Heavy Rainfall Associated with Double Low-Level Jets over Southern China. Part II: Convection Initiation

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ABSTRACT

Heavy rainfall that occurred at the south coast of China on 10–11 May 2014 was associated with a synoptic-system-related low-level jet (SLLJ) and a boundary layer jet (BLJ). To clarify the role of the double low-level jets in convection initiation (CI), we perform convective-permitting simulations using a nonhydrostatic mesoscale model. The simulations reproduce the occurrence location and mesoscale evolution of new convective cells as well as their small-scale wavelike structures at the elevated layers, which are generally consistent with radar observations despite some differences in their orientation. The nighttime BLJ over the northern South China Sea strengthens the convergence at ~950 hPa near the coast where the BLJ’s northern terminus reaches the coastal terrain. Meanwhile, the SLLJ to the south of the inland cold front provides divergence at ~700 hPa near the SLLJ’s entrance region. Such low-level convergence and mid-level divergence collectively produce strong mesoscale lifting for CI at the coast. In addition to the enhanced mesoscale lifting, the double low-level jets also provide favorable conditions for the superimposed small-scale disturbances that can serve as effective moistening mechanisms of the lower troposphere during CI. In a sensitivity experiment with coastal terrain removed, CI still occurs near the coast but is delayed and weaker compared to the control run. This latter experiment suggests that double low-level jets and their coupling indeed exert key effects on CI, while the BLJ colliding with terrain may enhance coastal convergence for amplifying CI. These findings provide new insights into the occurrence of coastal heavy rainfall in the warm sector far ahead of the fronts.

1. Introduction

The close relationship between low-level jets (LLJs) and heavy rainfall has been widely documented in previous studies (e.g., Stensrud 1996; Means 1952; Chen and Yu 1988; Rasmussen and Houze 2016). Generally, the LLJs provide favorable thermodynamic conditions for heavy rainfall (Astling et al. 1985; Tuttle and Davis 2006; Trier and Parsons 1993). The LLJs not only transport warm moist air and produce convergence at their termini to destabilize the environment (Trier and Parsons 1993; Higgins et al. 1997; Trier et al. 2006), but also can produce shear instability with strong wind shear (Mastrantonio et al. 1976; Sun and Zhai 1980). Some LLJs exhibit a diurnal cycle with a maximum at night and are closely associated with nocturnal heavy rainfall (Bonner 1968). For instance, the nocturnal LLJ in the southwesterly monsoon over southern China may enhance low-level convergence, moisture transport, and convective instability triggering several new MCSs at its northern terminus (G. Chen et al. 2017).

Convection initiation (CI) is one of the most difficult aspects of heavy rainfall to understand and predict because of its dependence on multiscale atmospheric processes (Trier 2003). LLJs may play a key role in CI through a combination of thermodynamic and kinematic factors, which is important for understanding their close relation with heavy-rain events (Marsham et al. 2011; Moncrieff and Liu 1999; Gebauer 2017). In particular, the LLJ’s high-$\theta_v$ air encounters boundaries (fronts, coastlines, and gust front) where enhanced lifting may help to initiate convection (Trier et al. 2017). Differential moisture and temperature advection by LLJs promotes a destabilization of the environment (Blackadar 1957). Low-level shear associated with an LLJ necessarily benefits deep lifting and convection initiation based on a theory of squall lines [the Rotunno–Klemp–Weisman (RKW) theory; Rotunno et al. (1988)].
In addition to LLJs, other factors also have a significant impact on CI such as topographic and land surface effects (Anthes et al. 1982; Lanicci et al. 1987), frontal boundaries (Koch and Kocin 1991; Trier et al. 1991), gust fronts related to cold pools (Wilson and Schreiber 1986; Schumacher and Johnson 2005), and gravity waves (Weckwerth and Wakimoto 1992; Trexler and Koch 2000; Richardone and Manfrin 2003; Wilson and Roberts 2006), etc. For instance, when an LLJ in the boundary layer encounters terrain or the coast (differential surface friction), it thus produces lifting that favors the generation of convection (Luo et al. 2017; Chen et al. 2014). Frontal secondary circulations near cold fronts may also produce ascent over a mesoscale area, further favoring CI (Shapiro 1981; Trier et al. 1991), and its lower branch may enhance the LLJ to transport warm moist air (Chen et al. 1998). Deep convection often occurs along a gust front interacting with a dryline or LLJ (Carbone et al. 1990). Vertical motions arising from low-level gravity wave disturbances can initiate convective storms (Tepper 1950; Abdullah 1955). Wang et al. (2013) observed gravity waves generated by an LLJ over Oklahoma City with a scanning Doppler wind lidar. Therefore, the interesting question of how these key factors, along with LLJs, jointly affect CI deserves further analyses.

At the coast of southern China, heavy rainfall usually occurs in the warm sector of a synoptic-scale disturbance (a few hundred kilometers away from cold or quasi-stationary fronts) during the early summer rainy season (Luo 2017; Luo et al. 2017). Since the relevant CI locally is far from the obvious synoptic forcing, rainfall forecasting has been a challenge and has attracted wide attention (e.g., Wang et al. 2014; Wu and Luo 2016). It has been found that coastal terrain, the land–sea contrast, cold pools left behind by previous day convection and boundary layer jets, are important for CI near the coast (Luo et al. 2017; Wu and Luo 2016; Wang et al. 2014). More interestingly, the boundary layer jet (BLJ; below 1 km) and the synoptic-weather-related jet (SLJJ; within 1–4 km) may coexist near the south coast of China, with different characteristics and formation mechanisms (Du et al. 2014; Du and Chen 2018, hereafter Part I). Part I found that the BLJ and SLLJ are closely related to the two rainbands in the coastal warm sector and in the inland frontal zone, respectively. However, how the double LLJs cooperate with each other and are influenced by the terrain, land–sea contrast, cold pools and gravity waves to affect CI is not well understood.

In this study, we conducted a mesoscale modeling study to examine the same case of heavy rainfall as in Part I, but with emphasis on the warm-sector rainfall at the coast of southern China on 10–11 May 2014. We focus on the mesoscale effects of the double LLJs in vertical and their small-scale effects (relating to wave-like disturbances and boundaries) on CI mechanisms. Section 2 briefly reviews the convective activities associated with the observed heavy rainfall and supplements Part I with more emphasis on the CI process. Section 3 describes the model configurations used in this study. The comparison between simulations and observations is also conducted in section 3. Section 4 presents the results from high-resolution simulations and further reveals the effect of the BLJ and SLLJ on the coastal CI. The sensitivity experiment with respect to topography is analyzed in section 5. Finally, section 6 includes discussions and summarizes our results.

2. Convective systems observed during heavy rainfall event at coast

To show the observed rainfall, the Climate Prediction Center morphing technique (CMORPH) rainfall data (Joyce et al. 2004) at high spatial (8 km) and temporal (30 min) resolutions was utilized in this study as in Part I. On 10–11 May 2014, there were two rainbands coexisting over southern China: inland rainfall and coastal rainfall (Figs. 1a–d). The inland rainband was oriented northeast–southwest and was collocated with the cold front (Fig. 2 in Part I). The coastal rainfall instead occurred in the warm-sector area several hundred kilometers south of the cold front. The coastal rainfall appeared in the afternoon hours as a result of isolated convection but it weakened toward evening (Figs. 1a–d in Part I). Coastal convection was reinitiated late at night and in the early morning (1800–2000 UTC) near Yangjiang, China (21.8°N, 111.9°E, cross in Figs. 1a–b). Unlike the afternoon convection, this nocturnal convection moved eastward slowly producing significant morning rainfall along the coastline (Figs. 1b–d). The present study focuses on the formation of the coastal warm-sector rainfall from 1800 UTC 10 May to 0000 UTC 11 May (Figs. 1a–d).

To examine the coastal convection in detail, the evolution of the radar composite reflectivity (mosaic) derived from operational weather radars is shown in Fig. 2. At 1200 UTC 10 May 2014, a mesoscale convective system (MCS, marked as “MCS-A”) was located to the east of Yangjiang (Fig. 2a), which was left over from the daytime. The MCS-A quickly weakened and moved eastward during 1200–1600 UTC (Figs. 2a–d). There was no radar composite reflectivity with greater than 40 dBZ near Yangjiang at 1600 UTC. Around 1700–1800 UTC, five banded new convective cells with southwest–northeast orientation began to develop near Yangjiang and exhibited a wavelike cluster with wavelength around...
FIG. 1. Hourly precipitation (mm) from (a)–(d) CMORPH and (e)–(h) WRF at (a),(c) 1800; (b),(f) 2000; and (c),(g) 2200 UTC 10 May 2014; and (d),(h) 0000 UTC 11 May 2014. The black cross indicates Yangjiang. The black line in (e)–(h) is used to define the cross section in Fig. 11.
FIG. 2. Radar composite reflectivity (dBZ) derived from operational weather radars in Guangdong province at (a) 1200, (b) 1400, (c) 1600, (d) 1700, (e) 1800, (f) 1900, (g) 2000, and (h) 2100 UTC 10 May 2014. The red ellipses indicate the location of the CI and the movement of MCS-B.
20 km (Fig. 2e). At the same time, the radar composite reflectivity became greater than 40 dBZ, which was regarded as a criterion of the CI. The convection cluster was further organized as MCS-B and moved eastward slowly during 1900–2100 UTC (Figs. 2f–h). The newly born MCS-B, which is the focus of this study, is responsible for producing intense coastal rainfall in the morning (Figs. 2f–h). Such nocturnal CI like MCS-B at the coast is one of the typical features that generate heavy rainfall in southern China (e.g., Luo et al. 2017).

The vertical cross sections of radar reflectivity along the white line in Fig. 2e are plotted in Fig. 3. Although the radar echo remained obvious below 2 km from MCS-A at 1700 UTC (Fig. 3a), it intensified significantly at 3–4 km aloft at 1730 UTC (indicated by red circle in Fig. 3b) and exhibited wavelike features. Such new convective cells at an elevated layer further developed and became deep convection (MCS-B) during 1800–1830 UTC (Figs. 3c,d). Radar observations thus suggest that the initiation of MCS-B is characterized by an obvious signature of the elevated convection.

We also note that a cold pool caused by MCS-A existed near the east of Yangjiang at 1200 UTC (Fig. 4a), but it declined to a much weaker temperature perturbation at 1800 UTC (Fig. 4b). The relevant gusty wind (2–3 m s$^{-1}$) was quite weak (Figs. 4c.d). Hourly evolution of surface temperature and winds also confirmed that the cold pool and associated surface boundary gradually became weaker from 1200 to 1800 UTC (not shown). The effect of the cold pool due to MCS-A on the CI of MCS-B seems to be minor. Therefore, it is worth further examining the mechanisms of the nocturnal elevated CI associated with MCS-B using a high-resolution model. As indicated in Part I, we will focus more here on the roles of the double LLJs and their effects combined with other forcings.

3. Model configuration and verification

To simulate the heavy-rainfall event, the Advanced Research version of the Weather Research and Forecasting (WRF-ARW) Model, version 3 (Skamarock et al. 2005), was utilized in this study. The model was...
initialized at 1200 UTC 10 May with the initial and lateral boundary conditions from the National Centers for Environmental Prediction Final (FNL) Operational Global Analysis data. The simulations used a one-way nested grid setup with two domains (Fig. 5). The grid spacing in the outer domain (D01) was 12 km, and that in the inner domain (D02) was 4 km. The vertical grid contained 51 levels with a stretched grid that involves finer grid spacing in the boundary layer. The pressure of the model top was 50 hPa, and the gravity wave-absorbing layer (5 km) was used near the model top. The Kain–Fritsch convection parameterization (Kain 2004) was applied in the 12-km mesh outer domain but not applied in the 4-km mesh inner domain. The other physical parameterizations used include the YSU boundary layer scheme (Hong et al. 2006), the Thompson microphysics scheme (Thompson et al. 2008), the RRTMG longwave and shortwave radiation scheme, the revised MM5 Monin–Obukhov surface layer scheme (Jiménez et al. 2012), and the unified Noah land surface-model scheme (Livneh et al. 2011). We also checked the sensitivity of the simulated convection to the initialization time regarding the forecast accuracy of CI. The simulation initialized at 1200 UTC 10 May outperforms those at 6 and 12 h earlier (figure not shown). The present simulation allows for a spinup of 6 h prior to CI, which is enough for the CI of MCS-B, but does not include the effect of MCS-A. The simulations initialized at 0600 or 0000 UTC 10 May might include the impact of cold pool driven by MCS-A, but they both failed to
capture the CI at the target location and times unfortunately. In the following sections, all discussion refers to the run initialized at 1200 UTC 10 May in the inner domain with the 4-km mesh.

First, we checked the model performance through comparing Figs. 1a–d from CMORPH with Figs. 1e–h from the WRF simulation (control run). It is shown that the northeast–southwest-oriented inland frontal rainband can be captured well. Both the location and movement speed of the frontal rainband are in good agreement with the CMORPH observations. The control run also reproduces well CI at the coast near Yangjiang at 2000 UTC. The location and time of CI are thus consistent with the observed, though the simulated rain rate is relatively weak. The simulated and observed radar reflectivity further confirmed that the simulated CI of MCS B occurs at similar location and timing (1–2 h delayed) in the mesoscale aspect indicated by red circles in Figs. 2e,f and 6b,c. Furthermore, the model can also capture some wavelike features for the coastal CI such as the wavelength, location, and occurrence time (Fig. 6) compared to the observation (Fig. 2). As shown Fig. 7, The convection generated at around 700 hPa aloft at 1900 UTC and further developed to deep convection, which is consistent with the observation (Fig. 3) Note that some small-scale (convective scale) errors exist in the mesoscale model simulation in which, for instance, the orientation of WRF-simulated wavelike convection zone (south–north) are somewhat different from the observations (southwest–northeast) (Figs. 2 and 6). Previous studies documented that the details of such local convection and related gravity waves are difficult to be reproduced well in the mesoscale model (e.g., Zhang et al. 2003). Nevertheless, simulations can generally reproduce the mesoscale evolution of new convective cells as well as their small-scale wavelike structures at the elevated layers.

Second, we examined the observed and simulated atmospheric conditions. Figures 8c and 8d show the evolution of the skew $T$–log$p$ diagram at the Yangjiang sounding station from the control run at 1800 UTC 10 May and 0000 UTC 11 May 2014. Compared with Figs. 8a and 8b [or Figs. 8c,d in Luo et al. (2017)], the simulated soundings at Yangjiang station exhibit similar
patterns and changes. It was relatively humid at low levels (below 700 hPa) but relatively dry at higher levels at 1800 UTC. The air at around 700 hPa became very moist and nearly saturated at 0000 UTC 11 May. Similar to the observation, the simulated low-level winds were veering with height, with two peaks of wind speed at about 900 and 700 hPa, respectively. In terms of convective instability, the simulated convective available potential energy (CAPE) for air originating near the surface increases from 1319 to 2835 J kg\(^{-1}\) during the 6-h period. The CAPE behavior agrees with the observations where it increased from 672 to 2511 J kg\(^{-1}\), suggesting a nocturnal convective instability. In addition to the near-surface high CAPE, there was an elevated high CAPE zone at around 800–700 hPa accompanying a low CIN zone during the CI hours.

Fig. 6. Radar composite reflectivity (dBZ) from the WRF simulation at (a) 1800, (b) 1900, and (c) 2000 UTC 10 May 2014.

Fig. 7. Vertical cross sections of simulated radar reflectivity (shading, dBZ) and vertical motion (contour, cm s\(^{-1}\), solid lines are updrafts, dash lines are downdrafts) along the black line in Fig. 1 at (a) 1800, (b) 1900, (c) 2000, and (d) 2100 UTC 10 May 2014.
providing favorable conditions for the growth of elevated convection, which will be further discussed in section 4b. Both observation and simulation show a shallow and weak cold layer near surface at 1800 UTC (Fig. 8). As mentioned in section 2, a cold pool remained near the east of Yangjiang at 1200 UTC but became much weaker during 1200–1800 UTC (Fig. 4) with the decayed MCS-A. In the simulation, the relevant cold anomaly existed at 1200 UTC from the initial condition of temperature and it also weakened but because of the absence of MCS-A (not shown).

Based on the above evaluation, the control run has a reasonable skill in simulating the CI particularly at the mesoscale. The model also captures the nocturnal elevated convective cells during CI and associated atmospheric conditions that are similar to the observations. Although some convective-scale errors exist, they are within an acceptable error degree. Therefore, it is reasonable to further investigate the mechanisms of CI using the output of the control run. In the present study, we will focus on the mesoscale features and mechanisms of CI, while we deemphasize the formation of convective-scale features.
4. Effect of double LLJs on coastal CI

a. Horizontal structure of double LLJs

Analyses of ensemble forecasts in Part I found that the BLJ and SLLJ coexist over southern China. The present study further identifies the structure and development of these double LLJs in our WRF simulations. Figure 9 shows the evolution of the horizontal distribution of winds at different levels. At 950 hPa (Figs. 9g–i), there are two enhanced wind cores over the ocean: one is over the Beibu Gulf (around 20°N, 108°E) and the other is over the northern South China Sea (SCS) (around 20.5°N, 112°E). The latter core is featured by a strong southerly wind, with speed increasing during 1400–1800 UTC 10 May and reaching a maximum of more than 18 m s⁻¹ (Fig. 9i).
The wind speed at 950 hPa is larger than that at the higher levels of 850 and 700 hPa (Fig. 9), which corresponds to a marine BLJ.

At 850–700-hPa levels (Figs. 9a–f), the strong southwesterly winds are accompanied by the northeast–southwest-oriented cold front and are situated to the south of the cold front. The strong wind area moves southward along with the southward-moving cold front. The wind at 850 hPa (>18 m s⁻¹) near the front (Figs. 9d–f) is generally stronger than that at 700 hPa (~14 m s⁻¹, Figs. 9a–c). Over the area west of Guangdong (around 23°N, 112°E), the wind at 700 hPa (~16 m s⁻¹) is slightly stronger than at 850 hPa (~14 m s⁻¹). Such a strong southwesterly wind at 850–700 hPa, associated with the cold front, belongs to a kind of SLLJ (Du et al. 2014).

The simulated BLJ and SLLJ are both consistent with...
the results in Part I and FNL reanalysis (Figs. 9c,f,i), suggesting a good performance in the WRF modeling. We can use the high-resolution WRF output to assess the fine structures and evolution of double LLJs and their impacts.

Usually, the entrance and exit of the LLJs are related to the horizontal divergence and convergence, respectively (Hastenrath 1985; Chen et al. 2014), although they also depend on the jet stream configurations (Keyser and Shapiro 1986). To better illustrate the entrance and exit of LLJs in this case, the envelope of 14 m s\(^{-1}\) is regarded as the core of LLJs (black curves in Fig. 10). With considering the direction of LLJs, the entrance (exit) of LLJs is located upstream (downstream)

Fig. 11. Vertical cross sections of wind speed (shading, m s\(^{-1}\)), equivalent potential temperature (K), and flow vectors (black vectors, wind in the y direction and 100 times vertical velocity) along the black line in Fig. 1 at 1600, 1700, 1800, 1900, 2000, and 2100 UTC 10 May 2014. The blue contours in (c) indicate wind speeds (14 and 18 m s\(^{-1}\)) from the FNL reanalysis.
of LLJ cores (indicated by red and orange circles in Fig. 10). The distribution of divergence/convergence at different levels is also shown in Fig. 10. There is a convergence zone at 950 hPa near the coast (Fig. 10a) due to the frictional contrast between land and ocean and coastal terrain (X. Chen et al. 2017). The convergence is relatively strong at the exit of the BLJ (Fig. 10a). Meanwhile, there is mesoscale divergence at 850–700 hPa at the entrance of the SLLJ (Figs. 10b,c, indicated by red circles). The convergence/divergence patterns from WRF and FNL reanalysis are quite similar (Figs. 10a–c and 10d–f). Yangjiang, the location of CI indicated by a black star, is situated in both the exit regions of BLJ (orange circles) and the entrance regions of SLLJ (red circles). Based on the analysis above, the coastal CI near Yangjiang may be affected by both the BLJ and SLLJ.

Using a similar method of ensemble-based sensitivity analysis in Part I, we further calculated the correlation coefficient between coastal warm-sector rainfall (Pc) during 0000–0600 UTC 11 May and full winds at 1800 UTC 10 May when the CI occurred (not shown). It is found that negative correlation at 925 hPa occurs near the coasts, whereas the negative (positive) correlation is at northeast (southwest) side of SLLJ. It might further suggest that stronger convergence (divergence) at the exit (entrance) of BLJ (SLLJ) are favorable for the rainfall at its early stage (or convection initiation).

b. Vertical structures of the double LLJs

We further examined the vertical structure of the BLJ/SLLJ and their relation with convection. Since CI occurs near Yangjiang (Figs. 1f and 2e), the vertical cross section indicated by a black line in Fig. 1 is chosen to study the evolution of the vertical structure of winds and equivalent potential temperature \( \theta_e \) (Fig. 11). The vertical cross section of wind speed clearly shows the double LLJs and their development. The FNL reanalysis validate the double LLJs (Fig. 11c). The BLJ mainly occurs below 900 hPa over the ocean, while the SLLJ mainly appears within 650–850 hPa over land and near the coast. Both the BLJ and SLLJ develop from midnight to early morning (1600–2100 UTC), with the maximum wind speed reaching up to 20 and 18 m s\(^{-1}\), respectively. Thus, the BLJ’s exit is well collocated with the SLLJ’s entrance near the coastal area. Interestingly, the double LLJs coexist near the coast.

To explicitly exhibit the impacts of double LLJs, the vertical structure of horizontal divergence and vertical motion are further shown in Fig. 12. Strong convergence occurs in the boundary layer near the BLJ’s exit whereas strong divergence occurs at 800–700 hPa near the SLLJ’s entrance. The collocation of the low-level convergence and midlevel divergence produces upward motion near the coast, which provides favorable conditions for CI. Previous studies documented that the coupling of the LLJ and the upper-tropospheric jet favors the development of severe convective storms (Uccellini and Johnson 1979). The present study further suggests that the coupling of the two types of LLJs can also produce lifting and favor convection when double LLJs coexist and cooperate to produce rising motion.

The vertical cross section of \( \theta_e \) shows two large \( \theta_e \) gradient zones coinciding with double LLJs that transport warm-moist air to the north (Figs. 11–12). At 1900 UTC, some small-scale disturbances of vertical motion and \( \theta_e \) appear near the entrance of the SLLJ (around 21°–22°N, Fig. 11d). The disturbances exhibit wavelike features with a wavelength of \( \sim 20 \) km that is similar to the wavelength of the convection in the cluster seen in the radar observations (Fig. 2). The amplitude of the wave perturbations becomes much stronger at 2000–2100 UTC (Figs. 11e,f).

To illustrate the moist processes, the vertical cross sections of cloud water mixing ratio is shown in Fig. 13. It is found that the cloud at low levels (\( \sim 950 \) hPa) is first produced near the coastal terrain and increased with
time, probably because of the effect of the coast terrain on the BLJ. At 1900 UTC when small-scale disturbances appear, some clouds appear at middle levels (~700 hPa, Fig. 13d) and become more and more significant in the following hours (Figs. 13e,f). The midlevel cloud is characterized by a wavelike structure with a wavelength of about 20 km, which coincides well with the perturbations of upward motion and $\theta_e$. The low-level cloud also occurs off the coast at 2000–2100 UTC and is also featured by wave structure. The levels of the cloud formation (950 and 700 hPa) are corresponding to two zones of high relative humidity (>90%, Fig. 13). The wavelike disturbances develop during 1900–2100 UTC with an increasing amplitude (Figs. 11d–f). To better illustrate the growth of convection, eastward-moving vertical cross sections of cloud water mixing ratio by
tracking convection were plotted (not shown). We found that the vertical extent of convection largely increased after CI and finally reached the top of 300 hPa level. The shallow convection thus gradually developed to deep convection. The growing disturbances are strongly coupled with convection condensation. Their relationship with the midlevel moistening will be discussed in section 4c. These small-scale disturbances are probably related to convective instability (Ching et al. 2014) or the activities of gravity waves (Chen 1982), which is discussed in the conclusions and discussion.

The simulated new convective cells exhibit elevated features that are generally consistent with the radar observation as also shown in section 3. Next, we check the mesoscale environment of convective instability for the elevated convection. The vertical cross sections of CAPE, CIN, and vertical motions are thus analyzed as shown in Fig. 14. Both CAPE and CIN are large at near-surface layers below 950 hPa (Figs. 14a–f). The vertical motion is too weak to overcome the convective inhibition to initiate convection from the near-surface parcels. In contrast, another high CAPE zone around 900 J kg\(^{-1}\) occurs near 800–750 hPa (Figs. 14a–c). At this elevated layer, CIN also decreases to a very small value of 2–4 J kg\(^{-1}\) as shown in Figs. 14d–f. Meanwhile, strong disturbances of vertical motion develop at 800–750 hPa and produce sufficient lifting to overcome CIN to trigger convective cells aloft (Figs. 14h,i). Therefore, such CAPE, CIN, and lifting conditions are thus favorable for the growth of elevated convection in the numerical simulations. The simulated and observed sounding at 1800 UTC (Figs. 8a,c) also confirmed the existence of convectively unstable layer at the elevated layer.

Fig. 14. Vertical cross sections of (a)–(c) CAPE (J kg\(^{-1}\)), (d)–(f) CIN (J kg\(^{-1}\)), and (g)–(i) vertical motion (m s\(^{-1}\)) and flow vectors (black vectors, wind in the y direction and 100 times vertical velocity) along the black line in Fig. 1 at (a),(d),(g) 1800; (b),(e),(h) 1900; and (c),(f),(i) 2000 UTC 10 May 2014.
c. The midlevel moistening associate with double LLJs

Around 1800 UTC when CI occurs, the water vapor mixing ratio at 700 hPa begins to increase rapidly over the region with collocated double LLJs (the blue box A in Fig. 5). Figure 15 shows that the midlevel moisture content reached a maximum of 10.2 g kg$^{-1}$ at 2000 UTC, which is 20% more than that at 1600 UTC (8.4 g kg$^{-1}$). The midlevel moistening leads to a high relative humidity for condensation and reduces dry air entrainment, which is favorable for the growth of moist convection (Peters and Bretherton 2006). Meanwhile, the variance of small-scale vertical motion over region A starts to increase at 1800 UTC and becomes largest at 2000 UTC, which suggests the development of wavelike disturbances. In contrast, in the adjacent area offshore (region B), the nocturnal increase of moisture has a much smaller amplitude than that in region A. More interestingly, the variance of vertical motion is virtually absent in region B where the wavelike disturbances do not develop.

To show the impacts of vertical motions on convective preconditioning, we estimate the vertical transports of water vapor. Figure 16 shows that the total vertical transport of water vapor ($\langle \tilde{q} \tilde{w} \rangle$) is mainly negative before midnight (1600 UTC) and becomes positive at late night and early morning. The vertical transport by small-scale wavelike disturbances ($\langle q' \tilde{w} \rangle$) becomes evident at late night and early morning ($q'$ and $w'$ are deviations from 650-hPa averages over blue box A in Fig. 5). The small-scale disturbances might be attributed to local absolute instability due to the midlevel moistening (Fig. 15). The vertical transports due to the small-scale disturbances further partly contributes to the increase of water vapor at middle levels, while the other part of midlevel moistening is attributed to the mean vertical vapor transport by mesoscale lifting. Such an effect may be more evident in the coastal area than in the offshore areas. It suggests that the joint effect of double LLJs and small-scale disturbance is important for midlevel moistening through both mesoscale lifting and small-scale disturbances. Although the 700-hPa divergence still exists in region B, the low-level convergence is much weaker in region B compared to region A (Figs. 10a and 10c). The weakened midlevel moistening in region B seems to occur in the background flow far away from the collocated zone of double LLJs (BLJ exit and SLLJ entrance). Therefore, the collocated double LLJs play a significant role in CI processes through enhanced low-level ascent and midlevel moistening.

5. Sensitivity experiment on topography

Previous studies have documented that topography or land–sea contrast may play a role in CI (Weckwerth and Parsons 2006). To explore the relative contribution of coastal terrain on CI, a sensitivity experiment with the topography replaced by ocean in the black box in Fig. 17a (NOTER run) is performed. Figure 17 shows that the NOTER run can capture the frontal rainfall, similar to the control run. The rainfall at the coast can be generated as well, but the rain rate is weaker than that in the control run. The rainfall near the coast occurs around 2200 UTC, which is later than observed. The rainy area moves to the east similar to the control run. These results suggest that the coastal terrain indeed has an impact on CI, particularly for the amplitude of convection. However, the terrain is not a critical factor since CI near coast still occurred in the NOTER run.
Figure 18 shows the spatial distribution of divergence and convergence at different levels in the NOTER run. Compared to the control run (Fig. 10), the divergence at 700 hPa near the coast is similar in the NOTER run, although the convergence at 950 hPa near the coast is much weaker (Fig. 18a). The location and development of the SLLJ in the NOTER run is similar to the control run (Figs. 19 and 11), resulting in a similar divergence at

**Figure 17.** Hourly precipitation (mm) from the WRF sensitivity experiment (NOTER) at (a) 1800, (b) 2000, (c) 2200 UTC 10 May 2014, and (d) 0000 UTC 11 May 2014. The black box in (a) indicates the region for topography replaced by ocean.
700 hPa (Figs. 18b,c and 10b,c). Without terrain, the BLJ can extend more to the north (Figs. 19 and 11), which means weaker convergence near the coast (Figs. 18a and 10a). As a result, the existence of the double LLJ can also produce convergence at low levels and divergence at high levels but the convergence at low levels is weaker because of the absence of coastal terrain. Thus, the low-level convergence and midlevel divergence patterns are associated with double LLJs that are not optimally superimposed to produce vertical lifting. Such convergence/divergence patterns produce upward vertical motion near the coast, similar to the control run, but are weaker (Fig. 20). Figure 20 shows the cross section of water vapor mixing ratio in the NOTER run. Compared to the control run (Fig. 13), the NOTER run can generate clouds at 700 hPa, but it cannot generate cloud near the coastal terrain because of the absence of the terrain. Therefore, the presence of SLLJ is important of

Fig. 19. As in Fig. 11, but from the sensitivity experiment (NOTER run).
the CI occurrence at its corresponding (elevated) layer. The growth of convective cells is clearly weakened and delayed in NOTER, which suggests that the convergence of BLJ plays a key role in enhancing convective updraft (CI amplification) as it supports the convectively coupled disturbances with strong low-level lifting. SLLJ (BLJ) may have a relatively large contribution in the occurrence (amplification) of CI. The midlevel cloud is also characterized by a wavelike structure in the NOTER run (Figs. 20e,f), but it is less significant than in the control run (Figs. 13e,f), which might suggest the impact of terrain on the generation of the small-scale disturbance.

6. Conclusions and discussion
In this study, we use the Advanced Research version of the Weather Research and Forecasting (WRF-ARW)
Model with 4-km horizontal grid spacing to investigate the mechanisms of convective initiation (CI) near the south coast of China in a heavy rainfall event (10–11 May 2014). The convection was newly generated at Yangjiang, China, located hundreds of kilometers to the south of an inland cold front. During the daytime, the BLJ over the northern South China Sea is relatively weak, whereas the SLLJ associated with the inland cold front is far away inland from the south coast. During the nighttime, with the development of the BLJ, the convergence at ~950 hPa strengthens at the exit of the BLJ. At the same time, the SLLJ to the south of the inland cold front provides divergence at ~700 hPa near the SLLJ’s entrance with the southward approach of the cold front. Such low-level convergence and midlevel divergence cooperatively produce strong mesoscale lifting for CI near Yangjiang.

1) Our analysis of the WRF simulation of the event indicates that double LLJs exert a significant influence on CI through mesoscale lifting. The double LLJs coexist near the coast; these jets involve the boundary layer jet (BLJ, below 1 km) and the synoptic-weather-system-related LLJ (SLLJ, at 850–700 hPa). During the daytime, the BLJ over the northern South China Sea is relatively weak, whereas the SLLJ associated with the inland cold front is far away inland from the south coast. During the nighttime, with the development of the BLJ, the convergence at ~950 hPa strengthens at the exit of the BLJ. At the same time, the SLLJ to the south of the inland cold front provides divergence at ~700 hPa near the SLLJ’s entrance with the southward approach of the cold front. Such low-level convergence and midlevel divergence cooperatively produce strong mesoscale lifting for CI near Yangjiang.

2) The moisture at midlevels rapidly increases during CI, which is favorable for the growth of moist convection. It is found that vertical transport of water vapor driven by mesoscale lifting and the vertical transports due to small-scale wavelike disturbances become evident at late night when CI occurs. Both the enhanced mesoscale lifting related to double LLJs and small-scale disturbances associated with convection serve as such effective moistening.

3) A sensitivity experiment on topography further reveals that coastal terrain is important but is not a critical factor for CI. Without coastal terrain, CI can occur but becomes weaker and delayed. The existence of the double LLJ can also produce convergence at low levels and divergence at high levels but the convergence at low levels is weaker due to the absence of coastal terrain and the northward movement of BLJ. The upward vertical motion near the coast is still produced but more weakly driven by such a convergence/divergence pattern. Therefore, the presence of SLLJ is important to the CI occurrence at its corresponding layer, whereas the coastal terrain and BLJ indeed has impact on CI particularly for the amplitude of convection (CI amplification) by enhancing coastal convergence.

Both radar observation and model simulations show that the newly generated convective cells are elevated and exhibit distinct wavelike structures with a wavelength of ~20 km, although they have some differences on orientation. The small-scale wavelike disturbances might be associated with convective instability (Ching et al. 2014) or gravity waves (Chen 1982). Midlevel moist unstable layers might develop the convectively induced secondary circulations (CISCs) that exhibit wavelike structure. Ching et al. (2014) suggested that models may generate unrealistic modeled CISCs (M-CISCs) due to
the model resolution at gray zone. To check whether the wavelike structures are due to M-CISCs, we further downscaled the simulation to 1-km grid spacing and confirmed that the wavelike disturbances with a wavelength of ~20 km still exist (not shown). It thus suggests that the wavelike disturbances are well resolved.

LLJs provide a favorable condition for the instability of gravity waves because of shear instability (e.g., Uccellini and Koch 1987; Stobie et al. 1983; Pecnick and Young 1984; Parsons and Hobbs 1983; Koch 1979; Koch and Dorian 1988) and geostrophic adjustment (Kaplan and Paine 1977; Van Tuyl and Young 1982; Uccellini et al. 1984; Koch and Dorian 1988). The local gradient Richardson number smaller than 0.25 indicates shear instability. Two small gradient Richardson number layers (~975 and 850 hPa) develop during CI (not shown), which are associated with double LLJs. A Lagrangian Rossby number, which describes the intensity of unbalance flow and implies geostrophic adjustment, increases rapidly at 950, 850, and 700 hPa and exceeds 0.5 when the CI occurred with the development of LLJs (not shown). Geostrophic adjustment occurs when the mass and momentum fields are unbalanced, which may develop gravity waves by which the atmosphere acts to recover balanced conditions (Blumen 1972). Further studies are needed to quantify the relative roles of various physical processes in producing these disturbances especially in the presence of LLJs.

This study is the second part of a two-part series on heavy rainfall over southern China associated with double LLJs. In Part I, we used the ensemble-based analysis method to identify the BLJ and SLLJ as key factors for the warm-sector heavy rainfall and frontal heavy rainfall, respectively. In the present study (Part II), we further examine the high-resolution simulation and find that the coupling of double LLJs (BLJ and SLLJ) plays a significant role on CI of warm-sector heavy rainfall through mesoscale lifting (convergence/divergence) and midlevel moistening. In the future, more cases near the coast of southern China will be studied to clarify the effect of double LLJs on CI and their roles, which may vary in the other atmospheric conditions. The combined roles of LLJs with other factors (e.g., cold pool, terrain, mesoscale vortex, etc.) also shall be examined. In addition to the mesoscale features, the coupling between convection and wavelike disturbances in the presence of LLJs is a very interesting issue that needs to be addressed in future work.

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