Diurnal Variations of Southerly Monsoon Surge and Their Impacts on East Asian Summer Rainfall

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ABSTRACT: Monsoon southerlies can be particularly active for days and produce substantial rainfall over East Asia. These multiday episodes of southerly monsoon surge may exhibit distinct diurnal variations due to regional forcings under given large-scale conditions. This study categorizes the southerly surges into two types with different wind diurnal variations to clarify their influence on rainfall over East Asia. In the summers of 1998–2019, there are 63 episodes of southerly surges with large wind diurnal cycles and 55 episodes with small diurnal cycles. The first type of southerly surges usually occurs with anomalous low-level warming over southeastern China related to the westward extension of the western Pacific subtropical high. The second type of southerly surges instead occurs with anomalous cooling due to the deepened midlatitude trough. They thus represent the different mechanisms downscaling from large-scale conditions to regional diurnal forcings. After the onset of the first type, the intensified monsoon southerlies at night lead to the northward displacement of large-scale ascent and northward water vapor transport with warm moist energy. The monsoon rainband tends to move to the north of 35°N with a robust response in precipitation systems, especially in the meso- α -scale rain events from midnight to morning. As a comparison, the rainband stays at 30°–35°N after the onset of the second type when the strengthened large-scale ascent and water vapor convergence are located relatively south. These differences between the two types of southerly monsoon surges highlight that the multiday large-scale conditions interact with subdaily regional forcings and greatly regulate the detailed evolution of summer rainband over East Asia.

KEYWORDS: Diurnal effects; Monsoons; Rainfall; Moisture/moisture budget

1. Introduction

East Asian summer monsoon (EASM) features strong low-level southerlies with a rainband extending from eastern China to Japan (Fig. 1a). The monsoon southerlies tend to march northward with the rainband in a stepwise manner from June to August (JJA) (Fig. 1c; Wang and LinHo 2002; Ding and Wang 2008; Zeng et al. 2012). The seasonal march of the EASM is closely related to the prevailing low-level southerlies at synoptic and subseasonal time scales (e.g., Ding and Wang 2008). In particular, strong southerlies may last for days or a few weeks, which is named as southerly monsoon surge (Zhang et al. 2002; Fang et al. 2007; Wang et al. 2009; Liu et al. 2017). The episodes of southerly monsoon surge are closely associated with the short-term variations of atmospheric circulations, such as the activities of the western Pacific subtropical high (WPSH), monsoon lows/troughs, and intraseasonal oscillations (ISOs) (Zhang et al. 2002; Wang et al. 2009; Liu et al. 2017).

They may also couple with the mei-yu front or tropical cyclones to provide favorable conditions for heavy rainfalls over East Asia (Liu et al. 2008; Ge et al. 2010; D. Zhao et al. 2021). Strong low-level southerlies during monsoon surge can act as effective conveyors of moisture and energy, thereby producing substantial rainfall at their northern terminus (e.g., Liu et al. 2017; Huang et al. 2019). The influence of short-term monsoon surge on summer rainfall is also observed over South America, South Asia, and northern Australia (Jones and Carvalho 2002; Goswami et al. 2003; Shrestha et al. 2012; Evans et al. 2014; Fujinami et al. 2014, 2017; Pai et al. 2016). Further understanding of the activities of southerly monsoon surge and their impacts help to improve our knowledge on the monsoon system and extreme precipitation.

Besides multiday variability, low-level winds over East Asia also exhibit pronounced diurnal variations with an accelerated southerly component at night (Fig. 1b; G. Chen et al. 2009; Yu et al. 2009; H. Chen et al. 2010; Yuan et al. 2013; Xue et al. 2018), whose amplitudes maximize in the summer season (Fig. 1d; Wang et al. 2013; Chen et al. 2021). Wind diurnal variations are usually regulated by dynamic and thermal forcings (Blackadar 1957; Holton 1967; Du and Rotunno 2014; Shapiro et al. 2016), which can be affected by large-scale conditions (Parish and Oolman 2010; Huang and Chan 2011; Sun et al. 2018). In the EASM regions, the amplitudes of monsoon flows are regulated by the strength of daytime boundary layer heating that may vary among the active monsoon days (Chen et al. 2009; Xue et al. 2018; Chen 2020). Anomalous large or small diurnal amplitudes in monsoon

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FIG. 1. (a) JJA daily mean rainfall (shading) and 925-hPa winds (vectors). (b) JJA midnight–morning rainfall (shading, 0000–1100 LST) and the diurnal amplitude of 925-hPa winds at 0200 LST (vectors; pink vectors are for the meridional components above 0.8 m s^{-1}). (c) Latitude–time diagram of daily mean rainfall (shading) and 925-hPa meridional winds (contours, m s^{-1}). (d) Latitude–time diagram of midnight–morning rainfall (shading) and the diurnal amplitude of 925-hPa meridional winds at 0200 LST (contours, m s^{-1}). In (a) and (b), the gray dashed lines denote the elevations with an interval of 1000 m. The box (22.5° – 35°N , 110° – 130°E) denotes southeastern China. In (c) and (d), the zonal average is made over 110° – 130°E .

southerlies can occur successively for several days or more (Chen 2020). It indicates that the multiday episodes of southerly monsoon surge may occur with the continuous subdaily regional forcings under given large-scale conditions. Further studies of the diurnal variations of southerly surge episode may help to clarify the interaction between large-scale conditions and regional forcings at diurnal and longer time scales.

It has been recognized that wind diurnal variations can greatly influence the rainfall diurnal variations with rainfall maximum at late night or early morning (Ohsawa et al. 2001; Hirose and Nakamura 2005; Tuttle and Davis 2006; G. Chen et al. 2009, 2017; Trier et al. 2014; X. Chen et al. 2016, 2017; Fujinami et al. 2017). The enhanced southerlies at night can convey abundant moisture northward to the frontal zone, where the convectively unstable environment is favorable for the growth of convective systems over East Asia

(Yamada et al. 2007; Yuan et al. 2010; Huang and Chan 2011; Sun and Zhang 2012; Fu et al. 2019; Shin et al. 2019). These processes are much more efficient in the active monsoon days with a large wind diurnal amplitude than in those with a small amplitude (Chen 2020). Meanwhile, the locations of maximum moisture convergence and precipitation are also distinct in the two types of active monsoon days, which is related to the interannual and interdecadal variations of the East Asian rainband (Guan et al. 2020; Chen et al. 2021). These studies suggested that the monsoon southerlies with different diurnal amplitudes may display differences in modulating the spatiotemporal patterns of moisture transport, atmospheric instability and thus precipitation. It calls for a detailed study of multiday surge episodes with different diurnal amplitudes to clarify the influence of multi-time-scale monsoon activities on rainfall patterns over East Asia.

Although the multiday episodes of southerly monsoon surge have been connected with the atmospheric circulations at subseasonal and synoptic time scales (Zhang et al. 2002; Wang et al. 2009; Liu et al. 2017), our knowledge on their relationship with the subdaily regional forcings is still lacking. This study aims to clarify the activities of the diurnal variations of monsoon surges and their impacts on rainfall patterns. To address this issue, an objective method is used to identify southerly monsoon surge episodes and classify them with the different diurnal amplitudes of low-level southerlies. Particular attention is given to the difference in the evolutions of large-scale conditions and associated diurnally varying processes that drive the rainfall patterns before and after the onsets of southerly surge episodes. The rest of this paper is organized as follows. Section 2 introduces the data used in this study and the method for identifying southerly surge episodes. Section 3 compares the characteristics and associated large-scale conditions of southerly surges with different diurnal variations. Section 4 examines the response of rainfall patterns to the different types of surge episodes. Section 5 shows the dynamic and thermodynamic processes that modulate precipitation systems during different types of surge episodes. Section 6 presents the conclusion and discussion.

2. Data and methods

a. Dataset used in this study

The Japanese 55-year Reanalysis (JRA-55) is used to describe the atmospheric conditions. It provides pressure-level data products with $1.25^\circ \times 1.25^\circ$ spatial grids and 6-hourly intervals (Kobayashi et al. 2015). JRA-55 is thought to depict faithfully the lower-tropospheric conditions with a vertical interval of 25 hPa below 750 hPa. This new-generation reanalysis has been widely used to analyze atmospheric conditions in summer over East Asia (e.g., Kudo et al. 2014; Horinouchi and Hayashi 2017; W. Li et al. 2021). JRA-55 is also shown to have a good quality in resolving the diurnal cycles of winds and other variables over East Asia, compared to the other reanalysis data (Chen et al. 2014a, 2021).

The merged microwave-infrared precipitation product from the Tropical Rainfall Measuring Mission (TRMM 3B42 v7) is used to analyze the activities of rainfall systems. TRMM offers a rain rate with a 3-hourly interval at a $0.25^\circ \times 0.25^\circ$ resolution since 1998 (Huffman et al. 2007). It presents well in reproducing diurnal variations of rainfall over East Asia, which is comparable to rain gauge measurements (Zhou et al. 2008; Shen et al. 2010). Besides, TRMM gives a good performance on reproducing the seasonal change of the rainband and individual rain episodes at a multiday time scale, which is of interest in this paper (e.g., He and Liu 2016; G. Chen et al. 2017; Guan et al. 2020).

All datasets used in this paper cover a 22-yr period in the summers (JJA) of 1998–2019. As the diurnal variations of active monsoon flows are usually regulated by daytime boundary layer heating (Chen et al. 2009; Xue et al. 2018; Chen 2020), the diurnal cycle is estimated from 1200 to 1100 LST of the next day. The four synoptic hours 1400, 2000, 0200, and 0800 LST (LST = UTC + 8 h) denote afternoon, evening, midnight, and

morning, respectively. The afternoon–evening hours refer to the time between 1200 and 2300 LST, while the midnight–morning hours refer to the time between 0000 and 1100 LST of the next day. Diurnal deviations for all variables at the synoptic hour are obtained by subtracting the daily mean. The anomalies are relative to the 22-yr climatological mean.

b. Definitions of southerly monsoon surges and their diurnal variations

Before identifying the southerly surge episodes, we estimate the activity of the EASM in each summer day at daily and diurnal time scales (Chen 2020). A key area is over southeastern China where the climate-mean diurnal variation is evident (the box in Fig. 1b). We obtain the anomalies of 925-hPa meridional winds averaged at 110° – 130° E, in terms of daily mean and diurnal deviation at 0200 LST, as shown by an example in Fig. 2. At the daily time scale (Fig. 2a), the strong monsoon day, named M(+) day, is defined when the following two criteria are met. First, a positive anomaly of daily mean southerly originates from the south of 25° N, ensuring that it represents the monsoon flows from the tropics. Second, the daily mean southerly anomaly can extend continuously in latitudes of at least 5° , indicating that the large-scale monsoon southerlies prevail over East Asia. The remaining days are considered as the weak monsoon days, named M(−) days. Similarly, at the diurnal time scale (Fig. 2b), the D(+) days denote the days with large diurnal amplitudes when the positive anomaly of southerly diurnal deviation covers at least 5° of latitude continuously over East Asia. The remaining days are considered as the days with small diurnal amplitudes, named D(−) days. We note that the results are insensitive to the threshold of continuous latitude in the definitions of M(+) and D(+).

An episode of southerly monsoon surge is identified when the same strong monsoon conditions can be sustained for at least 3 days in which 1-day intermittences are neglected (dark arrows in Fig. 2). Therefore, we can group the southerly surge episodes into two categories based on the anomalies of diurnal amplitudes of meridional winds. The southerly monsoon surges with large diurnal amplitudes are referred to as M(+)D(+), while those with small diurnal amplitudes are M(+)D(−). M(+)D(+) and M(+)D(−) are the focus of this paper considering the importance of strong monsoons. Similarly, the continuously weak monsoon conditions with large diurnal amplitudes are referred to as M(−)D(+), while those with small diurnal amplitudes are M(−)D(−) (light arrows in Fig. 2). The comparison among these groups helps us to interpret the various couplings of multiday monsoon activities and their diurnal variations.

3. Southerly monsoon surges and their associated large-scale atmospheric conditions

a. Climatology of the diurnal variation of southerly monsoon surges

Figure 3a shows the southerly surge episodes over southeastern China, with M(+)D(+) and M(+)D(−) as denoted by dark red and blue pixels, respectively. In the 22-yr summers, 63 M(+)D(+) and 55 M(+)D(−) are identified with an average of 5.4 surge episodes per year. The average duration

FIG. 2. (a) Latitude–time diagram of the anomaly of the daily mean 925-hPa meridional winds over 110°–130°E during July 2004. (b) As in (a), but for the anomaly of the diurnal amplitude at 0200 LST. The colors of circles denote the conditions of each day. The colors of arrows denote the conditions of each monsoon episode, which are the focus of this study.

of M(+)/D(+) is 6.1 days, longer than that of M(+)/D(-) (4.7 days). There are 10 extreme episodes that last for more than 10 days, with a maximum duration of 19 days. Eight of these episodes are M(+)/D(+) and the remaining two are M(+)/D(-). These extreme episodes usually occur with the active phases of the 30–60-day ISO from the tropics and the quasi-stationary frontal systems due to midlatitude westerly disturbances (not shown). The accumulated days of M(+)/D(+) and M(+)/D(-) are 385 days and 261 days, accounting for 19.0% and 12.9% of the summer days, respectively. The more activities of M(+)/D(+) than M(+)/D(-) agree with the previous studies in that active monsoon southerlies tend to occur with large diurnal variations (Chen et al. 2013; Chen 2020). The average daily mean wind speed of 925-hPa southerly during M(+)/D(+) is 4.7 m s^{-1} , which is comparable with that during M(+)/D(-) (4.3 m s^{-1}). However, the diurnal deviation of southerly at 0200 LST during M(+)/D(+) is 1.0 m s^{-1} , which is nearly twice that during M(+)/D(-) (0.53 m s^{-1}). The northern boundary of daily mean southerly anomaly is much more northward during M(+)/D(+) (44°N) than during M(+)/D(-) (38°N) (not shown), indicating that the northward march of monsoon southerlies is closely related to their nocturnal acceleration during surge episodes. As a comparison, the daily mean wind speed of southerly is 1.5 m s^{-1} during M(-)/D(+) with a diurnal deviation of 1.0 m s^{-1} , while it is 0.34 m s^{-1} during M(-)/D(-) with a diurnal deviation of 0.52 m s^{-1} . Therefore, the four groups are distinguishable from each other.

At the subseasonal time scale, M(-)/D(-) is dominant in early June when monsoon flows are weak (Fig. 3b). Surge episodes are active from mid-June to mid-July when about 54% of days are regarded as M(+)/D(+) or M(+)/D(-) (Fig. 3b). Specifically, M(+)/D(-) is active from mid- to late June and M(+)/D(+) from late June to mid-July. The active periods for

the two types of surge episodes correspond to the northward progress of the EASM (Fig. 1c). The transition from M(+)/D(-) to M(+)/D(+) in late June matches the intensification period of the diurnal variations of monsoon southerlies (Fig. 1d; Yuan et al. 2010), which may be supported by strong background flows and warm boundary layer (Xue et al. 2018; Chen 2020; Chen et al. 2021). These phenomena may reflect the subseasonal coupling of the monsoon flows and their diurnal variations from a new perspective. In late July, M(-)/D(+) becomes active due to the daily monsoon flows weakened by the WPSH and the large southerly diurnal variations maintained by strong solar heating (Chen et al. 2004; R. Zhao et al. 2021). In August, M(-)/D(-) becomes dominant under the declined solar heating. There are still several surge episodes in August (Fig. 3a), which may be related to tropical cyclones and monsoon revival (Chen et al. 2004; Ding and Chan 2005). Overall, the southerly monsoon surges with different diurnal variations represent well the subseasonal progress of the EASM.

The interannual variations of surge episodes with different diurnal variations are further examined (Fig. 3c). There are about 18 days of M(+)/D(+) per year with a standard deviation of 8.7, while there are about 12 days of M(+)/D(-) with a standard deviation of 5.5. The minimum of the accumulated day of surge episodes is observed in 2014, with only 10 days in total. In contrast, there are 54 days categorized as M(+)/D(+) or M(+)/D(-) in 1998 when intense summertime precipitation took place over East Asia (Zhang et al. 2002; Qian et al. 2004; Sun et al. 2016). Besides, in the summer of 2020 when the record-breaking mei-yu occurred (Ding et al. 2021; Fang et al. 2021), there are 56 days categorized as surge episodes. It implies that frequent episodes of southerly surge may be one of the key factors for extreme precipitation over East Asia.

FIG. 3. (a) Types of southerly monsoon episodes. (b) Subseasonal variations of the accumulated days of different types of episodes. (c) As in (b), but for interannual variations. Episodes in 2020 are excluded in (b).

In this study, we pay attention to the climatology of southerly monsoon surges in 1998–2019, while their influence on the heavy rainfall in 2020 will be studied in ongoing works. The monsoon activities in late June (the mei-yu period), when the transition from $M(+)D(-)$ to $M(+)D(+)$, show an evident interannual variation (Fig. 3a). The monsoon flows in this subseasonal period are mostly $M(+)D(-)$ from 1998 to 2003, change to $M(+)D(+)$ from 2004 to 2008 and return to $M(+)D(-)$ from 2009 to 2013. It suggests that southerly monsoon surges with different diurnal variations can represent well the large interannual variations of the EASM.

In the following sections, we focus on the impacts of southerly surge episodes with different diurnal variations by comparing the composites of $M(+)D(+)$ and $M(+)D(-)$. For composite analysis, we choose a reference date defined as day 0, denoting the onset of $M(+)D(+)$ or $M(+)D(-)$. Day 1 (–1) is defined as the day after (before) day 0, and so on. For instance, 5 July is day 0 of the $M(+)D(+)$ from 5 to 10 July 2004 (Fig. 2), while 6 July (4 July) is day 1 (–1).

b. Large-scale conditions associated with southerly monsoon surges

In this subsection, we examine the large-scale atmospheric conditions associated with the two types of southerly surge

episodes, and further clarify how they regulate the wind diurnal amplitude during surge episodes. Before the onset of $M(+)D(+)$, the WPSH (refer to the 5880-gpm contours) is located to the east of 127°E with an anomalous high growing over the western Pacific (Fig. 4a). After the onset, the anomalous high moves westward to southeastern China (Figs. 4b,c). Correspondingly, the WPSH strengthens and extends its western edge to the coastline of South China. The strengthened WPSH intensifies the low-level pressure gradient over southeastern China (not shown) and thus enhances the daily southerlies along its western flank. Meanwhile, a positive anomaly of low-level temperature up to 1.7 K is observed over southeastern China. The anomalous warming is supported by the strong diabatic heating under the WPSH and the advection of active southerlies (not shown). Such enhanced background flows and warm conditions facilitate enlarging the wind diurnal amplitude (Xue et al. 2018; Zeng et al. 2019; Chen 2020). In midlatitudes, a weak trough is intensified to the north of 35°N with an eastward-moving anomalous low.

As for $M(+)D(-)$ (Figs. 4d–f), the WPSH is also strengthened after the onset, but its westward extension is confined to the east of Taiwan and its rigid is farther south compared with that during $M(+)D(+)$. A deep midlatitude trough develops and intrudes to $\sim 30^{\circ}\text{N}$ with a southward anomalous low over eastern China. The approaching trough increases the low-level pressure gradient over southeastern China, thereby enhancing the prevailing daily southwesterlies. A negative anomaly of ABL temperature (–1.1 K) with a southwest–northeast orientation is observed in front of the trough (Figs. 4e,f), which is significantly different from that during $M(+)D(+)$ (Figs. 4b,c). The anomalous cooling is induced by diabatic cooling because of the cloudiness along the trough (not shown). The weak daytime heating in the lower troposphere is accompanied by small wind diurnal variations, in agreement with the results of Chen (2020).

Figure 4 shows that the surge episodes can form as the daily southwesterlies are intensified by the low-level pressure gradients over southeastern China, which can be caused by either the westward extension of the WPSH or the approaching midlatitude trough. These different large-scale conditions correspond to the different regional thermal forcings, as indicated by the differences of ABL temperature anomalies between the two types of surge episodes (cf. Figs. 4b,c,e,f). The thermal forcings over southeastern China play a key role in regulating the amplitudes of wind diurnal variations (Zeng et al. 2019; Chen 2020; Chen et al. 2021). Here, we highlight that such a mechanism downscaling from the multiday large-scale conditions to regional wind diurnal variations leads to the two different types of southerly surge episodes. We also checked the coupling of large-scale atmospheric conditions and wind diurnal variations in different subseasonal periods (see online supplemental material). The differences between $M(+)D(+)$ and $M(+)D(-)$ are highly similar to the above composites, so they are somewhat independent of the seasonal cycle. It is noted that a surge episode may be followed by another (Fig. 3a), with possible overlapping in their composite windows. We checked the composites of those episodes without any overlapping during their days –4 to 3 (not shown). The results are

FIG. 4. Composites of 500-hPa geopotential height (black contours with an interval of 30 gpm), its anomaly (blue contours with an interval of 3 gpm), 925-hPa temperature anomaly (shading), and the diurnal amplitude of 925-hPa meridional wind at 0200 LST (pink hatching, above 1 m s^{-1}).

consistent with those in Fig. 4, and thus the overlapping issue does not affect our conclusions.

4. Rainfall response to southerly monsoon surges with different diurnal variations

a. Spatiotemporal evolutions of low-level winds and rainfall patterns

The spatial evolutions of the daily mean low-level winds and rainfall anomaly before and after the onset of southerly surge episodes are shown in Fig. 5. During the onset of M(+)D(+) (Figs. 5a,b), the low-level southerlies prevail over southeastern China and increase substantially with enhanced diurnal amplitudes under the influence of the westward-extending WPSH (Figs. 4b,c). The rainfall anomaly changes from negative to

positive over most areas of North China (north of 35°N), while it decreases over southeastern China, implying the northward displacement of the monsoon rainband. The rainfall anomalies over the western Pacific and South China Sea are negative. During the onset of M(+)D(-) (Figs. 5c,d), the low-level southerlies also increase over southeastern China, but the increase is most evident to the south of 30°N, with reduced diurnal amplitudes. The low-level winds exhibit a large meridional gradient at 30°–35°N (Fig. 5d), suggesting an intensified mei-yu frontal zone in association with the deepening midlatitude trough (Figs. 4e,f). The positive rainfall anomaly is more confined to the frontal zone over Central China and southwestern Japan (30°–35°N). In contrast, the rainfall remains suppressed over North China, which is affected by the anomalous northerly behind the trough. The rainfall is also suppressed over the coast of southern China under the westward extension of the WPSH.

FIG. 5. Composites of daily mean rainfall amount anomaly (shading) and 925-hPa winds (vectors). The pink hatching denotes the diurnal amplitude of 925-hPa meridional wind at 0200 LST (above 1 m s^{-1}). The composites before and after the onset are averaged from days -4 to -1 and days 0 to 3 , respectively.

Figures 6a–f present the daily evolutions of low-level southerlies during the onset of surge episodes. The daily mean southerlies increase to greater than 5.0 m s^{-1} after the onset of both types of surge episodes (Figs. 6a,b). The two types of surge episodes show a small difference in the daily mean of meridional winds (Fig. 6c). However, the diurnal amplitudes of monsoon southerly are distinguishable between M(+)D(+) and M(+)D(-). The diurnal amplitudes of southerlies increase to $\sim 1.2 \text{ m s}^{-1}$ after the onset of M(+)D(+) (Fig. 6d). In contrast, the diurnal amplitudes decrease to $\sim 0.6 \text{ m s}^{-1}$ after the onset of M(+)D(-) (Fig. 6e). The relative difference between M(+)D(+) and M(+)D(-) could be greater than 0.6 m s^{-1} and last for 4 days after the onset of the episodes (Fig. 6f).

As a response to the wind diurnal amplitudes, the evolutions of the rainfall anomaly are significantly different during M(+)D(+) and M(+)D(-). The positive rainfall anomaly rapidly shifts from 27° – 35°N to the north of 35°N after the onset of

M(+)D(+) (Fig. 6g). In contrast, after the onset of M(+)D(-), the positive rainfall anomaly moves northward slightly from 25° – 32°N to 30° – 35°N (Fig. 6h). As a result, the positive rainfall anomaly tends to locate at North China during M(+)D(+), whereas it appears at Central China during M(+)D(-) (Fig. 6i). The difference in rainfall anomaly after the onset of M(+)D(+) and M(+)D(-) is up to 40% of the climate-mean rainfall. This rainfall difference appears at day 0 and becomes more pronounced in the following 4 days. Monsoon southerlies with a maximum near midnight tend to facilitate the early morning rainfall systems with durations longer than 6 h. Such a response of long-duration rainfall to wind diurnal variations is also noted in previous studies (Chen et al. 2009, 2013; Yuan et al. 2010; Li et al. 2020). In summary, the northward-displaced rainfall of M(+)D(+) compared with M(+)D(-) suggests that the southerly monsoon surges with different diurnal variations are closely related to the meridional movement of the monsoon rainband.

FIG. 6. Latitude–time evolutions of (a)–(c) the daily mean 925-hPa meridional winds, (d)–(f) the diurnal amplitude of meridional winds at 0200 LST, and (g)–(h) the daily mean rainfall anomaly averaged over 110° – 130° E during the onset of (left) $M(+)D(+)$ and (center) $M(+)D(-)$. (right) The relative difference between $M(+)D(+)$ and $M(+)D(-)$ [that is, $M(+)D(+)$ minus $M(+)D(-)$]. The hatching denotes the values above the 99.9% confidence level in (c) and (f) and those above the 95% confidence level in (i). Note that part of long-duration rainfall produced by the wind diurnal variation may be recorded on the following day based on the definition of the diurnal cycle since 1200 LST.

b. Diurnal variations of rainfall systems in response to southerly monsoon surges

The spatial distributions of the rainfall anomaly averaged on the duration days of both types of southerly surge episodes

(Figs. 7a,b) are similar to those averaged on days 0 to 3 (Figs. 5b,d). The spatial distribution of rainfall anomaly during $M(+)D(+)$ is different from that during $M(+)D(-)$, especially over Central China and North China (Figs. 7a,b). We wonder how rainfall systems in these two subregions respond to the two

FIG. 7. (a),(b) Composites of daily mean rainfall amount anomaly averaged on the durations of M(+D(+)) and M(+D(-)). The boxes denote Central China (27.5°–35°N, 110°–130°E) and North China (35°–42.5°N, 110°–130°E). (c)–(h) Areal-averaged rainfall amount with respect to area sizes. The number marked at the upper-left corner is the daily mean rainfall amount. The triangle marks a size threshold of $2 \times 10^4 \text{ km}^2$ for meso- α -scale rain events. The four numbers at the center are the sum of the rainfall amounts of small-medium/meso- α -scale rain events in afternoon–evening/midnight–morning hours. The solid (dashed) pink boxes in (c), (d), and (f) denote the primary (secondary) rainfall peak by meso- α -scale rain events.

types of surge episodes and whether the rainfall responses vary at the subseasonal time scale. To quantitatively analyze the regional response of rainfall systems, the organized and isolated rainfall systems are classified based on the contiguous rain area with a rain rate greater than 1 mm h^{-1} in satellite observations, which is in agreement with ground observations (Ebert and McBride 2000; Demaria et al. 2011; Chen et al. 2013). The organized rainfall systems are further classified into small-medium and meso- α -scale rain events

by an area threshold of $2 \times 10^4 \text{ km}^2$, equivalently to $\sim 150 \text{ km}$ in the horizontal scale.

During M(+D(+)), the rainfall anomaly is positive over both Central China and North China (Fig. 7a). The areal average rainfall amounts over these two subregions are 6.0 mm day^{-1} and 5.0 mm day^{-1} , respectively (Figs. 7c,f), while the latter one is significantly greater than the climate mean at the 99.9% confidence level (Fig. 7h). The increase in rainfall is primarily attributed to meso- α -scale rain events (an area size larger than

$2 \times 10^4 \text{ km}^2$). The rainfall induced by meso- α -scale events over Central China (5.3 mm day^{-1}) and North China (4.2 mm day^{-1}) is 11% and 24% greater than the climate mean (4.7 and 3.4 mm day^{-1}), respectively. Meso- α -scale events produce a major rainfall peak in the morning (0800 LST, solid pink box in Fig. 7c) and a secondary one in the afternoon (1400 LST, dashed pink box in Fig. 7c) over Central China, while the rainfall peaks at predawn (0500 LST, solid pink box in Fig. 7f) and late afternoon (1700 LST, dashed pink box in Fig. 7f) over North China. These events are more active from midnight to morning when the induced rainfall (2.2 mm day^{-1}) is 10% greater than that from afternoon to evening (2.0 mm day^{-1}).

During M(+)D(−), the rainfall anomaly is positive over Central China while negative over North China (Fig. 7b). Over Central China, the areal average rainfall amount is 6.7 mm day^{-1} (Fig. 7d), which is significantly greater than the climate mean at the 99.9% confidence level (Fig. 7e). The increase in rainfall is a result of active meso- α -scale events that produce 26% more rainfall during M(+)D(−) (5.9 mm day^{-1}) than in the climate mean (4.7 mm day^{-1}). Compared with M(+)D(+), the rainfall from meso- α -scale events during M(+)D(−) also exhibits double peaks in the morning and afternoon but with larger amounts. Similarly, meso- α -scale events produce 19% more rainfall in midnight–morning hours (3.2 mm day^{-1}) than in afternoon–evening hours (2.7 mm day^{-1}). Over North China, the rainfall amount is 3.9 mm day^{-1} (Fig. 7g), which is less than the climate mean of 4.1 mm day^{-1} (Fig. 7h). The rainfall is reduced by fewer activities of meso- α -scale events from midnight to morning. Previous studies also noted that the strong monsoon southerlies can bring a robust response in rainfall systems, particularly in the meso- α -scale rain events in midnight–morning hours (Hirose and Nakamura 2005; Chen et al. 2013; Chen 2020; W. Li et al. 2021). This study further categorizes the strong monsoons into two types of surge episodes and found that they can result in significantly different rainfall responses in the two subregions. It highlights the combined effects of large-scale conditions and subdaily regional forcings on rainfall systems.

Given that the rainfall over Central China and North China undergoes a notable subseasonal variation in climatology (Figs. 8a,b), we further examine whether the rainfall responses in these two subregions to the two types of monsoon surge episodes vary at the subseasonal time scale. During M(+)D(+), the positive rainfall anomaly over Central China is only evident in mid-June (Fig. 8c). However, the positive anomaly over North China is extensively observed throughout the day from mid-June to late July (Fig. 8d). During M(+)D(−), the positive rainfall anomaly over Central China is observed throughout the day from mid-June to early July (Fig. 8e). The positive rainfall anomaly over North China is observed from early to mid-July, which is mainly confined in afternoon–evening hours and less evident than that during M(+)D(+) (Fig. 8d). Overall, the rainfall response to M(+)D(+) is mainly observed over North China from mid-June to late July, while the response to M(+)D(−) is observed over Central China from mid-June to early July. These rainfall responses are thus evident in the periods when the monsoon rainband advances northward with active southerlies (Fig. 1c). Recent studies

noted that the strong southerlies coupling with large (small) diurnal variations contribute greatly to the summer rainfall over North (Central) China (Pan and Chen 2019; Chen et al. 2021). Here we further highlight that the diurnal variation of southerly monsoon surge is key for the meridional movement of the monsoon rainband. It is noted that some diverse features are observed from late July to August because of the small sample sizes in the 10-day composites (hatching in Fig. 8). For a better estimate of statistical significance, we make a composite of all episodes in late July–August and find that the differences between the two types of surge episodes are highly similar to those in mid-June–mid-July (not shown). It suggests that the composite results are less affected by the selection of period for study.

5. Atmospheric processes of southerly monsoon surges governing rainfall systems

In this section, we examine the specific processes of the two types of southerly surges modulating the rainfall in the EASM regions. We pay more attention to dynamic lifting, water vapor transport, and convective instability generation.

a. Dynamic lifting

Figure 9 shows the anomalous vertical motion before and after the onset of southerly surge episodes. Before the onset of M(+)D(+), the daily mean and midnight–morning deviations of anomalous vertical motion is small, suggesting that the large-scale ascent is similar to the climate mean (Fig. 9a). After the onset, the maximum anomaly of daily mean upward motion is $-2.7 \times 10^{-2} \text{ Pa}^{-1}$ at $\sim 34^\circ\text{N}$ at 500–600 hPa (Fig. 9c). The anomalous upward motion extends to the north of 35°N where the summer-mean vertical motion is generally downward (not shown), indicating the strengthening and northward displacement of large-scale ascent. The maximum of anomalous upward motion deviation in midnight–morning hours is $-1.3 \times 10^{-2} \text{ Pa}^{-1}$ at $\sim 33^\circ\text{N}$ at a lower layer of 600–700 hPa. The maximum latitude of the midnight–morning deviation almost overlaps with the daily value, implying a great impact of diurnal forcings on the large-scale ascent. The anomalous large-scale ascent is mostly supported by the upward motion in midnight–morning hours, related to the nocturnal accelerated southerlies, rather than those in afternoon–evening hours. Figure 9e shows the daily evolution of the anomalous vertical motion at 700 hPa. The anomalous upward motion extends to the north of 35°N and corresponds to rainfall anomaly after the onset (Fig. 6g), which is mainly attributed to the enhancement in the midnight–morning hours. It suggests that strong southerlies with large diurnal variations can strengthen the large-scale ascent with a northward displacement, especially from midnight to morning. Such an evolution of the large-scale ascent may support the positive rainfall anomaly over North China and result in the northward advance of the monsoon rainband (Fig. 6g).

Before the onset of M(+)D(−), the anomaly of daily mean upward motion is mainly observed at $23^\circ\text{--}30^\circ\text{N}$ with a maximum of $-3.2 \times 10^{-2} \text{ Pa s}^{-1}$ at $\sim 28^\circ\text{N}$ at 400–500 hPa (Fig. 9b). The maximum anomaly of upward motion deviation in midnight–morning hours is $-1.2 \times 10^{-2} \text{ Pa s}^{-1}$ at $\sim 31^\circ\text{N}$. After the onset,

FIG. 8. Subseasonal variations of (a),(b) the rainfall amount in climatology and (c)–(f) the rainfall anomaly during M(+D(+)) and M(+D(−)) averaged over (left) Central China and (right) North China. The hatching in (c)–(f) denotes the subseasonal periods when the accumulated day is less than 25 days, as shown in Fig. 3b.

the maximum of daily value stays at $\sim 29^{\circ}\text{N}$ while it is strengthened to $-5.5 \times 10^{-2} \text{ Pa s}^{-1}$ (Fig. 9d). The maximum of midnight–morning deviation overlaps with the daily value at $\sim 29^{\circ}\text{N}$, farther south than that during M(+D(+)). The anomalous upward motion to the north of 35°N is weak while the midnight–morning deviation exhibits as anomalous downward motion, which is different from those after the onset of M(+D(+)). Figure 9f shows that the anomaly of daily mean upward motion is confined at 26° – 32°N , persisting from day 0 to 3. The anomalous large-scale ascent during M(+D(−)) corresponds to the positive rainfall anomaly over Central China (Fig. 6h).

The results suggest that the diurnal variation of strong south-easterlies is closely related to the diurnal variation and location of large-scale ascent (Chen 2020). This relation is evident during southerly surge episodes and may efficiently modulate rainfall.

b. Water vapor transport

The spatial distributions of the water vapor flux and precipitable water before and after the onset of southerly surges are shown in Fig. 10. After the onset of M(+D(+)), the northward water vapor flux averaged over the EASM regions, including Central China and North China, is enhanced by 51%

FIG. 9. (a)–(d) Latitude–pressure sections and (e),(f) latitude–time evolutions at 700-hPa of anomalous vertical motion ($10^{-2} \text{ Pa s}^{-1}$) averaged over 110° – 130°E . The contours denote the daily mean and the shading denotes the deviation in midnight–morning hours (averaged at 0200 and 0800 LST).

to $\sim 122 \text{ kg m}^{-1} \text{ s}^{-1}$ (Figs. 10a,b). Correspondingly, the precipitable water is $\sim 3 \text{ kg m}^2$ greater than the climatological mean. As a comparison, after the onset of M(+)D(–), the enhancement of northward water vapor flux is only observed over southeastern China where the positive anomaly of precipitable water is up to $\sim 5 \text{ kg m}^2$ (Figs. 10c,d). During the two types of surge episodes, the moisture from tropical oceans is transported northward along the western edge of the WPSH. The water vapor can be further guided to North China when the rigid of the WPSH is located at $\sim 25^{\circ}\text{N}$ (Fig. 10b). It instead tends to stay in Central China when the rigid of the WPSH is relatively south (Fig. 10d). This difference suggests that the water vapor transport over East Asia is strongly regulated by the various westward extension of the WPSH

(Zhang et al. 2020; Chen et al. 2021; Y. Zhao et al. 2021; X. Li et al. 2021).

Figures 11a–c show the diurnal variations of precipitable water, water vapor flux and its horizontal convergence averaged over Central China. After the onset of M(+)D(+), the daily mean precipitable water increases to $\sim 47 \text{ kg m}^2$ (red bars in Fig. 11a). The daily mean water vapor flux enhances to $\sim 231 \text{ kg m}^{-1} \text{ s}^{-1}$ (Fig. 11b). The water vapor flux increases notably from 1400 LST ($\sim 199 \text{ kg m}^{-1} \text{ s}^{-1}$) to 0200 LST ($\sim 265 \text{ kg m}^{-1} \text{ s}^{-1}$). The induced daily mean flux convergence is strengthened to $-4.7 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$ with a peak at 0800 LST (Fig. 11c). After the onset of M(+)D(–), the precipitable water and water vapor flux have daily means comparable with those after the onset of M(+)D(+),

FIG. 10. Composites of daily mean column-integrated water vapor fluxes (vectors, above $120 \text{ kg m}^{-1} \text{ s}^{-1}$), precipitable water (shading with units of kg m^{-2}), and its anomaly (pink contours with units of kg m^{-2}). The green contours are the 5880-gpm line of 500-hPa geopotential height. The yellow arrows denote the maximum water vapor transport.

but their diurnal variations are smaller (blue bars in Figs. 11a,b). The daily mean flux convergence after the onset of M(+)D(-) is $-6.5 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$ with peaks at 0800 LST and 1400 LST (Fig. 11c), corresponding to the two peaks of rainfall in the morning and afternoon over Central China (Fig. 7d). The flux convergence after the onset of M(+)D(-) is $\sim 38\%$ greater than that after the onset of M(+)D(+) (Fig. 11c), which may play an important role in the greater response of rainfall over Central China (cf. Figs. 7c,d).

Figures 11d-f show the water vapor transport averaged over North China. After the onset of M(+)D(+), the daily mean precipitable water increases to $\sim 33 \text{ kg m}^{-2}$ (Fig. 11d). The daily mean water vapor flux enhances from ~ 71 to $\sim 103 \text{ kg m}^{-1} \text{ s}^{-1}$ with a midnight peak of $\sim 125 \text{ kg m}^{-1} \text{ s}^{-1}$ (Fig. 11e). The induced daily mean flux convergence is increased from -1.9×10^{-5} to $-3.5 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$ with a

midnight peak of $-4.5 \times 10^{-5} \text{ kg m}^{-2} \text{ s}^{-1}$ (Fig. 11f). The water vapor flux and its convergence at midnight are $\sim 37\%$ and $\sim 84\%$ higher than those in the afternoon, respectively, indicating a high efficiency of nocturnal monsoon flow in the water vapor budget (Chen et al. 2013). The flux convergence peaks at midnight, $\sim 6\text{ h}$ earlier than the morning peak of that over Central China (Fig. 11c), resulting in the predawn peak of rainfall (0500 LST) over North China that is earlier than the morning peak (0800 LST) over Central China (cf. Figs. 7c,f). Therefore, the supply of water vapor that maximizes at midnight can yield a significant rainfall response, though the rainfall is weaker than that over Central China because of the less precipitable water. As a comparison, the three variables after the onset of M(+)D(-) are similar to those before the onset except for the weakened flux convergence in midnight-morning hours (Figs. 11d-f). Overall, the southerlies with large

FIG. 11. Diurnal variations of (top) column-integrated precipitable water (PW), (middle) northward water vapor flux (qv) at the southern boundary, and (bottom) water vapor flux convergence $[-\text{div}(qv)]$ averaged over (left) Central China and (right) North China. The dashed lines denote the daily mean values.

diurnal variations can carry water vapor to North China while those with small diurnal variations tend to converge over Central China. The above results suggest that the water vapor transport in the EASM regions is jointly regulated by the large-scale conditions and subdaily regional forcings during southerly monsoon surges.

c. Convective instability

The cumulative distribution function of 925-hPa convective available potential energy (CAPE) is used to measure the convective instability (Moncrieff and Miller 1976; Emanuel 1994). Figure 12a shows that the 80th percentile of CAPE over Central China increases from ~ 480 to $\sim 660 \text{ J kg}^{-1}$ after the onset of M(+)/D(+). It decreases from ~ 430 to $\sim 390 \text{ J kg}^{-1}$ after the onset of M(+)/D(-). Figure 12b shows that the 80th percentile of CAPE over North China also increases from ~ 270 to $\sim 370 \text{ J kg}^{-1}$ after the onset of M(+)/D(+) while it decreases from ~ 290 to $\sim 240 \text{ J kg}^{-1}$ after the onset of M(+)/D(-). The convective instability is thus enhanced during M(+)/D(+) but suppressed during M(+)/D(-) over both Central China and North China. The ABL temperature is much greater during M(+)/D(+) than during M(+)/D(-) (cf. Figs. 4b,c,e,f), and so is the water vapor transport (cf. Figs. 10b,d). Both of them may affect the CAPE change that depends on the atmospheric temperature and humidity (Chen et al. 2014b).

It implies that the CAPE generation may be closely linked to the regional diurnal forcings modulated by large-scale conditions.

To examine the causes of the difference in CAPE change between M(+)/D(+) and M(+)/D(-), the budget of CAPE generation rate is estimated at the diurnal time scale. Based on Emanuel (1994) and G. Zhang (2002), the change of CAPE ($\partial \text{CAPE} / \partial t$) can be induced by the change of equivalent potential temperature in the boundary layer $[C_p(T_{vsl} - T_{vnb})\partial \ln \theta_e / \partial t]$ and the change of the convection-layer thickness $[-\partial(\phi_{lnb} - \phi_{sl}) / \partial t]$, while the latter one is small. Following Chen et al. (2014b), the CAPE change due to the boundary layer forcings can be decomposed into the rates by horizontal advection $(-u\partial \theta_e / \partial x - v\partial \theta_e / \partial y)$, vertical lifting $(-w\partial \theta_e / \partial z)$ and diabatic heating (Q).

Figure 12c shows the difference in CAPE generation rate between the two types of surge episodes averaged over Central China. Compared with M(+)/D(-), the CAPE generation after the onset of M(+)/D(+) is much greater at 1400 LST and smaller at 0200 LST. It is primarily caused by external diabatic heating, which may be related to solar radiation. In the less cloudy condition established by the westward extension of the WPSH during M(+)/D(+) (Figs. 4b,c), the solar radiation can strongly warm up the boundary layer in daytime while the radiative cooling is also strong in nighttime. Strong diabatic heating is also important to induce large wind diurnal

FIG. 12. (a),(b) Cumulative distribution function of 925-hPa CAPE. The dashed lines denote the results made before the onset (days -1 to -4) and the solid lines denote the results after the onset (days 0 to 3). (c),(d) The relative difference in the generation of CAPE by various processes between $M(+)$ $D(+)$ and $M(+)$ $D(-)$ [that is, $M(+)$ $D(+)$ minus $M(+)$ $D(-)$].

variations (Zeng et al. 2019; Chen 2020). However, the weak CAPE generation in midnight cannot efficiently couple with the dynamic lifting (Fig. 9c) and water vapor flux (Fig. 11b) to produce midnight–morning precipitation systems over Central China.

Figure 12d shows that the difference in the CAPE change over North China is greatly affected by horizontal advection. The horizontal advection is $\sim 7.3 \text{ J kg}^{-1} \text{ h}^{-1}$ after the onset of $M(+)$ $D(+)$, while it is $\sim 1.7 \text{ J kg}^{-1} \text{ h}^{-1}$ after the onset of $M(+)$ $D(-)$. The horizontal advection peaks at midnight because of the nocturnal accelerated southerlies. The convective instability over North China thus depends on the warm moist inflow, which is different from that over Central China. The strong horizontal advection during $M(+)$ $D(+)$ not only supports the CAPE generation in the afternoon with solar heating, but also compensates the nighttime CAPE reduction by radiative cooling, which is thought to facilitate the precipitation from midnight to morning. This difference in CAPE change between the two types of surge episodes can persist for 4 days, corresponding to the difference in the diurnal variations of southerlies (Fig. 6f) and the precipitation (Fig. 6i).

The above analyses clarify that the strong southerlies with large diurnal variations can strengthen the large-scale ascent with a northward displacement and also induce a strong advection of warm moist energy through enhancing moisture

transport. These favorable processes during $M(+)$ $D(+)$ work together to produce rainfall farther north than that during $M(+)$ $D(-)$. The meridional movement of the monsoon rainband has been linked to the strong monsoon southerlies (Zhu et al. 2011; Zeng et al. 2012; He and Liu 2016). We further highlight that it is also efficiently regulated by the southerly monsoon surges with different diurnal variations, which represent the multiscale interaction of the multiday large-scale conditions with subdaily regional forcings.

6. Summary and discussion

This paper presents a 22-yr climatology of southerly monsoon surges over East Asia during the summers from 1998 to 2019. The southerly surge episodes are classified into two types of large and small diurnal amplitudes [i.e., $M(+)$ $D(+)$ and $M(+)$ $D(-)$]. The differences between the two types of southerly surges in terms of the associated large-scale conditions, the induced rainfall response and the atmospheric processes governing rainfall are clarified. The major findings are summarized as follows:

- 1) There are 118 episodes of southerly surge observed over East Asia, of which 63 are $M(+)$ $D(+)$ and the rest 55 are $M(+)$ $D(-)$. $M(+)$ $D(+)$ and $M(+)$ $D(-)$ account for 19.0%

FIG. 13. Schematic summary of the diurnal variations of southerly monsoon surge and their impacts.

and 12.9% of the summer days with average durations of 6.1 and 4.7 days, respectively. While the daily mean speeds of low-level southerlies are comparable with each other, the diurnal amplitudes at 0200 LST during M(+)D(+) are almost twice of that during M(+)D(-). At the subseasonal time scale, M(+)D(-) is active from mid- to late June and M(+)D(+) from late June to mid-July, corresponding to the northward progress of the EASM and the intensification of its diurnal variation. The surge episodes with different diurnal variations also represent well the large interannual variations of the EASM.

- 2) The two distinct types of southerly surges with different diurnal variations are associated with a mechanism downscaling from multiday large-scale circulations to regional diurnal wind variations, as shown by the schematic summary in Fig. 13. During M(+)D(+), the daily mean southerlies increase due to the low-level pressure gradients over southeastern China under a westward extension of the strengthened WPSH. The extending WPSH also produces anomalous warming in the ABL, which facilitates enlarging the wind diurnal amplitude. During M(+)D(-), the pressure gradients and the induced daily mean southerlies are intensified by an approaching midlatitude trough while the rigid of the WPSH is relatively south. The approaching trough and its associated cloudiness lead to anomalous cooling over southeastern China and thereby suppress wind diurnal variations.
- 3) Southerly surge episodes with different diurnal variations are closely related to the meridional movement and diurnal variation of the monsoon rainband. The rainband tends to displace northward to North China (north of 35°N) after the onset of M(+)D(+), whereas it tends to appear over Central China (30°–35°N) after the onset of M(+)D(-). Such a rainfall difference in these two subregions can last for 4–5 days. The different rainfall response is mainly attributed to the meso- α -scale rain events, which produce a

dominant peak of rainfall in the early morning (0500–0800 LST) and a secondary one in the afternoon (1400–1700 LST).

- 4) The differences in rainfall responding to the two types of southerly surges are jointly modulated by the interaction of large-scale conditions and subdaily regional forcings. After the onset of M(+)D(+), the strong southerlies with large diurnal variations can strengthen the large-scale ascent with a northward displacement, especially from midnight to morning. The northward transport of water vapor by nocturnal accelerated southerlies also leads to a strong advection of warm moist energy at the western edge of WPSH, which can enhance the generation of convective instability over North China. The combined effects of these dynamic and thermodynamic processes are favorable for the northward advance of the monsoon rainband. As a comparison, during M(+)D(-), the large-scale ascent and moisture convergence by monsoon southerlies tend to be strengthened over Central China where they sustain a relatively south rainband compared with that during M(+)D(+).

The southerly monsoon surges are essential for the precipitation over East Asia in summer (e.g., Zhang et al. 2002; H. Liu et al. 2008; Y. Liu et al. 2017). For the first time, this paper categorizes the southerly monsoon surges into two different types regarding their wind diurnal variations. Their properties can represent the different interactions of multiday large-scale atmospheric circulations with regional diurnal forcings so that the influences of multiscale interactions are clarified. We note that there are ~85% days of surge episodes coexisting with the active ISO phases. Further analyses on the different types of southerly monsoon surges and their relationship with ISO help to improve our understanding of regional climate change. Case studies also show that the downscaling processes during multiday episodes coupled with diurnal

processes are crucial for the successive growth of mesoscale convective systems (G. Chen et al. 2017; Sun and Zhang 2012; Zhang et al. 2014, 2018; Zeng et al. 2019). Further studies on more specific cases from the synoptic-to-mesoscale aspect may shed insights into the cause of extreme heavy rainfalls.

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