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Abstract

Through the analysis of Doppler radar data, this study focuses on the characteristics and evolution of convection embedded within the principal band in Typhoon Morakot (2009) under the impingement of the intense southwesterly (SW) monsoonal flow. The intensity of the SW flow is comparable with the typhoon circulation at the third quadrant. The kinematic analysis shows that the northward component of the SW flow decelerates while approaching the rainband, creating significant convergence zones which results in the initiation and development of convective cells within the principal band.

The vertical kinematic characteristics of the rainband reveal two types of downdrafts namely inner-edge and low-level downdrafts. The inner-edge downdrafts coupled by the radially inward tilting convection were initiated by the precipitation drag. Dynamically, the existence of the perturbed high at 1.5 km altitude in the inner-edge downdrafts supported the finding. Furthermore, it is evident that the distribution of two perturbation highs in the vicinity of the rainband could lead to SW flow deformation locally and fortify the mechanism of convergence, resulting in the merging of convective cells into the rainband. The maximum vertical vorticity coupled with the horizontal wind maximum at the middle levels of the rainband was also observed.

Keywords southwesterly monsoonal flow; typhoon circulation; rainband; dual-Doppler synthesis; downdrafts

1. Introduction

Analyzing the structure and intensity of rainbands embedded within a tropical cyclone (TC) is considered an important research issue owing to their significant effect on quantitative and qualita-
tive rainfall forecasting. Compared with the notable improvements in tropical storm track forecasting in recent decades, more efforts are needed to enhance the prediction of the rainband intensity changes in a tropical cyclone. Willoughby et al. (1984) described a concept regarding stationary band complex (SBC) for an asymmetric hurricane rainband structure. This included the principal band, which is often regarded as an effective boundary between vortex-scale and environmental process. Barnes et al. (1983) schematized the cellar reflectivity structure on the upwind and central portion of a hurricane rainband via airborne radar reflectivity and in-situ measurements (Fig. 1). They inferred that the convective circulation within a rainband consisted of two patterns, including the updraft of high moist static energy and the simultaneous downdraft of low moist static energy. Their results were further confirmed by Powell (1990a, b) and Barnes et al. (1991) using pseudo-dual-Doppler synthesized data collected by the airborne Doppler radar with different viewing angles for a principal band. Ishihara et al. (1986) proposed an approach by implementing two ground-based Doppler radars to investigate the outer rainbands of a typhoon. The research discussed the convective structure of rainbands and made a comparison with an earlier conceptual model. The result showed that the structure of the convection embedded within the rainband was similar to previous studies. Yet the maximum updraft and downdraft in the outer rainband was 2 and −1 ms⁻¹, respectively, which were weaker than those of the conceptual model proposed by Barnes et al. (1983). By analyzing the airborne Doppler radar with high-resolution observations during the Hurricane Rainband and Intensity Change Experiment (RAINEX 2005), Hence and Houze (2008) validated the existence of a three-dimensional pattern of the convective structure embedded within rainbands. They also investigated how small-scale features, i.e., convection, affects the upscale storm circulation.

Although the principal band is the rainband dividing the hurricane inner core and environmental flow, the role that the environmental flow plays on the organized rainband and embedded convection has rarely been addressed in previous research. It seemed that background perturbation is negligible compared with the interaction between the vortex and convective cells. Over the past decade, however,
the interaction between a tropical cyclone and the Asian monsoonal system has been one of the significant mechanisms for triggering severe weather events over the northwestern Pacific area. The organized typhoon rainbands in these events induce heavy rainfall over land under the impingement of a marked environmental flow. Typhoon Morakot (2009) is the most prominent case delineating the interaction. Hong et al. (2011) concluded that Typhoon Morakot was associated with a large-scale convection region with monsoon circulation of a different time scale in the tropical western North Pacific. By analyzing the velocity azimuth display (VAD) of Doppler radar network observation, Wei et al. (2012) found that the southwesterly (SW) monsoonal flow, which exceeded 40 m s$^{-1}$, was more intense than the circulation of Typhoon Morakot (approximately 35 m s$^{-1}$). The strong environmental flow impinged on the typhoon
circulation and intensified the radial inward component. The confluence between the SW flow and typhoon circulation contributed to the development and maintenance of the asymmetric typhoon rainband structure.

The ability to timely predict sudden changes of the tropical cyclone intensity remains one of the most challenging tasks when monitoring a hurricane (Houze 2010). While many studies have deeply focused on the rainband characteristics in a vortex dominated-condition over the open ocean, there have been few thorough studies investigating the rainband and convective structures when a significant environmental flow strikes the typhoon circulation. Therefore, the goal of this study is to describe characteristics of the typhoon rainband and embedded convective cells when the influence of the environmental flow is no less than that of the inner-core vortex. The result will be compared with the concept delineated by Barnes et al. (1983) and other related research, where similarities and differences will be examined. Moreover, it is also feasible to employ consecutive observations from the ground-based Doppler radar network to discuss the evolution of the cellular convection in rainbands, which has been little discussed in previous studies, as well. The data sources and methodology are presented in Section 2. Descriptions of the synoptic environment are demonstrated in Section 3. The characteristics of the rainband and evolution of embedded convective cells are presented in Sections 4 and 5. Discussion and conclusions are shown in Sections 6 and 7, respectively.

2. Data source and methodology

The weather radar network in southwestern Taiwan includes three operational Doppler weather radars, which are respectively located at Makung, Chiku, and Kenting (Fig. 3). It is noteworthy that the Makung radar has dual-polarization capability (Table 1). The distances between Makung–Chiku and Chiku–Kenting radar sites are 64 and 158 km, respectively. Therefore, they can offer different areas of overlapping scan coverage, which are available for dual-Doppler radar wind synthesis. Wei et al. (2012) have discussed that the long baseline between Chiku and Kenting radar would smooth the synthesized...
wind field due to coarse sampling. The relatively short baseline between the Makung and Chiku radar, however, can improve the horizontal resolution of synthesized wind to 1.5 km. The radar data are interpolated to the horizontal plan via the Cressman interpolation method. Considering the enhanced sampling and limited velocity error variance, the grid setting for interpolating the radar data is also 1.5 × 1.5 × 0.5 km in the Cartesian coordinates (x, y, z) (Davies-Jones 1979). The consistency of the analyzed time sequence in this study suggests that the patterns of the synthesized wind field are reasonable. The retrieved vertical velocity is calculated by the continuity equation, setting w = 0 at the upper boundary with reflectivity less than 10 dBZ and at the lower boundary at 0.5 km level. The detailed methodology and processes for synthesizing wind in the study refer to Lee et al. (2008). It should be noted that the underestimation of the vertical velocity is an inevitable problem as there is no low boundary observation below the 0.5 elevation angle.

Once the three-dimensional kinematic field has been derived from the dual-Doppler synthesis, the pressure perturbation can be retrieved through the method proposed by Gal-Chen (1978) and Roux et al. (1984) in order to infer the dynamics responsible for the observed characteristics.

Because of the lack of radiosonde data in southern Taiwan, satellite retrieval temperature profiles are used to denote the thermodynamic characteristics of the environment field. The AIRS/Aqua Level 2 Standard physical retrieval (AIRX2RET) is a retrieved atmospheric profile, which contains several important parameters based on the Atmospheric Infrared Sounder (AIRS) observations installed on the Aqua satellite, including the atmospheric temperature and mixing ratio of water vapor and ozone content. The AIRX2RET is available over both land and ocean with a spatial resolution of 50 km. The accuracy of the temperature and humidity profiles derived from AIRS/AMSU/HSB is recognized as a way to improve forecasts made by meteorological prediction models. Researchers are using AIRS data products to validate climate models and test their representations of critical climate feedbacks. (NASA/AIRS 2013)

3. Synoptic conditions

Typhoon Morakot (2009) moved westward and passed Taiwan Island between 6 and 9 August 2009, the influence from the typhoon lasted for almost 64 h. Wu et al. (2011) divided the track of Morakot into three segments: 1) a general westward movement from 4 to 7 August, 2) moving across Taiwan and the Taiwan Strait from 7 to 9 August, and 3) making a second landfall over mainland China from 9 to 10 August. The corresponding mean translation speeds were 5.4, 2.6, and 4.4 m s⁻¹, respectively, for the three periods. Chien and Kuo (2011) pointed out two unique features of Morakot on the basis of the observations: 1) the slow translation speed of the system (~ 2.8 m s⁻¹) caused a long duration of typhoon-influenced rainfall; 2) the strong SW flow helped transport moisture-laden air to the northern South China Sea and southern Taiwan Strait.

The environment conditions for Typhoon Morakot have been deeply investigated by many researchers. Hsu et al. (2010) learned that the SW flow in the Typhoon Morakot case supplied abundant moisture and formed a large-scale convective region, which was much larger than the size of the typhoon.

<table>
<thead>
<tr>
<th>Radar Items</th>
<th>Makung Radar</th>
<th>Chiku Radar</th>
<th>Kenting Radar</th>
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<td>Wavelength</td>
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<td>10 cm (S Band)</td>
<td>10 cm (S Band)</td>
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<td>38</td>
<td>29</td>
</tr>
<tr>
<td>Type</td>
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<td>Meteor-1000C</td>
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<td>Dual Polarization</td>
<td>Horizontal Linear</td>
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<td>230(D), 460(ND)</td>
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<td>131(D), 32(ND)</td>
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<td>11.36</td>
<td>11.36</td>
</tr>
<tr>
<td>Nyquist velocity</td>
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<td>± 20.0</td>
<td>± 49.51</td>
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<tr>
<td>Pulse width (μs)</td>
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<td>0.4(D), 10.0(ND)</td>
<td>0.4(D), 10.0(ND)</td>
</tr>
</tbody>
</table>
vortex. Wei et al. (2012) demonstrated that Typhoon Morakot was embedded within a wide low pressure zone, which slowly moved westward under the extension of the subtropical ridge. The low pressure system introduced the intense SW flow from the Bay of Bengal reaching to the east of Taiwan carrying abundant moisture (Fig. 4a). The prevailing SW flow began to interact with typhoon circulation before Typhoon Morakot made landfall. Since the intersection between the SW wind and isobars was evident, as
seen by the 30–60° intersecting angles at low levels (Fig. 4b), it indicated that the interaction between the typhoon circulation and SW monsoonal flow might lead to the reinforcement of the typhoon intensity due to the confluent mechanism.

4. Mesoscale characteristics of typhoon rainbands

Radar reflectivity by integrating the islandwide weather radar observational data indicated that the eyewall of Typhoon Morakot (2009) dissipated after making landfall at 1557 UTC 7 August 2009, resulting in the typhoon to be downgraded from category 1 to a tropical storm. At 1200 UTC 8 August 2009, a wide rainband (denoted by B in Fig. 5a) located at the third quadrant with a distance of 300 km from the typhoon center was well pronounced. Another intense rainband (denoted by E) was located at the radially inner side of rainband B. The patterns of rainbands B and E corresponded to the concept of the SBC proposed by Willoughby et al. (1984), where the former was regarded as the principal band and the latter resembled the secondary band. Both bands

Fig. 5. The integrated column vector (CV) charts of reflectivity (dBZ) around Taiwan area at 1310 UTC on 7 August 2009. Letter A, B, C and D represents the eyewall, secondary rainband, principal rainband and quasi-linear rainband, respectively, related with Typhoon Morakot. The red spot at the center of circles stands for the position of the typhoon center. (adapted from the Central Weather Bureau of Taiwan)
extended from a distance 250 to 300 km away from the typhoon center, which was larger than that (100–150 km) suggested by the previous schematic concept, thus implying the high asymmetry of typhoon circulation. Since the principal band is often regarded as an effective boundary between vortex-scale and environmental processes (Barnes et al., 1983; Powell 1990a, b), the existence of rainfall band B with indistinctively spiral characteristics suggested that the environmental flow was impinging on the circulation of Typhoon Morakot (Fig. 5b). Liou et al. (2012) compared the radar observational data with rain gauge observations and showed that the major precipitation was primary, confined to the windward side of the mountains. Most of the moisture would have evaporated before the air in rainfall band B reached the leeward side of the mountain; therefore, rainfall band B weakened after 0000 UTC 9 August. Rainfall band E still maintained the same intensity and gradually replaced the upwind side of rainfall band B with a quasi-linear pattern (Fig. 5c). Eventually, rainfall band E sustained for more than 10 h after the typhoon moved further northward (Fig. 5d).

The dual-Doppler wind synthesis using Kenting and Chiku weather radar data is helpful to document the mesoscale environment where both rainbands B and E were developed. By 1801 UTC 8 August 2009 (Fig. 6a), the westerly flow coming from the typhoon circulation encountered the prevailing SW flow at which the rainfall band B was located. The SW flow subsequently extended further northward. At 2101 UTC (Fig. 6b), the convection embedded within rainfall band B weakened and dissipated, while rainfall band E became prominent and well-organized.

The influence of environmental flow will be determined by deducting the typhoon circulation. Since the Chiku radar site was located at the radially inner site of rainfall band E on 8 August 2009, its VAD derived wind field was selected to represent the mean velocity of typhoon circulation in the third quadrant. The flow pattern demonstrated that the southerly component of the SW flow could extend to rainfall band B (Fig. 6c). Its speed decreased and formed a convergence zone associated with rainfall band B. The maximum convergence was located at the radially inner edge of rainfall band with a magnitude of approximately $-1 \times 10^{-3}$ m s$^{-1}$. The flow pattern between rainbands B and E was rather weak and chaotic, suggesting transition between the SW environmental flow and westerly/west-northwesterly flow of typhoon circulation. After the dissipation of rainfall band B at 2101 UTC (Fig. 6d), the SW flow dominated the region and extended to the southern flank of rainfall band E. The boundary between the typhoon circulation and environmental flow thus identified was associated with rainfall band E.

The retrieved temperature and moisture fields from AIRX2RET data further suggested that the SW flow played a significant role from an atmospheric instability standpoint. At 1817 UTC 8 August 2009, the equivalent potential temperature ($\theta_\text{e}$) decreased with height between 1000 mb and 850 mb levels over the northern area of the South China Sea, upstream of the SW flow (Fig. 7). The estimated value of $\partial \theta_\text{e} / \partial z$ was larger than $-10$ K/1000 m, which indicated that the environment was unstable. Additionally, the specific humidity could exceed 22 g kg$^{-1}$ at 1000 mb and 20 g kg$^{-1}$ at 925 mb (figures not shown), which exceeded the average (15 g kg$^{-1}$) over the same area in summer (Willett et al., 2008). It was evident that the low level troposphere at the upstream of the SW flow was very warm and moist. The SW flow extended northeastward, transporting the markedly unstable air to the southwestern coast of Taiwan. The instability at low levels (<1500 m) provided an appropriate circumstance for the development and maintenance of convection embedded within rainbands.

5. Evolution of convection in rainfall band

Rainfall band E was regarded as the principal band after the dissipation of convection in rainfall band B. Its detailed feature could be delineated by the dual-Doppler wind synthesized from Chiku (RCCG) and Makung (RCMK) weather radar data. At 2139 UTC, rainfall band E was characterized as quasi-linear and east–west oriented. It consistently developed and maintained at the transition zone between the SW and westerly flows. The associated convergence zones triggered a number of convective cells, which were defined by the precipitation echo larger than 40 dBZ. Additionally, the convective core was identified by the precipitation echo larger than 45 dBZ (Fig. 8a). These convective cells propagated eastward along with the westerly flow of typhoon circulation with a speed of 25–28 m s$^{-1}$, while the convective cells in the further south of rainfall band E, which is denoted as E2, moved east-northeastward with a speed of 33 m s$^{-1}$. This indicated that the SW flow was more intense than the westerly flow. To describe the collision of the SW flow, the mean typhoon circulation at the third quadrant, which was defined by the VAD wind of the Makung radar, was also eliminated from the wind field in the vicinity of rainfall band E. It is obvious that the northward component of the SW flow decelerated while reaching the upband convective cell of rainfall band E, which is denoted as E1 (Fig.
Fig. 6. Dual Doppler synthesis wind from Kenting and Chiku radar data at (a) 1801 UTC and (b) 2101 UTC on 8 August 2009 at 2 km level. Panels (c) and (d) are at the same timeframe as (a) and (b), respectively. Contribution from the typhoon circulation has been removed, based on the VAD wind of Chiku radar. The axes indicate relative distances from the Kenting radar site (x = 0 km, y = 0 km). The Chiku radar is located at the position of (x = –81 km, y = 138 km). The arrows stand for the wind field and the color shading represents the rainfall echo (dBZ). The black contours represent the convergence field. The bold solid contours indicate convergence less than $-0.5 \times 10^{-3}$ s$^{-1}$. The gray solid line denotes the coastline. The dashed line in panel (d) indicates the location of the transition between the SW flow and westerly flow.
Fig. 7. The retrieved equivalent potential temperature $\theta_e$ (K) distribution from AIRX2RET data at 1817 UTC on 8 August 2009 at (a) 1000mb, (b) 925mb and (c) 850mb. The black solid lines represent the locations of the rainbands B’ and E. The arrow represents southwest airflow direction. The black box represents the upstream of the SW flow in a domain size of 400 km × 200 km.
Fig. 8. Dual Doppler synthesis wind field from Makung and Chiku radar data at 1.5 km altitude at 2139 UTC on 8 August 2009. (a) the wind field relative to ground and (b) the wind field which eliminates the typhoon circulation based on the VAD wind of Makung radar. The scale of wind vectors is shown at the right bottom of each panel. The color shading represents the precipitation echo (dBZ). The white contours indicate convergence field less than $-0.5 \times 10^{-3} \text{s}^{-1}$ with $0.5 \times 10^{-3} \text{s}^{-1}$ interval. The black contours represent the vertical wind field with 1 m s$^{-1}$ interval. The axes indicate relative distances from the Makung radar site ($x = 0 \text{ km}$, $y = 0 \text{ km}$). The Chiku radar is located at the position of ($x = 46 \text{ km}$, $y = -46 \text{ km}$). The black dashed line (K-K’) in panel (a) denotes the cross-section position, and the direction points to the typhoon center.
8b), generating a remarkable convergence with absolute value exceeding $1.5 \times 10^{-3}$ s$^{-1}$ (white contours in Fig. 8b). The upward vertical velocity associated with the convergence was larger than 2 m s$^{-1}$ (the solid black contour in Fig. 8b). In addition to the updraft, a significant downdraft appeared at the radially inner edge of E1, with a maximum velocity of −2 m s$^{-1}$ (dashed black contours in Fig. 8b). The coupling of the significant updraft and downdraft along the radial direction was comparable with the findings by Barnes et al. (1991). At 2154 UTC, the upband convergence zone enlarged since the northward component of the SW flow decelerated further significantly (Fig. 9). The northward component of the SW flow deflected slightly while reaching E1 and E2, also creating a narrow convergence zone between both convective cores. The relationship suggested that the convergence associated with the rainband was strengthened by the impingement of the SW flow. Meanwhile, in addition to a remarkable downdraft persisting at the radially inner edge of E1, there was also a moderate downward wind (~1 m s$^{-1}$) associated with E2, which finally merged with E1 and formed a new east–west oriented convective core, denoted as E3 at 2209 UTC (Fig. 10). Simultaneously, a new convective core denoted as E4 was initiated in the upband, and became more intense at 2216 UTC (Fig. 11). The absolute value of the convergence subsequently intensified, exceeding $1.5 \times 10^{-3}$ s$^{-1}$, because of the radially inward deceleration of the SW flow at 2224 UTC (Fig. 12), thus sustaining the intensity of E4. The convergence zone consistently moved eastward with a speed of 25 m s$^{-1}$, associated with E4.

Consequently, the dual-Doppler coverage by the Chiku and Makung radar sites provided a keen insight for the evolution of convective cells embedded within the rainband. Rainband E was the boundary between the westerly and SW flows. The upband deceleration of the northward component of the SW flow generated intense convergence zones, triggering the development of new convective cells. The pronounced convergence (<$−1.5 \times 10^{-3}$ s$^{-1}$), which was similar to the findings by Barnes et al. (1991) (<$−2.5 \times 10^{-3}$ s$^{-1}$), moved eastward associated with convective cells that constantly maintained rainband E. Moreover, the intensity of convective cells could be organized by merging the isolated convective cells.

6. Discussion

6.1 Vertical characteristics of convection

As mentioned in the Introduction section, both
Fig. 10. Same as Fig. 8b, except for 2209 UTC on 8 August 2009.

Fig. 11. Same as Fig. 9, except for 2216 UTC on 8 August, 2009. The black dashed line L-L' denotes the cross-section position pointing to the typhoon center, which will be presented in Fig. 18b.
earlier and recent observational studies via airborne radar have shown that the convective cells within rainbands were characterized by a radially outward slope. This section will further discuss the vertical characteristics of convective cells and kinematic field relative to the convective cell along the rainband E. Moreover, a comparison with the conceptual model (Fig. 2) regarding convection embedded within the principal rainband by Hence and Houze (2008) will be analyzed.

Figure 13 illustrates the precipitation echo that is larger than 30 dBZ within the convective core E1 along the F-F’ radial line with respect to the typhoon center (depicted in Fig. 9 at 2154 UTC) vertically extended to 8 km in height. The updraft was overturned at the same altitude by the intense radially outward flow, which was similar to the vertical characteristics in the principal band initially suggested by Barnes et al. (1983) and confirmed by Hence and Houze (2008) except that positions of the maximum vertical velocity were confined below the 4 km level and above the 9 km level. The reflectivity gradient was sharp at the radially inner side of E1, where a significant downdraft (>2 m s\(^{-1}\)) at 10–15 km in the horizontal coordinate originated below 6 km in altitude. The pattern suggested that the downdraft resembled an inner edge downdraft proposed by Didlake and Houze (2009), who explained that the inner-edge downdraft was driven by the buoyancy-induced pressure gradient acceleration originating at 6–8 km altitude, creating a significant reflectivity gradient along the inner boundary of the rainband. However, the convective core E1 tilted radially inward and updrafts tilted radially outward. This feature implies that the inner-edge downdraft might be forced by the precipitation drag instead of the buoyancy-induced pressure gradient acceleration.

Figure 14 depicts the vertical characteristics of E2 along the G-G’ radial line (depicted in Fig. 9 at 2154 UTC). Unlike E1, E2 was slightly outward slantwise. A pronounced radially inward flow penetrated E2 and subsided into the base of the convective cell. The flow pattern associated with the downdraft having a speed of 2 m s\(^{-1}\) was comparable with the low level downdraft, which emanated from 2–4 km levels, described by Didlake and Houze (2009).

Figure 15 shows the vertical characteristics of E4 along the H-H’ radial line with respect to the typhoon center (depicted in Fig. 12 at 2224 UTC). The radially inward flow protruded somewhat inward and decelerated at the base of E4, thereby generating a significant convergence. It lifted the weak radially outward flow...
at the inner edge of E4 between 2 and 5 km in altitude. Both radially inward and outward flows at the low levels generated a remarkable updraft with speeds exceeding 6 m s$^{-1}$ at the 5–6 km levels. Therefore, the reflectivity core and updrafts were tilted radially outward. This implies that the radially inward and outward flows at low levels played an important role in strengthening E4.

6.2 Merger of convective cells

The convective cell merger is a complex nonlinear process in which two identical convective cells tend to merge into a single cell (Simpson et al., 1980; Westcott 1984; Stalker and Knupp 2003). This is an
important process, which can further intensify the convective cells to become stronger than the previous unmerged counterpart. The merger of convective cells usually occurred during the development of the rainband, thereby strengthening the embedded convection. Figure 16a depicts the flow pattern relative to the movement of E2 (east-northeastward with a speed of 33 m$^{-1}$) at the 1.5 km level at 2139 UTC. It is evident that the flow in the vicinity of the core E2 turned radially inward at the low level during the developing state of E2. It induced a marked convergence ($< -0.4 \times 10^{-3}$ s$^{-1}$) in the area between E1 and E2. From 2146 to 2154 UTC, the convergent zone intensified and extended northeastward, enlarging the area of 40 dBZ that enclosed E1 and E2 with maximum convergence in absolute value larger than $1.0 \times 10^{-3}$ s$^{-1}$ (Figs. 16b, c).

In addition to generating convergence zones, the radially inward flow penetrating E2 seemed to be a channel to convey SW flow into E2 and its associated rainband efficiently. Basically, the wind field is primarily driven by the pressure gradient, especially for the tropical cyclone system. Therefore, the examination of the pressure perturbation in the vicinity of the rainband is significant for realizing the dynamic mechanism. Figure 17 depicts that a perturbation high pressure was located at the radially inner side of E1 at the 1.5 km level by 2150 UTC, coinciding with the location of the inner-edge downdraft (shown in Fig. 11 at $x = -10$ km and $y = -48$ km). The coupling of a perturbation high and marked downdraft also appeared at the low level in the analysis by Barnes et al. (1991). Furthermore, another perturbation high existed at the upwind side of the SW flow and E2 was just located between both the perturbation highs, where there was a wind deformation. The SW flow was thus distorted, turned into a southerly wind, and decelerated before reaching E1 owing to the weak pressure gradient between E1 and E2, which indicated the generation of convergence. It is evident that the distribution of two perturbation highs that occurred in the vicinity of rainband E could lead the SW flow deformation locally and fortify the mechanism of convergence, as well as the merger between E1 and E2 in the rainband.

6.3 Effect of SHWM in the vertical vorticity

While Barnes et al. (1983) and Barnes and Stossmeister (1986) first showed the tangential wind maximum associated with the hurricane rainband, Samsury and Zipser (1995) were the first to coin the term “second horizontal wind maximum” (SMWH). Hence and Houze (2008) confirmed the relationship between SHWM and convective cells by airborne observations. The radially inward flow penetrated the bottom of the SHWM and approached the convective-scale updraft region, where the low-level convergence of the radial wind stretched the vertical vorticity. The updraft located at the inner side of the SHWM would tilt the horizontal vorticity into

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**Fig. 15.** Same as Fig. 14 except for convective core E4 at 2224 UTC 8 August 2009. The white arrow indicates a streamline for the overturning updraft.
The maximum vertical vorticity along the vertical cross section $K-K'$ exceeded $1.6 \times 10^{-3}$ s$^{-1}$ at the height of 7 km as E1 was developing (Fig. 18a). The maximum vertical velocity, which was larger than 4 m s$^{-1}$, was superimposed on the intense vertical vorticity. The strong horizontal wind was located at the radially outer side of the convective cell at the height of 6 km. The same kinematic pattern also appeared in association with E4 at the developing stage (Fig. 18b). The maximum vertical vorticity was situated at the height of 6 km, somewhat lower than that at E1. The vertical velocity and SHWM could exceed 6 and 25 m s$^{-1}$, respectively, although the strength of the maximum vertical vorticity was also approximately $1.6 \times 10^{-3}$ s$^{-1}$ at that time.

While the pattern of the cross sections was similar to the conceptual model proposed by Hence and Houze (2009), the maximum vertical vorticity was weaker than that in the previous study. The relatively weak vertical vorticity may be attributed to two reasons. Since the typhoon circulation was disorganized and became an asymmetric pattern when sweeping across Taiwan’s topography, its vertical vorticity turned less intense because of the increase of surface friction. Moreover, the vertical shear below the SHWM was less than 1 m s$^{-1}$ km$^{-1}$, implying that the associated horizontal vorticity was mild.

### 7. Conclusions

The ability to understand the intensity changes of rainbands associated with a tropical storm remains one of the most challenging tasks. The airborne facility observations have examined the process between principal rainbands and accompanying convection in a tropical cyclone, confirming the conceptual model proposed by earlier studies. However, there are few systematic studies discussing the rainband and convective structures under the impingement of significant environmental flow on the typhoon circulation. During recent decades, the torrential rainfall associated with typhoon rainbands have been seen to inflict severe damage in the western Pacific region, especially in the southern Taiwan area.

This study analyzes the observational data collected by the ground weather radar network during the passage of Typhoon Morakot (2009) to explore the characteristics of convection embedded within the rainbands during the interaction between the typhoon circulation and SW flow. Furthermore, the evolution of these structures and their impact on local weather patterns will be discussed in detail.
of the cellular structure in the rainband is discussed.

The structure of Typhoon Morakot, as many studies have shown, was highly asymmetric. After the typhoon passed through the Taiwan area and moved further northward, the secondary band, which was characterized as a quasi-linear rainband with developing convective cells, replaced the upwind portion of the principal band. It was quasi-stationary and sustained for more than 10 h off the southwest coast of Taiwan. The dual-Doppler wind analysis revealed that the rainband was located at the transition zone between the westerly and SW flows. The transition zone may generate strong convergence to sustain the rainband.

The evolution of convective cells embedded within the rainband was also detected by dual-Doppler observations. While the rainband appeared stationary, the embedded convective cells propagated eastward with a speed of 25–28 m s⁻¹, which was accompanied with the westerly flow. The northward component of the SW flow decelerated at the upband portion of the rainband and generated intense convergence, triggering the development of new convective cells. The maximum absolute value of the convergence with an intensity of $1.5 \times 10^{-3}$ s⁻¹ moved eastward. The associated convective cells helped continuously sustain the rainband. Additionally, the isolated convective cells embedded within the SW flow gradually merged with the convective cells in the rainband. It is evident that the distribution of two perturbation highs that occurred in the vicinity of the rainband could lead SW flow deformation locally and fortify the mechanism of convergence, as well as the merger of convective cells in the rainband.

The vertical cross section of the rainband revealed that the overturning height of the updraft coincided with the top of the convective cell and two types of downdrafts (inner-edge and low-level downdrafts) were found. The feature is comparable with the conceptual model documented by previous studies. Nevertheless, the inner-edge downdraft coupled by the radially inward tilting convection was initiated by the precipitation drag, which was suggested by the overlap of the perturbed high pressure and downdrafts.

Fig. 17. Perturbation pressure (hPa) field from the Makung and Chiku radar data at 1.5 km altitude at 2150 UTC on 8 August 2009. The color shading shows the rainfall echo (dBZ). The contour represents the perturbation pressure field. The letter “H” indicates the perturbation high, while the letter “L” denotes the perturbation low.
at low levels.

The relationship between the horizontal wind maximum and vertical vorticity are also examined in this study. A key characteristic of the developing convective cells was that the strongest vertical velocity occurred at the middle levels, which was overlaid with intense vertical vorticity. Furthermore, the strong horizontal wind was located at the radially outer side of the convective cell. Although the dynamic and kinematic patterns revealed at the vertical cross section were similar to the conceptual model found by Hence and Houze (2008), the maximum vertical vorticity was weaker than that in the previous finding. The relatively weak vertical

Fig. 18. The vertical cross-sectional view of (a) the convective cell E1 along line K-K’ at 2139 UTC and (b) the convective cell E4 at 2216 UTC 8 August 2009 along L-L’. The color shading shows the vorticity field ($\times 10^{-3}$ s$^{-1}$). The white arrow indicates a streamline for the overturning updraft. The horizontal and vertical coordinates represent the radial distance and the height, respectively. The black and white contours denote the horizontal wind field (greater than 23 m s$^{-1}$ in 1 m s$^{-1}$ interval) and the vertical velocity (m s$^{-1}$), respectively.
vorticity possibly stemmed from two factors. First, the vertical vorticity brought by the typhoon circulation was weakened owing to the orographical destruction. Likewise, the vertical wind shear at the bottom of the SHWM was rather flat owing to the vertically stratiform feature of the SW flow. Hence, the vertical vorticity tube tilted by the updraft from the horizontal vorticity was mild, weakening its contribution to the strength of the SHWM.

In summary, the characteristics of the typhoon rainband and embedded convection, which developed in the circumstances that the environmental flow impinged on the typhoon circulation, were similar to previous concepts of the hurricane rainband except that the triggering mechanism of the convective downdrafts might not be the same. Additionally, the consecutive volume radar data showed the evolution of convective cells at different stages and maintaining mechanism of the rainband.

Although progress in the analysis of the kinematic and dynamic characteristics of convective cells embedded within the rainband has been made in this study, it should be remarked that some factors were not taken into account. For example, by analyzing the hydrometeor distribution from dual-polarimetric radar data collected by the Makung radar site, the origin of downdrafts in the rainband can be better recognized, thereby providing even better insight into the vertical cross section of cloud microphysics in the convective cells. In the future, the numerical model will be further tested to evaluate characteristics described in this study.

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