NOTES AND CORRESPONDENCE

The Role of Diurnal Solenoidal Circulation on Propagating Rainfall Episodes near the Eastern Tibetan Plateau

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ABSTRACT

In this case study, numerical simulations using the Weather Research and Forecasting model (WRF) are performed for 3–4 May 2002, in which two propagating rain episodes occurred in successive days with close ties to the terrain of the Tibetan Plateau (TP) in East Asia. Through sensitivity tests, it is found that the eastern TP not only facilitated convective development in the afternoon but that the solenoidal circulation between this region and its leeside lowlands (near the Sichuan basin) also contributed to the longevity and farther downstream propagation of the episodes under prevailing westerly winds. Reversed every 12 h, the thermally driven circulation induced ascending motion near the eastern edge of the TP during daytime but over the leeside at night. The episode propagation in this case, as often observed, was in phase with the ascent, from eastern TP in the afternoon to the lee at night, indicating both enhancing and modulating effects of the solenoidal circulation.

1. Introduction

The coherent behavior of warm-season precipitation (or cloud) “episodes,” defined as clusters of rain-producing systems in Hovmöller (longitude–time) space, to propagate to the lee of major mountains has been found previously in both North America (Carbone et al. 2002; Ahijevych et al. 2004) and East Asia (Wang et al. 2004, 2005). Because of the forcing from diurnal heating, episodes often develop over the terrain in local afternoon and then propagate eastward across the leeside overnight (see, e.g., Fig. 10 in Wang et al. 2005). This intimate tie of episodes in their phase to the diurnal cycle of certain mountains at a fixed location, together with the longevity of some events, suggests a potential to improve warm-season rainfall forecasts (Carbone et al. 2002), for which the present skill is relatively low (e.g., Olson et al. 1995; Fritsch and Carbone 2004). For East Asia, the major terrain is the Tibetan Plateau (TP), where leeside nocturnal convection near the Sichuan basin (SB) and surrounding peripheries, such as the southern slopes of the Himalayas, has been previously noted (e.g., Asai et al. 1998; Kurosaki and Kimura 2002; Barros et al. 2004). Although Wang et al. (2004, 2005) employed only satellite infrared (IR) brightness temperature $T_B$ data, the eastward shift in summer rainfall peaking time along the Yangtze River valley (YRV) near 30°N—from...
midnight at the upper reaches (the SB area) to the morning at the middle reaches and then to the afternoon at the lower reaches—is confirmed using rain gauge data (Yu et al. 2007) and Tropical Rainfall Measuring Mission (TRMM) 3B42 merged satellite observations (e.g., Johnson 2010). These studies, however, all deal with long-term, averaged characteristics of rainfall or convection, and the behavior of individual episodes to the lee of the TP has not been investigated. In this work, we study a two-day period during 3–4 May 2002 when repeated rain episodes occurred in succession. Although the month of May is premonsoon, thermally driven deep convection is already quite active over the TP (Fujinami and Yasunari 2001; Toumi and Qie 2004). The downstream propagation of episodes is also more pronounced in May, before upper-level westerly winds shift to the north around mid-June (e.g., Chen 1993), and events with scales greater than 900 km (in zonal span) occur roughly once per day, on average (Wang et al. 2005). In particular, the role played by the solenoidal circulation near the eastern periphery of the TP, not only to the development but also to the further propagation of rainfall systems, is examined through a modeling approach.

2. Episode propagation and its tie to the TP

Figure 1 presents the topography and IR $T_B$ cloud imageries during 3–4 May 2002 at 12-h intervals. At 0400 LST (at 105°E meridian) 3 May, relatively few clouds existed over the eastern TP near 97°–101°E (Fig. 1b), where cloud cluster C1 developed by 1600 LST (Fig. 1c). At 0400 LST 4 May, C1 traveled to about 104°E and weakened, leaving the area near 100°E with clearer skies, and cluster C2 moved to about 95°E (Fig. 1d). Again, C2 migrated to 100°E at 1600 LST and past 105°E 12 h later at 0400 LST 5 May as C1 moved farther downstream (Figs. 1e,f). With a distance of 800–1000 km apart, both clusters were enhanced when moving through 100°E over the eastern TP in the afternoon and then propagated.
across the lee (near 105°E) overnight (Fig. 1), consistent with the preferred phase and speed of propagation described in section 1. In the Hovmöller plots of IR hourly $T_B$ (Fig. 2a) and TRMM 3B42-derived rain rate (Huffman et al. 2007) at 0.25° latitude–longitude and 3-h resolution (Fig. 2b), this behavior with close ties to the TP is nicely shown (cf. Fig. 2c). Averaged over 30°–40°N (cf. Fig. 1b for location), both $T_B$ and rain data depict C1 and C2 clearly as two episodes oriented mainly from north to south (cf. Fig. 1), with a zonal phase speed of about 12.2 m s$^{-1}$ (Fig. 2). Thus, this period is chosen for a detailed numerical study on the role of the diurnal forcing on episode propagation.

3. WRF and experiment design

A nonhydrostatic version (version 2.2) of the Weather Research and Forecasting model (WRF; Skamarock et al. 2005) is used to simulate the present case. A total of three domains, including one coarse grid and two nested grids, at horizontal grid sizes of 45, 15, and 5 km (named D1, D2, and D3, respectively) are used with 31 sigma ($\sigma$) levels in the vertical and a model top of 50 hPa (Fig. 3; Table 1). It is noted that at least eight levels are placed in the lowest kilometer (through $\sigma = 0.89$), and the fine mesh of D3, at 4930 km $\times$ 3300 km, is large enough to cover from the western TP to 120°E. Two-way feedbacks are allowed between all domains, and a series of tests were first performed to determine the combination of physical schemes that produced the best result in cloud distribution. As listed in Table 1, these schemes include the Purdue–Lin six-class microphysics (Lin et al. 1983; Chen and Sun 2002) in all domains, Grell (1993) cumulus parameterization (CP) in D1 and D2, Yonsei University (YSU) planetary boundary layer (PBL) turbulence scheme (Hong and Pan 1996; Hong et al. 2006), and the Dudhia (1989) shortwave and Rapid Radiative Transfer Model (RRTM) longwave (Mlawer et al. 1997) schemes for surface radiation. The initial and boundary conditions are from the European Centre for Medium-Range Weather Forecasts (ECMWF) 6-hourly gridded analyses at 1.125° latitude–longitude resolution and 13 levels (at 1000, 925, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, and 50 hPa). The integration starts from 1900 LST (1200 UTC) 2 May 2002, with a forecast length of 72 h and an output frequency every 1 h.

4. Results of CT experiment

The results of the WRF control (CT) run are first compared with the observations using Hovmöller plots (averaged over 30°–40°N) in Figs. 4a,b, while the results of other runs in the sensitivity test will be described later in section 5. It is clear that the model reproduced the two episodes well in the CT run, and that their zonal phase and traveling speed (in relation to the terrain) are in good overall agreement with the observations (cf. Fig. 2) in both the derived cloud-top temperature (same as $T_B$; Fig. 4a) and rainfall (Fig. 4b). East of about 103°E, the phase of cluster C2 appears to be somewhat earlier, and...
its propagation speed appears to be slightly faster than those suggested by the TRMM data. The model also seems to produce more rain east of 105°E (Figs. 2b, 4b). Over the barren eastern TP (west of about 99°E), on the other hand, the convection apparently results in less rain because of the limited moisture supply in spring, in agreement with Tanaka et al. (2001) and Gao et al. (2003).

The cloud clusters produced in the WRF CT run are highly realistic, and their distribution and evolution over the area of interest closely resemble those of the observation over the case period (not shown). Figure 5 shows the D1 rainfall distribution in the model near the eastern edge of the TP at 6-h intervals from 1900 LST 3 May to 1300 LST 4 May. The downstream propagation of C1 across the SB area overnight and the enhancement of C2 near 100°E in the afternoon are both well captured, and in overall agreement with Fig. 1. Here, it is perhaps worthwhile to note that because of the two-way feedbacks between model domains, the coarse grid (D1) outputs are essentially the spatial averages of finer grids (D2 and D3). Since the D1 result contains fewer small, cloud-scale features and shows the overall patterns better, it is chosen here for presentation.

The D1 results of vertical velocity $w$, divergence, and horizontal winds at the level of $\sigma = 0.85$ from the CT run, at the same times as in Fig. 5, are shown in Fig. 6. At 1900 LST 3 May, rising motion at this level exists over most parts in the eastern TP (Fig. 6a), consistent with rainfall (cf. Fig. 5a). Six hours later at 0100 LST 4 May (Fig. 6b), the area of ascent appears farther east to the leeside near 105°E (about 28°–35°N), while sinking motion prevails over the eastern TP. At 0700 LST (Fig. 6c), the region of positive $w$ associated with C1 continues to move eastward together with rainfall (cf. Fig. 5). Finally, at 1300 LST (Fig. 6d), ascent over the eastern plateau (EP) has reappeared, especially along the steep slopes at 102°–104°E, and weak descent exists slightly farther east over the leeside (near 27°–32°N, 104°–106°E). Thus, the patterns of rising/sinking motion near the eastern edge of the TP are in good agreement with rainfall and are nearly reversed between day and night. Furthermore, the $w$ patterns lie generally parallel to the terrain and correspond well with low-level convergence/divergence and local winds over the slopes—that is, with downslope flow during the night (near 28°–32°N, 103°–104°E; Figs. 6b,c) and upslope flow during the day (Figs. 6a,d).

Table 1. (top) The basic domain configuration (nested grids) and (bottom) physical parameterization scheme options used in the WRF CT run in this study.

<table>
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<tr>
<th>Grid configuration</th>
<th>D1</th>
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<th>D3</th>
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<td>5</td>
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<td>Grid dimension (x, y)</td>
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<tr>
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<tr>
<td></td>
<td>0.980</td>
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<td></td>
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<td>Surface longwave radiation</td>
<td>RRTM</td>
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FIG. 3. The three WRF domains (D1, D2, and D3, at horizontal grid size of 45, 15, and 5 km, respectively) used in this study and the topography (m, gray shades, scale at bottom). The 3-km contour line is plotted in white.
Figure 7 shows the anomalies of potential temperature $\theta'$ and horizontal winds $u'/v'$, as well as the divergence at $\sigma = 0.94$ from the CT run, also at 6-h intervals. Here, the anomaly is defined as the deviation from the 2-day mean of 3–4 May 2002 (from 0000 UTC 3 May to 0000 UTC 5 May) and is computed independently at every grid point. When Figs. 5–7 are compared, it is obvious that the reversal of $w$, convergence, and wind patterns between day and night near the eastern TP is tied to the thermally driven diurnal cycle, with a daily range of about 6–9 K over the eastern plateau (Fig. 7). Such a diurnal cycle over the TP has been shown to be quite pronounced in late spring and early summer (e.g., Yang et al. 2006). Over 35°–37°N, 107°–110°E, $\theta'$ values drop considerably from Figs. 7a–d (by about 10 K), most likely linked to the passage of C1 (cf. Figs. 1d,e).

The D1 vertical cross sections along 32°N (line AA' in Fig. 6b) at two selected times—0100 and 1300 LST 4 May—of $w$, divergence, and wind vectors along the section plane are shown in Fig. 8. At night, again, low-level convergence and $w > 0$ exist over leeside lowlands (LLs) with downslope winds (Fig. 8a), while upslope flow appears (reaching 15 cm s$^{-1}$ near 103.5°E) during the day with broad descent and divergence to the lee (Fig. 8b). In both instances, at the leading edge of the downslope/upslope winds (i.e., along the mountain/valley breeze front at local scale), upward motion is enhanced. The distribution of potential temperature $\theta$ and equivalent potential temperature $\theta^e$ indicates that at 0100 LST (Fig. 8c), it is colder near the surface of the slopes than the leeside at a similar height (both $\theta$ and $\theta^e$), while the low levels (at about 4–5 km) over the eastern TP are stable and dominated by sinking motion. Over the leeside lowlands where deep convection exists (cf. Fig. 8a), there is potential/convective instability $(\partial \theta^e/\partial z < 0)$ at about 2–4 km. At 1300 LST, on the contrary, a deep
mixing layer (~2 km thick) develops over the eastern TP, with dry neutral and conditionally/convectively unstable conditions and significantly higher $u$ and $u_e$ values than the lee throughout about 3–6 km. Thus, a clear reversal in the solenoidal circulation over the eastern TP and its leeside exists between day and night in Figs. 6, 8a,b, in response to the diurnal thermal patterns shown in Figs. 7, 8c,d. A similar reversal in this circulation and low-level flow over the sloping terrain between 0100 and 1300 LST 4 May 2002 is also depicted in ECMWF 1.125$^\circ$ analyses along 31$^\circ$N (Fig. 9), although the circulation appears considerably broader and weaker because of the coarse resolution.

To examine the detailed evolution of the solenoidal circulation and its relation to convection and rainfall distribution at the full temporal resolution of the CT run, three regions are selected to represent the EP, LLs, and the leeside slopes (LSs) in between (Fig. 10), and the areal mean values of various parameters are computed from D3 outputs (and from ECMWF and TRMM data). As shown in Fig. 11, the TRMM hourly rainfall exhibits a pronounced diurnal cycle over both regions of the EP and LLs, with the most rain falling in the afternoon in the EP and overnight in the LLs. The diurnal variation is well captured in the CT run in its phase, while the nighttime rainfall is apparently overpredicted, as discussed earlier (cf. Fig. 4b). As expected, the rainfall cycle in the model is in phase with the (averaged) vertical motion at lower ($\sigma = 0.85$) and middle levels ($\sigma = 0.7$), with the strongest ascent near 1500 LST and descent near 0200 LST over the EP (Fig. 11b). Over the leeside lowlands, the time series of $w$ also shows a clear diurnal cycle with comparable amplitude, but its phase is shifted by 12 h (i.e., 180$^\circ$) from that in the EP, consistent with Figs. 6–8. Since at the same $\sigma$ level, the actual height (and $\theta$) is quite different in the EP and LLs, in Fig. 11c we present the difference in potential temperature anomaly $\Delta \theta'$ using the $\theta'$ value (deviation from the 2-day mean) of the EP minus that of the LLs. The largest $\Delta \theta'$ at $\sigma = 0.99$ (roughly 3–3.5 K) occurs near 1200 LST.
Fig. 11c) and about 2–3 h earlier than the peak in $w$ over the EP (cf. Fig. 11b). Since about dusk, the value of $D_u^{9}$ becomes negative until about 0800 LST, consistent with the low-level sinking motion over the EP (cf. Figs. 8c, 11b). At levels farther up from the surface, the $D_u^{9}$ cycle tends to decrease in its amplitude and eventually loses the diurnal characteristics at the level of $s = 0.7$, also in agreement with the depth of the mixing layer seen in Fig. 8d.

In Fig. 12, clear diurnal cycles in convergence/divergence within (or near the top of) the PBL ($\sigma = 0.92$) over both regions of the EP and the LLs as well as in the near-surface east–west ($u$) wind component ($\sigma = 0.97$) over the LSs, consistent with the solenoidal circulation (and Fig. 11), can be found in the CT run. Again, over the EP, low-level convergence (peaking at about $-3 \times 10^{-5}$ s$^{-1}$) appears during the day and divergence (peaking at roughly $1 \times 10^{-5}$ s$^{-1}$) exists during the night, and this pattern is reversed over the LLs (Fig. 12a). For the near-surface $u$ wind, the daytime upslope flow and nighttime downslope flow over the slopes are depicted nicely in the CT run and can also be seen in the ECMWF 6-h analysis of surface wind (Fig. 12b). The results of other experiments are also shown in Fig. 12 and will be discussed in section 5.

On the basis of these results, the development of the solenoidal circulation between the eastern TP and the
leeside lowlands during our 2-day case period is evident. Further, it can be hypothesized that such a circulation plays a role in episode propagation in the following manner. During daytime, upward motion reinforces and upslope flow also helps organize the convection near the eastern edge of the TP (Figs. 6–9, 11, 12). When rainfall systems travel downstream, presumably by the mechanism of self-propagation in a sheared environment (e.g., Rotunno et al. 1988), at a similar speed as seen here and also observed by Sato et al. (2007), across the lowlands at night, they remain in phase with the convergence/ascent as the circulation reverses (Figs. 6–9, 11, 12). Hence, the development and reversal of solenoidal circulation can contribute to the maintenance of episodes and thus their further propagation. In the next section, the earlier-mentioned hypothesis is tested through additional sensitivity experiments.

5. Sensitivity tests and discussion

Since the solenoidal circulation in the CT run can interact with and be influenced by moist deep convection (i.e., the rainfall systems themselves), the hypothesis in section 4 needs to be examined and solidified further.
Consequently, three sensitivity tests are carried out to assess the effects of surface heating, moist convection, and topography in our case. The first is a no-radiation (NR) run, where both surface shortwave and longwave radiation are turned off (so the net radiation is also zero). This experiment is designed to remove the radiative effects associated with the TP but retain all other effects. The second test is a no-microphysics (NM) run, in which neither any cloud microphysics nor CP scheme is used. In other words, phase change of water and the associated latent heating/cooling effects are removed from the atmosphere, and the development of solenoidal circulation, if any, would be excluded from the influence of moist convection. The third is a reduced-terrain (RT) run, in which the portion of the topography exceeding 2 km is lowered by half while everything else remains the same. For instance, a terrain height of 5 km (over 2 km by 3 km) would be lowered by 1.5 km (half of 3 km) and reset to 3.5 km. This experiment is to test the effects of a reduced topography while not only the thermal effects but also the dynamical effects of the TP are altered.

As shown in the Hovmöller plots, without surface radiation, the NR run (Figs. 4c,d) produces slightly warmer (i.e., lower) cloud tops but considerably less rain for both episodes—compared to the CT run—especially during...
the early stages near the TP (west of about 104°E). Nonetheless, the phase of the rainfall systems in general remains quite similar and is only shifted early by about 2–3 h, while such a shift is more evident for C2 (cf. Figs. 4a,b). In the AA′ cross section from the NR run (Figs. 13a,b), both the downslope wind and leeside convection weaken significantly at night, while the daytime upslope wind disappears, compared to the CT run (cf. Figs. 8a,b).

Over larger areas, when compared with the CT run, the time series of the near-surface $u$ wind indicates that the upslope wind (over the region LSs) is significantly weakened in the NR run when the surface radiation is turned off (Fig. 12b). A cycle in $u$ wind at the diurnal time scale but smaller amplitude, however, still exists and is in agreement with the zonal phase speed of the rainfall systems (cf. Figs. 4c,d, 13a,b). The vertical motion at $\sigma = 0.7$ in the NR run also exhibits diurnal variation (Fig. 14), but the daytime ascent over the EP is much reduced and peaks at least 3 h early, consistent with the forward shift in the phase of the episodes. The nighttime ascent over the LLs associated with C1 also weakens considerably but only slightly for C2 (Fig. 14). In the NR run, the curve of $\Delta \theta'$ at $\sigma = 0.99$ does not exhibit diurnal characteristics and is similar (i.e., in phase) to that at $\sigma = 0.7$ (Fig. 15a). This result suggests that the circulation in the NR run is not forced by the solenoidal effect, as expected. Since both the upward motion and rainfall associated with the cloud clusters reduced and their phase also shifted in the NR run (Figs. 4, 14), it can be confirmed that the solenoidal circulation both contributed and modulated the propagation of rainfall systems in the present case. Here, the eastern TP also acts as a heat source to facilitate afternoon convection, and this result is in agreement with Luo and Yanai (1983), Taniguchi and Koike (2007), and Wu et al. (2007).

In the NM run, as both cloud microphysics and CP scheme are not in use, there is neither cloud nor rainfall.

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**Fig. 9.** East–west vertical cross section along 31°N (from 99° to 107°E) of omega velocity ($\omega$, dPa s$^{-1}$, gray shades), divergence ($10^{-5}$ s$^{-1}$, contour), and wind vectors (m s$^{-1}$, reference vectors plotted) normal to the section plane from the ECMWF 1.125° analyses at (a) 0100 and (b) 1300 LST 4 May 2002. For negative $\omega$ (rising motion), white contours are added between shades. Contours for divergence are drawn every $0.2 \times 10^{-5}$ s$^{-1}$, with solid (dashed) lines for positive (negative) values.

**Fig. 10.** The three regions of EP, LLs, and LSs used for areal mean calculation. Both the size of EP and LLs are 8° × 4° (latitude × longitude), and the size of LSs is 5° × 2°.
In the AA' cross section, it is evident that a circulation pattern similar to that in the CT run, with ascent near the eastern edge of the TP during daytime and over the leeside at night, is still reproduced even when moist convection is eliminated from the model (Figs. 13c,d; cf. Figs. 8a,b). The near-surface upslope wind $(\mathbf{u}_0, s = 0.97)$ at day remains pronounced (peaking near 24 m s$^{-1}$) since about 0900 LST and reverses to downslope $(\mathbf{u}_0)$ during nighttime, with a variation and amplitude comparable to that in the ECMWF analysis (cf. Fig. 12b). Both low-level convergence/divergence (Fig. 12a) and midlevel vertical motion (Fig. 14) still exhibit a strong diurnal signal similar to that in the CT run, except that the nighttime ascent associated with C1 over the LLs is reduced more seriously (but a relative maximum still exists). Such a reduction in $w$ is quite reasonable and expected, since moist convection is not allowed. In Fig. 15a, the curves of $\Delta \theta'$ at low levels agrees well with the evolution of the $u$ wind component (over the LSs; Fig. 12b), with the strongest upslope wind in late afternoon (near 1600 LST) and downslope wind overnight. Without moist convection, the strength of the nighttime downslope wind is reduced but the effects on daytime upslope wind appear limited (Fig. 15b). Above the PBL at $\sigma = 0.7$, the time deviation of $\mathbf{u}'$ instead of $\mathbf{u}$ is also examined in Fig. 15b, so that the background mean westerly flow can be removed (cf. Figs. 8, 13c,d). Here, $\mathbf{u}'$ about 2.5 km aloft generally shows an out-of-phase (and slightly delayed) relationship to surface $\mathbf{u}$ over the LSs in both CT and NM runs, and it indicates the development of the return flow (i.e., the upper branch of the solenoidal circulation). A similar signature of return flow cannot be found in the NR run without the surface radiation (not shown). From the NM run results, it is therefore clear that the diurnal circulation indeed develops during our case period, and that the associated vertical motion can contribute to the rainfall episodes, as hypothesized in section 4.

When the terrain elevation is reduced in the RT run, the overall rainfall from C1 is altered in the Hovmöller plot, while the phase of C2 and the associated rainfall is much different (Figs. 4e,f). Consistent with Fig. 4, in the vertical cross section (Figs. 13e,f), the low-level ascent/convergence near the eastern edge of TP during daytime and those over lowlands during nighttime both decrease, and the leeside convection at night (at 0100 LST 4 May,
FIG. 13. As in Figs. 8a,b, but from D1 results of (a),(b) NR, (c),(d) NM, and (e),(f) RT runs.
associated with C1) also appears farther east. The faster propagation speed is apparently linked to a reduced blocking effect of the plateau and stronger prevailing flow in the RT run. The stronger westerly wind aloft can be seen in Figs. 13e,f, and a similar condition at the surface is also depicted in Fig. 12b. In Table 2, a brief summary between the areal mean \( w \) (at \( \sigma = 0.7 \)) and rain rate during daytime (1200–1800 LST) in the EP and nighttime (0000–0600 LST) in the LLs is given. Compared to the CT run, the values of mean ascent in the NM run decrease only by roughly 13% (over the EP) and 1/3 (over the LLs). On the other hand, both \( w \) and rain rate are significantly reduced in the NR and RT runs, by at least 44% and up to 87% (Table 2). In convectively unstable environments (cf. Figs. 8c,d), there is little doubt that the stronger background ascent from the solenoids in the CT run—when in phase with the rainfall systems—can lead to stronger convection and more rain, as demonstrated here. Thus, from the three additional runs, especially the NR and NM runs, the contributing and modulating role of the solenoidal circulation to rainfall systems can be confirmed. These results are in general agreement with the recent climatological study of warm-season rainfall in the United States by Carbone and Tuttle (2008). Of course, whether episodes propagate out from the TP or not still depends mainly on other crucial factors, including the presence of steering flow aloft (i.e., vertical wind shear) and the forcing from synoptic-scale waves (e.g., Tao and Ding 1981; Wang et al. 1993). Judged from 500-hPa geopotential height patterns (Fig. 1), the cloud clusters in our case were generally associated with a short-wave trough, and synoptic factors seemed quite favorable throughout the 2-day period. Nevertheless, we focus only on the role played by the terrain-lowland solenoid in this case as an illustrative example and isolate its effects through sensitivity tests. Obviously, more studies are needed to generalize our results and to evaluate the detailed effects of the solenoidal circulation on the behavior of rainfall systems near the eastern TP.

6. Conclusions

In this work, we study a 2-day period in which consecutive propagating rain episodes with close ties to the diurnal cycle of the TP occurred through WRF simulations. It is found that the eastern TP not only acts as

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<th>CT</th>
<th>NR</th>
<th>NM</th>
<th>RT</th>
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<td>0.93</td>
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<td>0.53</td>
<td>0.31</td>
<td>—</td>
<td>0.29</td>
</tr>
</tbody>
</table>
a heat source for convection, but the diurnal solenoidal circulation near its eastern slopes also contributes to the longevity and propagation of episodes. During the day, upslope winds prevail to favor convection near the eastern edge of TP, while downslope winds exist at night with rising motion over leeside lowlands. Under suitable synoptic conditions, the propagation of episodes (at about 12.2 m s⁻¹) in the present case was in phase with the ascent associated with the terrain-lowland solenoid at both day and night—namely, episodes over the eastern TP in the afternoon traveled to lowlands at night. Thus, the solenoidal circulation near the eastern TP contributes to the maintenance and downstream propagation of episodes. This role, when better understood, can help improve the warm-season rainfall prediction in East Asia.

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