Sensitivity of typhoon track to asymmetric latent heating/rainfall induced by Taiwan topography: A numerical study of Typhoon Fanapi (2010)

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1] Typhoon Fanapi (2010) traveled westward across the Central Mountain Range of Taiwan on 19 September and its rainfall shifted from a symmetric to an asymmetric pattern with convection mostly to the south and southeast. Meanwhile, the storm slowed down from 22 to 14 km h–1 for 12 h upon leaving Taiwan, and led to heavy rainfall (>800 mm) and serious flooding over the low-lying southwestern plains. Through simulation and sensitivity tests using the Cloud-Resolving Storm Simulator at 3 km grid size, this study shows that the sudden and temporary speed reduction was caused by the asymmetric latent heating (LH), not the environmental flow. Specifically, over a 9 h period, the model storm moved westward at 16 km h–1 in the control run, but increasingly faster and more toward the northwest when the moisture (and thus the LH effect and its asymmetry) is gradually reduced. Steering flow analysis and estimation using model results suggest an eastward motion vector of about 8 km h–1, consistent with the observation, is produced by the asymmetric LH effect, when the effects from the vertical wind shear and beta-drift are both taken into account. This result is further supported by the diagnosis on storm motion based on potential vorticity tendency. Although important, such feedback to typhoon track from rainfall asymmetry that is induced by the blocking effect of topography have not been reported or studied.


1. Introduction

[2] Located over the western North Pacific, Taiwan is hit by 3.7 typhoons annually on average [e.g., Wu and Kuo, 1999]. When a tropical cyclone (TC) makes landfall or moves close by, the steep mesoscale topography of Taiwan, especially the north-south elongated Central Mountain Range (CMR) peaking at 3952 m (Figure 1), often dictates the rainfall pattern (over windward slopes) and significantly increases the total amount through enhanced uplifting [e.g., Chang et al., 1993; Lee et al., 2006; Cheung et al., 2008; Wang et al., 2013]. More so for those making landfall in general, TCs can thus bring extreme rainfall to Taiwan, as exemplified by past cases such as typhoons Herb in 1996 (1736 mm in 24 h; Wu et al. [2002]), Mindulle in 2004 (1512 mm in 48 h; Lee et al. [2008]; Jou et al. [2010]), Haitang in 2005 (1880 mm in 48 h; Jian and Wu [2008]), and Morakot in 2009 (2855 mm in 96 h; Hendricks et al. [2011]; Wu et al. [2011]; Wang et al. [2012]). Therefore, accurate typhoon track forecasts are vital in Taiwan.

[3] The influences of Taiwan topography on TC tracks have been studied quite extensively in the past, including upstream deflection [Brand and Bleloch, 1974; Chang, 1982; Bender et al., 1985, 1987; Yeh and Elsberry, 1993; Wu, 2001; Lin et al., 2006; Huang et al., 2011], track discontinuity [e.g., Wang, 1980; Lin et al., 2005; Peng et al., 2012], and looping behavior prior to landfall [Jian and Wu, 2008; Yeh et al., 2011]. On the other hand, once becoming asymmetric to TC center, diabatic latent heating (LH) can also affect storm motion through potential vorticity (PV) generation and subsequent advection [e.g., Chan, 1984; Fiorino and Elsberry, 1989; Elsberry, 1995; Wu and Wang, 2001b; Chan et al., 2002]. As reviewed in detail by Elsberry [1995], the effects of condensational heating on TC motion have been addressed mostly through idealized simulation or case diagnosis without the presence of terrain in the past [also Chan et al., 2002]. Recently, a positive feedback from asymmetric rainfall/latent heating to further slow down the motion of the devastating case of Morakot during its post-landfall period in Taiwan, on 8 August 2009, was found by Wang et al. [2012]. Similarly, Hsu et al. [2013] examined the tracks of 84 westbound typhoons...
crossing Taiwan during 1960–2010, and suggested that a significant fraction of the slow-moving TCs might have been affected by the convection phase-locked to the topography upon their departure, such that their moving speeds were further reduced after they moved across the CMR. In individual cases, however, it is still unclear as to the extent and detailed mechanisms by which the asymmetric rainfall and latent heating can affect TC tracks, because case studies like Chan et al. [2002] and Wang et al. [2012] are rare. This issue is not only important for TC research and forecasting in Taiwan or in the western North Pacific, where typhoons can often develop an asymmetric rainfall pattern through interactions with monsoon circulation [e.g., Chien et al., 2008; Chien and Kuo, 2011; Wu et al., 2011], but also for those regions with significant terrain that are under threats of TCs around the world. Thus, through model sensitivity tests and further analysis and diagnosis, here we examine the feedbacks from asymmetric rainfall to the track of Typhoon Fanapi (2010), where the asymmetry was induced by terrain blocking of Taiwan, as will be shown in section 2. Such an asymmetry is different from that induced from monsoon interaction as in Morakot [Chien and Kuo, 2011; Wang et al., 2012] or by terrain enhancement [e.g., Hsu et al., 2013], and to our knowledge, the related phenomena and mechanism in real cases have neither been reported nor studied.

The remainder of this paper is arranged as follows. Section 2 provides a case overview of Typhoon Fanapi, and section 3 describes the Cloud-Resolving Storm Simulator (CReSS) model and experiment design. The model simulation is validated in section 4, and the sensitivity test results are presented in section 5. Further analysis and PV diagnostics using National Centers for Environmental Prediction (NCEP) data and model results are performed and discussed in sections 6 and 7. Finally, section 8 gives the summary and conclusions of this study.

2. Case Overview of Fanapi

As shown in Figure 1, Fanapi (TY1011) became a tropical depression (TD) to the east of Luzon island of the Philippines near 19°N, 129.5°E at 1800 UTC on 14 September 2010. It reached tropical storm and typhoon (TY, \( \geq 32.7 \text{ m s}^{-1} \)) status on 15 and 16 September, respectively, while moving slowly northward in general. Since 1200 UTC 17 September, TY Fanapi travelled nearly due west to first approach and pass through Taiwan, then entered the Taiwan Strait and made landfall again in China on 20 September (also Figure 2). Based on the Central Weather Bureau (CWB) best-track, the TC approached Taiwan at a speed of about 22 km h\(^{-1}\) but slowed down to only 14 km h\(^{-1}\) during 0600–1800 UTC 19 September, before restoring to \( \sim 24 \text{ km h}^{-1} \) afterward (Figure 1, solid curve). Thus, a significant reduction of about 8 km h\(^{-1}\) in its translation speed occurred during Fanapi’s exit from Taiwan, i.e., shortly after the TC moved across the CMR.

At sea-level, the synoptic-scale weather systems surrounding Fanapi during the period of 18–21 September 2010 were relatively weak in strength and did not evolve rapidly, except for a gradual development and expansion of the high pressure system near 29°N, 132°E to the southwest of Japan (Figures 2a–2d). This gradual intensification of the subtropical Pacific High was also visible at
850 hPa from 1200 UTC 18 to 1200 UTC 19 September (Figures 3a and 3d). Further aloft at 500 and 300 hPa, the high pressure center was located over central China (near 32°N, 113°E) with a ridge extending eastward to the north of Fanapi, and even showed a slight weakening during 18–19 September (Figures 3b, 3c, and 3e, 3f). Thus, the synoptic-scale systems near Fanapi did not show any rapid changes that might cause the sudden slow-down of the TC upon its departure from Taiwan, apart from a westward steering flow through the deep layer (Figures 2 and 3). As will be confirmed later in section 6.1, the above-mentioned temporary speed reduction was not due to changes in steering flow, and its reason is the focus of our study.

Figure 4 presents the rainrates derived from the polar-orbiting Tropical Rainfall Measuring Mission satellite observations at 1325 UTC 18 and 1408 UTC 19 September, one before and the other after Fanapi’s landfall in Taiwan. Figure 5 shows reflectivity composites of vertical maximum indicator from radars in Taiwan every 5–7 h from 1800 UTC 18 to 2300 UTC 19 September 2010. It is evident that Fanapi was compact with a clear eye roughly 40 km in diameter, and had a symmetric rainfall pattern before its landfall in Taiwan (Figures 4a and 5a, 5b, also cf. Figures 2a and 2b). However, its precipitation pattern became highly asymmetric when it moved across Taiwan due to the blocking (or rain-shadow) effect of the CMR on the northern half of the TC, and the deep convection was concentrated to the southern and southeastern sectors (Figures 4b, 5c, and 5d). Thus, the rainfall asymmetry in this small and compact typhoon developed due to a reason different from the interactions with southwesterly monsoon as in Morakot [Chien and Kuo, 2011; Wu et al., 2011; Wang et al., 2012] or by terrain enhancement [Hsu et al., 2013], as mentioned in section 1, and its subsequent impact on track has never been studied. The asymmetric rainfall pattern of Fanapi also persisted through 19 September, while a much larger eye (almost 200 km in diameter) reemerged near 1800 UTC after Fanapi had entered the Taiwan Strait (Figures 5e and 5f). Between 18 and 20 September, the daily rainfall over Taiwan was the most on 19 September with two distinct maxima: over the ridge of southern CMR (near 22.7°N) and the southwestern coastal plains (Figures 6a and 6b). The former center had a peak 24 h amount of 1080 mm while the latter (≥ 800 mm) was over the low-lying area and caused Kaoshiung, the second largest city in Taiwan (cf. Figure 6a for location), to be severely flooded. Just south of the track (cf. Figure 1), the heavy rainfall over the plains was clearly linked to the sudden slow-down of the TC upon leaving Taiwan. On the other hand, because the convection during this period was asymmetric and mostly to the left and rear of the storm, the possible impacts from LH on the track are examined below, mainly through numerical modeling and further analysis and diagnosis.

3. Model and Experiment Design

In this study, the Nagoya University CReSS v.2.3 is used [Tsuboki and Sakakibara, 2002, 2007]. This model employs nonhydrostatic and fully compressible equations with terrain-following coordinate. To properly simulate clouds at high resolution, an explicit bulk cold rain scheme based on Lin et al. [1983], Cotton et al. [1986], Murakami [1990], Ikawa and Saito [1991], and Murakami et al. [1994] with a total of six species (vapor, cloud water, cloud ice, rain, snow, and graupel) is adopted without the use of any cumulus parameterization (Table 1). This mixed-phase scheme includes microphysical processes of nucleation (condensation), sublimation, evaporation, deposition,
freezing, melting, falling, conversion, collection, aggregation, and liquid shedding [Tsuboki and Sakakibara, 2002]. Subgrid-scale turbulent mixing is parameterized using 1.5-order closure with turbulent kinetic energy prediction [Tsuboki and Sakakibara, 2007], and planetary boundary layer processes are parameterized following Mellor and Yamada [1974] and Segami et al. [1989]. Surface momentum/energy fluxes and radiation are considered with a substrate model [Kondo, 1976; Louis et al., 1981], but cloud radiation is neglected.

[9] The CReSS model adopts the Arakawa-C staggered (horizontal) and Lorenz (vertical) grid structure without nesting or moving domain. For computational efficiency, a time-splitting scheme [Klemp and Wilhelmson, 1978] is used with filtered leapfrog [Asselin, 1972] and the implicit Crank-Nicolson scheme for the integration at large and small time steps, respectively. The CReSS model has been used to study various types of convective systems including TCs [e.g., Wang and Huang, 2009; Hattori et al., 2010; Wang et al., 2011; Akter and Tsuboki, 2012].

[10] In this study, a single domain at 3 km grid spacing with a dimension of 480 × 450 × 50 is employed for all experiments (Figure 1 and Table 1). At lower boundary, terrain data at a horizontal resolution of 30 s (roughly 1 km) and weekly SST on a 1° × 1° latitude/longitude grid [Reynolds et al., 2002] are provided. Using the NCEP gridded analyses, also at 1° × 1° latitude/longitude resolution, at 26 levels every 6 h as initial and boundary conditions (IC/BCs), a three-day control run (CTL) starting from 0000 UTC 18 September is first performed to verify the model’s capability to reproduce the event realistically. Then, a series of sensitivity tests are carried out to access the impact of asymmetric LH on the track of Fanapi upon leaving Taiwan (starting from 0600 UTC 19 September), through gradually reducing the amounts of water vapor in the IC/BCs to 100%, 75%, 50%, 25%, and 0% of the true values in the NCEP data. These runs are named R01 to R05, respectively, and a similar strategy is also adopted in Wang et al. [2012]. Because the domain size is not
excessively large (1440 x 1350 x 25 km^3, Table 1, cf. Figure 1), the background environments farther away from the TC are not affected. Note also that few deep clouds existed inside the domain except for those associated with Fanapi during its departing period, as shown in Figure 4b (and other cloud imageries, not shown), so the reduction in moisture in R02–R05 will mainly affect only the LH in the TC vicinity (cf. Figure 1). Because the results of R04 and R05 (which has an initially dry atmosphere) are almost identical with no rain anywhere inside the model domain, most of R05 results will not be shown.

Figure 4. Tropical Rainfall Measuring Mission Precipitation Radar (PR) or Thermatic Mapping Imager rainrates (inch h^-1, color, scale at bottom) overlaid on the geostationary Multifunctional Transport Satellite infrared cloud imagery of Typhoon Fanapi at (a) 1325 UTC 18 and (b) 1408 UTC 19 September 2010. The time and type of Multifunctional Transport Satellite cloud imagery closest to Tropical Rainfall Measuring Mission data are labeled at the upper-left corner of each panel (all panels from Naval Research Laboratory).

Figure 5. Radar vertical maximum indicator reflectivity composites (dBZ, scale on right) and 850 hPa horizontal winds (m s^-1) in NCEP analyses at (a) 1800 and (b) 2300 UTC 18, and (c) 0600, (d) 1100, (e) 1800, and (f) 2300 UTC 19, September 2010. For winds, the analyses at 0000 and 1200 UTC are plotted in Figures 5b, 5f, and 5d, and pennants, full barbs, and half barbs represent 50, 10, and 5 m s^-1, respectively. The TY symbol depicts the TC center in the CWB best-track at each time.
4. Model Control Experiment (CTL)

[11] The track of the low-level (850–700 hPa) TC center, determined manually using both wind and pressure fields, as simulated in CTL agrees closely with the CWB best-track, and the storm slows down to about 17 km h\(^{-1}\) over 0900–1500 UTC 19 September (Figure 1 and Table 2). Here as in all later instances, the translation speed and direction are computed.

Table 1. CReSS Model Domain Confi guration and Experiment Design in This Study

<table>
<thead>
<tr>
<th>Experiment Name</th>
<th>CTL</th>
<th>R01 to R05</th>
</tr>
</thead>
<tbody>
<tr>
<td>Projection</td>
<td>Lambert Conformal, center at 120°E, secant at 10°N and 40°N</td>
<td>3 km × 3 km × 200–633 m (500 m)</td>
</tr>
<tr>
<td>Grid spacing (km)</td>
<td>480 × 450 × 50 (1440 km × 1350 km × 25 km)</td>
<td>Real at (1/120°) and weekly mean at 1° resolution</td>
</tr>
<tr>
<td>Dimension and size (x, y, z)</td>
<td>NCEP analyses (1° × 1° and 26 levels, 6 h)</td>
<td></td>
</tr>
<tr>
<td>Topography and SST</td>
<td>0000 UTC 18 September 2010</td>
<td>0600 UTC 19 September 2010</td>
</tr>
<tr>
<td>IC/BCs</td>
<td>72 h</td>
<td>42 h</td>
</tr>
<tr>
<td>Initial time</td>
<td>Integration length</td>
<td>Cloud microphysics</td>
</tr>
<tr>
<td>Planetary boundary layer parameterization</td>
<td>Surface processes</td>
<td>Soil model</td>
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</tbody>
</table>

An asterisk denotes that the vertical grid spacing (\(\Delta z\)) of the CReSS model is stretched (smallest at the bottom), and the averaged spacing is given in the parentheses.

Figure 6. Observed (a) 3 day total rainfall (mm) over 0000 UTC 18–21, and (b) 24 h rainfall (mm) over 0000–2400 UTC 19 September 2010. (c) As in Figure 6b but for model simulated 24 h total rainfall (mm) in CTL. (d–f) As in Figure 6c but for 18 h rainfall (mm) in R01 (Figure 6d), R02 (Figure 6e), and R03 (Figure 6f) over 0600–2400 UTC 19 September 2010. Color scales (shown on right) are the same for all panels. The two largest cities in Taiwan, Taipei and Kaoshiung, are marked by triangles in Figure 6a.
Table 2. Summary of Observed (CWB Best-Track) and Model-Simulated TC Translation Speeds (km h\(^{-1}\)) and Directions (°) From CTL and R01-R04 Experiments Over Different Time Periods in 18–20 September 2010 (Before, During, and After Speed Reduction). All Entries are Computed Using the Locations of TC Center at the Two Endpoints of Each Period

| Observation (CWB Best-Track) | | |
|-------------------------------|-------------------------------|-----------------|-----------------|-----------------|
| Period                        | 9/18 12 Z to 9/19 06 Z (18 h) | 9/19 06–18 Z (12 h) | 9/19 18 Z to 9/20 12 Z (18 h) |
| Speed                        | 21.8                          | 13.9             | 22.4             |
| Direction                    | 260                           | 282              | 279              |

Control experiment

<table>
<thead>
<tr>
<th>Period</th>
<th>9/18 12 Z to 9/19 09 Z (21 h)</th>
<th>9/19 09–15 Z (6 h)</th>
<th>9/19 15 Z to 9/20 12 Z (21 h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Speed</td>
<td>19.4</td>
<td>17.1</td>
<td>20.4</td>
</tr>
<tr>
<td>Direction</td>
<td>266</td>
<td>279</td>
<td>271</td>
</tr>
</tbody>
</table>

Sensitivity tests

<table>
<thead>
<tr>
<th>Expt.</th>
<th>9/19 06–15 Z (9 h)</th>
<th>9/19 15 Z to 9/20 12 Z (21 h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>R01</td>
<td>15.9</td>
<td>28.5</td>
</tr>
<tr>
<td>R02</td>
<td>18.0</td>
<td>27.8</td>
</tr>
<tr>
<td>R03</td>
<td>21.9</td>
<td>21.3</td>
</tr>
<tr>
<td>R04</td>
<td>21.9</td>
<td>23.8</td>
</tr>
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</table>

Figure 7. Comparison of CTL (thick solid, since 0000 UTC 18 September) and R01 (thick gray, since 0600 UTC 19 September) simulated (a) maximum surface (10 m) wind speed (m s\(^{-1}\)) and (b) central MSLP (hPa) with values from the CWB best-track estimates (solid line with open circles) and NCEP analyses (gray dashed).
using the TC center locations at two endpoints. Although not as dramatic as in the CWB best-track, this reduction in translation speed is better captured by the CReSS model than in the NCEP analysis, which is a more moderate 18 km h\(^{-1}\) (Figure 1).

[12] The maximum surface wind and central mean sea-level pressure (MSLP) in CTL simulation are compared with the CWB best-track estimates and NCEP analyzed values in Figure 7. The CWB has estimated a TC intensity considerably stronger than that shown in the NCEP 1\(^\circ\) x 1\(^\circ\) gridded analyses, typically by about 12.5 m s\(^{-1}\) and 20 hPa during 18–19 September. Without using any bogus technique [e.g., Davidson and Puri, 1992; Xiao et al., 2000; Nguyen and Chen, 2011], nonetheless, the TC vortex in the CTL simulation spins up rather rapidly after the initial time (at 0000 UTC 18 September) and the maximum surface wind speed increases to within a few meters per second from the CWB best-track estimates except for two periods: near 0000 UTC and after 1800 UTC 19 September. In both occasions, the model TC in CTL is close to make landfall (in Taiwan and China, respectively; cf. Figure 1) and exhibits an earlier and more rapid weakening (by about 3–6 h) compared to the best-track data (Figure 7a). Thus, the maximum wind speeds in CTL are considered to be in good agreement with the CWB estimates. On the other hand, the central MSLP values in CTL are quite close to the NCEP analyses throughout the simulation period, and only slightly lower (by a few hPa) during Fanapi’s departure period from Taiwan on 19 September (Figure 7b). As a result, the simulated MSLP is typically about 15 hPa too high compared to the CWB best-track on 19 September. Compared to the best-track data, the weaker TC intensity in CTL may partially contribute to the inadequate speed reduction upon the exit from Taiwan (cf. Figure 1). Overall, the model simulations of TC intensity are considered to be reasonable.

[13] The model-simulated rainrates and pressure and horizontal winds at about 1.5 km in CTL simulation at 1330 UTC 18 and 1415 UTC 19 September are shown in Figures 8a and 8b, and can be directly compared with Figures 4a and 4b, respectively. As shown in Figure 8a, the compact and rather symmetrical structure of Fanapi before landfall on 18 September is nicely captured by the CReSS model (cf. Figures 4a). Similarly, when the TC enters the Taiwan Strait and develops an asymmetrical rainfall pattern on 19 September, so does the model vortex in CTL (Figures 4b and 8b). The model column-maximum mixing ratio of total precipitation (Figure 9) also matches well with Figure 5, with the convection shifting from a more symmetric to a highly asymmetric pattern when Fanapi
makes landfall in Taiwan. Note that the rainbands to the south and east of the TC center (near southern Taiwan) are especially well reproduced during Fanapi’s departure (Figures 9c–9e, cf. Figures 5c–5e). As a result, the simulated rainfall distribution for 19 September also compares favorably with the observation (Figures 6b and 6c), with both maxima nicely captured. The rainfall center over southern CMR also reaches 1000 mm in the CTL simulation, and the one near Kaoshiung exceeds 600 mm with a somewhat smaller areal extent (Figure 6c), which may be partially attributed to the inadequate speed reduction of the TC in the model (cf. Figure 1). Over the coastal region near 23°C/120.2°E, there is some overprediction in rainfall in CTL. Nevertheless, the CReSS model reproduces the event realistically in the CTL experiment and its overall performance is quite satisfactory.

5. Model Sensitivity Tests (R01–R05)

As described, experiments R01–R05 are designed to test the influence of LH (including its asymmetry) on the track of Fanapi upon leaving Taiwan, and R01 serves as a benchmark for comparison for this purpose. All these runs start from 0600 UTC 19 September (Table 1) as the center of the storm was near the ridge of the CMR in the NCEP data (Figure 1), and the reason is mainly two fold. First, the impacts of Taiwan’s steep topography on the TCs during the approach and early landfall hours (as reviewed in section 1), different among the tests, can be avoided to focus on only the departing period. Second, the TCs in all runs are at the same location and have identical vortex structure and ambient flow at the initial time (i.e., at the beginning of their departure), so the variations of other influencing factors are kept to a minimum. Also adopted in Wang et al. [2012], this strategy allows us to compare storm tracks in different runs in a meaningful way and more easily, and note that this comparison will be limited to the first 9–12 h (i.e., the departing period). As shown in Figure 10a, the track in R01 also agrees well with NCEP and CWB tracks through at least 0000 UTC 21 September (cf. Figure 1). The maximum surface wind in R01 simulation is close to that in CTL, while the central MSLP is almost identical through 0000 UTC 21 September (Figures 7a and 7b).

As shown in Figure 8c, the asymmetrical rainband structure of Fanapi over the Taiwan Strait during its departure (at 1430 UTC 19 September) is again well captured in the R01 simulation as in CTL (cf. Figures 4b and 8b). Similarly, the rainfall structure after 0600 UTC 19 September as depicted by the column-maximum mixing ratio of total precipitation in R01 (Figures 11a–11d) also agrees well with the observations (Figures 5c–5e). Note that the intense convection just to the east and southeast of the TC center during this period is well captured. Due to this overall success in simulation, the two rainfall maxima on 19 September are also reproduced in R01 with lower amounts, peaking over 600 and 500 mm, respectively (Figure 6d), because the period of accumulation is only 18 h. Compared to CTL, the over-prediction in rainfall near 23°C/120.2°E is no longer seen in R01.

When the TC tracks among R01–R04 are compared (Figure 10a), evident and systematic differences exist especially during the first 9 h (0600–1500 UTC 19 September)
The reduction of moisture content and LH, of course, also leads to a more rapid weakening of the TC vortex in R02 to R05, as shown in Figure 12. In the next section, we further examine the evolution of the deep-layer mean (steering) flow and the detailed effects from LH in our model sensitivity tests, including those associated with a weakened vortex.

6. Further Analysis and Discussion


Using the 1° × 1° NCEP analyses, the mean flow at each layer from 1000 to 100 hPa inside different radii between 300 and 700 km (every 50 km) from the TC center during our case period is computed. It is found that the deep-layer mean flow over 300–700 hPa within a 450 km radius, as depicted in Figure 13, matches well with the TC motion both before and after the speed reduction (near 22 and 23 km h⁻¹, respectively, cf. Figure 1). Moreover, the mean flow remains nearly the same at 0600 and 1200 UTC 19 September, when the actual translation speed (of CWB best-track) reduced by about 8 km h⁻¹ (Figure 13, cf. Table 2). Therefore, the temporary speed reduction of the TC was not caused by changes in the environmental mean (steering) flow, but is hypothesized to be caused by the asymmetrical LH effect here. In R01, the slow-down is nicely reproduced during 0600–1500 UTC 19 September (16 km h⁻¹) with full moisture content, while there is nearly no rainfall and thus LH effect in R03 because the imposed moisture deficit can recover only very slowly (cf. Figure 10b). Hence, the difference of TC motion in R01 from R03 (during 0600–1500 UTC) can be taken as the motion vector from (total) LH effect in our model sensitivity test, which points toward the SSE (168°) at about 11 km h⁻¹, i.e., to the left and slightly rear of the TC motion at the time (Figure 13, cf. Figures 11c and 11f). The observation, however, indicates mainly a reduction in TC translation speed without much deflection in direction (cf. Figures 1 and 13), and we estimate the effects of LH in more detail and address this apparent difference in the next section.

6.2. Estimation of Detailed Effects From LH

As discussed in Wang et al. [2012a], when the moisture contents are lowered in the model (such as in R03), the reduction of total LH (cf. Figure 10b) leads not only to a decrease in the asymmetrical effect but also to a weakening of the vertical coupling mechanism from deep convection. As a result associated with such structural changes of the vortex, the upper- and lower-level TC vortices may become decoupled and drift separately [e.g., Wang and Holland, 1996; Wu and Wang, 2001a, 2001b]. Using NCEP data and the same 450 km radius as in Figure 13, the mean ambient flow at all levels from 1000 to 100 hPa over 0600–1500 UTC 19 September 2010 is plotted in Figure 14, which reveals a significantly stronger southerly flow component toward the lower levels. Thus, it is indeed possible that the northward TC motion in R03 during 0600–1500 UTC (on 19 September) is caused by the decoupling and vertical shear. Over this 9 h period, the TC motion in the NCEP analyses (at 5.1 m s⁻¹ and 276°, cf. Figure 1) closely matches the midlevel flow near 520 hPa, or the deep-layer averaged flow roughly over 700–350 hPa (Figure 14).
In the R01 experiment where the slow-down is largely captured, the time-mean hodograph also using a 450 km radius over 0900–1500 UTC (Figure 15a) differ somewhat from that in Figure 14 especially at midlevels near 3–7 km. This indicates that the ventilation flow [e.g., Chan, 1984; Wu and Wang, 2001b; Chan et al., 2002] experienced by the TC is altered in R01 when the LH effects are adequately simulated (section 5). The TC motion in R01 (at 4.4 m s$^{-1}$ and 274°, cf. Figure 10a and Table 2) follows closely the mean flow at 5.5 km (near 520 hPa also), or the deep-layer mean flow over 3–8 km (at 4.4 m s$^{-1}$ and 272°, Figure 15a). In response to the 50% reduction of moisture and almost no LH effect in the model (cf. Figure 10b), the vertical profile of the mean wind in R03 (Figure 15b) resembles Figure 14 more than that of R01 from the surface to about 7 km. On the other hand, the TC motion vector in this case matches in direction of the ventilation flow at a much lower level, near 4 km (~630 hPa), or the mean flow

![Figure 11](image1.png)

**Figure 11.** As in Figure 9, but for model-simulated column maximum mixing ratio of total precipitating hydrometeors (g kg$^{-1}$) and horizontal winds at 1530 m in R01 every 3 h from (a) 0900 to (d) 1800 UTC (left), and in R03 at (e) 0900 and (f) 1500 UTC (right), 19 September 2010. The low-level TC center is also depicted in each panel.

![Figure 12](image2.png)

**Figure 12.** Same as Figure 7, except for model-simulated MSLP (hPa, solid) and maximum surface (10 m) wind speed (m s$^{-1}$, dashed) from 0600 UTC 19 to 0000 UTC 21 September 2010 in experiments R01 to R05 (colors as labeled). For R03–R05, curves are drawn only until the time when the low-level TC center becomes diminished and beyond identification.

[20] In the R01 experiment where the slow-down is largely captured, the time-mean hodograph also using a 450 km radius over 0900–1500 UTC (Figure 15a) differ somewhat from that in Figure 14 especially at midlevels near 3–7 km. This indicates that the ventilation flow [e.g., Chan, 1984; Wu and Wang, 2001b; Chan et al., 2002] experienced by the TC is altered in R01 when the LH effects are adequately simulated (section 5). The TC motion in R01 (at 4.4 m s$^{-1}$ and 274°, cf. Figure 10a and Table 2) follows closely the mean flow at 5.5 km (near 520 hPa also), or the deep-layer mean flow over 3–8 km (at 4.4 m s$^{-1}$ and 272°, Figure 15a). In response to the 50% reduction of moisture and almost no LH effect in the model (cf. Figure 10b), the vertical profile of the mean wind in R03 (Figure 15b) resembles Figure 14 more than that of R01 from the surface to about 7 km. On the other hand, the TC motion vector in this case matches in direction of the ventilation flow at a much lower level, near 4 km (~630 hPa), or the mean flow
over about 2.5–5 km (at 4.3 m s\(^{-1}\) and 303°, Figure 15b). An examination of the TC center at different levels also confirms that the low-level vortex of Fanapi indeed becomes separated from the upper-level one in R03, but not so in R01 (not shown). Thus, the low-level TC motion in R03 is controlled by the flow over a much shallower layer near 4 km, rather than by the mean flow through the deep troposphere as in R01. The mean wind-shear vector from 630 to 520 hPa (i.e., from 4 to 5.5 km) can serve as an estimate on the difference in steering flow in R01 compared to R03 (associated with the decoupling effect with the presence of vertical wind shear), as depicted in Figure 14 (dotted arrow). It is also noted that radii from 150 to 600 km were also tested for Figure 15, but again the 450 km yielded the best overall agreement between the TC motion and the mean wind.

As shown in Figure 15, the TC in R01 (4.4 m s\(^{-1}\)) moves about 12% faster than the mean flow at 5.5 km (3.9 m s\(^{-1}\)). The storm in R03 (6.1 m s\(^{-1}\)) also moves faster compared to the mean flow at 4 km (4.3 m s\(^{-1}\)), by about 40%. Within the observed range [e.g., Carr and Elsberry, 1990] in both cases, such a westward and poleward propagation of TCs besides the advection by environmental steering flow is largely due to the effect from the advection of planetary vorticity \(f\), i.e., the beta-drift [e.g., Holland, 1983; Chan and Williams, 1987; Fiorino and Elsberry, 1989; Li and Wang, 1994; Wang et al., 1997]. Figure 16 shows the mean wind speed distributions with radius and time, averaged with respect to both azimuth and over the depth of 3–8 km in R01 and 2.5–5 km in R03. In R01, the TC vortex over the steering layer (3–8 km) exhibits only slight weakening with time during 0600–1800 UTC, and the radius of maximum wind (RMW) decreases from about 190 to 150 km (Figure 16a). The RMW of the vortex over a shallower layer (2.5–5 km) in R03, on the other hand, increases from about 180 to 270 km (Figure 16b). Although the mean vortex strength over the steering layer in R03 with nearly no LH is somewhat weaker than that in R01, the larger RMW and a weaker TC can presumably lead to a faster movement toward the west-northwest in response to the advection of \(f\) [e.g., Chan and Williams, 1987]. Thus, we estimate the difference in TC motion caused by the difference in the beta effect in R01 and R03 to be the difference between the departure of TC motion to the mean flow (at 5.5 km in R01 and 4 km in R03, respectively). This results in a relatively small vector of 1.2 m s\(^{-1}\) (or 4.5 km h\(^{-1}\)) at 133°, in R01 relative to R03.

From the above results, the estimates of detailed effects from total LH in R01 as compared to R03 can be summarized in Figure 17 (in km h\(^{-1}\)). With full moisture content and LH, the vortex in R01 (relative to that in R03) moves toward the SE (168°) at 11.3 km h\(^{-1}\) as in Figure 13 (short dashed arrow). Because the vortex is intact and moves with deep-layer mean flow, a southwestward component of 12.2 km h\(^{-1}\) at 224° is resulted in R01, as compared to R03 where a decoupling occurs. Thus, this component (dotted arrow) is labeled as associated with vertical wind shear (VWS, same as in Figure 14). In addition, the motion vector caused by the difference in the advection of planetary vorticity (FADV) (i.e., the beta effect) can be estimated as toward the SE (133°) at 4.5 km h\(^{-1}\) (long dashed arrow, Figure 17). Here both the VWS and FADV are linked to the structural changes of the TC vortex, and the former is
associated with the decoupling and the difference in steering flow while the latter is a more direct effect. Thus, the asymmetric LH effect on storm motion (ALH, solid arrow) can be estimated by $ALH = LH_{C0VWS}/C0FADV$, which yields a vector of 7.8 km h$^{-1}$ toward 87° (Figure 17).

Pointing eastward and to the rear of the TC motion on 19 September, this result agrees well with the speed reduction both in R01 run compared to the deep-layer steering flow and in the observation (cf. Figures 1 and 13).

### 7. Potential Vorticity Diagnostics

[23] In this section, a PV diagnosis following the method of Wu and Wang [2000] is performed and its results are discussed to further shed light on the reason of the sudden change in storm motion (i.e., the temporary slow-down) upon Fanapi’s departure from Taiwan. Adopted by Wu and Wang [2001b], Chan et al. [2002], and Hsu et al. [2013], this approach allows for a direct evaluation of the asymmetric components of LH, and is first reviewed briefly here.

[24] Following Wu and Wang [2000], the simplified forms of Ertel’s PV ($P$) and the asymmetric component (i.e., azimuthal wave-number one) of its inviscid tendency relative to a vortex center in a fixed frame can be expressed, respectively, as

$$P = \frac{\zeta + f}{\rho} \left( \frac{\partial \theta}{\partial z} \right),$$

$$\left( \frac{\partial P}{\partial t} \right)_1 = \Lambda_1 \left[ -\bar{V} \cdot \nabla \theta - \frac{\zeta + f}{\rho} \left( \frac{\partial Q}{\partial z} \right) \right],$$

where $\zeta$, $\rho$, $\theta$, $\bar{V}$, $\nabla h$, and $w$, are relative vorticity, air density, potential temperature, horizontal wind vector, horizontal gradient operator, and vertical velocity, respectively, while the subscript “1” indicates the wave-number one component and $\Lambda_1$ denotes the operator to obtain such a component. The variable $Q$ in equation (2) is diabatic heating (DH) and defined as

$$Q = \frac{d \theta}{dt} + \bar{V} \cdot \nabla \theta + w \frac{\partial \theta}{\partial z}.$$

[25] Thus, equation (2) states that the azimuthal wave-number one component of PV tendency that controls the

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**Figure 15.** As in Figure 14, but for the mean flow (m s$^{-1}$) computed from results in (a) R01 and (b) R03 experiments from the surface to 12 km (also using a radius of 450 km). The numbers indicate height levels (m) at nearby dots (closest to every km up to 10 km plus the one near 12 km) on the curve. The mean flow over 3–8 km is marked by a gray circle in Figure 15a, and that over 2.5–5 km is marked by an open circle in Figure 15b. The TC motion vector in R01 or R03 is also plotted (thick solid arrow) and labeled (speed in m s$^{-1}$ shown in parentheses).

**Figure 16.** Distributions of axisymmetrical horizontal wind speed (m s$^{-1}$) with radius and time, averaged over (a) 3–8 km in R01 and (b) 2.5–5 km in R03 during the first 12 h (0600–1800 UTC 19 September) of integration. The thick dashed lines mark the radius of maximum wind.
movement of the TC vortex (as all other wave numbers are symmetric and thus has no effect on vortex motion) is contributed by three terms: horizontal advection, vertical advection, and DH. Here, only the components involving the vertical gradient are retained in equations (1) and (2) for simplicity, because the other two components are relatively small [Hsu et al., 2013].

\[ \frac{\partial P}{\partial t} = -c_x \frac{\partial P_s}{\partial x} - c_y \frac{\partial P_s}{\partial y}, \]  

where \( P_s \) is the symmetric part of the vortex PV, and \( c_x \) and \( c_y \) are the two components of the TC’s motion vector. For an axisymmetric vortex, \( \frac{\partial P_s}{\partial x} \) and \( \frac{\partial P_s}{\partial y} \) in equation (4) both have zero mean and are uncorrelated, so the scalar speeds can be solved by the least square method as

\[ c_x = \frac{\sum_{i=1}^{N} \left( \frac{\partial P_x}{\partial x} \right)_i \left( \frac{\partial P_s}{\partial t} \right)_i}{\sum_{i=1}^{N} \left( \frac{\partial P_x}{\partial x} \right)_i^2}, \quad c_y = \frac{\sum_{i=1}^{N} \left( \frac{\partial P_s}{\partial y} \right)_i \left( \frac{\partial P_s}{\partial t} \right)_i}{\sum_{i=1}^{N} \left( \frac{\partial P_s}{\partial y} \right)_i^2}; \]  

where \( N \) is the total number of grid points inside a fixed area under consideration, and \( i \) denotes the \( i \)th grid point. Note that equations (4) and (5) can be used to obtain the motion vector from each of the right-hand-side terms in equation (2), if the wave-number one PV tendency \( \frac{\partial P}{\partial t} \) is from that term alone [Wu and Wang, 2000].

[27] Using the method described above, the diagnosis is again performed on R01 and R03 and compared for the 9 h period over 0600–1500 UTC 19 Sep 2010, mainly for two reasons. First, the R01 run captures the temporary slow-down of TY Fanapi better than CTL (cf. Figures 1 and 10a), and second, the results of the PV diagnostics then can be compared with our earlier analysis and estimates in sections 6. Each term in equation (2) is computed and then aggregated (from 3 km) to 15 km resolution to smooth out fine-scale features. Their wave-number one and symmetric
The actual TC motion and mean R03, respectively. The gray (black) cross and dots mark vectors of R01, R03, and their difference (R01 minus R03). The subscripts refer to different terms in equation (2) agree better with TC motions. Note that most of the intense rain-bands in and near the eye-wall during the departing period in R01 are enclosed by this radius (cf. Figure 9). Figure 18 shows the mean PV and wind fields averaged over the departing period and the above depths around the vortex centers in both runs. In R01, a distinct PV maximum (>15 PVU; 1 PVU = 10⁻⁶ K m² s⁻¹ kg⁻¹) appears about 75 km to the south-southeast of the center of Fanapi with a banded structure extending downwind (Figure 18a), and is clearly linked to the PV generation by LH near the eye wall and its subsequent advection (cf. Figures 11a–11c). A secondary PV maximum exists about 160 km to the north-northeast of the eye, and is likely associated with downward transport of PV (and vertical stretching) to the leeside of Taiwan topography [e.g., Hsu et al., 2013]. With almost no LH, the mean PV in R03 is considerably weaker and more smoothly distributed around the inner core, while a maximum in the eye wall is absent (Figure 18b).

The actual TC motion and layer-mean flow, as well as the diagnostic motion vector from the total PV tendency (TT) and diabatic heating (DH) on TC motion (km h⁻¹) over 0600–1500 UTC 19 September 2010 in R01 and R03 experiments. The results are obtained using a radius of 135 km and averaged through the depth of 3–8 (3–6) km for R01 (R03). The subscripts “1”, “3”, and “D” denote motion vectors of R01, R03, and their difference (R01 minus R03), respectively. The gray (black) cross and dots mark the actual TC motion and mean flow in R01 (R03) as in Figure 15 (except for the units).

Hence, the difference in DH, in R01 relative to R03, is 15.0 km h⁻¹ toward 159° (Figure 19). This differential vector, pointing at the left and rear of the TC motion, is from the direct generation of PV [e.g., Wu and Wang, 2001b] and can account for much of the difference in total tendency as well as in actual TC motion between R01 and R03 during the departing period of Fanapi.

As discussed by Wu and Wang [2001a] and Chan et al. [2002], the horizontal PV advection in equation (2) mainly consists of the advection of symmetric PV by the asymmetric flow (i.e., the ventilation flow, including both the steering and beta effects) and the advection of asymmetric PV by the symmetric TC vortex. The ventilation flow is estimated in section 6 for a radius of 450 km (VWS + FADV in Figure 17). The advection of asymmetric PV by the TC vortex includes the subsequent advection of PV generated by DH (cf. Figure 18a), which can be expected to shift the DH effect in a counterclockwise direction, i.e., more toward the rear of the TC motion. Thus, all three right-hand-side terms in equation (2) are linked together through nonlinear dynamics, and the DH effect is not limited only to the DH term even though its contributing motion vector (from PV generation) can be estimated directly through the wave-number one component of PV tendency. Overall, the PV diagnostics here gives results supportive to our earlier estimates of the asymmetric LH effect in section 6, and thus indicates that these estimates are quite reasonable.

Figure 19. The PV diagnostic results on the contributions from the wave-number one components of total tendency (TT) and diabatic heating (DH) on TC motion (km h⁻¹) over 0600–1500 UTC 19 September 2010 in R01 and R03 experiments. The results are obtained using a radius of 135 km and averaged through the depth of 3–8 (3–6) km for R01 (R03). The subscripts “1”, “3”, and “D” denote motion vectors of R01, R03, and their difference (R01 minus R03), respectively. The gray (black) cross and dots mark the actual TC motion and mean flow in R01 (R03) as in Figure 15 (except for the units).

8. Summary and Conclusion

Typhoon Fanapi (TY1011) moved westward and across the island of Taiwan on 19 September 2010 (Figure 1), and brought heavy rainfall peaking over 1000 mm to southern Taiwan. The low-lying coastal area in southwestern Taiwan also received up to 800 mm in total rainfall, causing the city of Kaoshiung to be severely flooded (Figures 6a and 6b). When crossing the CMR, the TC developed an asymmetric rainfall pattern due to terrain blocking (to the left and rear of the storm; Figures 4, 5), and subsequently showed a sudden and temporary slow-down in its translation speed over a 12 h period upon its departure (0600–1800 UTC), from about 22 to only 14 km h⁻¹ (Figure 1). This temporary slow-down was an important factor leading to the coastal flood in Taiwan, and whether it was caused by the asymmetrical latent heating around the storm is the focus of this study.
The Nagoya University CReSS model is employed to study the possible impacts of LH and its asymmetry to the motion of TY Fanapi. Using NCEP analyses (1° × 1° and every 6 h) as IC/BCs (Table 1), the control experiment (CTL) reproduced the event realistically at a grid size of 3 km (Figures 6c, 7, 8a, 8b, and 9). Then, starting from 0600 UTC 19 September, a series of sensitivity tests were carried out using both full (100%) and gradually lowered moisture content (75, 50, 25, and 0% of the true values) in the IC/BCs to investigate on the changes in storm track upon leaving Taiwan. Again, the slower moving speed (~16 km h⁻¹) toward the west (at 274°) over a 9 h period (0600–1500 UTC) is captured in the experiment using full moisture content (named R01, Figures 10 and 11a–11d). On the other hand, the TC moves faster and more northward with reduced moisture content, until at about 22 km h⁻¹ toward the WNW (at 304°) when the moisture is lowered to 50% or less (R03–R05, Table 2, Figures 8d, 10, 11e, and 11f). This difference in TC motion vector among model experiments is caused by the total LH effect, rather than just the asymmetric LH effect. Further analysis on steering flow (Figure 13) and model results suggest that when the effects from vertical wind shear and beta-drift are both taken into account (Figures 14 and 15), the asymmetric LH effect can be estimated to be about 8 km h⁻¹ toward the east (at 87°, Figure 17), in good agreement with the observation and the model control runs (CTL and R01). The diagnosis on storm motion based on PV tendency equation [Wu and Wang, 2000] also yields supportive results, and indicates that the TC slows down when strong diabatic heating occurs at its left-rear side during the departing period.

Thus, it can be concluded that TY Fanapi (2010) slowed down upon leaving Taiwan due to the asymmetric rainfall distribution and LH effect, which developed as the storm moved across the CMR. This track change allowed the rainfall over the coastal plains in southwestern Taiwan to accumulate longer and become more concentrated, thus reaching a higher total amount. Therefore, our results have important implications in the forecasts of TC and the associated hazards in Taiwan, as well as in other regions with significant terrain that are also under threats of TCs. To our knowledge, such feedback from rainfall asymmetry induced by mesoscale topography of Taiwan (or any other region) to typhoon track, as mentioned, has neither been reported nor studied in the literature.

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