cells to develop further. After the cells continues to move downstream in the leeward slope, they quickly weaken to dissipate.

In the above analysis, it has been shown that it is conducive to precipitation when the typhoon low-level airflow is lifted by the terrain, and the convective cells strengthened because the orographic effects in the rainband, which further intensify the rainfall. However, there are still unresolved issues in our analysis: how do rainfall particles grow and fall, and whether the stronger the convection in the rainband, the stronger the rainfall when passes through the terrain. These problems need to be further explored.

![Figure 12](image)

**Figure 12.** (a) Vertical cross-sections of radar reflectivity (shading, units: dBZ), vertical vorticity (blue solid and dashed lines indicate the positive and negative values, respectively, units: m s\(^{-1}\)) and the wind vectors along the black dashed line D-D1 in Fig. 11b; (b) is the same as (a), but for the black dashed line E-E1 in Fig. 11e.

**6 DISCUSSION AND CONCLUSIONS**

In this paper, a high-resolution numerical model is used to study the structure and evolution of the precipitation when the rainband about 100 km away from north of the typhoon’s center passed through the Dayao Mountain on August 23, 2017. The results show that the terrain has an obvious enhancement effect on the typhoon rainfall. The precipitation increases on the windward slope and the maximum precipitation appears at the mountain peak, so the precipitation is well correlated with the terrain height.
When the rainband passes through the Dayao Mountain, the upstream environment is humid and saturated with a large $F_u$ exceeding 1.5 and a small CAPE not exceeding 400 J kg$^{-1}$. Then the cyclonic flow is lifted in front of the windward slope and results in strong upward motion on the windward slope. Meanwhile, the warm and humid air is lifted to the steep slope, causing unstable energy to accumulate in the middle levels, which is conducive to the occurrence of rainfall on the windward slope. In addition, due to the strong low-level cyclonic flow, the precipitation particles will be advected to the downstream of the maximum updraft to a certain extent, and convective systems can be found in the vicinity of the main mountain peak. Although there is ascending motion on the leeward side, there is less precipitation on the leeward side due to less convective system propagating downstream.

At the same time, the convective cells generated in the upstream of rainfall cyclonically move to the foot of the mountain and encountered the upslope terrain, enhance the upward motion due to the lift, then strengthen and advect to the summit, and later weaken and die out in downstream areas. Therefore, convective activity in the rainband also contributes to rainfall under the influence of topography.

It should be noted that the above conclusions are obtained from the simulation of a single typhoon case, and there are still some model deficiencies. In this study, we only analyze from the dynamic effect of the terrain. However, microphysical processes are also important for understanding the mechanisms for the orographic enhancement of typhoon precipitation, and they are the focus of our future studies.

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