Diurnal Cycle of a Heavy Rainfall Corridor over East Asia

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ABSTRACT
Moist convection occurred repeatedly in the midnight-to-morning hours of 11–16 June 1998 and yielded excessive rainfall in a narrow latitudinal corridor over East Asia, causing severe flood. Numerical experiments and composite analyses of a 5-day period are performed to examine the mechanisms governing nocturnal convection. Both simulations and observations show that a train of MCSs concurrently developed along a quasi-stationary mei-yu front and coincided with the impact of a monsoon surge on a frontogenetic zone at night. This process was regulated primarily by a nocturnal low-level jet (NLLJ) in the southwesterly monsoon that formed over southern China and extended to central China. In particular, the NLLJ acted as a mechanism of moisture transport over the plains. At its northern terminus, the NLLJ led to a zonal band of elevated conditionally unstable air where strong low-level ascent overcame small convective inhibition, triggering new convection in three preferred plains. An analysis of convective instability shows that the low-tropospheric intrusion of moist monsoon air generated CAPE of $\sim 1000 \text{ J kg}^{-1}$ prior to convection initiation, whereas free-atmospheric forcing was much weaker. The NLLJ-related horizontal advection accounted for most of the instability precondition at $100–175 \text{ J kg}^{-1} \text{ h}^{-1}$. At the convective stage, instability generation by the upward transport of moisture increased to $\sim 100 \text{ J kg}^{-1} \text{ h}^{-1}$, suggesting that ascending inflow caused feedback in convection growth. The convection dissipated in late morning with decaying NLLJ and moisture at elevated layers. It is concluded that the diurnally varying summer monsoon acted as an effective discharge of available moist energy from southern to central China, generating the morning-peak heavy rainfall corridor.

1. Introduction
Torrential rainfall is one of the most devastating natural disasters over East Asia during the mei-yu season, which typically lasts from mid-June to mid-July (Fig. 1a; Ding 1992). Frequent torrential rains may occur along a quasi-stationary front during a period of several days or even 1–2 weeks (Ninomiya 2000; Qian et al. 2004; Sun and Zhang 2012, hereafter SZ12). The successive rainfall events are usually confined to a narrow latitudinal corridor ($\sim 300 \text{ km}$ wide but up to $\sim 1000 \text{ km}$ long) with little north–south drifts of their centroids ($\sim 200 \text{ km day}^{-1}$), similar to those in the central U.S. plains (Tuttle and Davis 2006; Trier et al. 2014). Cumulative rainfall can be substantial within the corridor, in a strong contrast to nearby regions. Such rainfall corridors have high impacts on regional hydrologic cycles at both short-term and seasonal scales (Carbone et al. 2002; Jiang et al. 2006; Chen et al. 2014). They are also associated with persistent extreme daily precipitation and cause severe floods in the east China plains (Chen and Zhai 2013; Sun et al. 2016). Therefore, an understanding of the preferred hour and locations of these corridor-type heavy rainfalls has valuable implication for quantitative precipitation forecast, disaster prevention, water resource management, and hydrological studies.

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In the past decades, numerous studies have been carried out to investigate mei-yu torrential rains and their associated large-scale circulations, synoptic-scale disturbances, low-level jet (LLJ), MCSs, as well as cloud–convection morphology (Chen and Yu 1988; Qian et al. 2004; Ding and Chan 2005; Xu and Zipser 2011; Luo et al. 2013, 2014). Recent studies show that the mei-yu rainfall exhibits a unique diurnal cycle with a morning peak, in contrast to the afternoon peak of rainfall far away from the front (Geng and Yamada 2007; Yuan et al. 2010). In particular, heavy rainfall corridors mostly occur in the hours from midnight to morning, which account for the large variation in the diurnal cycle of warm-season rainfall, as in the central U.S. plains (Jiang et al. 2006; Tuttle and Davis 2006; Szi12; Chen et al. 2014). It indicates that the mei-yu rainfall systems in such corridors can be strongly regulated by the regional forcing with a phase locked to diurnal cycle. Over East Asia, possible forcing may arise from the diurnal variability in prevailing monsoon, mei-yu front, thermally driven local circulation, and nocturnal LLJ (NLLJ) over complex terrains (Chen et al. 2009; Yin et al. 2009; Chen et al. 2010; Bao and Zhang 2013; Du et al. 2014). Although the large-scale circulation that regulates the mei-yu rainfall has been extensively studied, the regional forcing that further sharpens the rainband and determines its detailed features are less understood. Thus, the mechanism governing the morning-peak rainfall corridor should be addressed as a key scientific issue for understanding how regional forcing and large-scale circulation work together to trigger extreme precipitation during the mei-yu season.

A well-known diurnal mode of rainfall over East Asia is associated with convective systems that tend to grow at the foothills of the Tibetan/Yun-gui Plateau at midnight and propagate eastward in the late-night-to-morning hours (Miller and Fritsch 1991; Wang et al. 2004, 2012; Yu et al. 2007; Bao et al. 2011). This mode is analogous to the succession of nocturnal MCSs that propagate along similar paths producing extreme rainfall in a narrow corridor over the central United States (Tuttle and Davis 2006; Trier et al. 2006, 2014). The propagation speed over East Asia is slower than the counterpart over North America, probably because of a weaker mean flow over Asia (Carbone et al. 2002; Wang et al. 2004). Correspondingly, the propagating mode of rainfall is dominant in the middle reach of the Yangtze River valley (MYRV; 105°–112°E) from midnight to morning. In the lower reach of the Yangtze River valley (LYRV; 112°–120°E), there exists another mode of morning rainfall induced by local growth or revival of convection at late night (Chen et al. 2012; Luo et al. 2014). When both modes of the propagating and locally developed convective systems coexist, severe flooding is expected in the whole Yangtze River valley (YRV). The cause for such an extensive growth of nocturnal convection remains an open question.

Previous studies attributed nocturnal rainfall systems to the mountain–plains solenoid with cold pool dynamics, low-level moisture transport driven by pressure tides, low-level ascent enhanced by diurnally varying winds, favorable synoptic conditions with vorticity/moisture...
advection, and moist instability induced by monsoon flows at night (Augustine and Caracena 1994; Laing and Fritsch 2000; Chen et al. 2010; Trier et al. 2010; Huang and Chan 2011; Wang et al. 2012; Bao and Zhang 2013; Chen et al. 2014). SZ12 simulated idealized corridors and found that a combined effect of mountain–plains solenoid and NLLJ strengthens the eastward-moving mesoscale convective vortices to produce heavy rains at night. Trier et al. (2006, 2014) found that the growth and organization of nocturnal convective episodes often occur in a deep layer of conditionally unstable moist air along a quasi-stationary front. Luo et al. (2014) noted that convective initiation to the west of a nighttime MCS resulted from moist southwesterly flow overriding a mesoscale cold pool induced by the convective activities in previous afternoon–evening hours. More than being a steady mean flow, the East Asian summer monsoon has a large diurnal variation on meso-synoptic scales (sometimes in the form of NLLJ), and it becomes most efficient in moisture transport and convergence at night (Chen et al. 2009, 2013; Monaghan et al. 2010; Du et al. 2014, 2015). Although the monsoon–rainfall linkage is well recognized at seasonal and longer time scales (Ding and Chan 2005), our knowledge on the roles of diurnally varying monsoon flow in regulating the timing and location of MCSs (extreme rainfalls) is still lacking.

This study revisits a typical precipitation corridor that caused severe flooding over East Asia in the summer of 1998, with emphasis on its diurnal cycle and regional features. A series of numerical experiments and composite analyses of a 5-day period are conducted to examine the mechanisms that governed nocturnal MCSs in the corridor. The primary objective is to clarify the formation and evolution of MCSs in response to the environment, which itself varied both diurnally and regionally. The key question is how regional forcing factors collectively regulated low-level ascent, moisture supply, and convective instability to determine the corridor in this case study. Whereas SZ12 has shown the dynamics of local circulation, here we focus on the impacts of moist processes on the nocturnal MCSs in the corridor. In section 2, we describe datasets, case selection, and numerical model. In section 3, we present the activities of MCSs and relevant atmospheric diurnal variability in both observation and simulation. In section 4, we illustrate the impacts of diurnally varying winds on the NLLJ evolution, moisture transport–lifting, and mei-yu frontogenesis. In section 5, we analyze the convective instability responsible for convection initiation and further quantify the genesis of instability by various physical processes. Finally, the conclusions are presented in section 6.

2. Observational data, case selection, and numerical experimental design

a. Observational data

To present the activities of precipitation systems, we mainly use the multisatellite rainfall estimate (3B42 V7) of the Tropical Rainfall Measuring Mission (TRMM). TRMM 3B42 offers 3-hourly rain average at 0.25° resolution, covering 50°S–50°N from 1998 to the present (Huffman et al. 2007). This provides important information to detect finescale features of rain systems including their global diurnal cycles (Kikuchi and Wang 2008). Previous studies have shown that TRMM 3B42 performs well over East Asia, though it has some errors in capturing nocturnal rainfall in the low-lying areas of the YRV (Zhou et al. 2008; Chen et al. 2012; Yuan et al. 2012). Meanwhile, as TRMM 3B42 is mainly scaled to rain gauge data on a monthly basis, its capability representing rainfall corridor and individual rain events during a 5-day period of our interest should be validated. Therefore, we also use the daily data of rain gauge provided by the China Meteorological Administration (CMA). These quality-controlled rain gauges scattered over eastern China at a resolution of 80–100 km provide daily rainfall records at 2000 LT (LT = UTC + 8 h).

To show the observed atmospheric conditions, we use the GEWEX Asian Monsoon Experiment (GAME) reanalysis data during the warm season of 1998. A series of intensive observational data including 6-hourly soundings from the South China Sea Monsoon Experiment (Lau et al. 2000) have been assimilated using 3DVar method by the Japan Meteorological Agency (JMA) reanalysis team. The GAME reanalysis at 6-h interval has a resolution of 0.5°, covering the Asian region. These high-quality assimilated data are also used as initial and boundary conditions for the short-term forecast of mesoscale weather in this study.

b. Selected torrential rain events and their associated large-scale background

During the summer of 1998, the East Asian monsoon circulation was relatively weak, and the monsoon flow mainly affected the south regions of East Asia. The mei-yu rainfall became stationary over central China and caused catastrophic floods on a basin scale of the YRV (Qian et al. 2004; Sun et al. 2016). During 11–27 June, torrential rain events occurred repeatedly in both MYRV and LYRV (Fig. 1a). In particular, the centroids of successive rain events were mostly located at the latitudes of 27.3°–28.5°N during 11–16 June (not shown). A large amount of rainfall was recorded along a narrow band oriented nearly east–west, extending from the eastern foothills of the Tibetan/Yun-gui Plateau to the
East China Sea (Fig. 1b). Such rain events with little south–north drift over 5 days define a typical corridor of heavy rainfalls, as suggested by Tuttle and Davis (2006). Embedded in the corridor of special interest, four mesoscale areas of rainfall were centered at $\approx 110^\circ$, $\approx 114^\circ$, $\approx 117.5^\circ$, and $\approx 126^\circ$E (marked by A to D in Fig. 1b). These regional features were similar to the climate mean pattern with large rainfall amount at these locations over the YRV (rectangle in Fig. 1a).

During 11–16 June, the lower-tropospheric conditions were characterized by a quasi-stationary mei-yu front in the YRV, as shown by a strong gradient of moisture at 28°–32°N (Fig. 1b). To the south, the southwesterly monsoon with wind speed $\geq 12$ m s$^{-1}$ and humidity $\geq 15$ g kg$^{-1}$ at 850 hPa dominated southern China. Such a strong monsoon flow originated from the South China Sea and Bay of Bengal, and it played a vital role in transporting substantial moisture to the frontal zone in YRV, as in the climate pattern of the mei-yu season (Qian et al. 2004; Xu et al. 2008). Heavy rainfalls mainly occurred to the northern edge of monsoon flow where it encountered the mei-yu front. To the north of the front, two meso-synoptic vortices appeared at the Sichuan basin and east coast (marked as V1 and V2 in Fig. 1b), which corresponded to flow confluence and cyclonic wind shear in the frontal zone. The ridge of the Tibetan which corresponded to flow confluence and cyclonic energy and flow convergence coexisted (Figs. 2c,d). It suggests that the convection initiation in the corridor coincided with a well-defined surge of moist monsoon impinging on the mei-yu frontal zone at night.

In the morning (0500–0800 LT), the southwesterly NLLJ became the strongest and extended downwind (Fig. 2f). The maximum wind speed exceeded 17 m s$^{-1}$ based on the GAME reanalysis and reached 20 m s$^{-1}$ in the soundings near (25°N, 111°E) (not shown). In the LYRV downstream of the NLLJ, wind speed also increased to 12–14 m s$^{-1}$. The enhanced monsoon flow seemed to produce a train of MCSs that concurrently developed in the low-lying areas of the YRV, which are marked by A to C in Figs. 2c,f. In the LYRV, intense rainfalls were recorded in a zonal band at the northern edge of the NLLJ. In the MYRV, rainfalls were observed in the mei-yu rainband at $\approx 28^\circ$N and in the adjacent area south (at 24°–26°N). The rainy areas in the MYRV when tracked back in time (Figs. 2c–f) seemed to move from the Yun-gui Plateau and be coupled with the southwesterly NLLJ to the southeast of the quasi-stationary low-level vortex (V1). The high-$\theta_e$ air mass declined in the morning over southern China and the YRV, which indicates a depletion of static moist energy due to a moisture reduction and partly to the radiative cooling through the night. At 1100–1400 LT, the LLJ decayed with a gradual dissipation of rainfall, while the pool of the high-$\theta_e$ air mass was rebuilt south of $\approx 27^\circ$N (Figs. 2g,h).

**c. Observed diurnal variability in precipitation, monsoon, and mei-yu front**

Before discussing our numerical simulations, we examine the observed monsoon–rainfall linkage on the diurnal time scale during the 5-day period in 1998. Figure 2 shows the evolution of rainfall and lower-tropospheric conditions at subdaily intervals. In the afternoon and evening (1700 and 2000 LT), the southwesterly monsoon over southern China was in a suppressed phase with respect to daily mean (Fig. 2b). A pool of warm moist air with high equivalent potential temperature ($\theta_e \approx 348$ K, hatched in orange) was established over the landmass of southern China. The $\theta_e$ contours of $\approx 344$ K in the MYRV and 332–344 K in the LYRV roughly mark the mei-yu frontal zone, where the meridional gradient of $\theta_e$ and horizontal wind shear were maxima. Most of the rainfall systems tended to decay as they propagate eastward in the YRV, while new systems began to grow at the eastern slope of the Yun-gui Plateau with upslope winds (Figs. 2a,b).

In the midnight and late night (2300 and 0200 LT), the wind speed of monsoon increased by 2–4 m s$^{-1}$ in 6 h over southern China (Fig. 2d). A southwesterly NLLJ formed, with a mesoscale wind maximum exceeding 16 m s$^{-1}$ centered near (24°N, 110°E). Along with an enhanced monsoon flow, the high-$\theta_e$ air mass moved northward to the YRV. The meridional gradient of $\theta_e$ intensified, especially in the LYRV from $-4.6$ K (degree latitude)$^{-1}$ at 2000 LT to $-6.4$ K (degree latitude)$^{-1}$ at 0200 LT as shown by a narrow area hatched in green, indicating a nocturnal mei-yu frontogenesis (Chen et al. 2009). Rainfall began to grow in a west–east-elongated belt at the northern edge of the high-$\theta_e$ air mass (the southern edge of the intensified mei-yu front) where moist static energy and flow convergence coexisted (Figs. 2c,d). It suggests that the convection initiation in the corridor coincided with a well-defined surge of moist monsoon impinging on the mei-yu frontal zone at night.
FIG. 2. The 5-day composite of horizontal winds (short and long barbs denoting 2 and 4 m s$^{-1}$) and equivalent potential temperature (contours, hatched for 332–344 K in green and ≥348 K in orange) using the GAME re-analysis data, except (a), (c), (g) because reanalysis data are only available at 6-h intervals; and TRMM rainfall (mm, colored shading) at: (a)–(h) 3-hour intervals, 1700 to 1400 LT. The V1 and V2 symbols denote the centers of vortices. The A to C symbols mark the three major MCSs.
Li et al. (2007) also noted that most cloud clusters during the mei-yu periods of 1998 and 1999 had a convective peak between midnight and early morning. Thus, the selected period here is suitable for studying the diurnal cycle of rain systems that produced the morning-peak heavy rainfall corridor. To gain more insight into moist processes that diurnally modulated the MCSs in the corridor, we need to analyze the outputs from mesoscale model simulations.

d. Design of numerical experiments

The numerical simulations are performed using the JMA Nonhydrostatic Model (JMA-NHM). Details about this model can be found in Saito et al. (2006). Previous studies have compared results of JMA-NHM to those of the Weather Research and Forecasting (WRF) Model, and concluded that both models have similar performance in predicting precipitation and atmospheric conditions around Japan and Southeast Asia (Hayashi et al. 2008). More details on the operational and research applications of JMA-NHM were shown in Saito (2012). In this study, the model domain contains 504 × 364 horizontal grids at a resolution of 10 km, covering a large area of East Asia and the Tibetan Plateau (Fig. 1a). It has 38 vertical levels with variable intervals from 40 to 1060 m; the lowest level and model top are set at 20 and 19270 m above ground, respectively. The modified Kain–Fritsch convection scheme (Kain and Fritsch 1993), improved Mellor–Yamada level-3 model (Nakanishi and Niino 2004), and mixed-phase cloud physics are employed for parameterizations of atmospheric processes.

Using repeating atmospheric forcings, previous studies successfully simulated the idealized rainfall corridors with regular diurnal cycle (Trier et al. 2006; SZ12). However, large differences between simulation and observation often occur for individual rain events, as forecast errors for mesoscale systems become large in a long integration of 7–10 days. In the current study, to reproduce torrential rain events reliably, five simulations rather than a successive one are conducted for the period of 11–16 June 1998. They are initiated at 0800 LT (0000 UTC) of each day on 11–15 June for daytime spinup and last for 36 h, which is similar to the modeling of the NLLJ (Du et al. 2014). The results of 9–33-h forecasts are used to elucidate the life cycle of nocturnal MCSs. The 6-hourly assimilated data (pressure, horizontal winds, temperature, and humidity) from the GAME reanalysis at a resolution of ~50 km are used as initial and lateral boundary conditions in each experiment. The initial condition of cloud microphysical quantities is given by the prerun of 6-h forecast from 0200 LT. Based on realistic forcing on each day, these numerical simulations may present how atmospheric conditions (with a similar diurnal cycle but some variations among five days) determined the timing and regional features of MCSs.

It is recognized that the rain-bearing mesoscale disturbances are often embedded in large- and synoptic-scale conditions, which can be affected by huge terrains such as the Tibetan Plateau (Ding 1992; Qian et al. 2004; Wang et al. 2012). Here, we focus attention on the diurnal cycle of regional monsoon and its role in mesoscale convection along a synoptic-scale front. Choosing a large model domain over Asia (Fig. 1) with reasonably fine grids is a compromise in our simulation design. It allows for portraying the large- and synoptic-scale conditions associated with heavy rainfalls while having sufficient resolution to capture mesoscale disturbances as well. Qian et al. (2004) conducted a series of numerical experiments on heavy rainfalls during the same study period. A grid spacing ≈20 km seemed adequate to describe the mesoscale structures of rain-bearing systems. In SZ12, 9-km-resolution experiments produced a rainfall pattern similar to the convection-permitting simulations at 3-km resolution, implying that the mei-yu rainband may be insensitive to cumulus scheme and spatial resolution. In fact, we have tested a 4-km-resolution run in a small domain of eastern China without any cumulus parameterization, which produced similar results for the heavy rain case of 15–16 June (not shown). Other downscale studies also suggested that a horizontal resolution of 40 km or higher is capable to resolve the diurnally varying LLJ over southern China (Monaghan et al. 2010; Du et al. 2014).

3. Comparison of rainfall corridor between simulation and observation

a. Spatial pattern and diurnal cycle of the rainfall corridor

For validation, the 5-day cumulative rainfall of JMA-NHM simulations is compared with the observational data of rain gauge and TRMM 3B42. The simulation reproduces the west–east-oriented rainband at ~28.2°N (Figs. 3a–c). The area of heavy rainfall above 150 mm in both simulation and observations extended from ~107°E to ~120°E (~1300 km) in a narrow width of ~160 km. The numerical model reproduces the maximum intensity (above 300–400 mm) of the rain corridor observed by rain gauge (Figs. 3a,c). Both simulated and observed rainfalls were suppressed over the north China plains, southern China, and the South China Sea. In particular, the large contrast of rainfall along ~27°N between the rain corridor and the nearby area to its south is well simulated by the model. As a result, the
primary feature of substantial rainfall in the corridor, in contrast to the adjacent areas where precipitation was below normal, is well reproduced.

The satellite estimate presents smaller rainfall maxima in the corridor and a wider area of rainfall in the warm regions south of 27°N, compared to rain gauge observations and model simulations (cf., Figs. 3b and 3a,c). In particular, TRMM 3B42 tends to overestimate the rainfall in the eastern lee of the Yun-gui Plateau near (25.5°N, 110°E), where it reports false heavy rains for a fast-moving convective system with an extensive cirrus anvil in the early morning of 16 June (not shown). As we focus on the rain corridor in the YRV along 28°N, such errors in TRMM 3B42 over southern regions do not affect our conclusions. Overall, the simulations give a slight overestimate (~8%) of the 5-day total rainfall in the YRV compared to the satellite estimate. The model also produces more rainfall than TRMM 3B42 on the ocean surface east of 122°E.

Three mesoscale centers of rainfall at ~110°, ~114°, and 116°–118°E are clearly seen in both observations and simulations (Figs. 3a–c). All three maxima exceeded 300 mm and the strongest one in the east was greater than 400 mm. The rainfalls were located at the low-lying plains and had maxima on the windward slopes of local mountain ranges (Fig. 3d). The corridor of these extreme rainfalls reflected the footprint of convective systems that repeatedly developed over the plains and followed similar paths of eastward propagation. Sensitivity experiments using the numerical model with mountain ranges removed produced a zonal shift of maximum rain locations, resulting in weaker regional
features (not shown), which is similar to the difference of Figs. 3d,e in SZ12. The local topography thus can affect the mesoscale features of rainfall in the corridor, though it has little influence on the general pattern of the mei-yu rainband (SZ12). We also notice in Fig. 3 that some light rains appeared in the south mountains at ~25°N, but they were much weaker than the heavy rains in the corridor, indicating that terrain lifting was not the dominant factor determining the main location of the rain corridor. Overall, our numerical simulations capture reasonably well the general pattern, terrain-dependent mesoscale features, and magnitude of the heavy rain corridor.

Longitude–time (Hovmöller) diagrams of rain rate are used to evaluate the modeling of individual rain events and the mean diurnal cycle of the corridor. Three intense rain systems were observed in the MYRV in the morning of 13, 14, and 16 June (Fig. 4a). All three events initiated at the eastern foothills of the plateau near midnight and propagated eastward. Their timing, location, and propagation features are well simulated, despite an underestimate of rain intensity for the first event.
(Fig. 4b). Figures 4a and 4b show that intense rainfalls also occurred over the LYRV in the morning. They were relatively strong on 13, 14, and 16 June, implying diurnal and day-to-day variations of the rainband. It is clear that a train of MCSs developed nearly simultaneously over the MYRV and LYRV in the midnight-to-morning hours of those days. For instance, in the morning of 16 June, a MCS occurred at 108°–111°E, while the other two grew at 114°–115° and 118°–120°E, as marked by A to C. Such a corridor of nocturnal rainfall with three coexisting MCSs had a longitude extent of more than 1300 km, which was somewhat longer than that reported previously (Tuttle and Davis 2006; Trier et al. 2006, 2014; SZ12). Climatologically, the simultaneous growth of several MCSs can account for a large portion of heavy rain episodes over central China (Chen et al. 2012). Although the convective systems moving from the MYRV the previous day can be a precursor of the growth of nocturnal convection over the LYRV (SZ12; Luo et al. 2014), the coherent propagation of individual MCSs across the whole YRV is not obvious in Figs. 4a,b.

Figures 4c and 4d show a composite of diurnal cycle of the rainfall corridor. Two regimes of diurnal cycle with morning-peak rainfall occurred in the MYRV and LYRV. Between them was a gap at ~112°E with relatively weak rainfall. In the MYRV, rainfall initiated near 105°–106°E at midnight, reached its maximum at ~110°E in the morning, and dissipated in the afternoon. The propagation speed is estimated at about 14 m s\(^{-1}\), which is slightly slower than the climate mean (Wang et al. 2004; Chen et al. 2014). The peak time of simulated rainfall (0500–0800 LT) was somewhat earlier than that of the TRMM rainfall (0800–1100 LT). This difference was related to a bias in the TRMM-derived morning rainfall that usually had a delay of 2–3 h, which is different than the rain gauge observation over the MYRV (Figs. 3b and 5b in Zhou et al. 2008). In the LYRV, rainfall increased at late night and reached its peak in the morning, with two subregions of strong rainfall at 113°–114° and 116°–118°E. These morning-peak rains contributed greatly to the heavy precipitation over the low-lying plains between local mountain ranges (Figs. 3–4). Numerical experiments reproduce well the diurnal cycle of observed rainfall in terms of peak time, eastward-delayed diurnal phase, and regional differences in the corridor. Advanced data assimilation may be applied to further improve the forecast of heavy rainfalls, but considering that the purpose of this study is to examine the mesoscale meteorology during a 5-day period rather than the details of the convection forecast, these simulations are suitable for further composite analyses. Since a lot of averaging is being done over space and time, the conclusions are not especially dependent on the details of simulated convection in the model.

b. Atmospheric diurnal variability associated with rainfall corridor

We evaluate the diurnally varying atmospheric conditions by comparing our simulations with observations (Figs. 2 and 5). The simulations reproduce the convergent flow and growing rainfall on the slope of the Yungui Plateau in the afternoon and evening (Figs. 5a,b). From midnight to morning, the model captures the enhanced southwesterly monsoon over southern China (Figs. 5c–f). In particular, the formation of the NLLJ at the plateau’s foothills and the downwind movement of jet’s core are well simulated. The maximum wind speed reached ~20 m s\(^{-1}\), which is comparable to sounding observations but 2–3 m s\(^{-1}\) greater than the reanalysis data. The NLLJ brought warm moist air mass toward the YRV and enhanced mei-yu frontogenesis. The simulations also capture both the veering low-level wind vectors to the southeast of vortex “V1” (Figs. 5c–f) and the enhanced northerly wind to the west of vortex V2 (Figs. 5e,f). The vortices and strong monsoon flow acted to strengthen horizontal shear in the frontal zone in the morning. Along with the atmospheric diurnal variability, the convection initiation at 2300–0200 LT, the convective stage with intense rainfall at 0500–0800 LT, and the dissipating phase at 1100–1400 LT are all well simulated. More than the synoptic-scale patterns, three mesoscale areas of strong winds at 27°N can be seen near 109°, 113°, and 116°E (Figs. 5d,e). These localized maxima of wind speed reached 14–18 m s\(^{-1}\) and had a horizontal scale of 100–200 km, indicating the activities of mesoscale LLJs (mLLJs). They are not clearly seen in the GAME analysis, which has a relatively coarse resolution (Figs. 2d,f). The evolution and impact of these monsoonal mLLJs on convective growth are detailed in section 4.

Figure 6 shows the diurnal cycles of monsoon and rainfall that varied over the 5-day period. In the cross section for the MYRV, the rainfall systems initiated at ~28°N around 0200 LT on 12 and 13 June, which corresponded to an enhanced high-\(\theta_v\) inflow from the south (Figs. 6a,b). The intense rainfall occurred with a stronger moist inflow in the morning of 13 June than of 12 June. In the third diurnal cycle, rainfall system developed at midnight of 13 June, in relation to a relatively early enhancement of moist southerly flow. The model captures such an early shift of diurnal cycle, though it gives a northward displacement of rainfall. In the last diurnal cycle, the intense rainfall tended to propagate southward in the morning of 16 June. It was associated with the enhanced high-\(\theta_v\) inflow from the south since
Fig. 5. As in Fig. 2, but derived from the JMA-NHM simulations. The equivalent potential temperatures of 332–344 K and ≥350 K are hatched in green and orange, respectively; pennants on the wind barbs indicate a wind speed of 20 m s$^{-1}$. Dashed lines AB and CD in (h) mark the horizontal planes used in Fig. 6.
FIG. 6. Temporal variations of equivalent potential temperature (color shading for $\geq 344$ K), horizontal winds (short and long barbs denoting 2 and 4 m s$^{-1}$; pennants denoting 20 m s$^{-1}$) at 850 hPa, and rain rate (blue contours of 1, 2, 4, 8, and 16 mm h$^{-1}$) along the (a),(b) north–south line and (c),(d) the southwest–northeast line marked in Fig. 5h.
midnight and the developing northerly at the cold sector since morning due to the movement of vortex V1.

In the cross section for the LYRV, strong rainfall systems developed at ~28°N during 0200–0800 LT of 13 and 14 June (Figs. 6c,d). Their initiations coincided with the arrival of the high-θe air mass and an enhanced southwesterly around 0200 LT. They were earlier than the first event that occurred at 0800 LT of 12 June and had a north location (~29.5°N). Rainfall systems also developed in the morning of 15 and 16 June. The simulations reproduce the stronger rain event of 16 June with a larger supply of warm moist air. Rainfall systems tended to dissipate after 0800 LT usually, when the high-θe air was depleted at the frontal zone (Gale et al. 2002). The simulations present the perturbed northerly/easterly winds north of the rain events in the mornings of 13, 14, and 16 June (Fig. 6d), which are indicative of the formation and development of mesoscale convective vortices associated with organized convection (Davis et al. 2004; Schumacher and Johnson 2009; SZ12). Such local-scale features were not evident in the GAME reanalysis at coarse resolution (Fig. 6c). Overall, observations and model results agree on variations of diurnally varying monsoon flow over the 5 days could regulate the timing and intensity of individual MCSs in the corridor. These findings are consistent with previous studies, suggesting that the rainfall amount depends strongly on the supply of warm moist air into the mei-yu corridor. These findings are consistent with previous studies (Monaghan et al. 2010; Liu et al. 2012; Du et al. 2014; He et al. 2016). The localized areas of large diurnal variation in NLLJ clearly corresponded to the three monsoonal mLLJs over the plains, as shown by the dashed arrows.

Figures 7d–f show that, in the cold sector north of 28°N and east of 110°E, the diurnal component shifted from westerly to northerly wind (1–2 m s⁻¹) during 0200–0800 LT. The fast veering was related to a relatively short period of inertial oscillation and an initiation of daytime southerly vector. These diurnally varying winds with different phases between warm and cold regions of the LYRV, along with the vortex in the MYRV, accounted for the enhanced horizontal shear and convergence in an elongated belt. Such an enhancement by the synoptic-scale monsoon and vortices was more extensive than that shown in SZ12, which favored corridor convection with a longer span in the present study (Fig. 3). The horizontal convergence induced by wind diurnal variation had a magnitude of O(10⁻⁵) s⁻¹ and was comparable to that of daily mean. Note that the diurnal wind amplitude in the cold sector was much smaller than that in the warm sector. Thus, the speed-up of regional monsoon at night was vital for strengthening frontal convergence. In particular, the monsoonal mLLJs led to the localized maxima of convergence at their northern terminus. At 1100–1400 LT, the deviated northeasterly wind developed over southern China (Figs. 7g,h). The suppressed southwesterly monsoon led to a rapid decline of the LLJ, frontal shear, and convergence, in relation to the decaying rainfall in the corridor.
Figure 8 shows the vertical structures of low-level winds at the adjacent areas south of the mei-yu front. In the evening (morning), the anomalous easterly (westerly) wind appeared below 700 hPa, with the largest diurnal amplitude at 900–800 hPa at the plateau’s foothills (Figs. 8a–d). It was related to a large-scale contrast between the plateau and plains, while it was less dependent on the local mountains east of 110°E. Unlike the zonal wind, the meridional wind and its diurnal component were the largest at late night about 0.5–1 km AGL, in a close relation to the NLLJ (Figs. 8e,f). Three localized maxima of southerly wind appeared over the low-lying valleys or plains at 108°–110°, 111°–113.5°, and 115°–117°E, where the monsoonal mLLJs were established.
The south–north-oriented mountain ranges seemed to have a channeling effect that favored the local enhancement of the southerly wind. As a result, the mLLJs formed at the northern edge of the southwesterly monsoon as it penetrated the south–north-oriented plains over the YRV at night (Figs. 7–8). Since the mLLJs were collocated with mesoscale centers of intense rains at their terminus (Figs. 5d,e), they may play a role for bridging the synoptic-scale monsoon flow and mesoscale disturbances.

b. Diurnal cycles of water vapor flux and frontogenesis

Figure 9 shows that the diurnal variation of water vapor flux integrated in the lower troposphere. It is clear that the southwesterly monsoon transported a large amount of moisture from southern China to the frontal zone near 28°N. At 27°N, water vapor flux strengthened from 300 kg m\(^{-1}\) s\(^{-1}\) in the afternoon to ~450 kg m\(^{-1}\) s\(^{-1}\) at late night, with an increase rate of ~50%. In particular, the poleward transport was mostly confined to three low-lying plains, where the monsoonal mLLJs behaved like mesoscale conveyors of moisture. Such an enhancement of moisture transport was mainly induced by the increase of low-level winds, especially of meridional component rather than that of humidity (Fig. 8). On large and synoptic scales, summer monsoon is known as the major supply of moisture for the mei-yu rainband (Ding and Chan 2005). Recently, strengthened monsoon flow at night was found to be
most efficient for the poleward transport of moisture over China (Chen et al. 2009, 2013; Yuan et al. 2010). Here, such an impact of the diurnally varying monsoon flow is shown to possess evident local features over complex terrains, hence linking to the mesoscale rainfall systems.

Figure 9b also shows that the meridional gradient of water vapor flux intensified at the frontal zone at late night. It led to a large convergence of moisture similar to that in Figs. 7d–f. Along with the enhanced horizontal convergence and deformation of moisture inflow, the meridional gradient of $\theta_e$ is intensified at a large rate of $\sim 3 \, \text{K km}^{-1} \, \text{h}^{-1}$ at late night, indicating the mei-yu frontogenesis (Ninomiya 2000; Chen et al. 2009). On the synoptic scale, the mei-yu front generally determines the location and persistence of a rainband (Ding and Chan 2005). Here, it became most active with enhanced horizontal wind shear and mass convergence overnight, hence supporting the growth of nocturnal MCSs in an elongated zone (hatched in Fig. 9b). Note that the frontogenesis was the largest at the south portion of the mei-yu frontal zone due to the impact of monsoonal mLLJs. Therefore, the NLLJ that encountered the frontogenetic boundary helped define a narrow area of

![Figure 9](https://example.com/figure9.png)

FIG. 9. Composite water vapor flux (kg m$^{-1}$ s$^{-1}$, vectors and color shading) integrated in the lower troposphere below 700 hPa and averaged over five days at (a) 1700 LT and (b) 0200 LT. Red hatched and red enclosed areas denote total accumulative rainfall $\geq 50$ mm in the following 3 h.
enhanced low-level ascent for convective growth (Augustine and Caracena 1994).

5. Convective instability associated with convection initiation and evolution

a. Variation of convective instability

In this section, we investigate the diurnal cycle of convective instability and related genesis processes to understand the initiation and evolution of convective activities in the heavy rainfall corridor. The key question to be addressed is why the nocturnal MCSs were meridionally confined in a narrow zone. We examine the CAPE and CIN in the ambient inflow of the monsoonal air mass south of the mei-yu front. Over there, the CAPE variations derived from soundings data had been well reproduced by the model (not shown). The CAPE, CIN, and their changes are calculated for air parcels originating at all vertical grid levels. Following Trier et al. (2014), we pay more attention to the air parcels in the most unstable layer (the highest $\theta_e$ at a vertical grid point) and at the elevated layer (~1180 m AGL), which well represented atmospheric conditions for convective growth.

Figure 10a shows that the CAPE of the most unstable air parcel in a vertical column (MUCAPE) was large over the warm area south of 27°N in early evening, while it was relatively small in the corridor at 28.2°N. The CIN of the most unstable air parcel (MUCIN) was usually larger than 25 J kg$^{-1}$ over the domain except for some scattered areas in the south (Fig. 10b). These features corresponded to the warm moist air mass over southern China and the suppressed rainfall over the YRV. At late night, a significant change was an increased MUCAPE and decayed MUCIN in the corridor (Figs. 10b,d). The largest CAPE was tightly associated with three conveyor belts of moisture (Figs. 9b and 10b). The smallest CIN occurred in a narrow band at the northern edge of enhanced monsoon and was well collocated with intense rainfall in the following hours (Fig. 10d). Over there, the monsoonal NLLJ might provide the lifting needed to overcome weak CIN for initiating convection in a very confined zone (Fig. 7d). In contrast, the changes of CAPE and CIN were much small in nearby areas, implying a large south–north gradient in the genesis of convective instability.

To show the variation of moist instability in monsoon flow, Fig. 11 shows the diurnal cycles of MUCAPE, MUCIN, height of the most unstable layer, and 2-km-depth humidity at three locations south of the corridor. At the location ~200 km from the corridor, MUCAPE reached its maximum in the afternoon and declined from evening to the following morning (Fig. 11a). In contrast, the diurnal peak of MUCAPE appeared at late night at the location only ~50 km from the corridor. The CAPE maximum exhibited a northward delayed phase, in relation to the northward movement of warm moist air. Figure 11b shows that MUCIN decreased in the evening at the three locations. In particular, at the northernmost location MUCIN reached a minimum of ~5 J kg$^{-1}$ at midnight and late night, and it favored convective initiation with small lifting needed. Thus, the nocturnal presence of modest CAPE and small CIN south of the frontal zone were most conducive to corridor convection compared to the locations farther south, which explains the narrow band of rainfall. The feature is consistent with that in North America (Trier et al. 2014), but it occurs at a more extended span over East Asia under active monsoon conditions.

Figure 11c shows that the height of the most unstable layer increased from south to north and reached ~1 km AGL at the frontal zone. It indicates that the high-$\theta_e$ air was gradually elevated when moving northward. Figure 11d shows that the relative humidity above the most unstable layer increased from south to north and reached the highest value (~93%) at late night at the northernmost location. Such a nearly saturated condition was conducive to the growth of organized convection (Laing and Fritsch 2000). The vertical sections of $\theta_e$ and meridional wind show that a 2-km-depth layer of the high-$\theta_e$ air was established over southern China in the afternoon (Figs. 12a,c). At late night, it was transported by strong southerly wind and elevated in the frontal zone (Figs. 12b,d). At the northern edge of the high-$\theta_e$ tongue, convection developed with upward motion extending through the troposphere, because the elevated warm moist air helped sustain convection organization overnight (Trier et al. 2006). The vertical displacement of the high-$\theta_e$ air was most evident in a narrow zone, where the NLLJ encountered the frontal boundary to define strong ascent, as also noted in Peters and Schumacher (2015, 2016). Based on backward trajectory, previous studies also suggested that the high-$\theta_e$ air parcel may ascend 0.5–2 km along the NLLJ to produce the elevated conditionally unstable layer (Trier et al. 2006; Luo et al. 2014).

To show the convective instability that influences convection initiation, we estimate composites of CAPE and CIN for the air parcels from different vertical levels averaged over the inflow areas ~50 km south of the MCSs. Figure 13 shows that the inflow CAPE increased significantly for a period of ~6 h prior to convective initiation. The modest CAPE was formed in the lower troposphere below 2 km AGL. The CAPE maximum of 900–1100 J kg$^{-1}$ was located at 0.6–1.2 km AGL at the
FIG. 10. The 5-day composite spatial distributions of (a), (b) MUCAPE and (c), (d) MUCIN (J kg\(^{-1}\)) at 1700 and 0200 LT, respectively. Red hatched and red enclosed areas denote total accumulative rainfall ≥ 50 mm in the following 3 h. The wind barbs (a full pennant = 4 m s\(^{-1}\)) denote the horizontal winds at 710 m AGL. The three rectangles in (a), (b) mark the averaging area (100 km × 25 km) used for Fig. 11.
hours shortly after the convective onset, which suggests an elevated layer of conditionally unstable air. Meanwhile, the CIN decreased rapidly and became near zero at the hour of the convective onset. The height of CIN minimum was 0.4 km above that of CAPE maximum. It was analogous to that observed during the typical nocturnal rainfall episodes, in which convection initiation was favored by a large reduction in CIN despite a modest CAPE at the elevated layer (Chen et al. 2014; Luo et al. 2014; Trier et al. 2014).

Figure 13 also shows that at the mature stage of convection (3–6 h after onset), the CAPE decreased and CIN increased with the most intense rainfall. At 6 h after the convection onset, the small CAPE and large CIN turned to be unfavorable for sustaining the convection. Such a decline of convective instability, along with decayed low-level ascent and moisture supply, corresponded to the convective dissipation since morning. Thus, the large variations of CAPE and CIN at elevated layer might regulate both the initiation and dissipation of MCSs. In contrast, the CAPE (CIN) was much smaller (larger) near the surface during the life cycle of convection. The profiles of CAPE and CIN were very similar over the MYRV and LYRV, though the CAPE maximum and CIN minimum were located at a slightly higher layer over the LYRV where the frontal surface was more obvious.

b. Generation of convective instability by various physical processes

Given that convective instability at the elevated layer was a key factor for the evolution of nocturnal MCSs, its genesis processes were essential for governing the heavy rain corridor. Previous studies proposed a variety of processes that may regulate atmospheric instability at night: the vertical displacement of warm moist air in a frontal zone (Laing and Fritsch 2000; Trier et al. 2006, 2014; Luo et al. 2014), the vertical differential thermal advection due to diurnally varying circulation (Chen et al. 2010; Huang and Chan 2011), and the moisture advection and lifting enhanced by nocturnal monsoon (Chen et al. 2014). To clarify the relative importance of various processes in regulating the corridor convection in June 1998, we conduct quantitative analyses on the budget of CAPE generation.

Based on Emanuel (1994) and Zhang (2002), the total production of CAPE \( P_{\text{total}} \) can be induced by the production terms of the boundary layer forcing \( P_{\text{bl}} \) that regulates the air parcel’s \( \theta_e \) at source levels
\[
\frac{\partial}{\partial t} \left[ C_{p} (T_a - T_{\text{mb}}) \ln(\theta_e) \right]
\]
and of the free-atmospheric forcing \( P_{\text{fa}} \) that controls the thermal condition (geopotential thickness) of convection layer \( \frac{\partial (\phi_{\text{mb}} - \phi_{\text{fl}})}{\partial t} \).

Following Chen et al. (2014), the CAPE change due to boundary layer forcing can be further decomposed into the rates by low-level horizontal advection
\[
\frac{\partial \theta_e}{\partial x} \frac{\partial \theta_e}{\partial y} \left( -u \frac{\partial \theta_e}{\partial z} \right)
\]
and diabatic heating \( Q \) (i.e., \( P_{\text{adv}} + P_{\text{lifting}} + P_{\text{heating}} \)). Here, \( P_{\text{total}} \) is estimated as a local change of CAPE \( \frac{\partial \text{CAPE}}{\partial t} \) using the hourly data of model output and central difference scheme. To obtain \( P_{\text{adv}}, P_{\text{lifting}}, \) and \( P_{\text{heating}} \), the perturbations to \( \theta_e \) induced by each of the forcing terms are calculated at a specific level (the parcel’s source level), and the increments are added to the initial parcel. Then, the modified parcel is lifted assuming a pseudoadiabatic process in an unmodified free atmosphere to get a modified CAPE that can be compared to the original CAPE. To obtain \( P_{\text{fa}} \), virtual temperature or
geopotential thickness above the source level is perturbed, and then an unmodified parcel is similarly lifted in this modified environment to get the change of CAPE. The difference of $\Delta \text{CAPE}/\Delta t$ and $P_{bl} + P_{fa}$ indicates an error of budget estimate ($P_{\text{error}} = P_{\text{total}} - P_{bl} - P_{fa}$), which can be compared to the patterns and magnitudes of various CAPE generation rates to show the accuracy of calculation.

Figure 14 shows the latitude–time section of the CAPE and its change rates for the air parcel at the elevated layer. The CAPE maxima repeatedly appeared near 28°N at late night (Fig. 14a). These nocturnal peaks were associated with a large rate of CAPE increase (100–200 J kg$^{-1}$ h$^{-1}$) for several hours before late night. The increase of CAPE was not confined to the frontal zone, but occurred in a relatively wide area north of 26.5°N. The maximum and change rate of CAPE gradually shifted northward, due to the poleward movement of monsoonal air from southern China. As expected, the CAPE variation had a strong association with the frontal rainfall systems in terms of both locations and diurnal cycles over five days, as in North America (Augustine and Caracena 1994). Figure 14b shows that the generation rate $P_{bl}$ was highly analogous to the total change rate $P_{\text{total}}$ in Fig. 14a, in terms of magnitude, diurnal phrase, and northward shift. Thus, most of the generation of convective instability can be attributable to the boundary layer forcing that regulated the air parcel’s moist static energy. In contrast, the free-tropospheric forcing (warming or cooling aloft) generated a very small CAPE during the period (Fig. 14c). The term of $P_{fa}$ even became negative at the convection onset when the warm advection ahead of mid- to upper-level troughs might stabilize the stratification (Chen et al. 2014).

Figure 14d shows that, among the boundary layer forcing, the horizontal advection was a primary contributor to the large CAPE increase from evening to late night. The strong advection of monsoonal air, which could increase $\theta_e$ by a rate of about 4 K in 6 h, was the key to the precondition of convective instability for nocturnal MCSs. However, the CAPE production by horizontal advection became negative over southern

![Diagram](https://example.com/diagram.png)
China from late night to morning. It suggests that the relatively dry air after the passage of the high-$\theta_e$ air stabilized the stratification. As the reduction of CAPE arrived at the YRV in early morning, it led to the dissipation of frontal rainfall systems. Such strong variation of CAPE production by advection suggests that the inflow of warm moist air played a crucial role in regulating the diurnal cycle of moist instability for the corridor convection.

Figure 14e shows that the CAPE negative change from vertical motion (blue areas) occurred mostly where it rained (red contours in Fig. 14a). This feature corresponded to the strong updrafts in the convective zone that tended to remove the high-$\theta_e$ air elevated over the frontal cold air near 28°–30°N (Fig. 12b). Such a CAPE consumption was locally balanced by the CAPE generation from horizontal advection (cf., blue areas in Fig. 14e and red areas in Fig. 14d), so that high CAPE values could last for ~6 h supporting convection growth.

**Figure 14.** Latitude–time variation of CAPE (black contours, with an interval of 200 J kg$^{-1}$, color shading) and its generation rate by various processes (J kg$^{-1}$ h$^{-1}$, color shading) at 1180 m AGL. Generation rates by (a) all processes, (b) boundary layer forcing, (c) free-atmospheric forcing, (d) low-level horizontal advection, (e) lifting, and (f) diabatic heating. The red contours in (a) and (e) denote upward motion at 5330 m AGL (with an interval of 0.1 m s$^{-1}$) and 1180 m AGL (with an interval of 0.05 m s$^{-1}$), respectively. The rain rate is shown in Fig. 6d. An average over 112°–116°E is applied for all panels.
(Figs. 13 and 14a). Figure 14e shows also that, in locations slightly to the south of the rainy area, likely corresponding to the inflow regions for MCSs, the CAPE generation from vertical motion was positive. In particular, the CAPE generation was the greatest from late night to the following morning, when the low-level ascent of warm moist air was enhanced by monsoonal mLLJs and mesoscale convective disturbances, as discussed in previous sections. This CAPE generation by lifting over inflow regions partly offset the stabilizing effect of dry air advection (cf., red areas in Fig. 14e and blue areas in Fig. 14d). Overall, the change rate of CAPE by vertical motion had a relatively localized feature and small magnitude, compared to that by horizontal advection. On the other hand, diabatic heating at night generally led to a negative CAPE production at 28°–29°N (Fig. 14f), but at a magnitude smaller than advection and lifting (Figs. 14d,e).

To show the relationship of CAPE production with convection evolution, we make a composite of inflow air parcel at the elevated layer with respect to convective initiation. For a better presentative of inflow regions, the area average south of the MCSs is used for the calculation of CAPE change rates by various forcing terms. Figure 15a shows that the CAPE increased by a rate of 100–175 J kg⁻¹ h⁻¹ for 6 h prior to the convection onset and decreased by a rate of about −150 J kg⁻¹ h⁻¹ at 4–6 h after the convection onset. Such a large variation (diurnal cycle) of CAPE generation was primarily induced by boundary layer forcing, while the effect of free-atmospheric forcing was a relatively small and exhibited a semidiurnal cycle. Figure 15b shows that the horizontal advection was the major factor for the generation of CAPE prior to the convection and the reduction of CAPE at the convective stage. It highlights that the inflow of warm moist air was the dominant process regulating convective instability south of the corridor convection.

Figure 15b also shows that the low-level lifting increased at the hours of convective onset, probably due to the enhanced convergence by monsoon flow. Its CAPE production reached a maximum of ~100 J kg⁻¹ h⁻¹ at the mature stage of convection (~5 h after convective onset), which was likely related to the low-level ascent enhanced by mLLJs and mesoscale convective disturbances (Figs. 6d, 7, and 14e). It could partly compensate for the CAPE reduction by horizontal advection, hence supporting convection organization. Peters and Schumacher (2015, 2016) also noted that the increased isentropic ascent owing to the strong inflow responding to convection could destabilize the upstream region for further convective growth. The diabatic heating generally reduced CAPE, due to a joint effect of nocturnal radiative cooling and raindrop cooling. Figures 15c and 15d show that the relative importance of various processes in CAPE generation were similar over the MYRV, though the variation of CAPE production by lifting was relatively large, probably because of the mountain–plains solenoid and strong NLLJ. Finally, Figs. 15a–d show that the error of the budget estimate is quite small in terms of both magnitudes and variation amplitudes compared to the CAPE changes by various forcing terms, suggesting that the above results are reliable.

6. Conclusions

In this paper, we present numerical simulations and composite analyses of heavy rainfall events that were confined to a narrow latitudinal corridor over East Asia during five successive nights of the mei-yu season. We
investigated the moist processes that governed diurnal cycles and local features of the corridor convection. The major findings are summarized as follows.

1) A train of MCSs repeatedly occurred overnight over five days and produced morning-peak heavy rainfall (150–400 mm) in a narrow corridor (~1300 km long and ~160 km wide) in the lee of the Tibetan/Yun-gui Plateau. Our numerical experiments reproduced the observed heavy rainfall corridor, in terms of elongated pattern, diurnal cycle with morning peak, individual nocturnal MCSs, and rainfall maxima at plains and windward mountain slopes. They also captured the corridor-related atmospheric conditions, especially the large diurnal variations of low-level winds, vortices, high-$\theta_e$ air mass, and mei-yu front. Both the initiation and dissipation of MCSs in the corridor showed a close relationship to the nighttime speed-up of monsoon flow that determined the supply of the high-$\theta_e$ air mass to the south of the mei-yu frontal zone.

2) The NLLJ, with maximum wind speed exceeding 16 m s$^{-1}$ at late night, occurred in the southwesterly monsoon over southern China and extended downwind to central China. The strengthened monsoon flow, along with the meso-synoptic vortices in the YRV and the fast-rotating low-level winds in the cold region, led to the enhancement of horizontal wind shear and frontal convergence. As the monsoonal NLLJ extended northward, it was divided into three mLLJs over the low-lying plains between south-north-orientated mountain ranges. These monsoonal mLLJs strengthened the poleward transport of moisture by ~50% from the afternoon to late night. The strong meridional gradient of moisture flux by mLLJs also contributed to an intensified mei-yu front overnight. Through enhancing low-level ascent, moisture transport, and frontogenesis, monsoonal mLLJs collided with the frontal zone and defined three preferred locations in a zonal band favorable for the growth of nocturnal MCSs.

3) The initiation of corridor convection was tightly associated with a large increase of CAPE and a reduction of CIN in a narrow band at the northern terminus of the NLLJ, where enhanced low-level ascent might overcome a weak convective barrier. The CAPE maximum of ~1000 J kg$^{-1}$ was primarily established at the elevated layer of 0.6–1.2 km AGL, as a 2-km-depth high-$\theta_e$ air tongue was advected and lifted to the south of the mei-yu frontal zone. A quantitative estimate of the CAPE generation budget showed that the CAPE increase prior to convective initiation was mostly attributed to the low-level intrusion of moist monsoon air, while free-atmospheric forcing was much weaker. In particular, the horizontal advection of the high-$\theta_e$ air mass by the NLLJ could produce CAPE at a rate of 100–175 J kg$^{-1}$ h$^{-1}$ for more than 6 h south of the MCSs. At the convective stage, the enhanced vertical transport of moisture can produce CAPE at a rate of ~100 J kg$^{-1}$ h$^{-1}$ in the strong inflow region, which partly compensates for the energy consumption. This suggests that the convective-induced inflow ascent may play a feedback role for sustaining convection. In the morning of mature convection, the depletion of high-$\theta_e$ air at the elevated layer led to CAPE reduction, which along with decayed LLJ was followed by convection dissipation.

A similar phenomenon of heavy rain corridors has been documented at somewhat higher latitudes in the lee of the Rocky Mountains of North America (Tuttle and Davis 2006). Here, we showed that the elevated layer of conditionally unstable air south of a quasi-stationary front was favorable for the formation of such a corridor of extreme rains at night in East Asia, as in North America (Trier et al. 2006, 2014; Geerts et al. 2017). However, the current work pointed out some interesting differences between the two regions, which include the apparent stronger dependence on horizontal moisture transport to sustain CAPE during the heavy rains over East Asia. It is known that the mei-yu front over East Asia is quite shallow and has a small horizontal thermal gradient (Fig. 1b; Ding and Chan 2005), which may be sustained by the cold pool from convective systems (SZ12; Luo et al. 2014). Although the frontal lifting was present (Fig. 12), its production of CAPE was relatively small compared to that by horizontal advection (Figs. 14–15). This feature may reflect a strong differential effect of moisture advection by active monsoon flow over East Asia.

Another possible difference to that in North America is the modification to the LLJ associated with monsoon flow and smaller terrain features, and how this influences the heavy rains over East Asia. As the diurnal variation of monsoon flow was pronounced at the synoptic scale, it allowed for an extensive growth of several MCSs, producing a very long corridor of heavy rainfalls. The monsoonal NLLJ also displayed mesoscale features due to local terrains and explained the preferred locations of MCSs. The diurnally varying low-level winds thus could be viewed as a precursor for forecasting nocturnal MCSs (Augustine and Caracena 1994). Meanwhile, the high-$\theta_e$ air mass established over the heated landmass of southern China (Figs. 2 and 5) presented a recharge of available moist energy at
daytime (Shinoda and Uyeda 2002; Yamada et al. 2007). Its northward displacement by the monsoonal NLLJ at night thus may express a discharge of moist energy from southern to central China, producing a morning heavy rainfall corridor.

It is recognized that the mei-yu rainfall involved a hierarchy of space–time scales from large-scale circulation, synoptic-scale disturbances, to mesoscale systems (Laing and Fritsch 2000; Ninomiya 2000; Zhang et al. 2002), and from seasonal, intraseasonal, to daily variations (Qian et al. 2004; Liu et al. 2008; Sun et al. 2016). Our present study revealed an important process downscaling from synoptic-scale monsoon to mesoscale systems with evident diurnal variation. The diurnally varying monsoon flow was shown to effectively regulate three key factors (moisture transport, low-level ascent, and instability) that controlled the diurnal cycle and regional features of MCSs in the corridor. As large diurnal variability mainly occurs in active monsoon period, the diurnal variation of monsoon flow functions as a key regional forcing coupled with large-scale external forcings (Chen et al. 2013). Thus, more studies on the multiscale relationship of the summer monsoon with corridor convection are warranted to shed further light on extreme rains, which may be valuable for the prediction of severe flooding and climate anomalies.

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