Synoptic Conditions Associated with Propagating and Nonpropagating Cloud/Rainfall Episodes during the Warm Season over the East Asian Continent

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ABSTRACT

Using European Centre for Medium-Range Weather Forecasts (ECMWF) analyses, this study identifies common synoptic flow patterns associated with propagating (long lived) and nonpropagating (short lived) precipitation/cloud episodes in the warm season over East Asia (25°–40°N, 95°–125°E). Among 123 initial cases during May–July 1997–2003, 86 were classified into three categories (propagating cases with strong and weak forcing, and nonpropagating ones with strong inhibition) consisting of 10 types based on their 500-hPa flow patterns. For each type, composites at various levels when episodes moved across 100°E, 105°E, 110°E, 115°E, and 120°E were made and discussed to elucidate the influence of synoptic conditions.

For propagating episodes with strong 500-hPa forcing (46 cases), four types were identified. In their composites, favorable positive vorticity advection forcing from an approaching trough/short wave exists aloft, and concurrent development of southwest vortex and front/wind-shift line at low levels leads to enhanced southwesterly flow to provide vertical shear and moisture. For nonpropagating episodes with strong inhibition (25 cases), three types dominated by synoptic-scale subsidence from either negative vorticity advection or the subtropical high were identified. Thus, episodes weaken shortly after they move out from the Tibetan Plateau. There are also propagating episodes but with weak forcing (15 cases), for which another three types were found. The differences in their composites from nonpropagating types are mainly at low levels, where the southerly flow is stronger and can penetrate deeper inland. As stability conditions are similar for all type of episodes, the results suggest an important role of stronger low-level southwesterly winds in providing moisture and vertical wind shear to sustain episodes and their propagation.

1. Introduction

Using high-resolution radar data, Carbone et al. (2002) reported that warm-season (May–August) precipitation episodes, defined as clusters of rain-producing systems in Hovmöller (longitude–time) space, exhibit propagating characteristics to the lee of the Rockies over the central United States. Often composed of multiple mesoscale convective systems (MCSs), the longevity of some episodes can reach up to 3000 km and 60 h in space and time. Their phase in the diurnal cycle is also closely tied to the terrain, as episodes tend to develop over the eastern slopes of the Rockies in late afternoon, then propagate eastward across the Great Plains overnight, sometimes into the next day, in agreement with many previous studies of MCSs (e.g., Maddox 1983; Augustine and Howard 1991; Laing and Fritsch 1997). These coherent properties suggest an
intrinsic predictability and the potential to improve warm-season quantitative precipitation forecasts (QPFs) if they can be adequately captured in the operational models (e.g., Davis et al. 2003; Fritsch and Carbone 2004). Since the work of Carbone et al. (2002), there have been many studies examining similar patterns of behavior of propagating rainfall episodes in different areas around the world (e.g., Wang et al. 2004, 2005; Laing et al. 2008; Keenan and Carbone 2008; Liu et al. 2008; Levizzani et al. 2010; Pereira et al. 2010; He and Zhang 2010).

Over East Asia, a major terrain feature [i.e., the Tibetan Plateau (TP)] also exists, and nocturnal convection to its lee has been previously noted by Asai et al. (1998). The downstream (eastward) propagation of rainfall episodes (or cloud clusters) with a preferred phase in the diurnal cycle is confirmed using both satellite data (Wang et al. 2004, 2011c; Johnson 2011; Xu and Zipser 2011; Bao et al. 2011) and rain-gauge observations (Yu et al. 2007; Zhou 2004, 2011c; Johnson 2011; Xu and Zipser 2011; Bao et al. 2011) and case studies (e.g., Zhong et al. 1996; Tuttle and Davis 2006; Trier et al. 2010; Huang et al. 2010).

Although the diurnal phase of rainfall episodes is regulated by local solar effects and MPS circulations induced by the terrain, they are nonetheless also subject to the influences of their synoptic environments, as demonstrated by many earlier studies (e.g., Wang and Orlanski 1987; Wang et al. 1993; Tuttle and Carbone 2004; Trier et al. 2006). This is especially true for East Asia, where monsoon circulations are more prominent than in the United States. Consequently, the operational roles played by the MPS and low-level jet (LLJ, mainly nocturnal in the United States) have been closely examined, both climatologically (Chen et al. 2010; He and Zhang 2010; Bao et al. 2011) and in case studies (e.g., Zhong et al. 1996; Tuttle and Davis 2006; Trier et al. 2010; Huang et al. 2010).
closest synoptic times (at 6-h intervals) when it passed 100\degree 8, 105\degree 8, 110\degree 8, and 120\degree E (in the Hovmöller diagram) are identified as T1–T5, respectively. For T3, the year, month, date, and time in UTC are listed, while other times are given in hours relative to T3. Also, the times are in boldface if the cases are roughly in phase with the preferred propagating diurnal signal.

<table>
<thead>
<tr>
<th>Type</th>
<th>T1</th>
<th>T2</th>
<th>T3</th>
<th>T4</th>
<th>T5</th>
<th>T1</th>
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</tr>
</tbody>
</table>

Type T1 T2 T3 T4 T5 T1 T2 T3 T4 T5 T1 T2 T3 T4 T5

1 2 3 4 5 1 2 3 4 5 1 2 3 4 5

2. Data and methodology

The dataset of cloud streaks used in this study was compiled by the first author for warm seasons during 1997–2003, following the same method of Wang et al. (2004) for calculation and streak identification. The Hovmöller domain was 25\degree–40\degree N, 95\degree–125\degree E and cloud streaks in May–July were considered, as the propagation to the lee of the TP is more pronounced (Wang et al. 2005, 2011c; Xu and Zipser 2011; Bao et al. 2011). From Hovmöller plots of hourly infrared (IR) blackbody brightness temperature $T_{bb}$ data [see Wang et al. (2004) for details], a total of 125 cloud episodes was chosen subjectively for an examination of synoptic conditions. These include 73 propagating events that traveled across the East Asian continent and 50 nonpropagating events that migrated out from the TP only for short distances.

For synoptic examination of all chosen cloud episodes, the European Centre for Medium-Range Weather Forecasts (ECMWF) gridded analyses [Tropical Ocean Global Atmosphere (TOGA) advanced] were used. This dataset has a horizontal resolution of 1.125\degree latitude–longitude and provides geopotential height $z$, temperature $T$, $u$ and $v$ wind components, vertical velocity $\omega$, and relative humidity (RH) at 11 pressure $p$ levels (from 1000 to 100 hPa) every 6 h at 0000, 0600, 1200, and 1800 UTC. For each cloud streak, the 500-hPa $z$, wind, and relative vorticity $\zeta$ were plotted for the synoptic time closest to the time when it passes 110\degree E (in the Hovmöller diagram), and this flow pattern at 500 hPa was used to classify the cases into different synoptic types. For nonpropagating cloud streaks that did not reach 110\degree E, this time is estimated using extrapolation. Hence, the common types

*Table 2. Cases of propagating cloud episodes used for composites in the four types, with strong 500-hPa synoptic forcing (i.e., types listed in top third of Table 1). For each streak, the closest synoptic times (at 6-h intervals) when it passed 100\degree 8, 105\degree 8, 110\degree 8, and 120\degree E (in the Hovmöller diagram) are identified as T1–T5, respectively. For T3, the year, month, date, and time in UTC are listed, while other times are given in hours relative to T3. Also, the times are in boldface if the cases are roughly in phase with the preferred propagating diurnal signal.*
TABLE 3. As in Table 2, but for cases of nonpropagating cloud episodes in the three types with strong 500-hPa synoptic inhibition (i.e., types listed in bottom third of Table 1).

<table>
<thead>
<tr>
<th>Type</th>
<th>Time</th>
<th>Cases</th>
</tr>
</thead>
<tbody>
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<td>T1</td>
<td></td>
<td>12</td>
</tr>
<tr>
<td>T2</td>
<td></td>
<td>12</td>
</tr>
<tr>
<td>T3</td>
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</tr>
<tr>
<td>T4</td>
<td></td>
<td>0000 UTC 6 Jun 2001</td>
</tr>
<tr>
<td>T5</td>
<td></td>
<td>18</td>
</tr>
</tbody>
</table>

Using the method described, a total of 10 different types more frequently associated with propagating or nonpropagation episodes were identified and can be grouped into three categories as shown in Table 1, where the classification criteria are also given. For propagating episodes, four types are found to be associated with strong 500-hPa forcing of the synoptic scale (i.e., with clear favorable conditions), as will be illustrated shortly in section 3b. These types are called trough–downstream ridge (15 cases), ahead of a major trough (10 cases), moving short wave (10 cases), and confluence with short wave (9 cases), respectively. For nonpropagating episodes, three types of 500-hPa flow patterns are found to be common and exhibit strong inhibition (discussed in section 3c): ridge-downstream trough (10 cases), behind a major trough (10 cases), and strong subtropical high (5 cases). In addition to the above two categories, there are also propagating episodes that did not have favorable conditions at 500 hPa (discussed in section 3d). A total of three types are identified in this category of “weak 500-hPa forcing” (Table 1): strong subtropical high with short trough (six cases), westerly flow behind a trough (four cases), and weak westerly flow (five cases). Thus, a total of 86 (out of 123) cases were grouped into 10 common types of flow patterns, while each type consists of at least 4 and up to 15 cases. On the other hand, the remaining 37 cases (12 propagating and 25 nonpropagating) had patterns that are weaker and do not meet the criteria in Table 1 or were less common to produce meaningful composites and, thus, were not classified. Our focus is on the 10 identified flow types more commonly seen, and all of them will be further described in greater details in section 3.

To examine the evolution of synoptic patterns in relation to the propagation of episodes (or their termination) in the identified types, the (closest) synoptic times when each streak passed 100°, 105°, 115°, and 120°E (in the Hovmöller diagram) were also identified in addition to that for 110°E. These times are referred to as T1–T5 from west to east, and are listed in Tables 2–4 for all cases in the three categories, respectively. In these tables, the synoptic times for the streaks to cross 110°E (i.e., T3) are given in full, while other times are specified as relative to T3 for simplicity. Again, for nonpropagating episodes, these synoptic times were estimated using extrapolation if needed. Then, composites were made for T1–T5 using ECMWF data at p levels of 925, 850, 700, 500, and 300 hPa for each type. Thus, basic flow patterns and their evolution associated with propagating and nonpropagating cloud episodes at various levels throughout the troposphere can be discussed and compared, but...
later only selected plots will be shown. Finally, for the rainfall distribution of identified cases and their composites, we employ the Tropical Rainfall Measuring Mission (TRMM) 3B42 merged satellite rain rates, with a horizontal resolution of 0.25° × 0.25° at 3-h intervals (Huffman et al. 2007). Since the TRMM data are available only since 1998, cases in 1997 (accounting for 11% of the sample) were not included in the composites.

3. Types and composites of synoptic patterns

a. General results

As described in section 2, a total of 10 types of 500-hPa synoptic flow patterns in three categories were identified to be commonly associated with propagating or nonpropagating episodes in East Asia (Table 1). For propagating events, four types with strong forcing and another three types with weak forcing at 500 hPa were identified, and together they account for 61 out of 73 cases examined. Thus, as many as 83.6% of the propagating cases can be classified as having a common 500-hPa flow pattern, especially for those with strong forcing (46 cases, or 63%). The high percentage of propagating events occurring under certain types of synoptic flow patterns also implies the importance of such conditions and, thus, the need for the present study. For the 50 nonpropagating episodes before classification, 25 of them were grouped into three common types of flow patterns (Table 1). The identification rate is hence 50%, considerably lower than that of propagating events, as expected.

While the synoptic times used in all cases from T1 to T5 for composites of each flow type are listed in Tables 2–4, their preferences within the diurnal cycle are briefly examined in Table 5. From previous studies (section 1), the preferred (synoptic) times for propagating cloud/precipitation episodes to occur for T1–T5 are determined to be 0600–1200, 1800–0000, 0000–0600, 0600–1200, and 1800–0000 UTC, respectively. These time windows often receive a considerably higher share of cases (boldfaced in Table 5), and it is evident that the episodes tend to occur in phase with the preferred diurnal signal regardless of whether they propagate for long distances or not. Among T1–T5, the percentage for all cases is the highest (77.9%) near the eastern edge of the TP (at 100°E, T1) and gradually decreases toward the east to 59.3% (at 120°E, T5). If only the 61 propagating episodes are considered, all percentages of in-phase cases are increased (by roughly 2%–4%) and vary from 82.0% at T1 to 60.7% at T5. Obviously, the thermal effects of the TP still play an important role in the initiation of these episodes, in agreement with several recent studies (e.g., Huang et al. 2010; Trier et al. 2010; Bao et al. 2011). In the following
TABLE 5. Numbers of cases used for composites in the three categories of propagating and nonpropagating cloud episodes at T1–T5 (when the streaks passed 100°, 105°, 110°, 115°, and 120°E in the Hovmöller diagram, respectively) vs the synoptic times (UTC, every 6 h) within the diurnal cycle. The total case number for each category is given in parentheses, and boldface font indicates the times roughly in phase with the preferred propagating diurnal signal (at 0600/1200 UTC for T1, 1800/0000 UTC for T2, 0000/0600 UTC for T3, 0600/1200 UTC for T4, and 1800/0000 UTC for T5, respectively). The percentages of propagating and all cases roughly in or out of phase at T1–T5 are given at the bottom.

<table>
<thead>
<tr>
<th>Category</th>
<th>Synoptic time (UTC)</th>
<th>T1 (100°E)</th>
<th>T2 (105°E)</th>
<th>T3 (110°E)</th>
<th>T4 (115°E)</th>
<th>T5 (120°E)</th>
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<td>7</td>
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<tr>
<td>Percentage of 61 propagating cases only (%)</td>
<td>In phase</td>
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<td>72.1</td>
<td>70.5</td>
<td>65.6</td>
<td>60.7</td>
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<td>Out of phase</td>
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<td>27.9</td>
<td>29.5</td>
<td>34.4</td>
<td>39.3</td>
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<tr>
<td>Percentage of all 86 cases (%)</td>
<td>In phase</td>
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<td>70.9</td>
<td>68.6</td>
<td>61.6</td>
<td>59.3</td>
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<td></td>
<td>Out of phase</td>
<td>22.1</td>
<td>29.1</td>
<td>31.4</td>
<td>38.4</td>
<td>40.7</td>
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</table>

subsections, the composites of common types of 500-hPa flow patterns associated with propagating and nonpropagation episodes from T1 to T5 are presented and discussed.

b. Patterns with strong 500-hPa forcing

In the category of strong 500-hPa forcing, four types of flow patterns are identified to be commonly associated with propagating episodes (Table 1, top third). These four types have a total of 46 cases, by far the most among all three categories, and here their composites and evolution (from T1 to T5) are presented and discussed. The first type (type 1) is called a trough–downstream ridge, as the episodes are associated with a clear trough and its downstream ridge inside the study domain at 500 hPa, and contains 15 cases and the most among all 10 types (Table 1). In type 1, the trough–ridge pair exhibits appreciable amplitude at least 10° in all cases, and a cutoff low even appears (from the trough) in some cases (not shown). In Fig. 1, the Hovmöller diagrams of hourly IR $T_{BB}$ data for the 15 type-1 propagating episodes with strong 500-hPa forcing are shown as examples.

Figure 2 presents the composite flow patterns from the 15 type-1 (trough–downstream ridge) propagating episodes with strong 500-hPa forcing at 500 and 850 hPa at T1, T3, and T5. At 500 hPa, a pair featuring a well-defined trough and its downstream ridge (roughly 10°–15° apart) amplifies as it moves across the study domain (25°–40°N, 100°–120°E) in the composites (Figs. 2a–c). The mean trough exhibits peak $\zeta$ values of 4–5 × 10$^{-5}$ s$^{-1}$ and the moving episodes near or ahead of the trough are thus generally located in regions with strong west-southwesterly (WSW) flow (up to about 35 kt; 1 kt = 0.5144 m s$^{-1}$) and positive vorticity advection (PVA) at 500 hPa. At 850 hPa, the development of a low near 30°N, 105°E, often referred to as the southwest vortex (e.g., Kuo et al. 1986, 1988), under the 500-hPa trough and its development with an eastward-extending trough are both clear (Figs. 2d–f). Due to the significant tightening of the (horizontal) height gradient (i.e., frontogenesis) east of the deepening low, the southwest vortex (e.g., Kuo et al. 1986, 1988), under the 500-hPa trough and its development with an eastward-extending trough are both clear (Figs. 2d–f). As isotherms are nearly perpendicular to the SW edge and near the episodes (Figs. 2d–f). As isotherms are nearly perpendicular to the SW edge and near the episodes (Figs. 2d–f). As isotherms are nearly perpendicular to the SW edge and near the episodes (Figs. 2d–f).
The second type (type 2) of propagating episodes with strong 500-hPa forcing is “ahead of a major trough.” Its pattern (at T3) differs from type 1 in that the downstream ridge is unclear and the trough extends into the midlatitudes and thus is typically deeper and has a longer wavelength (Table 1). At 12, type 2 has the second-most cases among all types. Their Hovmöller plots of $T_{BB}$ data are similar to those in Fig. 1 and, thus, not shown here (as are the other types in this section). The composite flow patterns at 500, 850, and 700 hPa are presented in Figs. 3 and 4. Associated with greater curvature and $\zeta$ maxima (about $3.5 \times 10^{-5}$ s$^{-1}$) near 30°N, the 500-hPa trough merged with the main trough farther north after T1 while moving eastward (Figs. 3a–c). Ahead of the 500-hPa trough, strong WSW flow (up to about 40 kt) prevails near the episodes. At 850 hPa, a southwest vortex also appears immediately ahead of the upper-level trough, and the SSW flow strengthens to about 20 kt along with an increase in height gradients (Figs. 3d–f). Compared to type 1 of the same category, the region of strong SSW flow at 850 hPa in type 2 appears broader and extends into the northwestern quadrant of the subtropical high (STH, east of about 125°E) at T5, but the resulted convergence is less concentrated and weaker (cf. Figs. 2d–f). The evolution at 700 hPa (Fig. 4) reveals that the southwest vortex is located near 30°N, 100°E at T1, and becomes migratory with concurrent development when the 500-hPa trough moves across the region. The enhancement of the SW flow ahead of the trough/low is also evident at 700 hPa, and the maximum wind speed reaches about 22 kt at T5 (Fig. 4).

The third type (type 3) in the category of propagating episodes with strong 500-hPa forcing is a “moving short wave” (10 cases). As indicated by the criteria in Table 1, the flow pattern of type 3 has no significant trough–ridge system but rather a moving short-wave trough inside the study domain often near 30°N, sometimes with another short-wave trough about 20° downstream. For this type, Figs. 5 and 6 give the composites at 500 and 850 hPa, respectively. At 500 hPa, the composite short-wave trough is the strongest in terms of $\zeta$ (reaching $5 \times 10^{-5}$ s$^{-1}$) at T2, and roughly moves from 100°E at T1 to 110°E at T4 across the central part of the study domain (Fig. 5). The trough farther downstream also appears in the composite east of 120°E, and a short ridge exists between the two troughs. At T5, the moving short wave weakens in the composite (not shown). At 850 hPa (Fig. 6), the low over southwestern China (near 28°N, 103°E) develops toward the
FIG. 2. Composites of type-1 flow patterns (trough–downstream ridge) with strong 500-hPa forcing in association with propagating episodes. Geopotential height ($z_F$, gpm, contour), horizontal wind (kt, arrow lines), and relative vorticity ($\zeta$, $10^{-5}$ s$^{-1}$, shaded) at 500 hPa at (a) T1, (b) T3, and (c) T5. (d)–(f) As in (a)–(c), but showing $z_F$ (gpm), horizontal wind (kt), and divergence ($D$, $10^{-5}$ s$^{-1}$, shaded) at 850 hPa. Contour intervals for $z_F$ are 20 gpm in (a)–(c) and 10 gpm in (d)–(f), and troughs (thick dashed) and ridges (thick dotted) are drawn. For $D$ at 850 hPa, only negative values (i.e., convergence) are plotted in (d)–(f).
FIG. 3. As in Fig. 2, but for composites of type-2 flow patterns (ahead of a major trough) with strong 500-hPa forcing in association with propagating episodes at T1, T3, and T5.
east-northeast (ENE) from T1 to T4 with an extending trough, although its center remains stationary. Coincident with the development of this low (also clear at 700 hPa but not shown), the southerly flow to the east also gradually strengthens and extends eastward. At the leading edge of the southerly flow and near the extending trough, areas of convergence over $-1 \times 10^{-5}$ s$^{-1}$ are found at 850 hPa (Fig. 6).

The fourth and final type (type 4) of propagating episodes with strong 500-hPa forcing is "confluence with a short wave" (nine cases), indicating a large-scale confluent flow pattern in which a moving disturbance is embedded without a significant trough–ridge system (Table 1). For type 4, the composites at 500, 850, and 700 hPa are shown in Figs. 7 and 8. In the 500-hPa composites, a cold vortex is located over northeastern China and the confluent pattern is evident inside the figure domain (Figs. 7a,b). From T1 to T5, two disturbances move across from the west, the first from 101° to 118°E and the second from 96° to 105°E. Linked to the PVA forcing from the approaching disturbances and the large-scale confluent flow aloft, the southwest vortex (near 26°N, 101°E) deepens and frontogenesis takes place near the domain center at 850 hPa, over the southern side of the confluent zone (Figs. 7c,d; e.g., Carlson 1991, chapter 14; Keyser and Shapiro 1986; Trier et al. 2006). The accompanying SSW flow over southern China also intensifies to about 25 kt from T1 to T3, with strong convergence (over $-2 \times 10^{-5}$ s$^{-1}$) at its leading edge near 28°N. This same pattern is maintained through T5 with little change at 850 hPa (not shown). At 700 hPa (Fig. 8), a ridge and high pressure region develop under the 500-hPa confluence zone, and the eastward development of the southwest vortex is clearer than at 850 hPa. The strengthening of the SW flow southeast of the vortex is also more pronounced at 700 hPa, from less than 20 kt at T1 to over 25 kt at T5 (Fig. 8).

In Figs. 2–8, there are some common ingredients among the four types of flow patterns most frequently associated with propagating episodes. One factor is the PVA forcing provided by migratory disturbances associated with $\zeta$ maxima in the upper levels (at 500 hPa). At low levels, concurrent development of the low/vortex over southwestern China and a tightening of the height gradient to its southeast, and subsequently an increase in SSW/SW wind speed and convergence, are also seen in all four types. Even in the composites, the low-level wind speed can reach at least 20 kt, close to the 25-kt criterion typically used for LLJs (e.g., Chen and Yu 1988; Chen and Chen 1995; Chen et al. 2005). Thus, a coupling in the upper-level and low-level lows/troughs provides dynamic forcing and favorable conditions for the development and maintenance of propagating, rain-producing episodes.

The composites of TRMM total rainfall from T1 to T5 in this category of strong 500-hPa forcing (Fig. 9) indicate that widespread precipitation occurs inside the study domain in all four types, as expected. The rainfall maxima often exhibit an east–west-elongated shape along about 30°N, with a higher peak amount in type 1.
 (>35 mm; Fig. 9a) and lower in types 3 and type 4 (<30 mm; Figs. 9c,d). When compared with 850-hPa patterns, the rainfall areas are located near and south of the front/wind-shift line and/or over the convergent zone near the leading edge of the strong SW flow (cf. Figs. 2d–f, 3d–f, 6, and 7c,d).

c. Patterns with strong 500-hPa inhibition

In this section, the three common types in the category of strong 500-hPa inhibition associated with non-propagating episodes are presented (Table 1, bottom third), as two of them have patterns (and criteria) that are nearly opposite to those in section 3b. The three types together have 25 cases, accounting for half of the 50 nonpropagating episodes examined. This classification rate is considerably lower than those for the other two categories (also sections 2 and 3a), as 500-hPa flow patterns of nonpropagating events at T3 tend to be more diverse and dissimilar from each other (not shown). The first type (type 1, 10 cases) is ridge–downstream trough and consists of a clear ridge with a downstream trough at 500 hPa (Table 1). As shown in Fig. 10 as examples of Hovmöller plots, these cases (as well as the remaining ones in the same category) originate near or west of 100°E over the eastern TP, but disappear and cease to propagate mostly by 105°E (and all by about 110°E). Thus, using
As if they continue to propagate through extrapolation (as described in section 2), the composites can reveal flow patterns that are not only unfavorable but also can suppress and terminate episodes propagating out from the TP.

The composite type-1 flow patterns (ridge–downstream trough) of nonpropagating episodes with strong 500-hPa inhibition at T1 and T3 are shown in Fig. 11. Opposite to Fig. 2, there exists a prominent ridge–downstream trough pair at 500 hPa (Figs. 11a,b). Although the deep trough exhibits large $\zeta$ values (peaking near $5 \times 10^{-5}$ s$^{-1}$), it is too far east and the episodes are under the control of the ridge and the northerly winds with negative vorticity advection (NVA). At 850 hPa, a pair of high–low systems corresponding to the 500-hPa ridge–trough also develops, leaving the entire area of southern China with weak flow conditions (Figs. 11c,d). After T3, similar ridge–trough and high–low patterns remain (while traveling slowly downstream) through T5 at both 500 and 850 hPa (not shown). Thus, the suppression of upper-level dynamics (i.e., synoptic-scale sinking motion) and a lack of a moisture supply at low levels cause the episodes moving off the eastern TP to weaken and dissipate quickly, and these conditions are in sharp contrast to
those seen in Fig. 2 for propagating episodes with strong 500-hPa forcing.

The next type (type 2, 10 cases) is behind a major trough, in which the episodes (if they would not dissipate) constantly stay behind a deep trough at 500 hPa (Table 1), in contrast to ahead of a major trough for propagating events in Fig. 3. In the composite flow pattern at 500 hPa at T3 (Fig. 12a), there exists a slow-moving deep trough, and the episodes would be located under the northwesterly flow behind the trough, again over regions with NVA and synoptic-scale sinking motion. At 850 hPa, a high pressure area with a northward-extending ridge develops over 25°–40°N just behind the upper-level trough (Fig. 12b) and migrates slowly eastward (not shown). So, the region surrounding the episodes, if they remained in existence, is controlled by high pressure with weak winds in type-2 composites at 850 hPa, similar to the conditions in type-1 nonpropagating episodes (cf. Figs. 11c,d).

The third type (type 3) of the nonpropagating episodes with strong 500-hPa inhibition is strong subtropical high, where the ridge of STH in each case extended deep into the study domain (Table 1). The 500-hPa composite at T3 shows a quasi-steady STH much stronger than all of other types so far, with its ridge extending westward to 108°E along about 26°N (Fig. 13a). Near 32°N, 103°E,
a weak short trough also exists but moves toward an area with strong westerly winds about 10° north of the ridge. At 850 hPa, the composite at T3 indicates that SW flow of about 10–18 kt prevails over much of southern China, but the resulting convergence is either too far northeast (along about 34°N) and mostly over the ocean, or too far south along the coast near 20°N (Fig. 13b). As at 500 hPa, this low-level pattern evolves very slowly with time (not shown). Thus, although the low-level SW flow is not weak, the 500-hPa STH and the associated sinking motion seem

![Fig. 8](image_url)

**Fig. 8.** As in Fig. 7, but showing $z_{500}$ (gpm, contour), horizontal wind (kt, arrow lines), and wind speed (shaded) at 700 hPa, for composites of type-4 flow patterns (confluence with a short wave) with strong 500-hPa forcing in association with propagating episodes, at (a) T1 and (b) T5. Contour intervals and plotting styles are the same as in Fig. 4.

![Fig. 9](image_url)

**Fig. 9.** The averaged TRMM total accumulative rainfall (mm) from T1 to T5 among all cases in type- (a) 1, (b) 2, (c) 3, and (d) 4 flow patterns of propagating episodes with strong 500-hPa forcing.
effective at suppressing the development of deep convection (and thus rain-producing episodes).

From Figs. 11–13, some common ingredients are also found in nonpropagating episodes with strong 500-hPa inhibition. Even though convective systems develop over the eastern TP, they soon move into unfavorable environments dominated by sinking motion either from upper-level dynamical forcing of NVA or from a strong, westward-extending STH. At low levels, the flow is consequently weak (types 1 and 2) with limited moisture flux (not shown), or remains quasi-steady with little convergence (type 3) in the vicinity of the episodes (if they would not dissipate). The TRMM total rainfall composites for this category indicate only a limited amount of rain over the focus area and, thus, are not shown here.

d. Patterns with weak 500-hPa forcing

In addition to the two categories with strong upper-level forcing and inhibition, there are also propagating episodes that do not seem to have favorable conditions at 500 hPa, as mentioned in section 2. Thus, this third category is referred to as weak 500-hPa forcing and it is less frequent (with a total of 15 cases) than the other two categories, especially the one with strong upper-level forcing. However, three types can still be identified, with at least four cases in each category (Table 1, middle third).

The first type (type 1) of propagating episodes with weak 500-hPa forcing is strong subtropical high with a short trough, and the Hovmöller plots of its six cases are shown in Fig. 14. When compared with similar plots for events with strong 500-hPa forcing (e.g., Fig. 1), many episodes in Fig. 14 (and those for the other two types in this section), not surprisingly, appear more intermittent and less well organized due to the weak forcing aloft. At first glance, the composite 500-hPa flow pattern at T1 for this type (Fig. 15a) also exhibits a strong STH similar to that for nonpropagating type-3 episodes (cf. Fig. 12a), except with an approaching short trough from the west in addition (cf. Table 1). However, the ridge of the STH is more to the south (near the coast of China), thus allowing for a stronger height gradient and SW flow (about 15–20 kt) over southern China, and the short wave strengthens and continues to move eastward at T3 (Figs. 15a,b). At 850 hPa, the southwest vortex develops and strong southerly flow (with a peak near 35 kt) appears over southern China from about 20° to 32°N, in the NW quadrant of the STH (Figs. 15c,d). Along the northern edge of the southerly flow, strong convergence (over $-2 \times 10^{-5} \text{s}^{-1}$) also exists near the trough, extending from the vortex at 850 hPa. In composites at 700 hPa, strong SW winds also prevail over southern China with a maximum speed over 25 kt since T2 (not shown).

The second type (type 2) in the category of weak 500-hPa forcing is westerly flow behind a trough and has the fewest cases (four) among all types. Using identical criteria (Table 1), its pattern at 500 hPa (Figs. 16a–c) is very similar to that of type 2 with strong upper-level inhibition (the behind a major trough type; cf. Fig. 12a) and also dominated by westerly-to-northwesterly flow behind a deep trough. On the other hand, a short trough approaches from the west and large-scale confluence also exists near this trough (Figs. 16a–c), so the pattern is also somewhat reminiscent to that of type-4 propagating
episodes with strong 500-hPa forcing (confluence with a short wave; cf. Figs. 7a,b). The moving short wave grows from the time of T1, and eventually becomes the dominant trough in the study domain by T5. At 850 hPa, a high pressure system develops under the region of 500-hPa confluence and NVA behind the trough by T1 (Fig. 16d), and its circulation converges with the SW monsoonal flow (up to 15 kt) along a wind-shift line (near 26°N) extending from the low over southwestern China through T5 (Figs. 16d–f). Hence, despite a similar pattern aloft, the low-level conditions are quite different from those in type 2 with strong upper-level inhibition especially over southern China (cf. Fig. 12b).

The third and final type (type 3) among the propagating episodes with weak 500-hPa forcing is weak westerly flow (five cases). As revealed in Figs. 17a,b, the flow pattern of this type at 500 hPa is characterized by zonal flow with a small trough or ridge, brought about by a STH centered near 18°N, 120°E, while the wind speed is often relatively slow within a zone of about 5°–8° in latitude (near 30°–37°N). However, weak disturbances still exist and travel eastward (Figs. 17a,b). Thus, the 500-hPa pattern also bears some similarity to that of type 3 with strong upper-level forcing (moving short wave; cf. Fig. 5). At 850 hPa, the STH is located farther to the east and is not nearly as strong, and the SW monsoonal flow

Fig. 11. As in Fig. 2, but for composites of type-1 flow patterns (ridge–downstream trough) with strong 500-hPa inhibition in association with nonpropagating episodes at 500 hPa at (a) T1 and (b) T3, and at 850 hPa at (c) T1 and (d) T3.
(peaking at about 20 kt) can still penetrate into southern China with appreciable convergence near its leading edge (Figs. 17c,d).

The TRMM total rainfall composite for the type-1 propagating cases with weak 500-hPa forcing has the largest maxima (>55 mm) among all 10 types (Fig. 18a), and the rain is concentrated along the low-level front/convergence zone near 32°N and again at the leading edge of the southerly flow (cf. Figs. 15c,d). Over the area controlled by the STH south of 30°N, on the other hand, little rainfall exists. This indicates the importance of an abundant moisture supply in causing heavy rainfall. The averaged total rainfall for types 2 and 3 (Figs. 18b,c) is considerably less compared to type 1 and not as widespread compared to the cases with strong 500-hPa forcing (cf. Fig. 9). The latter statement is especially true for type

**Fig. 12.** As in Fig. 2, but for composites of type-2 flow patterns (behind a major trough) with strong 500-hPa inhibition in association with nonpropagating episodes at (a) 500 and (b) 850 hPa at T3.

**Fig. 13.** As in Fig. 2, but for composites of type-3 flow patterns (strong subtropical high) with strong 500-hPa inhibition in association with nonpropagating episodes at (a) 500 and (b) 850 hPa at T3.
3 where the rain is confined to the south of 28°N (Fig. 18c) but again in agreement with the extent of SW flow at 850 hPa (cf. Figs. 17c,d).

4. Pattern comparison and discussion

a. Pairs with reversed patterns at 500 hPa

In section 3, some types of propagating and non-propagating episodes are found to associate with synoptic patterns nearly opposite to each other. Two pairs of such instances between the categories with strong forcing and inhibition at 500 hPa are apparent: the two type-1 patterns (trough–downstream ridge and ridge–downstream trough) and the two type-2 patterns (ahead of a major trough and behind a major trough). The differences in the dynamical forcing aloft in both pairs are obvious, and lead to much stronger low-level southerly flow (by about 20 kt in maxima) from the ocean into southern China in the composites of propagating types, as shown earlier in sections 3b and 3c (Figs. 2–4, 11, and 12). The contrast between these cases and the resultant rainfall is also evident. Therefore, favorable factors such as upper-level forcing and low-level moisture supply both seem important for episodes to propagate for long distances over the East Asian continent.

b. Pairs with similar patterns at 500 hPa

In addition to cases that are nearly opposite, there are also 500-hPa patterns that are similar but associated with propagating (section 3d) and nonpropagating episodes (section 3c). In such propagating types with weak forcing, the upper-level conditions are quite similar to those with strong inhibition and not particularly favorable, but somehow the episodes can be sustained and even produce large rainfall amounts (in type 1; cf. Fig. 18a). Therefore, it seems that a strong forcing aloft is not always required and the differences between these types, especially at low levels, can play a key role in determining the maintenance and further propagation of episodes. Two pairs of such types are discussed here: the first is type-2 weak forcing (westerly flow behind a trough) and type-2 strong inhibition (behind a major trough), and the second is type-1 weak forcing (strong STH with a short trough) and type-3 strong inhibition (strong STH).

Figure 19 presents the differences between the patterns of type-2 weak forcing and type-2 strong inhibition at 500 and 850 hPa. Although the two patterns share identical criteria (Table 1), and are both characterized by strong northwesterly-to-westerly flow behind a deep trough at 500 hPa and seem quite similar (Figs. 12a and 16a–c), subtle differences exist between propagating and nonpropagating events (Figs. 19a,b). The phase of the trough in type-2 weak forcing is farther east (by about 5°–10°) and there is an incipient short-wave trough (cf. Figs. 12a and 16b for T3), so a high pressure system (anticyclone) appears near 110°E at T1 (and later moves to 122°E at T4) followed by a low/trough with cyclonic flow (Figs. 19a,b). Thus, a (differential) low–high pattern with southerly winds in between emerges at 500 hPa. At 850 hPa, likewise, additional southerly–southeasterly flow is found behind the high/ridge but ahead of the incipient low/trough to channel the moisture supply toward southern China (Figs. 19a,b). Similar features also appear at 700 hPa and the southerly flow can reach almost 20 kt (not shown). So, differences exist throughout the lower troposphere and are not just confined within the shallow layers.

The second pair with similar 500-hPa patterns is between type-1 weak forcing and type-3 strong inhibition, both with strong STH to the south. The differences in their patterns at T3 are shown in Fig. 20. Except that the ridge of the STH in the propagating cases (type-1 weak forcing) is located at lower latitudes (south of 25°N) as noted earlier, the STH itself is also more prominent and thus an east–west-oriented ridge (with positive $z_{dp}$ differences) appears near the coast of southern China.
(Fig. 20a). Also, the incipient disturbance (of type-1 weak forcing) near 35°N, 104°E is stronger with a southward-extending trough. Between the ridge to the south and low/trough to the NW, strong SW differential flow (nearly 20 kt) results (Fig. 20a). At 850 hPa, a ridge also appears south of 25°N (Fig. 20b), indicating a stronger (composite) STH in the type-1 weak forcing (cf. Figs. 13b and 15d). Thus, additional southerly flow also exists over southern China (Figs. 14d and 20b). Again, similar differences are observed at 700 hPa (not shown). Thus, although the upper-level forcings may seem similar between the two pairs examined above, discernible differences exist in their corresponding patterns at low levels. Specifically, propagating episodes are associated with significantly stronger southerly flow through the deep layer, which transports moisture from the south to their vicinity (e.g., Chen and Chen 1995), resulting in heavy rainfall (cf. Fig. 18a). The stronger vertical shear at low levels can also help organize and sustain convective systems (e.g., Rotunno et al. 1988; LeMone et al. 1998; Doswell 2001; Tuttle and Carbone 2004).

c. Thermodynamic conditions and discussion

The thermodynamic and stability conditions of the three categories are examined by a comparison of the mean maximum convective available potential energy (CAPE)
FIG. 16. As in Fig. 2, but for composites of type-2 flow patterns (westerly flow behind a trough) with weak 500-hPa forcing in association with propagating episodes at (a)–(c) 500 and (d)–(f) 850 hPa at T1, T3, and T5.
and convective inhibition (CIN; Colby 1984) computed from the ECMWF 1.125° analysis and averaged over different longitudinal bands (each 5° in width) inside the Hovmöller domain and over all cases from T1 to T5 in Fig. 21. The averaged CAPE values in all categories are limited west of 105°E (<100 J kg⁻¹) and increase toward the east to reach a peak near 115°E with at least a moderate amount of 400–750 J kg⁻¹ (Figs. 21a–c). For the two categories of propagating episodes, this basic distribution agrees well with the location of the 850-hPa LLJ (section 3b). Their CAPE values over 105°–115°E are all higher at T1 but decrease afterward (Figs. 21a,b). This initially higher value can be viewed as a favorable precondition for convection (e.g., Johnson and Mapes 2001) while the decrease after T1 is most likely linked to the release of instability by active convection, in agreement with Figs. 9 and 18 and the findings of Trier et al. (2010). For the nonpropagating category, the CAPE east of 100°E is consistently higher than those of propagating cases, especially toward T4 and T5, clearly due to accumulation. The mean CIN values are limited west of 110°E and are all below 50 J kg⁻¹ across the region (Figs. 21d–f). Although the differences are small in general, CIN values are nonetheless slightly higher for nonpropagating episodes (often by 10–20 J kg⁻¹), as expected, which is likely due to the strong dynamical suppression (section 3c).
There are two necessary ingredients for deep convection to occur: instability (CAPE) and a convective trigger (i.e., forcing to uplift air parcels to overcome CIN and achieve free convection), which can be at upper or low levels such as a front, wind-shift line, or a convergent zone/area (e.g., Johnson and Mapes 2001). Once initiated, outflow from old convection can also trigger new development (e.g., Doswell 2001; Tuttle and Carbone 2004; Wang et al. 2009, 2011b). To organize and sustain the propagation of MCSs through cold-pool dynamics for long hours, low-level vertical wind shear (at the propagation direction) and continuous supply of moisture are both needed as well (e.g., Rotunno et al. 1988; LeMone et al. 1998; Tuttle and Carbone 2004; Trier et al. 2006). For the four types of propagating episodes with strong 500-hPa forcing, which provides synoptic-scale ascent and can serve as a convective trigger (section 3b), other ingredients of CAPE (Fig. 21a), strong vertical shear provided by the LLJ, and moisture supply clearly also exist in the composites (Figs. 2–8 and section 4a). Developed possibly in response to the forcing aloft, the frontal zone and low-level convergence often present at 850 hPa correspond well with rainfall (Fig. 9) and obviously also act to initiate new convection. For nonpropagating episodes with strong inhibition at 500 hPa, synoptic-scale descent (from NVA or the STH) dominates the region of our interest (section 3c). Although CAPE is comparable (Fig. 21c), there is a lack of low-level shear and moisture supply due to weak flow and, likely, also a trigger for types 1 and 2 (Figs. 11 and 12). Likewise for type 3, a triggering mechanism at sufficient strength to overcome the sinking from the STH is apparently not in place, although the 850-hPa SW flow is stronger (Fig. 13). Therefore, at least one or two of the necessary ingredients (i.e., instability, trigger, vertical shear, and moisture supply) for long-lived propagating convective systems is missing for each of the three types with strong upper-level inhibition.

For the three types of propagating episodes with only weak forcing aloft, the corresponding flow patterns at low levels (850 and 700 hPa) are more favorable than those of nonpropagating episodes, although the forcing at 500 hPa is limited (sections 3d and 4b). A common feature in Figs. 15–17, 19, and 20 is the much stronger southerly-to-southwesterly flow (i.e., LLJ) that penetrates into southern China and provides moisture and vertical shear. In type 1 (strong STH with a short trough), this is brought about by an additional height gradient mainly from a stronger STH at low latitudes with a ridge extending E–W (Figs. 15 and 20). It can play a role similar to that of the nocturnal LLJ with high equivalent potential temperature $\theta_e$, as examined by Trier et al. (2006) and Tuttle and Davis (2006). Thus, new convection is more easily triggered (and so happened) in propagating cases with weak 500-hPa forcing (as compared to nonpropagating ones) through low-level convergence at the leading edge of the LLJ (as suggested by rainfall; cf. Fig. 18a) and outflow boundaries of old convection, and all four ingredients are in place. The differences in low-level flow here are in general agreement with those found recently by Wang et al. (2011a, their Fig. 5).

Climatologically, the mei-yu rain belt starts around mid-May in southern China (and Taiwan) then migrates northward near mid-June to the YRV, and again to central China north of the YRV in about mid-July during the transition period of the East Asian summer monsoon (e.g., Tao and Chen 1987; Lau et al. 1988; Chen et al.

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**Fig. 18.** As in Fig. 9, but for type- (a) 1, (b) 2, and (c) 3 flow patterns of propagating episodes with weak 500-hPa forcing.
Along with this shift in the primary rain belt is the gradual retreat of upper-level westerly flow to the north of the TP as has been shown in several studies (e.g., Wang et al. 2004, 2005; Xu and Zipser 2011; Bao et al. 2011). In the midsummer during the post-mei-yu season, the STH also strengthens and moves north, with its ridge extending deep into southern/central China (e.g., Chen 2004; Chen et al. 2004; Xu and Zipser 2011). Such a pattern is quite similar to Fig. 13a while the low-level winds tend to be weak (or from the south) as in Figs. 11c,d and 12b for the nonpropagating cases (also Bao et al. 2011). As shown previously (Wang et al. 2004, 2005, 2011c; Xu and Zipser 2011), in midsummer the propagation of rain-producing systems significantly weakens or even ceases (especially over southern China). Based on our results and the above discussion, this weakening of propagation to the lee of the TP in midsummer is most likely due to a general lack of westerly vertical wind shear and moisture supply at low levels (brought by the SW LLJ) and/or the suppression by the STH, since instability and some
means of a convective trigger (such as local circulations) still exist. It is also noted that rainfall over central and southern China in midsummer has a predominately late-afternoon peak that is essentially nonpropagating (e.g., Xu and Zipser 2011; Bao et al. 2011), in agreement with and supportive of our assessment here. For the LLJ that provides moisture, warmth, and westerly low-level shear in East Asia, its diurnal oscillation in intensity and location is not yet known as well as the Great Plains LLJ in the United States (e.g., Zhong et al. 1996; Trier et al. 2006), and deserves further research.

5. Conclusions and summary

In an attempt to better understand the roles played by the synoptic environment, the ECMWF gridded analyses at 1.125° latitude–longitude and 6-h resolutions were used to identify 500-hPa flow patterns commonly associated with propagating (long lived) and nonpropagating (short lived) cloud/rainfall episodes during the warm season to the lee of the Tibetan Plateau over the East Asian continent (over 25°–40°N, 95°–125°E). A total of 123 cases (73 propagating and 50 nonpropagating) during May–July 1997–2003 were inspected based on their patterns when they moved across 110°E (domain center), and 86 of them were classified into 10 frequently observed types in three categories, that is, propagating episodes under strong and weak (synoptic scale) forcing, and nonpropagating ones under strong inhibition at 500 hPa, respectively. Each type contains at least 4 and at most 15 cases, and the propagating cases have a higher classification rate (61/73 or 83.6%) than nonpropagating ones (25/50 or 50%). Then, composites of flow patterns at various levels in each type when the episodes passed 100°, 105°, 110°, 115°, and 120°E (referred to as T1–T5) were made, presented, discussed, and compared to further elucidate the relation between the evolution of synoptic conditions, the propagation of episodes (or the cessation of further propagation), and the associated rainfall obtained from TRMM satellite observations.

For the category of propagating episodes with strong 500-hPa forcing (section 3b), four types were identified with the largest number of cases (9–15 in each type, 46 total), and they are named trough–downstream ridge, ahead of a major trough, moving short wave, and confluence with a short wave. In their composites, clear and favorable dynamical forcing of PVA from either an approaching trough or a migratory shortwave disturbance exists at 500 hPa. Concurrent development of the southwest vortex and (mei-yu) front/wind-shift line with an increased height gradient also leads to strong SSW/SW flow and convergence over southern/central China (near the YRV) at low levels, thus providing the moisture and vertical shear needed to sustain the episodes and their propagation.

For the category of nonpropagating episodes with strong inhibition at 500 hPa (section 3d), three types are identified with a total of 25 cases: ridge–downstream trough, behind a major trough, and strong subtropical high. As indicated by their names, the first two types
have 500-hPa patterns that are almost opposite to those of the first two types with strong forcing mentioned above, and the study area is dominated by descent from NVA with weak winds at low levels. In the third type, the STH at 500 hPa extends well into southern/central China to suppress the convection, despite stronger 850-hPa winds that can reach 15–18 kt in the composite. Thus, after their initiation over the eastern TP, convective systems soon move into unfavorable environments downstream and dissipate.

In addition to the two categories described above, there are also propagating episodes (15 cases total) with apparently only weak forcing aloft in three types: strong subtropical high with a short trough, westerly flow behind a trough, and weak westerly flow. The differences in their composites from those of nonpropagating episodes are primarily at low levels, where the southerly/southwesterly flow in propagating cases is stronger and can penetrate at least into southern China.

Since our analysis of thermodynamic conditions indicates that instability exists in all types, the cessation of nonpropagation cases must be a result of a lack of other crucial ingredients. Through a comparison of low-level patterns, these factors are determined to be the vertical shear and moisture supply provided by a stronger southerly-to-southwesterly low-level flow (i.e., LLJ) and/or a convective trigger of sufficient strength. The result that the vertical shear is essential to episode propagation in many cases here is consistent with the findings of several earlier studies that the propagation characteristics of rainfall to the lee of the TP weaken in midsummer when prevailing upper-level westerly flow shifts to the north.

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**Fig. 21.** (a)–(c) Maximum CAPE (J kg\(^{-1}\)) averaged over different longitude bands of 5° width centered at 100°, 105°, 110°, 115°, and 120°E (abscissa) and 25°–40°N, and among all cases with (a) strong forcing, (b) weak forcing, and (c) strong inhibition at the synoptic scale at 500 hPa from T1 to T5 (curve styles indicated). (d)–(f) As in (a)–(c), but for CIN (J kg\(^{-1}\)). The number of cases in each category is given in parentheses.
It is also noted that additional types may be identified in the future using more cases and help further our understanding of warm-season propagating episodes in East Asia.

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REFERENCES


