Collaborative Effects of Cold Surge and Tropical Depression–Type Disturbance on Heavy Rainfall in Central Vietnam

SATORU YOKOI
Center for Climate System Research, University of Tokyo, Chiba, Japan

JUN MATSUMOTO
Department of Geography, Tokyo Metropolitan University, Tokyo, and Institute of Observational Research for Global Change, Japan Agency for Marine-Earth Science and Technology, Kanagawa, Japan

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ABSTRACT

This paper reveals synoptic-scale atmospheric conditions over the South China Sea (SCS) that cause heavy rainfall in central Vietnam through case study and composite analyses. The heavy rainfall event discussed in this study occurred on 2–3 November 1999. Precipitation in Hue city (central Vietnam) was more than 1800 mm for these 2 days. Two atmospheric disturbances played key roles in this heavy rainfall. First, a cold surge (CS) northerly wind anomaly in the lower troposphere, originating in northern China near 40°N, propagated southward to reach the northern SCS and then lingered there for a couple of days, resulting in stronger-than-usual northeasterly winds continuously blowing into the Indochina Peninsula against the Annam Range. Second, a southerly wind anomaly over the central SCS, associated with a tropical depression–type disturbance (TDD) in southern Vietnam, seemed to prevent the CS from propagating farther southward. Over the northern SCS, the southerly wind anomaly formed a strong low-level convergence in conjunction with the CS northeasterly wind anomaly, and supplied warm and humid tropical air. These conditions induced by the CS and TDD are favorable for the occurrence of heavy orographic rainfall in central Vietnam. The TDD can be regarded as a result of a Rossby wave response to a large-scale convective anomaly over the Maritime Continent associated with equatorial intraseasonal variability.

Using a 24-yr (1979–2002) reanalysis and surface precipitation datasets, the authors confirm that the coexistence of the CS and TDD is important for the occurrence of heavy precipitation in central Vietnam. In addition, it is observed that CSs without a TDD do not lead to much precipitation.

1. Introduction

Vietnam is located along the eastern coast of the Indochina Peninsula (ICP) and is separated from other parts of the peninsula by the Annam Range, which lies along the coast (Fig. 1). The seasonal march of precipitation in Vietnam differs markedly from place to place because of its substantial latitudinal extent from 8° to 22°N (Matsumoto 1997). In the northern part, the main rainy season is September–October, while that in the southern part is November–December. In other words, the maximum precipitation area shifts southward from a northern location in September to a southern location in December (Yokoi et al. 2007).

The central part of Vietnam is characterized by a narrow lowland plain with a width narrower than 100 km. The main rainy season in this region is the October–November period. Climatologically, precipitation in Hue city (16.4°N, 107.7°E; see Fig. 1), for example, is more than 1400 mm in this 2-month period, which is about half of the annual rainfall.

The October–November period is also known as the transition phase from boreal summer to winter monsoon seasons. Over the South China Sea (SCS), boreal summer southwesterly monsoon winds retreat in September, followed by the appearance of northeasterly monsoon winds around 20°N in early October (Fig. 2). The northeasterly winds blow into the eastern ICP against the Annam Range, bringing monsoon rainfall...
through orographic rainfall processes. The northeasterly wind axis then shifts southward to reach 10°N in December in concert with the maximum precipitation area in Vietnam.

During the rainy season in central Vietnam, frequent heavy rainfall events often menace the local population. One of the heaviest rainfall events occurred on 2–3 November 1999. Figure 3 shows the precipitation amount for these 2 days obtained from the Tropical Rainfall Measurement Mission (TRMM) 3B42 dataset (Huffman et al. 2007). Large amounts of precipitation exceeding 600 mm (2 days) were recorded in central Vietnam. This heavy rainfall was also observed at surface weather stations. For example, at the Hue observation site more than 1800 mm of precipitation was recorded in these 2 days (Fig. 4), which was more than 60% of the mean annual precipitation and was the largest amount on record in the previous 50 yr. In addition, precipitation amounts greater than 700 mm (2 days) were recorded at Da Nang (16.0°N, 108.2°E) and Nong Son (15.7°N, 108.0°E), which are located southeast of Hue city. This heavy rainfall event caused several rivers to flood, resulting in severe destruction in downstream cities. Hereafter, this event is called NOV99 for brevity.

It is well known that synoptic- and large-scale atmospheric disturbances cause precipitation variability in the ICP during the summer monsoon season, with a wide range of time scales from several days to several weeks (e.g., Yokoi and Satomura 2005; Yokoi et al. 2007). It is quite possible that such disturbances also play some role in precipitation variability in the eastern coastal area during the monsoon transition phase. As will be shown in this paper, the following two atmospheric disturbances played important roles in the NOV99 event: a cold surge (CS) and a tropical depression–type disturbance (TDD) associated with tropical intraseasonal variability (ISV) with a 30–60-day time scale.

The CS is known as an episodic southward progression of cold, midlatitude air along the eastern flank of the Tibetan Plateau in the lower troposphere, and is observed mainly in the November–March period. Spatiotemporal structures of the CS have been revealed by a number of studies. The CS northerly wind signal first appears near 40°N in northern China in association with a trough–ridge system aloft (Joung and Hitchman 1982; Lau and Lau 1984; Wu and Chan 1997; Chen et al. 2002) after the intensification of the Siberian High (Ding 1990; Zhang et al. 1997). The CS signal then propagates southward into the tropics within a few days (Chu and Park 1984; Sumi 1985; Wu and Chan 1995; Compo et al. 1999). The arrival of the CS is characterized by a sharp increase in surface pressure, strong northerly winds, and a decrease in temperature (Chang et al. 1979; Johnson and Priegnitz 1981; Chang et al. 2005; Wu et al. 2007).

The ISV prevails in convection and tropospheric circulation anomalies in the tropics. It is dominated by the Madden–Julian oscillation (MJO; Madden and Julian 1994), which is characterized by eastward-propagating convective and zonal wind signals with a zonal wave-number of 1 or 2. Its characteristics exhibit pronounced seasonality (e.g., Knutson and Weickmann 1987). During boreal summer, the southwesterly monsoon winds and other seasonal conditions in south and Southeast Asia modify the MJO structure and behavior (Lau and
The convec-
tive signal propagates northward in the Asian monsoon region and western Pacific, as well as eastward along the equator in the Eastern Hemisphere (Yasunari 1979; Murakami and Nakazawa 1985; Krishnamurti et al. 1985; Lau and Chan 1986). The convective and low-level zonal wind anomalies in south and Southeast Asian monsoon regions form a band structure oriented from a west-northwest to east-southeast direction (Murakami and Nakazawa 1985; Lau and Chan 1986). Cyclonic circulation anomalies aligning along this band move northwestward because of Rossby wave dynamics (Kemball-Cook and Wang 2001).

Although these two atmospheric disturbances have been studied in many papers as introduced heretofore, the disturbances that occur during the monsoon transition phase have yet to be described in detail. Most of the studies dealing with CSs have targeted the mature winter monsoon season, although CSs occur most frequently in November (Zhang et al. 1997), and the CSs that occur during the monsoon transition phase exhibit features that are different from those during the winter monsoon season (Compo et al. 1999). In addition, some studies on ISV exclude the monsoon transition phase in order to present its exemplary features. However, for understanding the causal mechanisms of heavy rainfall events in central Vietnam, it is necessary to study the behavior of disturbances during the monsoon transition phase. Therefore, the purpose of this study is to describe the synoptic-scale atmospheric conditions and disturbances during the NOV99 event in detail, and to discuss the mechanisms responsible for heavy rainfall in central Vietnam. To confirm the mechanisms, we will also conduct composite analyses using 24 yr (1979–2002) of reanalysis and precipitation data.

The rest of this paper is organized as follows: Section 2 describes the datasets used in this study. The structure and behavior of the atmospheric disturbances during the NOV99 event are presented in section 3, and composite analyses are discussed in section 4. Section 5 presents a summary and discussion.

2. Data

The atmospheric circulation field dataset used in this study is the Japanese 25-yr Reanalysis (JRA-25; Onogi et al. 2007) provided by Japan Meteorological Agency (JMA) and Central Research Institute of Electric Power Industry (CRIEPI). Its horizontal resolution is 1.25° in longitude and latitude, and its temporal interval is 6 h. There are 12 mandatory levels in the tropical troposphere (1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, and 100 hPa). This paper will focus primarily on the 925-hPa wind field. Note that there are seven levels below the nominal 850-hPa level in the JRA-25 assimilation model, so the 925-hPa level is well resolved. To eliminate the diurnal cycle, a 1–2–2–1 temporal filter is applied to the reanalysis data in section 3 and a daily average is used in section 4. The analysis is performed over the 24-yr period from 1979 to 2002. To describe synoptic-scale disturbances, we also analyze anomaly data, which are constructed by removing the first three harmonics of the climatological seasonal cycle.

This study also uses daily station rainfall data in the eastern ICP. These data were routinely collected by the Department of Meteorology and Hydrology of Laos (at 8 stations), the Department of Meteorology of Cambodia (at 4 stations), and the National Hydrometeorological Service of Vietnam (at 52 stations).

In addition, daily mean interpolated outgoing longwave radiation (OLR) data provided by the National Oceanic and Atmospheric Administration (NOAA;
3. Synoptic-scale atmospheric conditions during NOV99

In this section, we will examine the synoptic-scale circulation field in the lower troposphere during the NOV99 event. The horizontal wind field at the 925-hPa level at 1200 UTC 2 November (Fig. 5a) was characterized by a strong northerly wind blowing from the Yellow Sea into the northeastern part of the ICP along the eastern coast of the Eurasian continent. While the wind direction was considerably similar to the climatological northeasterly monsoon wind in early November (Fig. 5b), the wind speed over the northern SCS during the NOV99 event was 22 m s$^{-1}$, which was twice as strong as the climatological wind speed of 11 m s$^{-1}$. The equivalent potential temperature field (Fig. 5c) shows that a warm and humid tropical air mass existed south of the northern SCS and a front-like zone with a large horizontal gradient lay in a west-southwest to east-northeast direction. A part of this zone was also reported as a stationary front in a surface weather chart issued by JMA, which is represented by a thick solid line in Fig. 5a. Because the maximum wind speed was observed very close to the 335-K iso-equivalent potential temperature line, the stronger-than-usual northeasterly wind transported relatively thermally unstable air into the ICP, creating conditions capable of activating orographic rainfall processes over and upstream of the Annam Range.

The stronger-than-usual northeasterly wind was associated with a CS from the midlatitudes. A time–latitude cross section of meridional wind at the 925-hPa level averaged over 110°–120°E (Fig. 6a) shows that a northerly wind signal stronger than −6 m s$^{-1}$ originated near 40°N at 0600 UTC 31 October, which was associated with a trough–ridge system aloft (figure not shown). The signal propagated rapidly southward to 25°N within 18 h, and then slowly reached the northern SCS south of 20°N by 0000 UTC 2 November. During the southward propagation, the northerly wind signal was accompanied to its south by a horizontal convergence anomaly (Fig. 6b) and to its north by a negative temperature anomaly (Fig. 6c). These characteristics are consistent with well-known CS features derived by...
many researchers, as given in section 1. It is worth noting that over the SCS only a weak convergence anomaly was observed at the 850-hPa level, whereas a strong divergence anomaly was found to the west of the 925-hPa convergence (figure not shown). This seemed to represent the thin vertical structure of the CS (Chen et al. 2002).

One of the distinctive features of this CS was observed when the northerly wind signal reached the northern SCS. While ordinary CS signals propagate farther southward to at least 10°N, and sometimes across the equator (Chang et al. 1983; Love 1985; Wu et al. 2007), the signal of this CS stopped propagation and lingered there until 0000 UTC 5 November (Fig. 6a). Concurrently, an easterly wind began to strengthen at the same latitude (Fig. 6d). As a result of this lingering feature, stronger-than-usual northeasterly winds continued for a couple of days.

To discuss the reason for this lingering feature, Fig. 7a shows the meridional wind anomaly at the 925-hPa level at 1200 UTC 2 November. A southerly wind anomaly with a 6 m s⁻¹ magnitude was found over the central SCS centered at 11.25°N, 112.5°E. This southerly wind anomaly seemed to prevent the CS from propagating farther southward, possibly through changing the dynamic balance on the meridional propagation of the CS. Because the southward propagation speed of the CS in 20°–25°N can be estimated from Fig. 6a at about 5 m s⁻¹, the southerly wind anomaly of 6 m s⁻¹ could block the southward propagation. However, it is difficult to discuss the mechanism in more detail, because there is no consensus regarding the dynamics of the southward propagation of CS signals in the subtropics.

In addition to the blocking effect mentioned above, the southerly wind anomaly seemed to contribute to the heavy rainfall in the following two ways: First, the anomaly formed a strong horizontal convergence in the lower troposphere over the northern SCS (Fig. 6b) in collaboration with the northeasterly wind of the CS. Second, the anomaly tended to transport the warm and humid tropical air into the northern SCS, particularly

![Fig. 6](image-url)

FIG. 6. (a) Time–latitude cross section of meridional wind at the 925-hPa level averaged in 110°–120°E; contour interval is 2 m s⁻¹, and dark (light) shading indicates values higher than +6 m s⁻¹ (lower than −6 m s⁻¹). (b) Same as (a), except for the horizontal divergence anomaly; contour interval is 5 × 10⁻⁶ s⁻¹, and dark (light) shading indicates values lower than −1 × 10⁻⁵ s⁻¹ (0 s⁻¹). (c) The same as in (a), but for temperature anomaly; contour interval is 1 K, and dark (light) shading indicates values higher than +1 K (lower than −1 K). (d) The same as in (a), but for zonal wind; contour interval is 2.5 m s⁻¹, and shading indicates values lower than −10 m s⁻¹.
prior to the arrival of the CS. These effects made the lower troposphere more unstable and favorable for convection. To confirm the second effect, Table 1 shows horizontal fluxes of equivalent potential temperature at the 925-hPa level across the boundaries of a 10°–20°N, 110°–120°E box in the northern SCS. Because of the northeasterly monsoon wind, climatological flux is inward at the northern and eastern boundaries and outward at the other two boundaries. As for the NOV99 event, on the other hand, the flux at the southern boundary was inward and horizontal convergence in the box was stronger than that of the climatology. The differences in the flux at the southern boundary and convergence were primarily due to the advection of the climatological equivalent potential temperature by the southerly wind anomaly, shown in the last row of Table 1.

To confirm the unstable condition over central Vietnam, Fig. 8 shows a time series of convective available potential energy (CAPE) at Xisha Dao station (Fig. 1), calculated from the sounding data. During 29–30 October, CAPE fluctuated around 900 J kg⁻¹; then CAPE increased dramatically on 31 October to reach a maximum of more than 2000 J kg⁻¹, and maintained high values until 2 November. This increase of CAPE represents that the atmosphere became more unstable. On 3 November, CAPE fell to nearly zero and the atmosphere became stable, probably because of the approach of the CS bringing cold air into the lower troposphere.

This southerly wind anomaly was a component of a cyclonic circulation anomaly centered at the southeastern ICP, which is marked as B in Fig. 7b. This circulation anomaly had a horizontal scale of about 2000 km and appeared to combine with the northeasterly wind anomaly of the CS in its northeastern sector. The vertical structure of this circulation anomaly was almost standing below the 300-hPa level, with a slight inclination to the west with height (figure not shown). The longitude–time cross section of the meridional wind anomaly (Fig. 9) shows that the southerly wind anomaly had appeared at 140°E by 28 October and then propagated westward to reach 110°E on 2 November. These spatiotemporal structures are considerably similar to that of the TDD (Takayabu and Nitta 1993).

In addition to disturbance B, there were two other TDDs: one was over the eastern coast of the Indian subcontinent and the other was southeast of the Philippines (marked as A and C, respectively, in Fig. 7b). The southerly wind anomaly associated with disturbance C had appeared by 1 November and propagated westward (Fig. 9). Likewise, disturbance A had been generated in the southern SCS by 28 October and prop-

Table 1. Horizontal fluxes of equivalent potential temperature at the 925-hPa level passing across northern, southern, eastern, and western boundaries of, and horizontal convergence integrated over, the 10°–20°N, 110°–120°E box in units of 10⁸ K m² s⁻¹. Values of early November climatology and just before the NOV99 event (31 Oct–1 Nov average) are shown. Positive values indicate inward flux and convergence. The difference between the NOV99 event and climatology, and fluxes of the climatological equivalent potential temperature by the wind anomaly (u' Θe) and resultant convergence anomaly are also shown.

<table>
<thead>
<tr>
<th></th>
<th>Northern</th>
<th>Southern</th>
<th>Eastern</th>
<th>Western</th>
<th>Convergence</th>
</tr>
</thead>
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<td>21.5</td>
<td>-22.4</td>
<td>+7.6</td>
</tr>
<tr>
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<td>-24.3</td>
<td>+13.0</td>
</tr>
<tr>
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<td>+14.5</td>
<td>-0.7</td>
<td>-1.9</td>
<td>+5.4</td>
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<tr>
<td>u' Θe</td>
<td>-6.5</td>
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agated northwestward along the southern coast of the ICP. These three disturbance centers aligned in a west-northwest to east-southeast direction, accompanied by a westerly wind anomaly band to their south (Fig. 7b).

In the equatorial region, the large-scale circulation field was characterized by a MJO-type signal. The velocity potential anomaly field at the 150-hPa level (Fig. 10a) exhibited a zonal wavenumber 1 structure. Its negative anomaly was found in Southeast Asia and the western Pacific, with a negative maximum over the Maritime Continent. This implies the presence of a center of large-scale divergence and an upward branch of a tropical circulation anomaly. Accompanying this velocity potential anomaly was a negative OLR anomaly, which indicates active convection. These negative anomalies originated over the Indian Ocean in late September and propagated eastward at a speed of about 3 m s\(^{-1}\), with a slight weakening in amplitude just east of 90\(^\circ\)E (Fig. 10b). These characteristics are consistent with those of a typical MJO, as reported by many studies (e.g., Murakami and Nakazawa 1985; Krishnamurti et al. 1985; Weickmann and Khalsa 1990). To the east of 150\(^\circ\)E, the negative OLR anomaly vanished and the negative velocity potential anomaly weakened, while the latter propagated farther eastward in the Western Hemisphere at a faster speed of about 13 m s\(^{-1}\). The differences in amplitude and propagation speed between the Eastern and Western Hemispheres are also well-known MJO features (Knutson et al. 1986; Hendon and Salby 1994; Kikuchi and Takayabu 2003). The relationship between the MJO-type convective signal centered over the Maritime Continent and the low-level circulation anomaly field in the subtropics shown in Fig. 7 was in good agreement with that of the ISV during boreal summer (Knutson and Weickmann 1987; Kemball-Cook and Wang 2001). This accordance suggests that the TDD train in the lower troposphere shown in Fig. 7b can be regarded as a result of an off-equatorial Rossby wave response to the equatorial convective anomaly of the ISV, as discussed in Kemball-Cook and Wang (2001).

The collaborative effects of the CS and TDD associated with the equatorial ISV can be summarized as follows: Because of these two disturbances, the lower troposphere over central Vietnam became thermally unstable. The stronger-than-usual northeasterly wind caused by the CS blew this unstable air into central Vietnam against the Annam Range, resulting in vigorous orographic rainfall processes. Because of the lingering feature of the CS, which was probably caused by the TDD, the northeasterly wind lasted for a couple of days, leading to a large amount of precipitation.

4. Composite analyses

To confirm the importance of the collaborative effects for the heavy rainfall in central Vietnam, which were suggested through the case study in section 3, we will perform composite analyses of CSs and TDDs found during the monsoon transition phase.

There have been a variety of CS definitions proposed by previous studies, most of which were based on a sharp drop in the surface or lower-tropospheric temperature, meridional pressure gradient, strong northerly wind, or a combination thereof. Because we focus on the CS behavior over the SCS, we set the CS definition as follows: 1) a 3-day mean meridional wind anomaly at the 925-hPa level averaged over 110\(^\circ\)–120\(^\circ\)E and 7.5\(^\circ\)–12.5\(^\circ\)N, with a slight weakening in amplitude just east of 90\(^\circ\)E (Fig. 10b). These characteristics are consistent with those of a typical MJO, as reported by many studies (e.g., Murakami and Nakazawa 1985; Krishnamurti et al. 1985; Weickmann and Khalsa 1990). To the east of 150\(^\circ\)E, the negative OLR anomaly vanished and the negative velocity potential anomaly weakened, while the latter propagated farther eastward in the Western Hemisphere at a faster speed of about 13 m s\(^{-1}\). The differences in amplitude and propagation speed between the Eastern and Western Hemispheres are also well-known MJO features (Knutson et al. 1986; Hendon and Salby 1994; Kikuchi and Takayabu 2003). The relationship between the MJO-type convective signal centered over the Maritime Continent and the low-level circulation anomaly field in the subtropics shown in Fig. 7 was in good agreement with that of the ISV during boreal summer (Knutson and Weickmann 1987; Kemball-Cook and Wang 2001). This accordance suggests that the TDD train in the lower troposphere shown in Fig. 7b can be regarded as a result of an off-equatorial Rossby wave response to the equatorial convective anomaly of the ISV, as discussed in Kemball-Cook and Wang (2001).

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poral minimum in the 5-day period from 2 days before to 2 days after the meridional wind minimum. The second condition can exclude strong TDDs and typhoons over the western Pacific or Philippines that do not accompany CSs, which can also cause northerly wind anomalies over the northern SCS. In the 24-yr (1979–2002) October–November periods, we identify 45 CS cases (hereafter ON-CS). Although the CS during the NOV99 event also meets the above definition, we exclude this case from composite analyses. In addition, we identify 40 CS cases in the 24-yr January–February periods (hereafter JF-CS) for evaluating the seasonality of CS features.

Figure 11 shows the lag–latitude cross section of the composite meridional wind anomaly averaged at the 925-hPa level over 110°–120°E. In this figure, a lag of zero days corresponds to the day when the CSs satisfy the first condition of the definition. Composite meridional wind anomalies of JF-CS and ON-CS over the Eurasian continent (north of 20°N) are qualitatively similar to each other; statistically significant northerly wind anomalies propagate southward from the midlatitudes. Other atmospheric conditions (e.g., temperature and pressure anomalies and upper-tropospheric circulation fields) of both CS groups are also qualitatively similar to each other. A notable difference between JF-CS and ON-CS is found to the south of 15°N. While the JF-CS northerly wind signal propagates farther southward to near the equator, the ON-CS meridional wind anomaly hardly exhibits any statistically significant signal south of 10°N. To examine the difference in more detail, the composite meridional wind anomaly averaged over lags from +1 to +3 days is shown in Figs. 12a,b. In the JF-CS composite (Fig. 12a), a northerly wind anomaly with statistical significance is extensively found over the SCS and western Pacific; this is consistent with previous studies, such as that of Zhang et al. (1997). On the other hand, as for the ON-CS composite (Fig. 12b), a significant northerly wind anomaly is found only over northern SCS and the ICP, whereas the anomaly over most of the central and southern SCS, where we found the southerly wind anomaly during the
NOV99 event, is not statistically significant. In addition, the standard deviation of the meridional wind anomaly of the ON-CS cases (Fig. 12c) exhibits a local maximum in the central SCS; this is consistent with the statistical insignificance of the composite anomaly.

These results imply that the direction as well as the magnitude of the meridional wind anomaly in the central SCS is different among individual ON-CS cases. In fact, the meridional wind anomaly averaged over 10°–15°N, 110°–115°E (hereafter $v_{scs}$) ranges from $-12$ to $+8$ m s$^{-1}$.

To clarify the relationship between $v_{scs}$ and precipitation over the ICP, we categorize the 45 ON-CS cases into the following three groups: CSs where $v_{scs}$ is negatively larger than $-7$ m s$^{-1}$ [CS with strong northerly wind (CS-NW); 6 cases], CSs where $v_{scs}$ is positively larger than $+4$ m s$^{-1}$ [CS with southerly wind (CS-SW); 6 cases], and other CSs (CS-Normal; 33 cases). The composite horizontal wind anomaly over and around the SCS for the CS-SW at the 925-hPa level (Fig. 13) is quite similar to Fig. 7b, showing a TDD centered over the central SCS and a northeasterly wind anomaly blowing into the ICP. Low-level horizontal convergence and the northward transport of the warm and humid tropical air over the northern SCS are also observed as in the NOV99 event. On the other hand, the CS-NW and CS-Normal composites have horizontal wind fields that are qualitatively similar to those of the JF-CS composite over and around the SCS, although the wind speed is different. In addition, horizontal convergence at the 925-hPa level over central Vietnam is not found in these two composites.

Figures 14a–c show the composite precipitation in eastern ICP of the three above-mentioned groups averaged over lags from +1 to +3 days. Composite precipitation of CS-SW (Fig. 14c) is more than 60 mm day$^{-1}$ in the coastal areas between 13° and 16°N, which is much higher than that of CS-NW (Fig. 14a) and CS-Normal (Fig. 14b). In particular, precipitation averaged over the nine coastal stations shown in Fig. 14f is 67.4 mm day$^{-1}$ for CS-SW, which is about 3 times higher than that for CS-NW (18.9 mm day$^{-1}$) and CS-Normal.
(24.1 mm day$^{-1}$); this difference is statistically significant at a 99% confidence level. In addition, composite precipitation of CS-NW and CS-Normal is only comparable to the October–November climatology (Fig. 14e). Thus, we can conclude that a CS cannot independently cause heavy rainfall in the eastern coast of the ICP.

To examine the importance of the CS, we compare precipitation of CS-SW to that under the condition of a southerly wind anomaly over the central SCS that does not accompany a CS. We identify 62 such cases where $v_{cscs}$ is higher than $+4$ m s$^{-1}$ and is a temporal maximum, and there is no CS over the SCS and China (hereafter SW-alone). Composite horizontal wind and geopotential height anomaly maps of SW-alone at the 925-hPa level (not shown) indicate that the southerly wind anomaly is a component of a TDD centered over the central SCS. Composite precipitation of SW-alone (Fig. 14d) is less than that of CS-SW, although it is much higher than that of CS-NW and CS-Normal. In fact, precipitation averaged over the nine stations in the coastal areas (Fig. 14f) for SW-alone is 40.8 mm day$^{-1}$, which is less than that for CS-SW with statistical significance at a 95% confidence level.

Therefore, we can conclude that the coexistence of the CS and the southerly wind anomaly associated with the TDD is a more favorable condition for heavy rainfall in the eastern coastal areas of the ICP than the existence of either the CS or the southerly anomaly.

The NOV99 event described in section 3 is a typical example of such a collaborative effect.

It should be mentioned that one-half of the CS-SW cases are associated with the equatorial ISV, which has the upward branch and convective anomaly over the Maritime Continent, as in the NOV99 event. Therefore, the equatorial ISV is one of the major candidates, but of course not a necessary condition, for the generation and maintenance of the TDD that induces southerly wind anomalies over central SCS.

5. Summary and discussion

This study revealed synoptic-scale atmospheric disturbances in the East and Southeast Asian regions associated with the record heavy rainfall event that occurred in central Vietnam on 2–3 November 1999 (NOV99 event). During NOV99, a lower-tropospheric northeasterly monsoon wind over the northern SCS was strengthened by a CS and a TDD. The CS northeasterly wind signal originated near 40°N in northern China, propagated southward along the eastern coast of the Eurasia continent, and then lingered over the northern SCS for a couple of days. A southerly wind anomaly in the central SCS associated with the TDD seemed to play a role in this lingering feature. In addition, this anomaly formed a strong convergence over central Vietnam in association with the CS northeast-
erly wind anomaly, and supplied warm and humid tropical air that destabilized the lower atmosphere. As a consequence, stronger-than-usual northeasterly winds transported the unstable air into central Vietnam for a couple of days, resulting in vigorous orographic rainfall processes over and upstream of the Annam Range.

We performed composite analyses to confirm that the coexistence of the CS and TDD is important for heavy rainfall. In the October–November periods of 1999–2002, we identified 6 CSs that accompanied the strong northerly wind anomalies over the central SCS (CS-NW), 6 CSs that accompanied the southerly wind anomalies (CS-SW), and 33 CSs for which meridional wind anomalies were between the former two extreme cases (CS-Normal). The CS-SW corresponded to the coexistence cases of the CS and TDD. We also identified 62 cases in which a southerly wind anomaly associated with a TDD existed over the central SCS but did not accompany a CS (SW-alone). We showed that composite precipitation of CS-SW was much more than that of CS-NW, CS-Normal, and SW-alone, thus confirming the significance of the collaborative effects of the CS and TDD. In addition, we found that CSs without TDD could not cause much precipitation; composite precipitation of CS-NW and CS-Normal was only comparable to the October–November climatological precipitation amount.

The TDD during the NOV99 event could be regarded as the result of the Rossby wave response to the large-scale convective anomaly over the Maritime Continent associated with the equatorial ISV. In addition, the TDDs in one-half of the CS-SW cases also seemed to be associated with the ISV convective anomaly. Therefore, the ISV is one of the major candidates for the generation and maintenance of the TDD that contributes to the collaborative effect.

In addition to the collaborative effect, other factors may contribute to the occurrence of heavy rainfall. This may be explained by taking into account the fact that the amount of precipitation during the NOV99 event was the highest in the previous 50 yr, as mentioned in section 1. In addition, precipitation averaged over the nine coastal stations shown in Fig. 1f during 2–4 November 1999 was 203 mm day$^{-1}$, which is 3 times higher than that for the CS-SW composite (67.4 mm day$^{-1}$). One possible factor is an anomalous sea surface temperature (SST) in the central SCS. The NOAA Optimum Interpolation SST (OISST) dataset (Reynolds et al. 2002) reveals that the SST in the northern and central SCS during the NOV99 event was higher (by a maximum of 1 K) than the climatological value in early November. The warmer SST supplies higher amounts of sensible and latent heat to the lower troposphere, causing the convective precipitation processes to be more vigorous. Quantitative estimates are necessary for discussing this subject and will be dealt with in our future study.

As presented in section 4, we found that in the October–November period, meridional wind anomalies over the central SCS associated with CSs differ markedly among individual cases because of the existence and location of TDDs, whereas most CSs in the January–February period are accompanied by a northerly wind anomaly over the central SCS and seem to be less subject to the TDD. This seasonality can be attributed to the seasonal march in the activity of the TDD over the central and southern SCS. As a proxy for this activity, Fig. 15 shows variance of the 3–20-day filtered relative vorticity, which is calculated with the use of a Lanczos bandpass filter (Duchon 1979), which has cutoff periods of 3 and 20 days and a filter length of 41 days, at the 925-hPa level averaged in $5^\circ$–$15^\circ$N, $110^\circ$–$120^\circ$E. The variance is highest in October and lowest in January–March. Because TDDs occur more frequently in environmental trough zones, the seasonal march in the variance is consistent with the southward shift of the northeasterly monsoon wind and monsoon trough shown in Fig. 2. Because the CS occurs frequently during November–March, we can conclude that the collaborative effect of the CS and TDD studied in this paper is peculiar to the October–November period. Compo et al. (1999) noted that the composite structure of the CS during the October–November period is different from that during December–March. The difference may possibly be attributable to the diversity in the meridional wind anomaly over the central and southern SCS. To fully understand the behavior of the CS during
the October–November period, we also have to consider the effect of TDDs.

The interaction of CSs with tropical atmospheric systems during the winter monsoon season was also reported by Chang et al. (2005), who emphasized the effects of the Borneo vortex whose center was located south of 10°N. They showed that the horizontal distribution of cloud activity over the southern SCS, Malay Peninsula, and Maritime Continent during CS events depended on the existence of the Borneo vortex. The results of Chang et al. (2005) and the present study suggest that tropical cyclonic vortices, regardless of their latitude, considerably affect the distribution and strength of convection and precipitation induced by CSs.

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