An Observational Study of a Coastal Barrier Jet Induced by a Landfalling Typhoon

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ABSTRACT

In this study, the structure and evolution of a coastal barrier jet (CBJ) along the east coast of Taiwan is documented using operational Doppler radars. The formation of the CBJ was controlled by the flow regime associated with the approaching Typhoon Haitang (2005). The CBJ persisted for 6 h and was approximately 140 km long and 25 km wide. The northern branch of the CBJ had stronger winds with maximum wind speed 49–52 m s⁻¹, a greater vertical extent with jet core between 1.0 and 2.5 km in height, and a more persistent jet signal than the southern branch with maximum wind speed 43–46 m s⁻¹ and jet core between 1.0 and 2.0 km. We investigated the terrain blocking effect leading to the CBJ formation using an idealized simulation. A vortex resembling Haitang is constructed based on circulation retrieved from generalized velocity track display (GVTD) technique. The result of a no-terrain simulation reveals wind speed 10–22 m s⁻¹ lower than the observed Doppler velocity. The difference suggests the enhanced wind speed along the coast was most likely due to the terrain blocking effect.

1. Introduction

A barrier jet is a low-level terrain-parallel strong wind induced by the blocking of airflow by orography (Schwerdtfeger 1975; Parish 1982). Barrier jets are frequently observed adjacent to quasi-two-dimensional topographic barriers worldwide.

Schwerdtfeger (1975) explained the barrier jet formation process. Whenever a stratified synoptic-scale flow blows steadily toward a mountain, the airflow is forced to lift along the windward slopes of said mountain. Because the rising airflow is stable, this forced ascent is resisted and decelerated. Deceleration of the airflow leads to air mass accumulation over the windward slope of the mountain and a local high pressure is established. Thus, a horizontal pressure gradient

directed away from the mountain is created. If this gradient persists for a few hours, geostrophic adjustment begins to occur and the Coriolis force eventually balances the pressure gradient force, resulting in geostrophic flow parallel to the mountain. Overland and Bond (1995) suggested that the strength and characteristics (i.e., length, width, and depth) of a barrier jet are dependent on the stability of the upstream flow, normal-terrain wind speed component, and dimensional scale of the mountain in question.

Taiwan is a mountainous island with complex terrain, and more than 70% of its area is at least 1000 m above sea level. The Central Mountain Range (CMR) has northeast–southwest orientation through the island and a length of approximately 340 km, width of 80 km, and average elevation of more than 3000 m. In early summer, a barrier jet is frequently created along the northwestern coast of Taiwan in the prefrontal southwesterly flow regime. Chen and Li (1995a,b) discovered locally strong winds at low levels along the northwestern coast of Taiwan and suggested that these

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strong winds were due to blocking of the southwesterly flow by the island obstacle. Li and Chen (1998) defined a barrier jet as a strong southwesterly flow along the northwestern coast of Taiwan when the wind speed has a local maximum (10 m s^{-1}) between 0.5 and 1.5 km above sea level. By analyzing the sounding data of 14 barrier jet occurrences during the Taiwan Area Mesoscale Experiment (TAMEX), Li and Chen (1998) demonstrated that jet core was located at a height of approximately 1 km and had an average wind speed of 14 m s^{-1} . Because northwestern Taiwan has few sounding stations, identifying the dimension of a barrier jet is difficult.

Using TAMEX IOP13 data, Jou and Deng (1992) and Li et al. (1997) have found that the barrier jet along the northwestern coast of Taiwan plays an important role on triggering deep convections over the cyclonic shear side of the jet and causes heavy rainfall. Furthermore, by analyzing the TAMEX IOP4 data, Trier et al. (1990) discovered after surface front passed over NE Taiwan, northeasterly cold-air stacking over the windward slope led to intensification of the northerly wind parallel to the coastal mountain over East Taiwan. Although the winds parallel to the steep topography along the east coast were only moderate in strength ($5\text{--}7 \text{ m s}^{-1}$) and ship data over the adjacent ocean were lacking, Trier et al. (1990) suggested the presence of a barrier jet parallel to the mountains along the east coast of Taiwan. This barrier jet caused rapid southward propagation and distortion of the surface front along the east coast.

Barrier jets can be induced by landfalling typhoons. Lin et al. (1999) and Lin et al. (2005) have conducted idealized numerical experiments to examine track deflection when a vortex approaches an idealized terrain. A strong low-level northerly jet was found between the terrain and vortex center. Research suggested that this jet had been due to blocking and channeling effects and the low-level northerly jet was considered the key factor causing southward vortex deflection.

Jian and Wu (2008) used Weather Research and Forecasting (WRF) Model (horizontal resolution 4 km) to simulate the landfall behavior of Typhoon Haitang (2005). The results showed that the west side of the eyewall circulation was intensified and a low-level northerly jet had formed owing to blocking and channeling effects when Typhoon Haitang approached the east coast of Taiwan. This asymmetric inner-core circulation induced a southward advection flow that caused the typhoon to turn sharply southward. Jian (2011) used the same simulation data to further examine the asymmetric eyewall convection evolution of Haitang. An enhanced rainband along the southern

quadrant of the eyewall was attributed to low-level convergence between the circular orographic flow and low-level northerly jet. This enhanced rainband eventually merged into the eyewall and caused the asymmetric convection evolution of the eyewall and generated a looping track. Huang et al. (2011) conducted a high-resolution numerical experiment (horizontal resolution 3 km) to investigate the track deflection of Typhoon Krosa (2007). Their results further supported the channeling effect concept.

Mashiko (2008) constructed a high-resolution numerical model (horizontal resolution 2 km) to simulate Typhoon Ma-on, which landed at the Kanto Plain of Japan on 9 October 2004. A strong cold northeasterly wind was observed on the Kanto Plain as Ma-on moved north-northeast into Tokyo Bay. A warm southerly wind from the south sector of Ma-on encountered a strong northeasterly wind, resulting in a mesoscale front on the Kanto Plain while Ma-on underwent extratropical cyclone transition. A north-south low-level northerly jet formed between the Kanto Mountains and the typhoon's center because of mountain blockage. This jet was accompanied by considerable cold airflow with a depth of approximately 400 m. The airflow's axis was located at an altitude of approximately 250 m and the maximum wind speed was higher than 60 m s^{-1} . The acceleration of the low-level northerly jet was supported by strong pressure gradient force in the north-south direction parallel to the Kanto Mountains. The duration of the low-level northerly jet was very short ($\sim 40 \text{ min}$) because of the rapid moving speed (70 km h^{-1}) of Ma-on.

Typhoons are the most threatening weather system in the area around Taiwan. On average, three to four typhoons affect Taiwan every year, and strong winds and torrential rainfall that accompany typhoons often cause considerable loss of human life and property. Because of complex interaction between typhoon circulation and terrain, dramatic changes in trajectory, intensity, circulation, and rainfall structures are expected when typhoon approaching the island and render precise forecasting of typhoons highly challenging (Wu and Kuo 1999). According to previous numerical studies, a barrier jet is a mesoscale phenomenon that can crucially affect typhoon behavior during the landfall period. However, detailed structural characteristics, which include maximum wind speed, height of jet core, length, and width, have not been well documented and jet evolution involving formation timing, strength, and spatial structural change relative to typhoon motion and structural change remains unclear. Furthermore, a detailed analysis from the observational perspective is rarely performed.

The long-range weather radar network of Taiwan was established in 2001 and can continuously survey typhoon

FIG. 1. Best track of Typhoon Haitang (0505) and various observations used in this study. The terrain height (in MSL) is indicated by shading. The Snow Mountain Range, Central Mountain Range, Coastal Mountain Range, and East Rift Valley are labeled. Blue line is the best track from the CWB of Taiwan with dates indicated (date/hour in UTC). Hua-lien (RCHL) and Wu-Fen-San (RCWF) Doppler radars are denoted by radar symbols. Surface-observing stations are denoted by red dots (surface station name abbreviations are listed in Table 1). The sounding station at Hua-lien (466990) is denoted by a solid square.

activity at distances of up to 460 km when using low-level reflectivity and within approximately 230 km when using both the volumetric reflectivity and Doppler velocity data. Typhoon Haitang, a super typhoon and the fifth typhoon of the 2005 northwest Pacific typhoon season (0505), made a loop before its landfall on the east coast of Taiwan (Fig. 1). During this period, a pronounced coastal barrier jet (CBJ) was identified along the coast, and dramatic wind and rainfall structure changes were detected from the high spatial and temporal resolution Doppler radar data. This study aims to document the CBJ associated with Typhoon Haitang and to identify the physical processes of formation mechanism and structure evolution from an observational perspective for the first time. Moreover, in order to quantify terrain effect on the CBJ formation, an idealized vortex experiment is conducted as well. The results of this study can be used as a reference for numerical simulation studies.

Section 2 describes the data and methodology. Section 3 presents an overview of Typhoon Haitang. The structure and evolution of the CBJ is documented in section 4. The possible CBJ formation mechanism and the terrain effect on the CBJ formation is also examined. Section 5 provides a summary and the conclusions.

2. Data and methodology

a. Data

The datasets used in this study are summarized in Fig. 1. Observations from the Central Weather Bureau (CWB) operational S-band Doppler radar located at Hua-lien (RCHL) on the east coast of Taiwan is used to document the CBJ and rainfall structure of Typhoon Haitang. Data from another CWB operational S-band Doppler radar (WSR-88D; RCWF) located at Wu-Fen-San near the northeast coast is also employed. The characteristics of these two radars are provided in

appendix A. A total of 154 volumes of RCHL data obtained between 2103 UTC 16 July and 2233 UTC 17 July 2005, are analyzed. Because of strong winds, RCHL was shut down at 2233 UTC 17 July 2005. Regarding RCWF, because of the blockage of the Snow Mountain Range (Fig. 1), the radar could not detect the main portion of the CBJ, especially below 2 km.

In addition to radar data, hourly wind observations at surface stations along the east coast of Taiwan are used to describe the surface flow during Typhoon Haitang's landfall. Twice daily sounding data from Hua-lien (466990) is used to verify the environmental low-level vertical wind profile retrieved from the Doppler radars and to estimate the Brunt–Väisälä frequency for Froude number calculation.

The sea surface wind vector product of QuikSCAT (<https://manati.star.nesdis.noaa.gov/datasets/QuikSCATData.php>) is used to define the outer circulation of Typhoon Haitang. QuikSCAT, which was equipped with the SeaWinds instrument, was designed to retrieve the surface winds over the ocean and had a 1800-km-wide measurement swath. Generally, QuikSCAT had twice-daily coverage with a spatial resolution of approximately 0.25° latitude.

The global atmospheric reanalysis European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim) dataset is used to analyze the synoptic environment, including the steering flow, vertical wind shear of Typhoon Haitang, and Froude number. The spatial and temporal resolution of the ERA-Interim data is 0.75° latitude and 6 h, respectively.

b. Methodology

The horizontal wind field derived from dual-Doppler synthesis of RCWF and RCHL radial Doppler velocity (e.g., Ray et al. 1980) is the best way to depict the kinematic structure of CBJ. However, because of geometric limitation and blockage of mountain, the horizontal winds of CBJ cannot be obtained directly from dual-Doppler synthesis. Nonetheless, dual-Doppler synthesis is still applied in this study to deduce part of primary circulation of Haitang before it made landfall.

V_r is wind velocity along the radar beam, also called the radial velocity. The cross-beam component of the horizontal wind cannot be detected by Doppler radar. With the assumption of linear variation of the horizontal winds, Browning and Wexler (1968) has developed the velocity–azimuth display technique (VAD) to retrieve vertical profile of the horizontal wind with one single elevation sweep. For retrieving more complicated structure of the flow, Srivastava et al. (1986) developed the extended VAD (EVAD) method. Deng and Jou

(1994) performed an error analysis of EVAD technique using an idealized circular velocity pattern resembled a tropical cyclone. The result showed that the error of retrieved wind speed and wind direction is less than 5% and 3° , respectively.

An idealized vortex experiment with circulation similar to Haitang was constructed to examine the possible terrain effect. Techniques to retrieve TC circulation and intensity using ground-based single Doppler radar have been developed in late 90th [ground-based velocity track display (GBVTD; Lee et al. 1999, 2000); Generalized-VTD; Jou et al. 2008]. The retrieved TC circulation includes axisymmetric tangential wind, axisymmetric radial wind, and high wavenumber tangential and radial winds. Error analysis indicates reasonably accurate retrieved circulations can be obtained (Jou et al. 2008). Recently, Shimada et al. (2016) applied the technique to landfall typhoons over Japan area and found reasonably accurate intensity estimation were obtained.

In current study, the retrieved axisymmetric tangential wind (V_T) is adopted to construct the idealized vortex. It is worth noting that the retrieved V_T is sensitive to the position of TC center and wind speed of the environmental mean flow (V_M). If the uncertainty of TC center is less than 5 km, the retrieved speed error of V_T is less than 20%. On the other hand, if the speed uncertainty of V_M is less than 50%, the retrieved speed error of V_T is less than 10%. During the analysis period, radar reflectivity image of Haitang revealed a clear eye pattern and a steady movement. It suggests the uncertainties of center locations and the environmental mean flow are all small.

3. Typhoon Haitang

Typhoon Haitang originated from a tropical disturbance at subtropical latitude (23°N) and subsequently strengthened to a tropical storm (17 m s^{-1}) at 0000 UTC 12 July 2005. It intensified into a typhoon (33 m s^{-1}) at 1800 UTC 13 July and reached peak intensity as a category 5 super typhoon at 1200 UTC 16 July with an estimated minimum mean sea level pressure of 898 mb and maximum sustained wind of 140 kt ($1\text{ kt} \approx 0.5144\text{ m s}^{-1}$) based on the Joint Typhoon Warning Center best track. On 17 July, Haitang headed west-northwestward and weakened while approaching the east coast of Taiwan. As Haitang approached the coast, it made a cyclonic-looping track and then made landfall at approximately 0650 UTC 18 July (Fig. 1). It is interesting to note that the CBJ started to occur while Haitang center still 250 km away from the coast and disrupted before Haitang made a looping track.

FIG. 2. (a) QuikSCAT sea surface winds at 2148 UTC 16 Jul; wind vector is plotted using standard wind barbs, with the IR satellite image overlaid. Color shading represents the reflectivity (dBZ) at the 0.5° elevation PPI of RCHL. Red dot is the location of HL surface station, blue line is the track of Typhoon Haitang from the CWB of Taiwan, and red circle is the 600-km radius from the typhoon center. (b) Azimuthally averaged sea surface wind radial profile from QuikSCAT within a 600-km radius of the typhoon center (black line), and hourly averaged wind speed at HL surface station (blue line). Abscissa is the distance from the typhoon center (and also time, labeled at the top of the panel), and ordinate is the wind speed. Double-headed lines indicate the outer circulation (400 km) and core (180 km) regions. (c) Wind profile below 5-km height, obtained at Hua-lien (466990) sounding station. Blue (red) line indicates 0000 UTC (1200 UTC) 17 Jul.

A typhoon's wind field can be considered circular. As typhoon moves toward a mountain barrier, the incident angle to the coastal terrain and the wind speed of the incoming branch of the cyclonic vortex circulation changes considerably. Multiple flow regimes are usually associated with an approaching typhoon (Lin et al. 2005). Therefore, identify size of the outer and core

circulation of Haitang is helpful to clarify the evolving aspect of the CBJ.

At 2148 UTC 16 July, when Haitang was approximately 550 km from the east coast of Taiwan, QuikSCAT passed over Taiwan, and most of Haitang circulation was covered except for the northeastern periphery (Fig. 2a). This single swath of QuikSCAT data is used to define

outer circulation extent of Haitang. The criterion of the outer circulation of a typhoon is wind speed of more than 17 m s^{-1} (Merrill 1984). The azimuthally averaged radial wind profile within the 600-km-radius range in Fig. 2b gives a maximum surface wind speed of 27 m s^{-1} at a 90-km radius and 17 m s^{-1} at a 400-km radius. QuikSCAT quality is better away from strong precipitation. The data are considered quite accurate for wind speed between 3 and 20 m s^{-1} in rain-free areas with the bias lower than 20% (Zeng and Brown 1998; Hoffman and Leidner 2005; Zou et al. 2015). Because the periphery of a typhoon has less rain and relatively weaker wind speed, defining the radius of the outer circulation region as 400 km is reasonable.

For inner-core circulation, Willoughby (1988) defined the core region of TC 100–200 km away from the storm center where the local Rossby number (Rossby number, $\text{Ro} = V/fr$, the ratio of centrifugal to Coriolis force) is greater than 1. In the study, we followed Tang et al. (2014), the radius of core region of Haitang is defined as twice the radius of maximum wind (i.e., 180 km).

The wind speed information at the coastal surface station Hua-Lien (HL; a list of surface station names and abbreviations is shown in Table 1) and sounding station (466990) are examined to tentatively identify the CBJ signal. The blue line in Fig. 2b represents the hourly surface wind speed at HL. The wind speed exhibits an increasing tendency when Typhoon Haitang approached the coast. Between 0600 and 1500 UTC 17 July (when the distance between the typhoon center and HL was 200–380 km), the wind speed increased at a rate of approximately $6.36 \times 10^{-2} \text{ m s}^{-2}$. The wind speed increasing rate at HL is significantly higher than the increasing rate of the mean tangential wind speed ($\sim 3.02 \times 10^{-2} \text{ m s}^{-2}$) over the open ocean estimated from the QuikSCAT. It is noted that from surface pressure observations, there was no obvious sign of intensity increase of Haitang in this period. The result seems to suggest that coastal terrain may play role on it. It is also noted that vertical profiles of the horizontal winds of HL soundings at 0000 and 1200 UTC (Fig. 2c) showed no jet signals. The jetlike structure wind appeared after 1200 UTC.

4. CBJ

a. Spatial structure

Figure 3a presents an image of the composite reflectivity at 1910 UTC 17 July 2005, whereas Figs. 3b and 3c show Vr images at the 2.5-km height CAPPI of RCHL and RCWF at 1913 UTC, respectively. The horizontal winds (u and v) derived from the dual-Doppler synthesis method at 2.5-km altitude is overlaid in Fig. 3b

TABLE 1. List of surface stations name abbreviations.

Surface station name	Abbreviation
Peng-jia-yu	PJY
Yi-lan	YL
Su-ao	SA
Huo-chung	HC
Hua-lien	HL
Kuang-fu	KF
Feng-bin	FB
Wu-he	WH
Cheng-kung	CK
Tai-tung	TT
Da-wu	DW
Lan-yu	LY
Yonaguni island	YNG

as well. In this moment, Haitang's center was approximately 120 km offshore east of RCHL. It is noted the locations of the inbound and outbound maximum Vr of the inner-core region of Haitang (Fig. 3b) are consistent with the inner and outer eyewall in the reflectivity field (Fig. 3a). Nevertheless, in addition to the eyewall signature of strong Vr, there existed another strong Vr signature located at the coastal region. The coastal signature of strong radial velocity showed a negative maximum (approaching) wind on the northbound of the radar and a positive maximum (receding) wind southbound of the radar indicating the existence of a northerly jetlike wind feature. The magnitude of the jet was comparable to that near the eyewall region. The RCWF image (Fig. 3c) shows a relative maximum outbound velocity area located in the north of RCHL. This observation further corroborates a jet signal over HL area. Here, we define this locally strong Vr signal along the east coast which far from the eyewall region of Haitang as CBJ.

Because of the inherent limitation of the RCHL–RCWF dual-Doppler horizontal wind synthesis algorithm, only part of the northwestern circulation of Haitang's outer eyewall could be retrieved. Nevertheless, the result provides further evidence for the existence of a CBJ. By examining the northwestern portion of the outer eyewall in Fig. 3b, the area where the retrieved horizontal winds are $>40 \text{ m s}^{-1}$ is found to be consistent with the area of inbound radial velocity $\text{Vr} > 40 \text{ m s}^{-1}$, indicating that the single Doppler radar radial velocity (along beam component only) has captured most of the wind information. This result also supports the direction of the CBJ being mostly from north to south along the east coast.

Figure 4 displays the time–height cross section of the horizontal winds retrieved from RCHL by using EVAD technique. The time series covers the period 2103 UTC 16 July to 2233 UTC 17 July 2005, almost 26-h period

FIG. 3. (a) Composite reflectivity field of operational radar network of CWB at 1910 UTC 17 Jul. (b) Horizontal CAPPI Doppler radial velocity image of RCHL overlaid with dual-Doppler synthesis wind (black contour represents wind speed over 40 m s^{-1}) at 2.5-km height at 1913 UTC 17 Jul. Cool (warm) colors indicate inbound (outbound) flow; gray gradient represents terrain; abscissa and ordinate are the relative distance (km) to RCHL. (c) As in (b), but for RCWF.

with 154 data points. These horizontal winds can be treated as local winds averaged over an area with a radius of 30 km. Wind speed larger than 40 m s^{-1} appeared after 1400 UTC 17 July, especially below 3-km height. This strong wind signature lasted more than 6 h between

1400 and 2100 UTC 17 July before eyewall moved over the coastal region. This result provides a direct evidence of the existence of CBJ.

To define the CBJ more objectively, several criteria are employed to select a jetlike profile during the

FIG. 4. (top) Time–height cross section of horizontal wind retrieved from RCHL using the EVAD technique from 2103 UTC 16 Jul to 2233 UTC 17 Jul. Color shading is wind speed and area of wind speed over 40 m s^{-1} is overlaid by black contours. Bold blue contour is divergence (10^{-5} s^{-1}). (bottom) Red color bars indicate vertical wind profiles which meet the criterion of a jetlike profile (see text).

analysis period. The criteria of a jetlike profile include a maximum wind speed below 5-km height of $>30 \text{ m s}^{-1}$ (30 m s^{-1} is the average maximum wind speed below 5-km height in the 154 analysis data points) and a vertical wind shear above (below) the height of maximum wind speed $>1.4 \times 10^{-3} \text{ s}^{-1}$ ($2.0 \times 10^{-3} \text{ s}^{-1}$). Vertical wind shear above (below) the height of maximum wind speed is calculated by searching for the minimum wind speed altitude between the maximum wind speed altitude and 5-km height (0.5-km height). According to these criteria, 40 jetlike profiles are identified as red bars in bottom panel of Fig. 4. Through this definition, the CBJ period can be identified from 1423 to 2003 UTC 17 July and 31 profiles are included.

The average profile of 31 jetlike horizontal winds profiles during the CBJ period reveals the general vertical characteristic of the CBJ (Fig. 5a). The average maximum wind speed indicates the height of the maximum wind is at 2.5 km and with a magnitude of 41.9 m s^{-1} a standard deviation 1.9 m s^{-1} . We thus define the core of CBJ is at 2.5 km. The average vertical wind shear is $6.3 \times 10^{-3} \text{ s}^{-1}$ below and $2.2 \times 10^{-3} \text{ s}^{-1}$ above the CBJ core, and the wind speed threshold of the CBJ at 2.5-km height is defined as 40 m s^{-1} (average maximum wind speed minus one standard deviation).

Figure 5b displays the occurrence frequency of V_r over 40 m s^{-1} at 2.5-km height during the CBJ period (i.e., the number of pixels with $V_r > 40 \text{ m s}^{-1}$ and the maximum number is 31). The generally feature of the CBJ is clearly depicted. The CBJ is located along

the east coast in a terrain-parallel direction and is approximately 140 km long and 25 km wide. The CBJ occurrence frequency is higher in the northern branch implying greater persistence over time. The composite vertical cross section of the maximum V_r along the axis of the CBJ (Fig. 5c) shows stronger wind speeds (maximum wind speed $49\text{--}52 \text{ m s}^{-1}$) and a greater vertical extent (core of the jet is located at 1.0–2.5 km in height) along the northern branch. By contrast, weaker wind speeds (maximum wind speed $43\text{--}46 \text{ m s}^{-1}$) and a smaller vertical extent (core of the jet is located at 1.0–2.0 km in height) are noted along the southern branch.

Jian and Wu (2008) conducted a numerical experiment demonstrating that the formation of the CBJ on the western side of the eyewall of Haitang (i.e., between the typhoon center and east coast terrain) during the landfall period was related to the orographic blocking and channeling effects. Backward trajectory analysis of air parcels from the low-level atmosphere (between 900- and 850-mb pressure levels) arriving at the western part of Haitang reveals that these low-level air parcels accelerated when moving from a wider area to a narrower channel while experiencing substantial confluence and gradual ascent. Estimation of the divergence by using EVAD shows clear convergence associated with the CBJ signal (bold blue contours in Fig. 4).

The maximum magnitude of the convergence is approximately $6 \times 10^{-4} \text{ s}^{-1}$ below 2 km [similar to the numerical simulation result of Jian and Wu (2008)]. The higher wind speed and greater vertical extent at

FIG. 5. (a) Vertical wind profile during the CBJ period (black line is average, light blue shading shows the standard deviation of the 31 jetlike profiles). Average maximum wind speed (standard deviation) is $41.9 (1.9) \text{ m s}^{-1}$ at 2.5-km height. (b) Frequency of horizontal distribution of Doppler radial velocity (V_r) over 40 m s^{-1} (defined as a wind speed criterion of a CBJ, see text), at 2.5-km altitude. The rectangular box indicates the analysis domain of the temporal evolution of the CBJ (145 km long, 70 km wide). Black lines $\overline{on}$ and $\overline{os}$ indicate the direction of the cross sections in (c). (c) Composite maximum V_r (color bar on right-hand side, unit m s^{-1}), along the axis of the CBJ [line $\overline{on}$ and $\overline{os}$ in (b)] during the CBJ period (31 analysis data points). Abscissa is the distance from the radar, and ordinate is the altitude.

the northern branch of the CBJ are likely due to the convergence associated with the channeling effect. The stronger prevailing northeasterly wind impinged upon the northeast of the CMR with higher terrain height, resulting in a strong channeling effect associated with convergence that forced a stronger vertical velocity.

By contrast, the weaker wind speed and smaller vertical extent of the southern branch of the CBJ might have been due to the topographic features of the east coast. The Coastal Mountain Range has a lower altitude and a northeast–southwest orientation along the east coast south of HL, with a 1-km average mountain height. The East Rift Valley located beside the Coastal Mountain Range (see Fig. 1 for the location) is long and narrow ($\sim 150 \text{ km}$ long, 1–3 km wide). The southern branch of the CBJ may have diverged because of the Coastal Mountain Range. The wind speed and vertical extent were therefore smaller than those of the northern branch.

The surface wind at stations HC and FB (located along the coast line, see Fig. 1 for the location) and KF

and WH (located in the East Rift Valley, see Fig. 1 for the location) are examined to verify that the CBJ split here (Fig. 6). During the CBJ period, the persistent northerly wind substantially increased in speed at HC station (located north of RCHL, corresponding to the CBJ northern branch). Conversely, the northerly wind at FB station (located south of RCHL, corresponding to the CBJ southern branch) first increased and then decreased in general (this phenomenon, which is related to the evolution of the CBJ, is explained in section 4b) and was weaker than the wind at HC station. This difference in surface wind speed between HC and FB is consistent with the V_r analysis.

By further comparing the surface wind between the coastline (FB station) and East Rift Valley (KF station), a similar variation is discovered between these two stations. At the beginning of the CBJ period (before 1600 UTC 17 July), both KF and FB experienced northerly winds of increasingly high speed (although the wind speed at KF was lower because of greater surface

FIG. 6. Time series of surface wind at coastline stations (HC and FB) and East Rift Valley stations (KF and WH, see Fig. 1 for the locations). Blue lines are wind speed, and light red shading area indicates the CBJ period. Abscissa is time (time arrow is from right to left), and ordinate is wind speed.

friction), implying that the southern branch of the CBJ was split by the Coastal Mountain Range.

b. Evolution

The evolution of the CBJ is investigated by examining the temporal variation in the maximum V_r at 2.5-km altitude along the long axis of the selected domain (black rectangular box in Fig. 5b) covering the CBJ horizontal extent between 2103 UTC 16 July and 2233 UTC 17 July (Fig. 7a). Four distinct periods of the CBJ can be identified according to the pattern and rate of change of V_r .

Period I was the formation stage and corresponds to before 0503 UTC 17 July, that is, before the outer circulation of Haitang reached the east coast. In this period, V_r slowly increased (i.e., the rate of change of V_r was $\sim 1.0 \times 10^{-4} \text{ m s}^{-2}$). The analysis domain was affected by the passage of several outer rainbands (Fig. 7b, 0103 UTC). Intermittent strong wind was the main feature, with passing rainbands associated with higher wind speed ($> 20 \text{ m s}^{-1}$). Period II was the intensifying stage and lasted from 0503 to 1413 UTC 17 July. In this period, the distance between Haitang's center and the coast was 220–410 km. V_r increased dramatically, and the rate of change of V_r was approximately $7.0 \times 10^{-4} \text{ m s}^{-2}$. Similar to Period I, intermittent higher

wind was associated with passing rainbands (Fig. 7b, 1403 UTC); nevertheless, the major feature of Period II was the considerable increase of V_r over time. After 0800 UTC 17 July, V_r began to exceed 30 m s^{-1} in some areas. This feature is consistent with the surface wind increase at HL.

Period III was the mature stage, lasting from 1423 to 2003 UTC 17 July, and the CBJ period, as identified in section 4a. In this period, the core of Haitang covered the analysis domain, and the rate of change of V_r was low and steady ($\sim 0.4 \times 10^{-4} \text{ m s}^{-2}$). The CBJ signal with $V_r > 40 \text{ m s}^{-1}$ began to dominate. Two distinct features can be identified in this period. Before 1613 UTC, the CBJ extended in the meridional direction both north and south and with much stronger V_r in the northern branch than the southern branch. V_r began to exceed 45 m s^{-1} in some areas. After 1613 UTC, the area with $V_r > 45 \text{ m s}^{-1}$ became persistent in the northern branch. However, the CBJ signal south of HL disappeared. Once the core region of Haitang had reached the coast (1803 UTC), the meridional extent of the CBJ began to decrease, that is the $V_r > 30 \text{ m s}^{-1}$ area shrank. The shrinkage was associated with formation of an east–west rainband south of Haitang's center (Fig. 7b, 1933 UTC). This rainband merged into the eyewall and distorted the inner-core precipitation structure of Haitang.

Period IV was the final stage, lasting from 2023 UTC 17 July. In this period, the analysis domain was influenced by the eyewall (Fig. 7b, 2203 UTC), and the CBJ signal decayed significantly. V_r shows typhoon-strength wind at all levels. No jet structure was observed in this period. Notably, in the far south, the area with weak southerly wind expanded rapidly over time, especially below 2.5-km, indicating a recirculating flow around the southern tip of the island.

The V_r patterns at individual analysis times are examined to clarify the detailed aspects of the CBJ evolution (Fig. 8). At 1543 UTC 17 July (Fig. 8a), Haitang's center was located approximately 180 km from the coast, and the CBJ had a terrain-parallel orientation. It is interesting to note that in the northern branch of the CBJ has a wide extend and stronger intensity at 2.5-km than 1.0-km height. However, in the southern branch, the opposite feature was observed. By 1853 UTC (Fig. 8b), Haitang's center had moved to 130 km from the coast and the inner-core region had reached the coast. The northern branch of the CBJ had strengthened, especially at 1.0-km height. The southern branch of the CBJ had weakened and shifted off the coast and finally become southerly.

Several main features of the CBJ also can be found based on surface station observations. Figure 9 presents the time series of surface winds at eight stations along Taiwan's east coast (see locations in Fig. 1).

FIG. 7. (a) Temporal variation of maximum Vr (shading) in direction normal to the long axis of the rectangular box marked in Fig. 5b at 2.5-km altitude between 2103 UTC 16 Jul and 2233 UTC 17 Jul. Abscissa is time, and ordinate is distance to RCHL, positive (negative) indicating to the north (south) of RCHL. Black contour is maximum reflectivity in direction normal to the long axis over a 40-dBZ area. Maximum wind speed over the four periods (Period I–IV) is depicted in the lower panel. Distance between the typhoon center and HL is depicted in the upper panel. (b) Reflectivity (dBZ) images of Typhoon Haitang at 0.5° elevation PPI of RCHL (from right to left) at 0103, 1403, 1933, and 2203 UTC 17 Jul. Bold black circles indicate 400- and 180-km radii from the typhoon center.

The time window is the same as that in Fig. 7a. First, the wind speed was quite high ($>13 \text{ m s}^{-1}$) along the coast between SA and CK during the CBJ period (see the rectangular black box). To the south of CBJ, there existed low wind speed area between CK and DW, the wake zone. This feature implies the length scale of the

CBJ. The distance between SA and CK is approximately 130 km, which is consistent with the length of the CBJ estimated using radar observations. Second, the surface wind evolution patterns at SA, HL, and CK during the CBJ period are similar to the Vr evolution pattern (Fig. 7a), implying that the surface

FIG. 8. CAPPI Doppler radial velocity image for RCHL at (top) 1-km height and (bottom) 2.5-km height during the CBJ period for (a) 1543 UTC and (b) 1853 UTC 17 Jul.

observations also reflect the major features of the CBJ evolution.

Before the CBJ period, the surface wind speeds at HL and CK exhibited large fluctuations because of the outer rainband passage (corresponding to Period I in Fig. 7a) but tended to increase because of Haitang's circulation influence (corresponding to Period II in Fig. 7a). For the CBJ period, indicated by the rectangular box in Fig. 9, the wind speed change is similar to that for Period III in Fig. 7a. After 1600 UTC 17 July, the wind speeds at HL and CK decreased over time (the wind direction at CK even shifted from northerly to southerly), corresponding to the southern portion in Fig. 7a after 1800 UTC 17 July. The wind speeds at SA tended to increase slightly, similar to the northern portion in Fig. 7a. By comparing the V_r evolution (Fig. 7a) with the surface wind speed evolution (Fig. 9) in the CBJ period, a time difference ~ 2 h between decrease of V_r and HL surface wind is noted.

The above analysis suggests the behavior of flow near coast was strongly influenced by the approaching typhoon. In the following, the possible relation between the circulation structure of Haitang and the formation timing and evolution of CBJ is explored. The outer circulation flow of Haitang was associated with a small curvature and quasi-steady state (smaller variation in wind speed and direction with relative weaker vorticity), whereas the flow in the core region was corresponded to a large curvature, with considerable variation in the wind speed and direction (relative stronger vorticity). When the CBJ developed, the outer circulation with a vorticity of approximately $0.64 \times 10^{-4} \text{ s}^{-1}$ deduced from GVTD technique was under the influence of the coastal terrain, thereby generating persistent strong northeasterly wind at the coast and a terrain-parallel orientation of the CBJ.

When the core region approached the coastal terrain, the considerable curvature of core-region circulation

FIG. 9. Time series of surface wind at eight stations along the east coast of Taiwan. Shaded area indicates wind speed, and dashed contour lines indicate divergence (10^{-4} s^{-1}). Rectangular box indicates the CBJ period and location. Abscissa is time. Relative latitude positions of the typhoon center are indicated by typhoon symbols. Distance between the typhoon center and HL is depicted in the upper panel.

started to affect the coast, and stronger vorticity ($\sim 1.23 \times 10^{-4} \text{ s}^{-1}$) changed the CBJ structure, with the southern branch of the CBJ shifting offshore. This shift resulted in both disappearance of strong wind on the south side of the radar after 1803 UTC 17 July and a decrease in surface wind speed after 1600 UTC 17 July. The wind speed began to decrease at 2.5-km height in the southern branch of the CBJ approximately 2 h later than the surface wind, likely because the southern branch of the CBJ axis moved into an offshore direction near the surface more significantly (Figs. 8a,b). The time difference of wind speed weakening between surface and 2.5 km altitude can be explained by the differential friction.

c. CBJ formation mechanism

Although many earlier studies suggested the formation of barrier jet is due to flow blocking, it is still worthwhile to explore the forcing mechanism quantitatively caused by an approaching typhoon. For the purpose of examining terrain blocking effect, estimation of the Froude number and balance-of-forces analysis are conducted in this section.

The Froude number ($\text{Fr} = U/NH$, where U is the upstream wind speed normal to the mountain, N is the Brunt–Väisälä frequency, and H is the terrain height) can be calculated to assess the flow blocking potential.

According to earlier theoretical analyses, flow over the mountain under a blocking regime is favored when $\text{Fr} < 1$ (Smolarkiewicz et al. 1988; Smith 1989; Sun et al. 1991).

Figure 10 presents a schematic diagram for the Froude number estimation. Accounting for the position of maximum frequency of wind speed over 40 m s^{-1} (Fig. 5b) and because the CBJ feature was more persistent to the north of HL, the topography in the rectangular box in Fig. 10a is selected as a target for estimating the Froude number. The maximum terrain height (2686 m) in the selected box is used to represent H . The sounding information obtained at 1200 UTC 17 July at HL is used to estimate N (10.3×10^{-3} ; the detailed calculation is provided in appendix B). In upstream flow estimation, because the YNG surface station was located approximately 150 km from the target terrain and the typhoon circulation had considerable curvature, thus the surface wind at YNG is not appropriate to represent the upstream flow. Instead, the ECMWF ERA-Interim wind at point B is employed because it is more appropriate to represent the upstream flow.

Comparing the ERA-Interim 10-m wind at point A (the nearest grid point to YNG) with the surface wind at YNG reveals that wind speed and direction are in agreement over time, except for a discrepancy in the wind speed increase after 1200 UTC 17 July (blue and

TABLE 2. Comparison between the northern and southern branch of the CBJ. The discrepancy in wind speed between point A and YNG is added to the normal component of incident flow estimation at points B and C.

Variable definition	Variable name	Northern CBJ	Southern CBJ
h_m	Max mountain height	2686 m	1682 m
N_m (s^{-1})	Moist static stability	7.54×10^{-3}	7.54×10^{-3}
U ($m s^{-1}$)	Normal component of incident flow	15.28	11.87
$h_f(m) = U/N_m$	Gravity height	2028	1574
$Fr = U/N_m h_m$	Froude number	0.75	0.93

between SA and HL (~ 75 km), V_s is the mean along-coast wind between SA and HL, and ∂V_s is the difference of along-coast wind between SA and HL. The calculation of term I is $4.02 \times 10^{-4} m s^{-2}$.

To reasonably estimate the along-barrier pressure gradient (Term II), the pressure field associated with Haitang is removed from SA and HL first. By removing the axisymmetric pressure field associated with Haitang, the pressure gradient between SA and HL is calculated as $6.85 \times 10^{-3} m s^{-2}$ (e.g., the calculation procedure is given in [appendix C](#)). The Coriolis force term (Term III) is $4.72 \times 10^{-5} m s^{-2}$. The frictional force is computed to be a residual. The analyses show that the pressure gradient force term is the dominant term in Eq. (1), which suggests that the acceleration of flow downstream along the northeastern coast was mainly caused by the pressure gradient force. The budget analysis reinforces the conclusion that blocking is the major mechanism for the formation of CBJ in this case.

d. Idealized vortex experiment

To gain more quantitative information about the terrain-induced CBJ during landfall of Haitang, an idealized vortex experiment is conducted. The model is frictionless and the vortex is designed based on the retrieved mean tangential wind only. Details of idealized vortex structure is presented in [appendix C](#). [Figure 11a](#) shows observation of Vr at 2.5-km height at 1933 UTC 17 July and [Fig. 11b](#) is an image of Vr for the idealized vortex overlaid with mean flow ($6.4 m s^{-1}$ and 107°) at the same time.

The idealized vortex is assumed to move along the best track of CWB, and vortex strength and structure does not change and no terrain encountered during the analysis period. It is clear the two images show similar Vr patterns including the double dipole signature of the concentric eyewall and the zero wind line. However, there exists significant difference (i.e., the lack of local strong Vr signature over the coast for the idealized vortex experiment). The result again suggests coastal terrain does play important role on the formation of CBJ.

[Figure 12a](#) presents time–latitude section of observed Vr at 2.5-km height over the domain shown in [Fig. 5b](#).

The analysis period is from 1403 to 2233 UTC 17 July including the major CBJ period (period III, 1403–2023 UTC). The maximum Vr of each latitude is taken for analysis. [Figure 12b](#) presents Vr from the idealized vortex no-terrain experiment. No signature of CBJ is found in the time–latitude section. By subtracting each other, the difference of Vr between observed value and the vortex-no-terrain is presented in [Fig. 12c](#). The difference of Vr shows a pronounced CBJ signal in northern branch. The magnitude of the differential wind speed can be 10 – $22 m s^{-1}$, a very large number during the CBJ period. It is interesting to note that the differential wind speed over the eyewall period (period IV, 2023–2233 UTC) is relatively small (i.e., most of the difference is smaller than $10 m s^{-1}$). The result suggests the idealized vortex well resemble the circulation structure of Haitang and the inner-core circulation did not change significantly during the analysis period.

5. Conclusions and discussion

The study documents the spatial and temporal evolution of the CBJ formed over the shore of east Taiwan as Typhoon Haitang approached. The formation timing and evolution of CBJ was controlled by different flow regimes associated with Haitang. When outer circulation of Haitang started to touch the coastal terrain, CBJ formed. A persistent prevailing strong northeasterly wind impinging the mountain led to the formation of CBJ parallel to the shore. After the core region of Haitang had reached the coastal terrain, the considerable curvature of the core-region circulation caused the change of CBJ, with its southern branch shifted offshore and cannot be detected well by RCHL Doppler radar.

The CBJ persisted for almost 6 h and was approximately 140 km long and 25 km wide. The northern branch of the CBJ had a stronger wind (maximum wind speed 49 – $52 m s^{-1}$), a greater vertical extent (the core of the CBJ was between 1.0 and 2.5 km in height), and a more persistent jet signal than the southern branch (maximum wind speed 43 – $46 m s^{-1}$ and the core was between 1.0 and 2.0 km). An idealized vortex simulation experiment is conducted to quantify the possible

FIG. 11. (a) Observed V_r (shading) with reflectivity (over 35 dBZ area as indicated by contour) overlaid at 2.5 km altitude at 1933 UTC 17 Jul. (b) V_r of idealized vortex circulation (Fig. C1b) superimposed on mean flow estimated using the GVTD technique (6.4 m s^{-1} and 107°).

terrain effect on CBJ formation. A differential wind speed of $10\text{--}22 \text{ m s}^{-1}$ is estimated and it is about 25%–50% of the CBJ strength.

Earlier studies indicate that jetlike vertical wind profile appears also in the core region of tropical cyclone. Comparing the vertical wind profiles of the two hurricanes studied by Kepert (2006a, b), obvious differences can be found. 1) The CBJ began to appear when center of Haiyang located $\sim 200 \text{ km}$ away from the coastline, namely CBJ occurred far from the core region of Haiyang. This differs from Hurricane Georges and Mitch in which jetlike profiles appeared within 100 km or less from the center. 2) The core of CBJ appeared at 2.5 km in altitude which is higher than those of Georges and Mitch which are at 250–700 m in altitude. 3) The CBJ appeared in the moving front-left (southwest) quadrant of Haiyang and persisted for a long duration ($\sim 6 \text{ h}$).

According to Samsury and Zipser (1995), their results of 787 flight legs of 20 Atlantic basin hurricanes, outside the core region, the azimuthal wind maximum mostly associated with rainbands and with large variability in the location and magnitude. The preferred location of azimuthal wind maximum is in southeast quadrant (30%) than that in southwest quadrant (19%). From above discussion, the result supports that CBJ associated with Haiyang is different from jetlike structures found near the inner-core region or the rainbands of the hurricanes.

Under the southwesterly flow regime during May and June, a barrier jet frequently forms along the northwestern coast of Taiwan (Jou and Deng 1992; Li and Chen 1998). Although both types of barrier jet are

caused by terrain blocking, the strength and stability of upstream flow and the scales (height, width, and length) of the mountain range are different, as are the flow characteristics of these barrier jets. A comparison of these two barrier jets is of interest, and the results are summarized in Table 3.

Because of relatively unstable oncoming flow, higher wind speed, and higher terrain altitude, the CBJ associated with Typhoon Haiyang possess a higher maximum wind speed and higher jet core altitude. It also exhibited a more two-dimensional structure possibly because of the coastal topography in east of Taiwan. The vertical wind shear (including above and below the jet core) of both mountain-parallel jets are similar, with shear magnitudes below the jet core approximately twice those above the core.

Examining the reflectivity images, the rainfall structure of Haiyang changed from quasi-symmetric to highly asymmetric after CBJ formation. What is the role of the CBJ in contributing to Haiyang's inner-core structure change? Several factors, including the storm track, large-scale environment, and topography all can induce asymmetric convection distribution in association with a landfalling tropical cyclone (Powell 1982). The differential surface friction between the land and sea induces a frictional convergence on the right of a landfalling tropical cyclone in the Northern Hemisphere (Dunn and Miller 1960). A stronger convection within the inner core of the tropical cyclone preferentially occurred in the downshear-left quadrant (Corbosiero and Molinari 2002). This asymmetric convection distribution is suggested caused by the tilting of the vortex and its associated thermal adjusted

FIG. 12. Temporal variation of maximum V_r in direction normal to the long axis of the rectangular box marked in Fig. 5b between 1403 UTC 16 Jul and 2233 UTC 17 Jul. (a) Observational maximum V_r (absolute value) at 2.5-km altitude. (b) Maximum V_r of idealized vortex at 2.5 km altitude. (c) Difference between (a) and (b). Abscissa is time, ordinate is distance to the radar, and positive (negative) indicates to the north (south) of RCHL.

vertical circulation (Jones 1995). Tang and Chan (2014) conducted an idealized experiment to simulate the passage of a tropical cyclone over the CMR of Taiwan. They indicated that considerable asymmetric diabatic heating was induced in the southwest area of the TC. Diabatic heating distribution was suggested to be the result of convergence induced by a terrain-altered low-level airflow.

A pronounced east–west rainband formed at the southwest sector before Haitang made landfall.

Asymmetric convection located at the moving left and downshear-right quadrant (vertical wind shear is approximately 3.12 m s^{-1} and directed to southeast). This observation indicates that the topographic effect is not only the essential factor affecting CBJ formation but the formation of asymmetric rainfall distribution before Haitang made landfall. It was suggested by earlier studies that the CBJ and the asymmetric rainfall distributions are caused for a cyclonic looping track before Haitang made landfall

TABLE 3. Comparison between the CBJ associated with Typhoon Haitang and the barrier jet during TAMEX.

Parameters	CBJ (typhoon Haitang)	BJ (TAMEX)
Maximum wind speed (m s^{-1})	~44	~14
Core altitude (km)	~2.5	~1.0
Length (km)	~140	<200
Offshore width (km)	~25	60–120
Location	Parallel to the east coast	Parallel to the NW coast
Vertical wind shear ($\times 10^{-3} \text{ s}^{-1}$)		
Above	~2.2	~3
Below	~6.3	~6
Topography		
Max height (m)	2686	2116
Orientation	N–S (2D)	NE–SW (3D)
Formation mechanism	Blocking of typhoon circulation by the island obstacle	Blocking of the SW flow by the island obstacle
Stability (N) of upstream flow ($\times 10^{-3} \text{ s}^{-1}$)	~10	~10–18

(Jian and Wu 2008; Jian 2011; Huang et al. 2011; Tang and Chan 2014).

The existence of CBJ is highly dependent on the typhoon track and circulation structure as suggested from this study. More case analyses of CBJ associated with landfall typhoon are recommended to further clarify the relation between CBJ and landfall typhoon structural changes.

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APPENDIX A

Characteristics of Operational Doppler Radars RCHL and RCWF

The detailed characteristics of two operational Doppler radars Hua-Lien (RCHL) and Wu-Fen-San (RCWF) are described herein (Chang et al. 2009). RCHL operates in two scan modes with a 1-min interval between them. The first mode is designed for surveillance; a low pulse repetition frequency (PRF) is employed to obtain a 460-km unambiguous range in the lowest two tilts, and only reflectivity data are obtained. The second mode has a scan range of 230 km; it makes nine scans and uses a high PRF to achieve a high unambiguous Doppler velocity. A total of 11 PPIs

TABLE A1. Characteristics of RCHL and RCWF.

Parameter	RCHL	RCWF
Type	METEOR 1000	WSR-88D
Location	121.62°E 23.99°N	121.77°E 25.07°N
Altitude	63 m	766 m
Wavelength	10 cm	10 cm
Peak power	750 kW	750 kW
Frequency	2.7–3.0 GHz	2.7–3.0 GHz
Antenna gain	45.6 dB	45.5 dB
Beamwidth	0.95°	0.95°
Pulse repetition frequency	250–1302 Hz	318–1304 Hz
Unambiguous velocity	$\pm 34.60 \text{ m s}^{-1}$	$\pm 26.55 \text{ m s}^{-1}$ $\pm 30.95 \text{ m s}^{-1}$
Max range	230–460 km	230–460 km
Pulse duration	0.5 μs	1.6 μs
Min detectable signal	–112 dBm	–133 dBm
Range resolution	1 km (reflectivity) 250 m (radial velocity)	1 km (reflectivity) 250 m (radial velocity)
Elevation (°)	0.5, 1.4, 2.4, 3.4, 4.3, 6.0, 9.9, 14.6, 19.5	0.4, 1.4, 2.3, 3.3, 4.2, 6.0, 9.8, 14.5, 19.5

TABLE B1. Sounding information of 466990 at 1200 UTC 17 Jul.

Altitude (m)	T (°C)	T_d (°C)	RH (%)
19	29.8	21.5	61
26	30.1	20.9	58
267	29.5	19.7	56
621	26.1	19.0	65
956	23.1	18.3	74
1011	22.6	18.3	76
1163	22.9	16.1	66
1361	21.8	14.5	63
2352	14.6	11.2	80
3019	11.2	8.1	81

of these two scan modes comprise the radar volume data, and the volume update cycle is 10 min. RCWF takes 6 min to complete a volume scan with nine elevation angles. The observational range of reflectivity (Doppler radial velocity) is 460 km (230 km). The characteristics of RCHL and RCWF are summarized in Table A1. RCWF was upgraded in 2014 to possess polarimetric capability.

APPENDIX B

Calculation of Brunt–Väisälä Frequency

The Brunt–Väisälä frequency in unsaturated (N) and saturated (N_m) air can be expressed as

$$N^2 = \frac{g}{T} \left(\frac{dT}{dz} + \Gamma_d \right), \quad (\text{B1})$$

$$N_m^2 = g \left[\frac{1 + \left(\frac{Lq_s}{RT} \right)}{1 + \epsilon L^2 \left(\frac{q_s}{C_p} \right) RT^2} \times \left(\frac{d \ln \theta}{dz} + \frac{L}{C_p T} \frac{dq_s}{dz} \right) - \frac{dq_w}{dz} \right]. \quad (\text{B2})$$

Equation (B2) is an accurate approximation equation proposed by Durran and Klemp (1982). The parameters in Eqs. (B1) and (B2) are defined as follows: G is the gravitational acceleration, $\sim 9.8 \text{ m s}^{-2}$; T is the sensible temperature (K); Γ_d is the dry adiabatic lapse rate, $\sim 0.00976 \text{ K m}^{-1}$; L is the latent heat of vaporization, $2.5 \times 10^6 \text{ J kg}^{-1}$; R is the ideal gas constant of dry air, $287 \text{ J K}^{-1} \text{ kg}^{-1}$; q_s is the saturation mixing ratio; q_w is the total water mixing ratio; C_p is the specific heat capacity at constant pressure, $1004 \text{ J K}^{-1} \text{ kg}^{-1}$; ϵ ($\text{J K}^{-1} \text{ kg}^{-1}$) is equal to R/R_v , where R is the gas constant of dry air and R_v is the gas constant of water vapor; and θ (K) is the potential temperature of dry air, $\theta = T(P_0/P)^{R/C_p}$.

The sounding information of 466990 at 1200 UTC 17 July (Table B1), which is below the terrain peak height (2686 m) of the target domain, was used to calculate N and N_m for Froude number estimation. T in Eqs. (B1) and (B2) was substituted with the average temperature between

200507171933 UTC RCHL PPI

FIG. C1. (a) GVTD-retrieved circulation within 100 km radius of typhoon center at 1933 UTC 17 Jul. Contour is wind speed, and shading is reflectivity of RCHL. (b) Axisymmetric tangential wind radial profile of a composited vortex (blue line) And corresponding perturbation pressure field (red line) computed using the gradient wind balance equation. The GVTD-retrieved axisymmetric tangential wind component is employed inside 100-km radius, and the idealized Rankine vortex profile is employed outside the 100-km radius.

19 and 2352 m, and q_s is estimated by $0.622[e_s/(P - e_s)]$, where e_s is the saturation vapor pressure, which is estimated by $e_s(T) = e_{s0}\exp[17.27(T - T_0)/(T - 35.86)]$ where $e_{s0} = 6.11$ mb and $T_0 = 273.15$ K. dq_w/dz in Eq. (B2) is approximated by dq_s/dz . The results of estimation of N and N_m are 10.3×10^{-3} and $7.54 \times 10^{-3} \text{ s}^{-1}$, respectively.

APPENDIX C

Calculation of Pressure Gradient Force

The axisymmetric pressure field associated with Haitang is estimated using an idealized vortex and gradient wind balance relation. Figure C1a illustrates the retrieved primary circulation of Haitang at 1933 UTC 17 July using the GVTD technique (Jou et al. 2008). The retrieved circulation includes an axisymmetric tangential and radial wind, the asymmetric wavenumbers 1–3 components of the tangential wind, and the mean flow. Due to the radar observation coverage limitation, only circulation within 100-km radius of the typhoon center can be successfully retrieved by GVTD technique. However, the distances from Haitang's center to SA and HL are ~ 190 and ~ 180 km, respectively. Thus, we extend the retrieved circulation by adopting a Rankine vortex outside 100-km radius. The blue line in Fig. C1b is the tangential wind radial profile, which is composed of the axisymmetric component of the GVTD-retrieved circulation within a 100-km in radius and an idealized Rankine vortex outside 100-km radius. The red line in Fig. C1b is the retrieved perturbation pressure field followed Lee et al. (2000). Based on the CWB's estimation of the sea level pressure at Haitang's center at 1600 UTC 17 July (925 mb) and the perturbation pressure, the axisymmetric pressure field associated with Haitang at SA and HL is 974.6 and 974.1 mb, respectively.

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