ANALYSIS OF THE COLD AIR EFFECT ON AN EXTREME PRECIPITATION EVENT TRIGGERED BY AN INVERTED TROUGH OF TYPHOON HAIKUI (1211)

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Abstract: Based on intensive automatic weather station data, satellite cloud imagery, NCEP reanalyzed data, and the simulation results from mesoscale numerical models, this study analyzes the characteristics and formation mechanisms of the mesoscale convection system (MCS) during the extreme precipitation event that was triggered by a weakened low-pressure inverted trough of Typhoon Haikui on August 10/2012. The results of this study show that cold air at the rear of a northeastern cold vortex creates thermodynamic conditions favorable to the development of extreme precipitation. The main body of the cold air is northward located so that the cold air invades only the middle layer of the periphery of the inverted trough. Thus, the cold air minimally affects the lower layer, which results in a vertically distributed structure of the temperature advection that augments the formation and development of convective instability stratification. In the middle troposphere, the cold air encounters the convergent, ascending, warm moist air from the low-pressure inverted trough, leading to frontogenesis. The frontogenesis enhances wind convergence which, in turn, further enhances the frontogenesis, and the positive feedback between these two forces augments the development of meso-scale and small-scale convection systems in the rainstorm region and its vicinity, which strengthens the upward transportation of water vapor from low layers and thickening of water vapor convergence and results in local heavy rains.

Key words: heavy rain; cold air; frontogenesis; mesoscale convection system (MCS)

1 INTRODUCTION

With approximately 7-8 typhoons making landfall every year, China is one of the countries most severely affected by typhoons (Chen and Ding [1]). The extreme precipitation triggered by typhoon landings usually causes flash floods, reservoir collapse, and inundation, which threaten human lives and property. The distribution of the precipitation from a typhoon strongly depends on the structure and evolution of the storm (Tao et al. [2]; Chen and Luo [3]; Zhu et al. [4]; Zhou et al. [5]). In addition, a typhoon’s structure is usually affected by the underlying surface, topographic friction and convergence, and elevation, which can also contribute to extreme precipitation (Zhu and Zhao [6]; Wang [7]; Luo [8]; Zheng et al. [9]). On the other hand, a landing typhoon interacts with its surrounding middle-latitude systems to trigger a peripheral rainstorm, which is usually associated with cold air (He et al. [10]; Wei et al. [11]) and the corresponding transitional phase (Li et al. [12]). Sometimes, extreme precipitation systems may exhibit semi-tropical characteristics (Jiang et al. [13]), resulting in more complicated rainstorm formation mechanisms. It has also been shown that invasion of weakened cold air can increase the magnitude of the extreme precipitation events triggered by typhoons (Sun and Zhao [14]). Weakened cold air increases to the greatest degree and releases the most effective potential energy on the periphery of the typhoon (Ding and Chen [15]), and cold air invading the center of a typhoon results in rapidly decreased diabatic heating, reduced rainfall in the center, and increased rainfall at both the periphery and in the inverted typhoon trough (Niu et al. [16]). Previous studies of distant typhoons (Li et al. [17-19]) show that the instability of non-geostrophic gravity may be one of the dynamic mechanisms for underlying extreme precipitation.

Although previous studies have improved our understanding of the formation of extreme precipitation related to landing typhoons, the evolution of the internal structure of landing typhoons is still unclear due to the
extremely complicated formation mechanisms and limited amount of observational data. Thus, further investigation of the physical processes related to landing typhoons is required, including the occurrence and development of the mesoscale convection system (MCS).

On August 10, 2012, a local, extreme precipitation event occurred on top of a low-pressure inverted trough of the weakened typhoon Haikui. The rainfall in northeastern Jiangsu often exceeded 100 mm and reached as high as 511 mm in 24 h at the Xiangshui weather station, resulting in severe urban waterlogging and property loss. This study analyzes the effects of both cold air and the development of the MCS in triggering extreme precipitation events based on multiple types of observational data and mesoscale numerical simulations. We hope to improve our understanding of the different stages of evolution of such extreme precipitation and thus improve our forecasting ability.

2 DATA SOURCES

The data analyzed in this study include rainfall data from automatic weather stations (AWSs) in Jiangsu from 00:00 August 10 to 08:00 August 11, 1°×1° data reanalyzed by the National Centers for Environmental Prediction (NCEP), satellite cloud imagery, black body temperature (TBB) data, and the Weather Research and Forecasting model version 3.0 (WRFV3.0) mesoscale numerical simulation data based on the reanalyzed data. Unless noted otherwise, time is in Beijing time.

3 RAINFALL OBSERVATIONS AND CIRCULATION BACKGROUND

3.1 Observed precipitation

As typhoon Haikui (1211) made landfall and started to move northward from the west, heavy and extreme local precipitation events occurred in northern Zhejiang, southern Jiangsu, and southern Anhui from August 8 to 9, 2012. During the day of August 9, the low pressure of the typhoon apparently weakened, precipitation decreased, and the heavy precipitation region gradually shrank to within the area surrounding the low-pressure center of the typhoon (southern Anhui and southwestern Jiangsu). At midnight on August 10, the mesoscale convection cloud was formed and quickly developed at the top of the low-pressure inverted trough, which resulted in a sudden increase in weakened and local extreme precipitation in the northeast of Jiangsu province. From 05:00 August 10 to 05:00 August 11, the 24-h precipitation measured at the Xiangshui weather station reached 510.9 mm. Meanwhile, from 08:00 August 10 to 08:00 August 11, the 24-h precipitation levels at four weather stations exceeded 100 mm during the extreme precipitation event in northwestern Jiangsu. For reference, the precipitation at Xiangshui weather station was 487.4 mm at that time. The precipitation was higher than 100 mm at 11 AWSs and greater than 250 mm at six (Fig.1a). The distribution of this extreme precipitation was asymmetric (Fig.1a) with the extreme precipitation area concentrated in two regions: one was near the low-pressure center in southern Anhui, and the other was on the low-pressure inverted trough in northeastern Jiangsu. The second region had apparent mesoscale characteristics; the major extreme precipitation area was within a 200-km radius, and the extreme precipitation region was within a 50-km radius. In addition, there was another extreme precipitation center of weaker intensity that occurred in a small region at the boundary of Jiangsu and Anhui near Tianchang and Xuyi. Fig.1b shows the variation in the timing of the precipitation at Xiangshui station, the center of the extreme event, during the extreme precipitation period. According to this figure, the heavy precipitation mostly occurred from 08:00 to 15:00 August 10 with a 6-h cumulative precipitation of 460.7 mm. Furthermore, on August 10, the hourly precipitation at Xiangshui was close to or even exceeded 50 mm from 09:00 to 13:00, and the hourly precipitation was 115.4 mm and 100.9 mm from 09:00-10:00 and 12:00-13:00, respectively. Thus, this extreme precipitation event was notable because it resulted in extremely high precipitation that lasted for a long period of time.

3.2 Background circulation

Typhoon Haikui (1211) formed on the sea surface in the eastern Philippines at 08:00 August 3 (Fig.2a) and gradually became stronger while moving westward. It landed in the coastal area of Xiangshan, Zhejiang at 03:20 August 8 while maintaining its severe typhoon rating with a maximum wind at Level 14 in the center. After landing, Haikui crossed through northern Zhejiang and continued to move northwestward with rapidly decreasing intensity and was downgraded to a severe tropical storm at 16:00 August 8. Later, when it arrived in southeastern Anhui at 20:00, it further downgraded to a tropical storm. At 12:00 August 9 on arrival at Chizhou, Anhui, it became a tropical depression that barely moved in southern Anhui with a stationary, low-pressure inverted trough that reached northern Jiangsu. In the morning of August 10, precipitation suddenly developed in the convergent area of the inverted trough, resulting in an extreme precipitation event. On August 12, the low pressure zone gradually merged with the low-pressure zone and disappeared.

At 08:00 August 9 at 500 hPa (figures not shown), the high latitudes in East Asia showed two troughs and one ridge. A weaker high pressure ridge was on Lake Baikal in the north; two vortices were on the east and west of Lake Baikal; a thick cold vortex was maintained from the Okhotsk Sea to Northeast China, and the western trough extended from the center of the cold vortex to North China and gradually moved east. At 20:00 August 9 (Fig.2b), the westerly trough passed 120°E with its bottom at the northern end of the Shandong Peninsula. It is noteworthy that the rear of the transversal...
trough of the cold vortex moved downward and resulted in south-moving cold air. At the same time, a subtropical high was distributed across north-south trending blocks with its main body over the sea. The north side of the high extended to the northern Sea of Japan, and the continental high expanded further eastward. The weakened Haikui westerly tropical low stopped between the two highs in southern Anhui, and its low-pressure inverted trough top extended to northeastern Jiangsu. As the warm moist air brought by the southeasterly jet encountered the southerly cold air on top of the inverted trough, it caused rapid development of convection in this region, resulting in an extreme precipitation event.

Figure 1. The 24-h precipitation distribution observed from 08:00 August 10, 2012 to 08:00 August 11, 2012 (a) and the hourly precipitation at the Xiangshui weather station from 00:00 to 20:00 August 10 (b). Unit: mm.

Figure 2. The 850-hPa height field (solid line, unit: dagpm), wind field, and the path of typhoon Haikui (a) and the 500-hPa height field (solid line, unit: dagpm) and wind field (b) at 20:00 August 9.

4 FORMATION AND EVOLUTION OF MESOSCALE CONVECTION CLOUDS

Why did the extreme precipitation form again after the apparent decrease in precipitation? Fig.3 (see next page) shows the characteristics of the evolution of TBB of the convective cloud related to the heavy precipitation. The west side of the typhoon’s spiral band quickly became weaker and weaker with the landing and westward movement of the typhoon (figure not shown) while the weakening of the convection on the east side was slower, which can be explained to a certain extent by the southeasterly jet on the east side of the typhoon that continued to provide vapor and energy. Afterwards, the typhoon’s cloud structure became loose; its internal convection became less active, and the precipitation apparently weakened. At 20:00 August 9 (figure not shown), the westerly trough in the northeast-north part of China started to move eastward, and the tail of the westerly trough gradually approached the northerly trough cloud of the weakened typhoon. At 04:00 August 10 (Fig.3a), a number of small- to meso-scale convective clouds were formed in northeastern Jiangsu. The scale of the newly formed convection clouds was relatively small, but the clouds were merging with each other. At 06:00 August 10 (Fig.3b), the convection
cloud region increased from its initial 20 km × 20 km to 100 km × 100 km scale. Meanwhile, the TBB was 55°C; the cold cloud (< -32°C) area was larger than 1 000 km, and the convection clouds became a well-developed MCS. Three hours later (Fig.3c), the MCS evolved into its strongest stage and continued to expand. At this time, the TBB in its center exceeded -65°C with regular cloud boundaries. The TBB gradients along the edge were quite crowded with sharp cloud boundaries, which furthered the maintenance and evolution of the MCS. By comparison, the precipitation in northeastern Jiangsu evolved quickly, and these convection clouds lasted for 6 h. At 16:00 August 10 (Fig.3d), the MCS gradually became weaker and merged with the clouds in the center of the typhoon low. Based on the formation and evolution of the precipitation, we found that heavy precipitation primarily occurred in the mature stage of the MCS.

Figure 3. The evolution of TBB at (a) 04:30; (b) 06:00; (c) 09:00 and (d) 16:00 August 10, 2012. Unit: °C.

5 COLD AIR INVASION AND DISTRIBUTION CHARACTERISTICS OF CORRELATED TEMPERATURE ADEPTION

Based on the discussion above, we found that the mesoscale convection clouds that resulted in local extreme precipitation were formed where the westerly trough merged with the low-pressure inverted trough cloud system. Did cold air invade the low-pressure inverted trough at the rear of the trough that was moving eastward? How did the cold air invade? How did the cold air interact with the original environment of the typhoon low?

According to the evolution of the vertical distribution of the temperature advection, the maintenance and development of the low-pressure inverted trough were primarily associated with warm advection when the typhoon first landed. At 08:00 August 9 at 500 hPa (Fig. 4a), the low pressure circulation of the typhoon was primarily warm advection consisting of two parts: one part corresponded to the southeasterly jet on the east side of the low pressure area, and the other major precipitation region was related to the release of the latent heat of the condensation. During this time, the main, northward cold advection region was under the northwesterly wind field at the rear of the northeastern vortex. In the lower layer (figure not shown), the typhoon low was consistently and completely within a warm advective region. At 20:00 August 9 at 500 hPa, cold air started to move southward with a transversal trough at the rear of the northeast vortex. Although it was northward located, the cold air at the rear of the trough continued to move southward along the northerly wind in front of a North China high. After it reached the Yangtze-Huaihe Rivers
Basin, a relatively strong cold advective center with an intensity of $-6 \times 10^{-5} \degree C \cdot s^{-1}$ was formed on the north side of this region (Fig. 4b). Due to the cold advection effect at 08:00 August 10 at 500 hPa, the 24-h temperature variation in northern Jiangsu was $-2\degree C$. At the same time, the warm advection associated with the east side of the southeasterly jet of the inverted trough apparently increased; cold and warm advections merged along the Huaihe River, and the extreme precipitation region was within the dense area of the temperature advection gradient. These conditions were ideal for frontogenesis.

6 COLD AIR AND FRONTOGENESIS IN THE MIDDLE TROPOSPHERE

Frontogenesis can easily form in extreme precipitation regions due to the thermodynamic environment. In this case, (1) the saddle field was low in the north-south direction and high in the east-west direction (Fig. 2b), and (2) in the merging region the cold air from the mid-

In the lower troposphere, by comparison, the cold air was even further north; there was no apparent invasion by the cold air, and the lower layer in the extreme precipitation region continued to maintain warm advection (Fig. 4c). Thus, the cold advection layer stayed on top of the warm advection layer in the extreme precipitation region, which promoted the development of unstable stratification in the extreme precipitation region along the Yangtze-Huaihe Rivers Basin. The evolution of the vertical structure of the average temperature advection in the extreme precipitation region (Fig. 4d) over time shows that, at the initial stage of typhoon landing, most of the future extreme precipitation region consisted almost entirely of warm advection. At 20:00 August 9, weak, cold advection appeared in the middle convection layer and expanded downward with increasing intensity. Throughout the heavy precipitation period (08:00-14:00 August 10), the cold advection region stayed within the middle convection layer; its layering was thin, slightly expanded, and maintained between 400-600 hPa. In contrast, the lower convection layer maintained warm advection, which evolved with the evolution of the precipitation. During the period of heaviest precipitation, a $5 \times 10^{-5} \degree C \cdot s^{-1}$ warm advection center was formed at 950-800 hPa in the extreme precipitation region. Therefore, we conclude that the increased intensities of the cold and warm advections in the middle and lower layers helped destabilize the atmospheric stratification in this region and led to the enhanced precipitation.
The troposphere encountered the warm and moist air from the inverted trough. Therefore, was the occurrence and evolution of the extreme precipitation associated with frontogenesis, and what were the characteristics?

Figure 5a shows the vertical cross section of the potential pseudo-equivalent temperature $\theta_e$ along 120°E at 08:00 August 9. It is clear that the front was near 38°N. The frontal area was relatively wide and almost vertical, and the south side of the front was almost completely controlled by warm moist air. Within the boundary layer, the center $\theta_e$ value reached 362 K at approximately 31°-33°N, so it was strongly convectively unstable, and local extreme precipitation occurred in this region. After 09:20 August 9, the cold air in the middle layer started to move south and expanded to lower layers due to the southerly cold air at the rear of the cold vortex. At 08:00 August 10 (Fig.5b), the front became narrower; one relatively dry and cold region between 500 and 600 hPa in the middle layer expanded towards the south; the area of the middle-layer front was inclining southward, and the $\theta_e$ gradient in the middle layer increased. Thus, to a certain degree, the invasion of cold air in the middle layer promoted the frontogenesis that occurred there.

Frontogenesis or frontolysis is defined as the variation in a horizontal temperature gradient, $\nabla \theta$. If $\nabla \theta$ increases with time, it is considered frontogenesis; otherwise, it is called frontolysis. As a frontal region is affected by both temperature and moisture gradients, the variation in $\nabla \theta_e$ is used in this study to represent frontogenesis or frontolysis. The latitude-time evolution plots of the 500-hPa potential pseudo-equivalent temperature as well as its gradient along 120°E (Fig.5c) show the characteristics of the frontogenesis in the middle troposphere. At 20:00 August 8, the region of high $\nabla \theta_e$ values (the frontal area) was located at approximately 40°-42°N and expanded southwards with the cold air at the rear of the northeastern cold vortex. The relatively low $\nabla \theta_e$ region gradually moved southwards along with the frontal area, so frontogenesis was always associated with the frontal area, which aided its maintenance and further development. From 20:00 August 9 to 08:00 August 10, the frontogenesis belt of the middle layer stayed at 34°-35°N and continued to increase in intensity. For reference, the mean $\nabla \theta_e$ value increased from 0.06 K/m to 0.1 K/m. Extreme precipitation occurred more frequently on the warm side of the strong frontogenesis zone in the middle layer, indicating that the increase in the intensity of the middle-layer frontal region promoted the formation and development of the
heavy rain.

According to the time evolution of the mean $\nabla \theta_e$ value of all of the layers in the extreme precipitation region (Fig.5d), we can see that, during the development of extreme precipitation from 20:00 August 9 to 14:00 August 10, the corresponding frontal region at 500 hPa clearly evolved with frontogenesis being most obvious at 02:00 August 10 when the $\nabla \theta_e$ value reached 0.10 K/m. However, in the convective lower layer, the frontal regions at 700 and 850 hPa were relatively stable without any obvious increases in intensity. In the Yangtze-Huaihe Rivers basin, the main body of the cold air is northerly located when it was moving southward, which mainly affected the middle layer while the lower layer remained unaffected, which resulted in differences in the frontogenesis in different layers.

7 CHARACTERISTICS OF VAPOR DISTRIBUTION AND EVOLUTION

Figure 6a shows the 850-hPa wind field and the total distribution of the $\leq 500$-hPa vapor flux divergence at 20:00 August 9. The vapor moved along the low-pressure inverted trough convergence region and formed a northeasterly-southwesterly flux divergent zone from southern Anhui to northern Jiangsu with a center value of $-50 \times 10^{-5}$ g/(cm$^2$ hPa s). This corresponded to the low-pressure center of southern Anhui, and the heavy precipitation area was located around the vapor convergent area. At 08:00 August 10 (Fig.6b), the vapor convergent belt associated with the low-pressure inverted trough remained stationary. It is worth pointing out that, in addition to the corresponding strong vapor convergence center in the low-pressure center, another strong vapor convergence center appeared near the inverted trough top, whose center value was $-35 \times 10^{-5}$ g/(cm$^2$ hPa s). This convergent center corresponded to the large extreme precipitation region in northeastern Jiangsu.

To understand the evolutionary characteristics of the dynamic field and the vapor in the center of the extreme precipitation region, latitude-height cross section plots of the vapor flux divergences, wind field, and vorticity at different altitudes were plotted along the longitude through which the extreme precipitation center passed. At 02:00 August 10 (Fig.6c) in the wind field of the lower and middle layers, the south wind was stronger and greatly decreased in the north and formed a wind-speed divergent area between 35$^\circ$ and 37$^\circ$N. The vapor divergent region was mainly concentrated in the lower layer (below 800 hPa) in this area and combined with the relatively low ascending region to create precipitation in southern Shangdong. At 08:00 August 10, the more northerly flow in the middle layer (500 to 400 hPa) became stronger on the north side of the extreme precipitation area and gradually moved south to 35$^\circ$N where it was stopped by the more southerly jet in the extreme precipitation region (34$^\circ$ to 35$^\circ$N). Finally, the northerly flow helped to strengthen the wind field convergence of the middle layer and develop its cyclone.
vorticity with a value of $4 \times 10^{-5}$ s$^{-1}$ in its center. Thus, the vertical movement quickly increased, and the abundant vapor from the lower layer started to be transported upward through the extreme precipitation region with this relatively strong flow. As a result, a thick vapor flux divergent area was formed in the extreme precipitation region from the surface to 500 hPa, and a strong vapor convergent center appeared close to the land. In addition, another vapor convergent center appeared in the 700 hPa to 800 hPa layer with a center value of $-6 \times 10^{-5}$ g/(cm$^2$ hPa s), which promoted the development of precipitation.

8 ANALYSIS OF THE NUMERICAL SIMULATION RESULTS

8.1 Numerical simulation methods and results

The development of the MCS in the extreme precipitation region in northeastern Jiangsu from August 9 to August 10, 2012 was analyzed using the WRFV3.0 mesoscale numerical model. The 20:00 FNL reanalyzed data for August 9 were used as the initial conditions. The horizontal resolution in the WRFV3.0 model was 5 km; the model center point was at 119°E and 33°N, and the number of grids was 241×191. The area covered included Jiangsu and its surrounding provinces and cities. The vertical dimension in the model was divided into 35 layers; the integral step length was 20 s; the simulation time was 48 h, and the results were exported at different time steps. The main model parameterization methods included the following: the Lin microphysical methods, the Noah land process model, the YSU boundary method and the BMJ cumulus cloud parameterization method.

Figure 7a shows that the 24-h precipitation from 08:00 August 10 to 08:00 August 11, whose peak of the extreme precipitation event and shape of the heavy precipitation distribution are generally consistent with the observation (Fig.1a). The location of the extreme precipitation in the simulation is from 119.2° to 120°E and from 33.2° to 34.5°N, which is very close to the observed location. However, the 24-h precipitation calculated for the extreme precipitation center by the simulation (193 mm) is much lower than the observation (511 mm). Based on the hourly precipitation simulation in the heavy precipitation center (Fig.7b), it can be observed that, although the precipitation intensity is lower than the observational data, the evolution of the simulated precipitation trend and the concentrated time periods of the extreme precipitation are consistent with the observation (Fig.1b). In addition, although the simulated amount of rainfall at the extreme precipitation center is lower, the predicted extreme precipitation region and range, the tendency toward the evolution of heavy rain, and the simulated extreme precipitation center in northeastern Jiangsu are all very robust with respect to the results of local extreme precipitation simulations in previous studies. Furthermore, the simulated weather system before and after the extreme precipitation also agrees with the observations (figure not shown). Thus, we can further analyze the development of the mesoscale convection in the extreme precipitation region based on the simulation results.

Figure 8a shows the simulated 500-hPa stream field. In the middle troposphere, the low-pressure typhoon was still completely centered in southern Anhui, and the east side of the low pressure area was accompanied with the more southerly flow, which carried enough energy and vapor to maintain the low pressure. The inverted trough extended from the low-pressure center to the northern Huaihe River and was within the frontogenesis belt formed by the merging of southerly cold air and warm moist air on top of the low-pressure inverted trough. According to the variation of the 500-hPa divergent field over time, the wind field convergence near the inverted trough in the northern Huaihe River was more evolved; precipitation along this convergence line developed quickly, and the mesoscale rain groups gradually formed a mesoscale extreme precipitation belt. During this time period, the development of convection disturbance was quite systematic, and this
frontogenesis belt was the most active region in which the MCS developed.

The vertical cross-section of the $u-w$ simulated wind field along the surface of the frontogenesis in the extreme precipitation center (Fig. 8b) clearly shows that $119^\circ$–$122^\circ$E was a relatively wide, ascending area. A strong ascending flow with a speed of 1.8 m s$^{-1}$ that was almost vertical on the convection scale occurred near $120^\circ$E in the center of the extreme precipitation region. It was in the middle troposphere (400 to 500 hPa) and combined with the upward jet in the inverted trough of the low troposphere, resulting in a thicker ascending flow. During the development of the extreme precipitation, the upward motion gradually increased with the frontogenesis in the extreme precipitation region, and the flow accumulated, decreased in size and increased in intensity. The convergence uplift in the middle layer and the release of latent heat caused by cumulus clouds led to the development of this ascending flow.

Figure 8c shows the latitudinal vertical cross section of the divergent field along the extreme precipitation center. Corresponding to the wide upward-moving region, the lower-layer wind field was convergent and the layer above was divergent. In contrast, near the extreme precipitation region ($120^\circ$E), the convergent region in the lower layer was thicker and moved upward to a region at approximately 500 hPa. The region above was a divergent area centered at 300 hPa with a value of $3\times10^{-4}$ s$^{-1}$. This vertical distribution of the divergent field helped to develop and strengthen the ascending motion. In addition, the vertical structure of the vorticity field (Fig. 8d) shows that the cyclonic vortex regions in the lower and middle layers were mainly within the extreme precipitation region and along its east side. Additionally, cyclonic vorticity was the main characteristic near the extreme precipitation center, and the region of intense and positive vorticity matched the upward-moving region. The center of cyclonic vorticity was approximately at 600 hPa and a smaller scale anti-cyclonic center was above it, which favored the development and maintenance of the ascending flow in the extreme precipitation center. A narrow region of anti-cyclonic vorticity was adjacent on the west side of rainstorm area in the low troposphere and the layer above was a region of cyclonic vorticity, which showed a nearly symmetric structure, similar to that of the vorticity field near the extreme precipitation center. This means that the convection should have occurred on a smaller scale in this

Figure 8. The simulated 500-hPa wind field and frontogenesis area at 08:00 August 10 (a, shaded area, unit: K/km); the latitudinal vertical cross section of the $u-w$ wind field and vertical velocity (b, shaded area, unit: $10^4$ m/s) along the rainstorm center, the vertical cross sections along rainstorm center of the divergence (c, unit: $10^5$ s$^{-1}$) and vorticity (d, unit: $10^5$ s$^{-1}$).
region but did not show up clearly in the wind field and the divergent field due to its small scale.

To summarize, the invasion of the middle layer by cold air from the north and the merging of the cold air with the warm, moist flow in the low-pressure inverted trough resulted in a clear region of frontogenesis near the zone of extreme precipitation. The frontogenesis maintained and strengthened the wind field convergence in the inverted trough. In turn, the wind field convergence further promoted the frontogenesis. This positive feedback greatly increased the convergence and cyclonic vorticity near the frontogenesis region, enabled the occurrence and development of the mesoscale and small-scale convection system and finally brought about the extreme precipitation event.

9 CONCLUSIONS AND DISCUSSION

A local extreme precipitation event that occurred in the low-pressure inverted trough of the weakened typhoon Haikui on August 10, 2012 was analyzed using observational data and numerical simulation.

(1) The local extreme precipitation event occurred on top of the low-pressure inverted trough of a weakened typhoon. The cold air at the rear of the cold vortex expanded southward, promoted the convective instability of the stratification in the extreme precipitation area and stimulated the MCS in this area.

(2) As the southward-moving cold air was located northward, it only invaded from the middle troposphere to the north of the low-pressure inverted trough while maintaining warm advection in the lower layer in the extreme precipitation region. Thus, it led to a vertical structure in which the cold advective layer lay above the warm advection layer and promoted the development of the convective instability stratification in this region. This led to the formation and development of the MCS in this region, and the heavy precipitation started at the MCS mature stage.

(3) In the lower and middle layers, vapor accumulated in the low-pressure inverted trough and the extreme precipitation area corresponded to the vapor convergence center. During the initial development of the heavy rain, the developing and moving southward of the northerly wind met with warm and moist flows near the extreme precipitation area in the middle layer, which strengthened the wind field convergence. Thus, the cyclonic vorticity and ascending flow were more evolved, and abundant vapor from the low pressure area moved upwards with the ascending flow, which resulted in the thickening of the convergent vapor layer near the center of the extreme precipitation and heavy rainfall.

The increased perturbation of the frontogenesis in the middle troposphere promoted the organization and development of the MCS in the inverted trough. The south-moving cold air in the middle layer merged with the convergent, ascending, warm and moist air in the inverted trough, resulting in the middle-layer frontogenesis effect. The frontogenesis effect then maintained and strengthened the wind field convergence of the inverted trough, which further promoted the frontogenesis. This positive feedback resulted in the occurrence and development of the MCS and the extreme precipitation.

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