Onset of the Summer Monsoon over the Indochina Peninsula: Climatology and Interannual Variations*

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ABSTRACT

The temporal and spatial structures of the atmospheric circulation associated with the climatology and interannual variations of the summer monsoon onset over the Indochina Peninsula were studied using the observed daily rainfall at 30 stations and the NCEP–NCAR reanalysis from 1951 to 1996. The climatological monsoon onset over Indochina is on 9 May, with a standard deviation of 12 days. The monsoon onset is characterized by the pronounced northeastward progression of the low-level southwesterlies over the Indian Ocean and the intensification and northward extension of the tropical convection from Sumatra. It coincides with the weakening of the midlatitude westerly over south Asia, and the westward propagation of the intraseasonal oscillation (ISO) originated in the South China Sea (SCS) and the western Pacific with a dominant timescale of 12–25 days.

A close relationship between the interannual variations of the monsoon onset and El Niño/La Niña was identified. Years with warm (cold) sea surface temperature (SST) anomalies in the western Pacific and cold (warm) SST anomalies in the central-eastern Pacific in the preceding spring have an early (late) onset. For an early onset year, strong convective activities appear over the southern Indochina Peninsula and the southern SCS in the preceding winter and spring. Associated with the changes of the Walker circulation and the local Hadley circulation related to La Niña, strong convective activities were maintained by the convergence between the anomalous southwesterlies in the Indian Ocean and northeasterlies over the northern SCS. The anomalous southwesterlies in the Indian Ocean were induced by both the anomalous Walker circulation associated with La Niña and anomalous land–sea thermal contrast. The anomalous northeasterlies over the northern SCS were originated in northern winter due to the combined effects of the cold east China land and the warm Philippine Sea, and further maintained by a positive thermodynamic air–sea feedback mechanism related to La Niña. An opposite scenario is found for a late onset year with warm SST anomalies in the central-eastern Pacific (El Niño).

1. Introduction

As one of the dominant circulation systems in the tropical troposphere, the Asian monsoon exhibits an abrupt seasonal transition from winter to summer (Yeh et al. 1959; Matsumoto 1992). It is generally recognized that the initial onset stage of the Asian summer monsoon (ASM) is characterized by the establishment of strong convection and the change of direction of the prevailing wind over the Bay of Bengal, the Indochina Peninsula, and the South China Sea (SCS) in early and middle May (He et al. 1987; Matsumoto 1992; Murakami and Matsumoto 1994; Lau and Yang 1997; Wu and Zhang 1998; Hsu et al. 1999). Following this initial stage, the evolution of the Asian summer monsoon undergoes diverse directions in association with the establishment of the east Asian monsoon (EAM) and south Asian monsoon (SAM) subsystems. The EAM, whose evolution is characterized by the northward movement of the elongated rain belt in a stepwise fashion, migrates from the SCS to the southern coast of China in late May, reaches the Yangtze Valley and southern Japan in mid-June, and finally arrives at its northernmost location in northern

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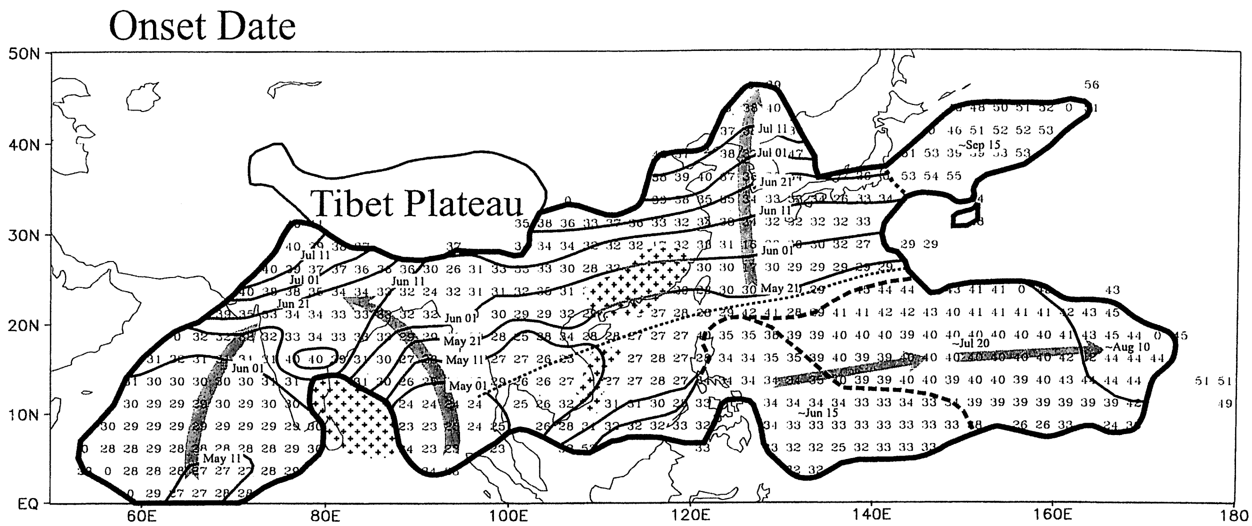

FIG. 1. Mean onset date of the Asian summer monsoon adopted from Wang and LinHo (2002). The number is the onset (Julian) pentad. The thick dashed lines denote discontinuities (merger of three or more contours). The arrows point to the directions of rain-belt propagation. The thin dashed line divides subtropical monsoon and oceanic monsoon regions. The Tibetan Plateau is outlined by the 3000-m contour.

China in late July (Tao and Chen 1987). The establishment of the SAM, on the other hand, follows two routes (Wang and LinHo 2002): the northward march of convection from south of the Arabian Sea toward the west coast of India, and the northwestward march of convection from the southeastern Bay of Bengal toward the Indian subcontinent. The former route reaches the southern tip of the Indian subcontinent by the end of May, and reaches 20°N by 10 June. The rainy season starts at Kerala, the southern tip of India, around 1 June (Joseph et al. 1994). Along the latter route, convection reaches the northern Bay of Bengal and the northeast coast of India around 6–10 June (Rao 1976).

The appearance of strong convection over the Indochina Peninsula, indicative of the earliest start of the summer monsoon over the Asian continent, has been documented in some recent literature (e.g., Lau and Yang 1997; Webster et al. 1998; Wang and LinHo 2002). Figure 1 shows the map of onset date of the Asian summer monsoon derived by Wang and LinHo (2002), in which the monsoon onset is defined by the difference between the pentad mean and the January mean precipitation rates. It is exhibited that Indochina is the region where the continental Asian monsoon begins around early May through mid-May on average. So far, it is not clear what weather regimes are responsible for the outbreak of the strong convection in this region. Matsumoto (1997) investigated the onset of the summer rainy season over the Indochina Peninsula by using 15-yr rainfall data and found that during late April to early May, the beginning of the summer rainy season, Indochina is under a midlatitude westerly regime, and full summer monsoon circulation begins to establish in mid-May. A question remains as to whether or not the tropical influence is necessary in triggering the monsoon onset over the Indochina Peninsula. As we know, Indochina is a unique region where the monsoon activity reflects a transitional feature of two distinct monsoon subsystems, the south Asian monsoon and east Asian monsoon. The former is a typical tropical climate system, whereas the latter is a combined tropical–midlatitude system (Chen and Chang 1980). The processes controlling the weather over Indochina are complex during late spring to early summer. Figure 2 shows the climatological 5-day mean horizontal wind field at 850 hPa derived from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) daily reanalysis from late April through mid-May. It is seen that during mid-April to mid-May, three branches of the prevailing winds with different sources and properties may affect the weather regime over Indochina. One is the subtropical westerly stretching from the northern Indian subcontinent to the Indochina Peninsula. Another is the southeasterly associated with the subtropical ridge over the western Pacific. The third is the westerly over the equatorial eastern Indian Ocean, which stretches northeastward in early May as the southwesterly flows. It is interesting to see what influence these three branches of the prevailing winds have on the development of the heavy monsoon rainfall over Indochina, respectively.

Much effort in the past has focused on the interannual variability of the overall monsoon strength measured by the summer season total rainfall. The variability of the monsoon onset date has received less attention, even though it is of importance to agricultural practices. Various previous studies have documented a relationship between the early/late onset of the Asian summer monsoon and El Niño–Southern Oscillation (ENSO; e.g., Joseph et al. 1994; Ju and Slingo 1995; Lau and Yang 1997; Wu and Wang 2000). In general, a late onset is associated with the warm SST anomaly (SSTA) in the

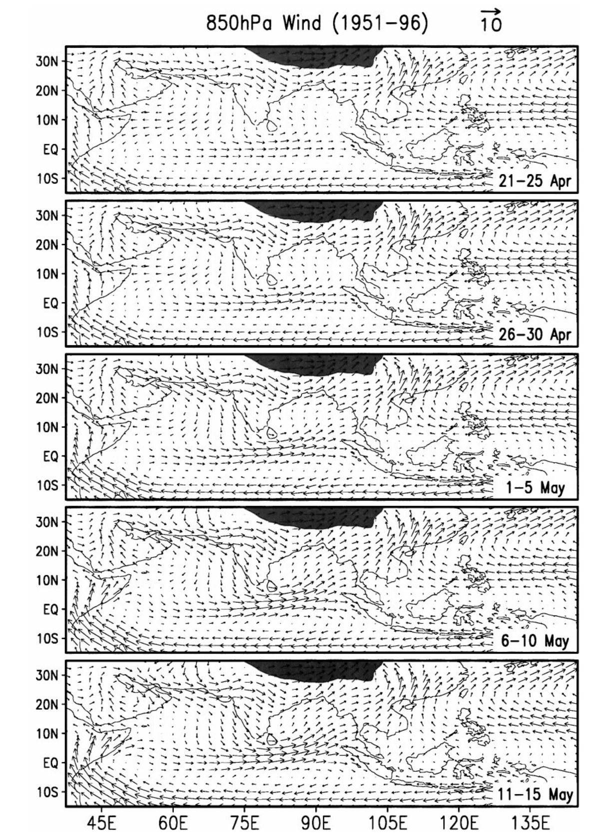

FIG. 2. Evolution of the 5-day mean wind fields at 850 hPa for 46-yr climatology in the transitional season (m s^{-1}). The shading area denotes the terrain above 3000 m.

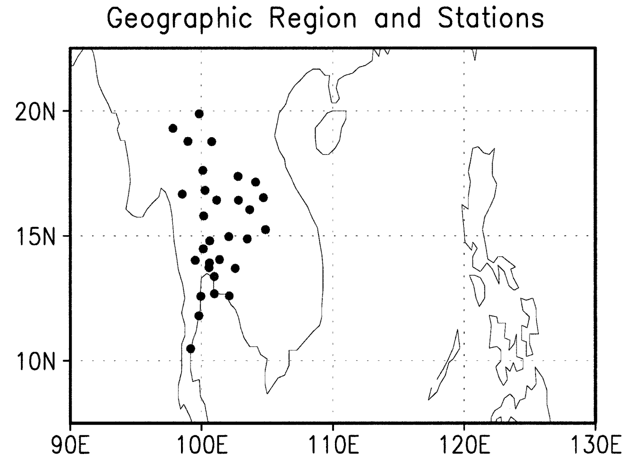

FIG. 3. Map denoting the rainfall stations used to construct the daily rainfall index.

equatorial Indian Ocean and the eastern Pacific Ocean and the cold SSTA in the western Pacific, and vice versa. Soman and Slingo (1997) and Wu and Wang (1999) pointed out that during El Niño events the warm SSTA in the equatorial eastern-central Pacific plays a major role, while during La Niña events the local warm SSTA in the western Pacific is more important. It was proposed that the influence of ENSO is through the regulation of the northward (southward) latitudinal position of the tropical convective maximum (TCM) over Indonesia and the western Pacific in the preceding spring (Ju and Slingo 1995; Soman and Slingo 1997). While these studies have raised some important issues on the monsoon–ENSO relationship, the detailed physical processes through which the eastern Pacific SSTA affects the monsoon onset deserve further investigation.

The objective of this study is to reveal the processes that are involved in the initial onset stage of the Asian summer monsoon over Indochina on annual and interannual timescales. To achieve this goal, the long-term observed daily rainfall from 30 stations over the central Indochina Peninsula is employed to construct a domain-averaged index. Then, the onset date is defined based on the seasonal evolution of this rainfall index. The daily NCEP–NCAR reanalysis data are used to compose the atmospheric circulation associated with the onset. The precursory signals in the preceding winter and spring for the early/late onset years are investigated, and the possible linkages among these signals are proposed. The rest of paper is organized as follows. Section 2 describes the data. Section 3 describes the definition of the monsoon onset. Section 4 presents the temporal and spatial structures of the atmospheric circulation during the monsoon onset and the associated intraseasonal oscillation (ISO) activities. The precursory signals associated with the interannual variations of the monsoon onset are investigated in section 5, and their underlying processes are discussed in section 6. Finally, conclusions are given in section 7.

2. Datasets

In this study, the rainfall index is represented by the daily area-averaged precipitation from 30 stations over the central Indochina Peninsula from 1951 to 1996. The geographic locations of these stations are displayed in Fig. 3. The ratio of missing data for each station is less than 1% during late spring to early summer (1 April–30 June). These rainfall gauge stations are collected from the Southeast Asian daily surface observations from NCAR during 1951–85 and the global summary of daily observations from the Climate Prediction Center (CPC) of NCEP during 1986–96.

NCEP–NCAR reanalysis for 1948–99 is the primary dataset used in this study. Monthly and daily means of the variables are created by the average of the 6-hourly data from the analysis fields. The monthly sea surface temperature (SST) is taken from Reynolds SST (Reynolds and Smith 1994) from 1950 to 1999. Daily data of outgoing longwave radiation (OLR) are derived from twice-daily National Oceanic and Atmospheric Administration (NOAA) satellite series from 1979 to 1996. It is reinterpolated, with gaps filled by the Climate Diagnostics Center (CDC).

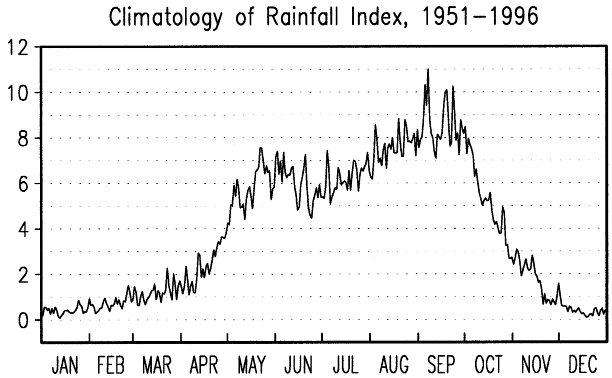

FIG. 4. Daily climatology (5-day running mean) of rainfall index for 46 years in units of mm day^{-1}.

TABLE 1. The onset date of the Asian summer monsoon over Indochina.

Year	Onset date	Year	Onset date	Year	Onset date
1951	20 May	1967	15 Apr	1983	23 May
1952	7 May	1968	23 Apr	1984	11 May
1953	22 Apr	1969	25 May	1985	25 Apr
1954	28 Apr	1970	12 May	1986	5 May
1955	11 May	1971	2 May	1987	26 May
1956	29 Apr	1972	31 May	1988	13 Apr
1957	26 May	1973	1 May	1989	12 May
1958	13 May	1974	15 May	1990	16 May
1959	18 May	1975	2 May	1991	24 May
1960	16 May	1976	28 Apr	1992	4 Jun
1961	24 Apr	1977	4 May	1993	13 May
1962	16 May	1978	10 May	1994	5 May
1963	1 Jun	1979	17 May	1995	3 May
1964	3 May	1980	18 May	1996	22 Apr
1965	4 May	1981	1 May	Mean	9 May
1966	3 May	1982	12 May		

3. Definition of the monsoon onset

Figure 4 displays the 5-day running mean climatology of the rainfall index for 46 years. The wet season over Indochina is identified from early May to mid-October while the dry season is from November to late April. The transition from a dry to a wet season is characterized by a sudden increase of the daily rainfall from 2 to 6 mm day^{-1} sometime from late April to mid-May.

In previous studies, various methods have been proposed to determine the date of the monsoon onset. Mostly, local rainfall or convection activity indicated by satellite-observed data is used (e.g., Tao and Chen 1987; Ananthakrishnan and Soman 1988; Lau and Yang 1997; Wang and LinHo 2002). Others use either the change of prevailing winds (Holland 1986) or the combined wind–convection criteria (Wang and Wu 1987; Matsumoto 1997; Xie et al. 1998). As shown in Fig. 2, the wind patterns over Indochina are very complicated. For this reason, we use only the rainfall index to define the monsoon onset. It is notable that although the monsoon onset is defined by rainfall only, the circulation fields also show clear changes during the onset as will be shown in section 4a.

Based on the strength and persistence of the rainfall over Indochina, the onset date for each individual year is defined as the day on which the 5-day running mean rainfall index satisfies the following conditions:

1) The amount of the daily rainfall index exceeds 5 mm day^{-1} and persists continuously for 5 days;
2) Within the consecutive 20 days, the number of the days with rainfall greater than 5 mm day^{-1} exceeds 10 days.

The above definition is determined among many tests in order to best reflect the change of the rainfall situation from a dry to a wet season. Compared to the method proposed by Wang and LinHo (2002), which is based on the rainfall difference between a specific pentad and the corresponding winter, we found that for most years the results from different definitions are almost the same, except for several years in which the contrast between a wet and dry season is not sharp.

According to the above definition, the monsoon onset dates for each individual year are determined objectively (Table 1). The mean onset date for the 46 years is 9 May. It is consistent with the existing results of 10 May and 6–10 May, which were reported by Webster et al. (1998) and Wang and LinHo (2002), respectively. The earliest onset year is 1988 when the monsoon onset begins on 13 April. The latest year is 1958, when the monsoon onset begins on 4 June. The standard deviation is 12 days.

Since the definition is based on the domain-averaged rainfall index, we further check whether it is representative for the individual stations. We compared the 10-day rainfall difference after onset and prior to onset (figure omitted). Over most of the region, the rainfall amount increases more than 40% after the monsoon onset, implying that the definition based on the domain-averaged rainfall index well represents the general onset over this area.

4. The change of general circulation associated with the monsoon onset

a. *The influence of the tropical circulation*

In order to reveal the large-scale circulation and convection associated with the monsoon onset, we compose daily fields based on the onset dates (by centering the onset day). The onset day is denoted as day 0 while the signs of "$-$" and "$+$" denote prior to and after the onset day, respectively. Then, the pentad mean fields are obtained with pentad 0 representing the average from day 0 to $+4$ and pentad -1 denoting the average from day -5 to day -1. Similar procedures are done for other successive pentads. To minimize the difference of the seasonal background in the extreme early and late onset years, the cases with onset dates earlier than 25 April

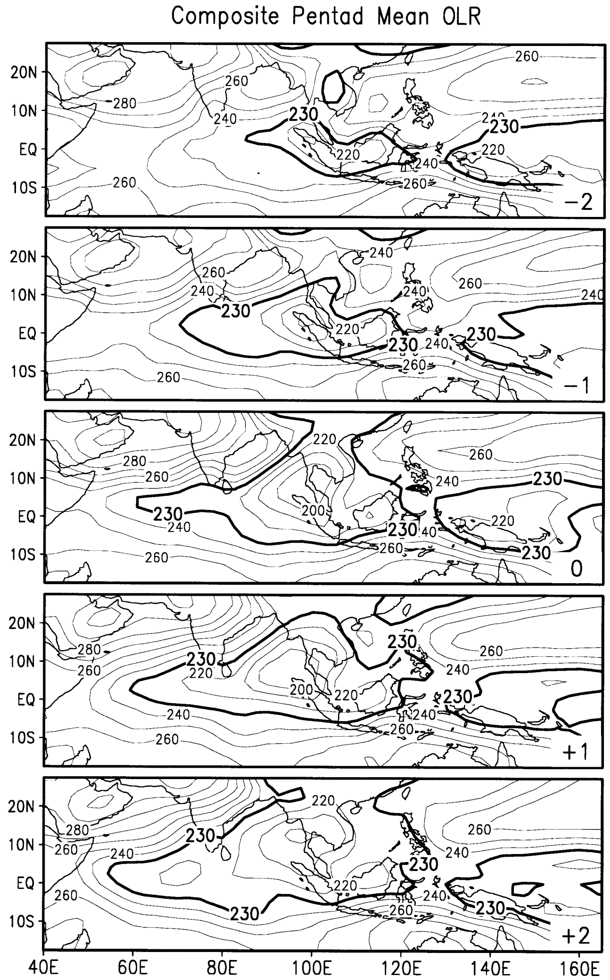

FIG. 5. Pentad mean evolution of the composite OLR with centering the onset dates. Here ''0'' denotes onset pentad and ''−'' and ''+'' refer to prior to and after onset pentad, respectively. See text for detail about the definition of the onset pentad. The interval of the contour is 10 W m^{-2} with the 230 W m^{-2} contour denoted by the thick solid line. To minimize the seasonal background differences in the extreme early and late onset years, the years of onset dates earlier than 25 Apr or later than 25 May are excluded from the composite.

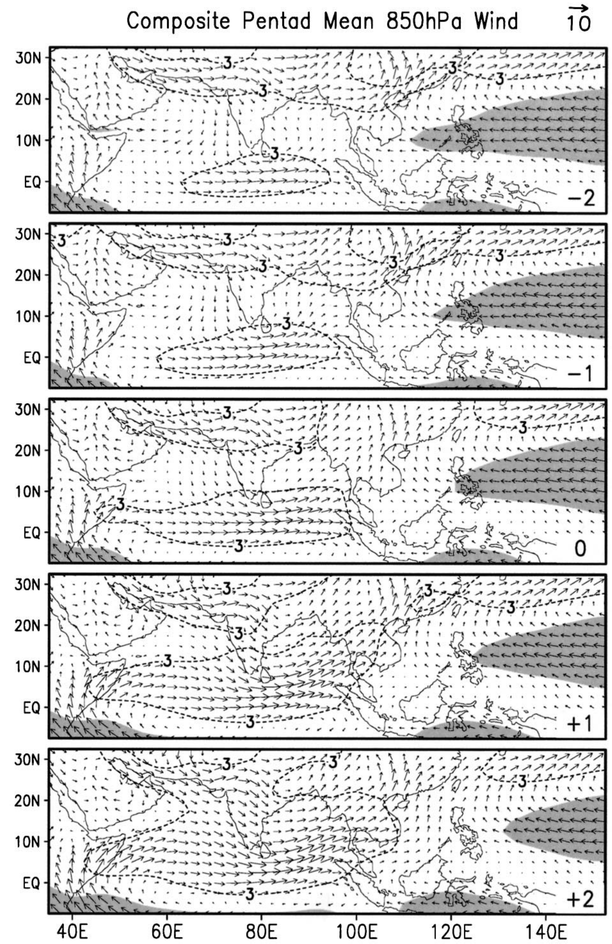

FIG. 6. Same as the Fig. 5, but for the 850-hPa wind vector. The dashed line refers to the contour of -3 m s^{-1} zonal wind and shading indicates the easterly jet with its value exceeding 3 m s^{-1}.

or later than 25 May are excluded in composite. Overall, 8 of 46 years are removed.

To show the evolution of the tropical convection associated the monsoon onset, we compose the OLR field (Fig. 5). Before the monsoon onset (at pentad -2 and -1), two minimum OLR centers (which represent the strong convection) appear over Sumatra and the western Pacific, respectively. The Bay of Bengal and the SCS are under control of the maximum OLR centers (which denote the suppressed convection). At pentad 0, the minimum OLR center over Sumatra intensifies and extends northward noticeably. The contour of 230 W m^{-2} controls most of the Indochina Peninsula. At pentad $+2$, the contour of 230 W m^{-2} covers the northern SCS and the Bay of Bengal, indicating the onset of strong convection over the two regions. It suggests that the northward extension and intensification of strong convection center over Sumatra might be responsible for the outburst of the strong monsoon rainfall over the Indochina Peninsula.

To reveal the changes of the associated lower-level circulation, we compose the 850-hPa wind field shown in Fig. 6. In the preceding pentads (-2 and -1), the strong westerly, as part of the midlatitude zonal westerly belt ranging from northwestern India to southeastern China, is controlling the central part of the Indochina Peninsula. During monsoon onset (at pentad 0), the strength of this westerly jet decreases remarkably over Indochina and the southeastern China regions. Concurrent with the rapid intensification of the westerly over the equatorial Indian Ocean, the southward wind component increases in the Indochina Peninsula. At pentad $+1$, the northeastward progression of the southwesterly flow in the equatorial Indian Ocean is clearly identified. The southwesterly jet stretches into the central Indochina from the southeastern Bay of Bengal. In contrast, the midlatitude branch of the westerly flow continuously

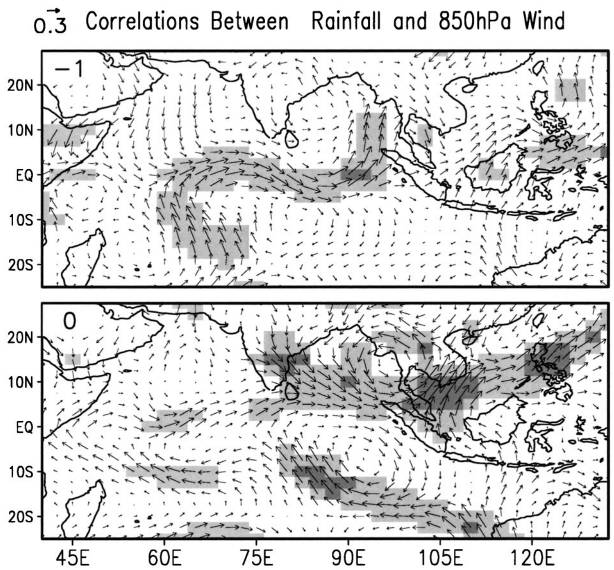

FIG. 7. Vectors of the correlation coefficient value of the rainfall at the onset pentad (pentad 0) with the 850-hPa wind at the pentad 0 and pentad −1, respectively. Slight shading represents only zonal or meridional wind, which satisfies the 5% significance level, and heavy shading indicates both the zonal and meridional component, which passes the 5% significance level test.

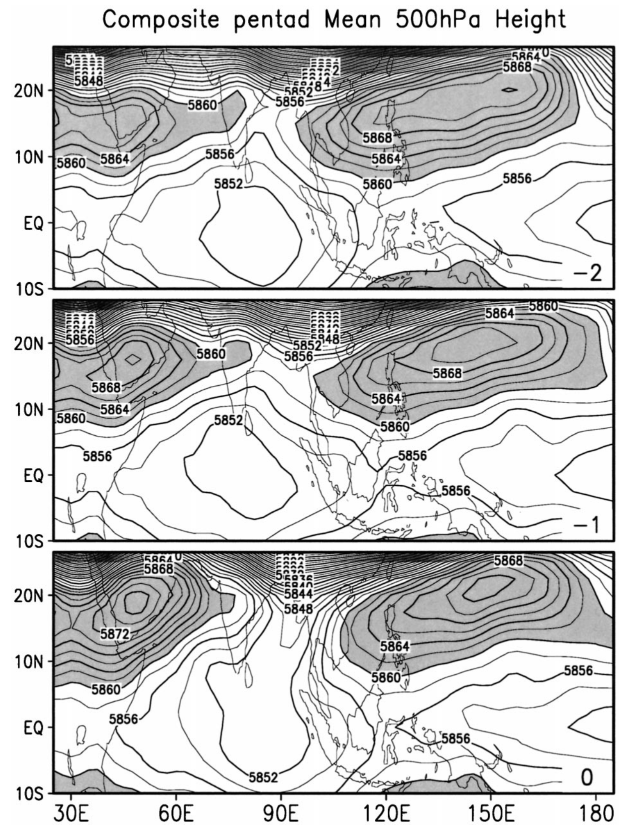

FIG. 8. Same as Fig. 5, but for 500-hPa height with shading denoting the value more than 5860 m.

decreases and retreats westward. Meanwhile, the easterly trade wind associated with the western Pacific subtropical high over the SCS and the western Pacific notably retreats eastward.

The results above indicate that the summer monsoon onset over Indochina is closely associated with the development of the southwesterly over the equatorial Indian Ocean. To further investigate this, we calculate the correlations of the rainfall at the pentad 0 with the 850-hPa wind field at the pentad −1 and 0, respectively. Figure 7 provides the vector maps of the correlation coefficients of zonal and meridional wind components. The rainfall during the onset pentad is closely correlated to a strong cross-equatorial flow in the Indian Ocean in the preceding pentad (top panel). This cross-equatorial flow provides necessary moisture supplies for the development of the strong convective activities over the southeastern Bay of Bengal and Indochina. The simultaneous correlation map (bottom panel) shows a clear anomalous cyclonic circulation centered over the western coast of Indochina, reflecting the Rossby wave response of the circulation to the strong convective heating over this region, which is presented in Fig. 5. It is also noticeable that the monsoon onset is coincident with deepening of the cyclonic circulation over the northern SCS. This may link to the ISO activities that we will discuss in section 4b.

To examine the change of the upper-tropospheric circulation during the monsoon onset, we compose the 200-hPa winds and the wind tendency fields (figures not shown). Enhancement of the northeasterly cross-equatorial flow over the eastern equatorial Indian Ocean and Sumatra during the monsoon onset is noticeable. This, along with the increase of the low-level southeasterly cross-equatorial flow, reflects the intensification of the local monsoon circulation.

Figure 8 presents the evolution of the composite 500-hPa height field during the monsoon onset. The pronounced features are the rapid intensification of a trough over the Bay of Bengal and the eastward retreat of the western ridge of the western Pacific subtropical high (shaded area) from Indochina. The intensification of the trough coincides with the strengthening of the western Pacific and northern African subtropical highs, implying a strong feedback between the convective heating over the Bay of Bengal and the western Pacific/northern African subtropical high.

The association of monsoon onset over the Indochina Peninsula with an increase of the southwesterlies over the tropical northern Indian Ocean agrees with the result of the case study of Wu and Zhang (1998), who found that the occurrence of the strong tropical southerly flow over the eastern coast of the Bay of Bengal is responsible for the earliest monsoon onset over Southeast Asia. It is clear that the summer monsoon onset differs from the early raining event discussed by Matsumoto (1997), which is primarily caused by the strengthening of subtropical westerlies.

TABLE 2. The peak of the power spectrum period that passes the 5% significance level test (red noise). The data are tested from 1 Apr to 30 Sep. Here *R* denotes the domain-averaged Indochina rainfall index and *V* denotes the domain-averaged 850-hPa meridional wind (m s^{-1}) over 5°–10°N, 102.5°–107.5°E.

Year	*R*	*V*	Year	*R*	*V*	Year	*R*	*V*
1951	22.0, 15.0	20.0, 15.0	1967	12.0	30.0, 12.0	1983	18.0	
1952	15.0	20.0, 13.8	1968	30.0, 12.9	22.5, 13.8	1984	30.0, 18.0	15.0
1953	22.5, 13.8	22.5, 15.0	1969	13.8	25.7, 16.4	1985	12.9	25.7, 15.0
1954	12.9	12.0	1970	20.0	20.0, 13.8	1986	16.4	22.5, 15.0
1955	20.0	15.0	1971	12.0	12.9	1987	12.9	25.7
1956	12.9	15.0	1972	12.9	12.9	1988	15.0	13.8
1957		16.4, 12.0	1973	25.7		1989	15.0	20.0
1958	20.0, 15.0	20.0	1974	12.0	16.4	1990	20.0, 13.8	18.0
1959	30.0, 20.0	22.5	1975	22.5, 13.8		1991		20.0, 12.0
1960	16.4	16.4	1976	30.0, 13.8	36.0, 20.0	1992		12.9
1961		30.0, 15.0	1977	15.0	15.0	1993	16.40	18.0
1962	30.0, 18.0	36.0, 12.9	1978	18.0	13.8	1994	16.4	20.0
1963	16.4	16.4, 12.0	1979	25.7	13.8	1995	36.0, 20.0	12.9
1964	15.0	15.0	1980	36.0, 12	15.0	1996	12.9	18.0
1965	20.0	30.0, 12.9	1981	15.0	18.0			
1966	15.0	25.7	1982	18.0	20.0			

Whether or not the subtropical westerlies around the Tibetan Plateau affect the monsoon onset is not conclusive here because the weakening of the subtropical westerlies is not necessarily a result of northward propagation of the southwesterlies over the Indian Ocean. In the climatological 850-hPa wind fields (Fig. 2), the subtropical westerlies over the northern coast of the Bay of Bengal do not change much during late April to mid-May, while the tropical southwesterlies increase dramatically. This may link to the potential problem of the NCEP–NCAR reanalysis data for subtropical Asia during the early years (the 1950s and 1960s) as discussed in Yang et al. (2002).

b. *The activities of the ISO associated with the monsoon onset*

When we examine the seasonal cycle of the rainfall in Indochina and the wind fields for each individual year, we find that the appearance of the strong rainfall usually exhibits certain periodic oscillations and the monsoon onset is often associated with the activities of the ISO. Actually, the activities of the ISO with time bands of 10–20 and 30–40 days have been considered as an important factor modulating the timing of the monsoon onset in India (Krishnamurti and Ardanuy 1980; Murakami 1976; Chen and Chen 1993) and the SCS (Chen and Chen 1995; Wu and Zhang 1998; Xie and Saiki 1999). To investigate what time periods are dominant in the rainfall and wind fields, the power spectrums of the daily rainfall index and the meridional wind index, which is obtained by averaging 850-hPa meridional winds over the southern SCS (5°–10°N, 102.5°–107.5°E), are analyzed. This domain is selected because the simultaneous correlation between the 850-hPa meridional wind and rainfall during the monsoon onset pentad exceeds 0.5 (Fig. 7). Table 2 lists the peak periods that pass the 5% significance level (red noise). The analysis period is from 1 April to 30 September. For the rainfall index, the 12–25-day period is identified as being significant in 43 years except in 1961, 1991, and 1992. For another period greater than 30 days, only 7 years are identified as being significant. For the meridional wind index, the 12–25-day period is dominant in 42 years. For the period greater than 30 days, only 5 years are identified. Thus, the 12–25-day period is a dominant frequency for both the rainfall and associated wind fields.

In this study a 12–25-day bandpass filter is applied to the rainfall index, NCEP–NCAR daily reanalysis, and OLR data. The relationships between the monsoon onset and the ISO activities are examined for each individual year. We find that for most years the monsoon onset is closely related to the 12–25-day ISO activities. Figure 9 shows the evolution of filtered 850-hPa vorticity and OLR averaging over 10°–20°N in 1979, 1980, and 1994. For these 3 years, the monsoon onset occurs on 17, 18, and 5 May, respectively. It is seen that the monsoon onsets are in phase with the westward propagations of positive vorticity and that the low OLR originated over the SCS and the western Pacific. This suggests that the low-frequency wave identified in the low-level cyclonic circulation and convective activities may trigger the sharp increase of rainfall in Indochina.

Figure 10 shows the composite of the filtered rainfall index and the meridional wind index (by centering the onset dates) for the 46 years. In general, a wet phase is associated with the anomalous southerly flow while a dry phase is concurrent with the anomalous northerly flow. To describe the general temporal–spatial features of the low-frequency oscillation in association with the monsoon onset, a composite study is further conducted for those 22 years in which the rainfall and meridional wind are in phase. Figure 11 presents the composite daily 850-hPa wind. It is seen that during the monsoon onset period (from day 0 to +4), there is an anomalous

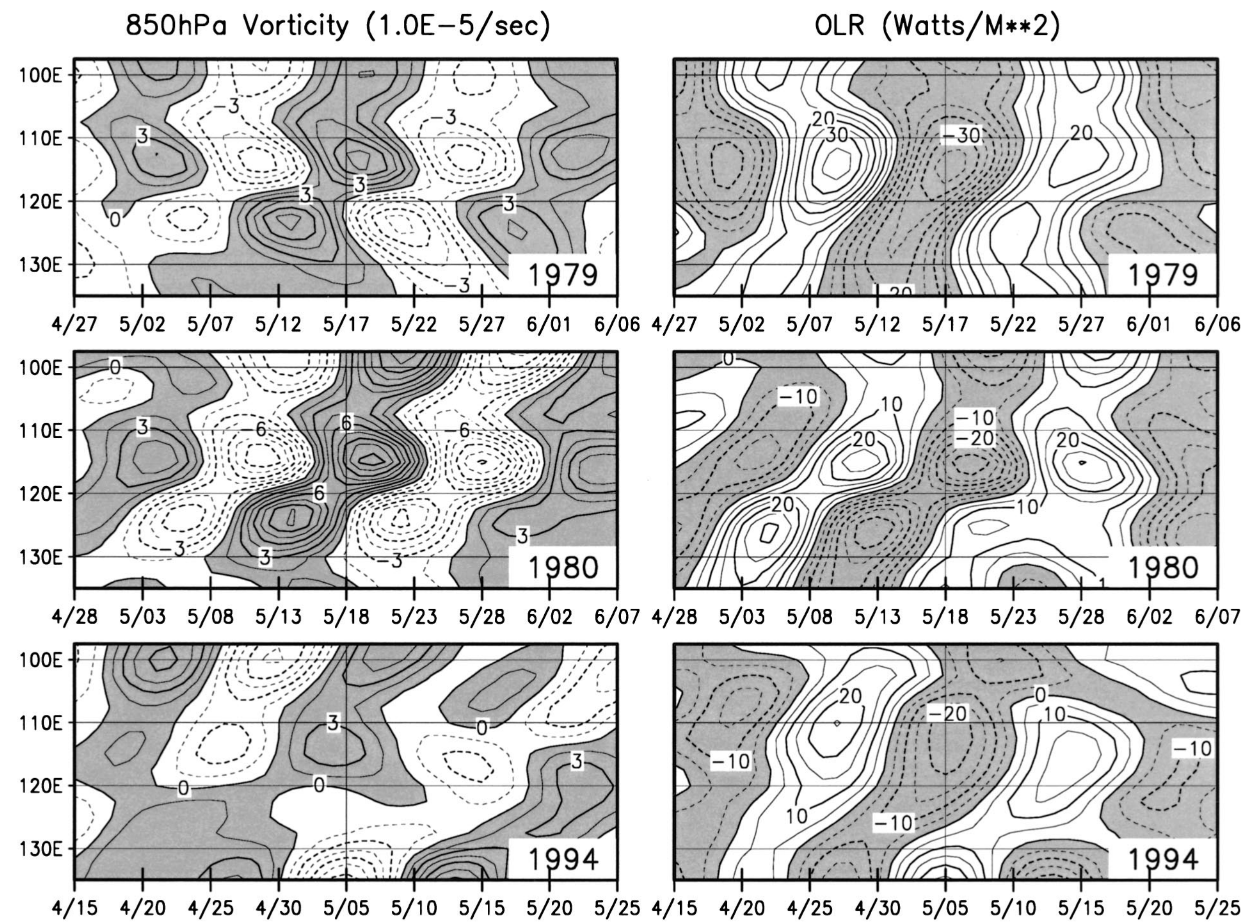

FIG. 9. The longitude–time cross sections of the 12–25-day bandpass-filtered (left) vorticity and (right) OLR at 850 hPa along 10°–20°N in 1979, 1980, and 1994, in which onset dates are 17, 18, and 5 May, respectively.

cyclone that propagates westward from the SCS. Before that, an anticyclone moves across Indochina from day −6 to day −2. Strong southerlies blow into Indochina when the cyclone arrives in the western coast of the Indochina Peninsula. The cyclone (anticyclone) has a mirror image south of the equator, implying the possible linkage of the ISO to the westward-propagating equatorial Rossby waves.

The westward propagation of the ISO can be further seen from the composite 700-hPa vertical velocity field (Fig. 12). The monsoon onset is associated with a local transition from a sinking to a rising phase. This rising (sinking) branch can be traced back to east of 140°E. It moves across the SCS, Indochina, the Bay of Bengal, and finally reaches the Indian subcontinent.

From the above analysis, it is concluded that the timing of the monsoon onset over Indochina is regulated by both the background basic flow and the activity of the 12–25-day ISO. On one hand, the establishment of low-level westerlies over the southern Bay of Bengal (Fig. 6) and convective activities over Sumatra (Fig. 5) provide the favorable backgrounds for the monsoon onset. On the other hand, the seasonal transition of the atmospheric circulation is accelerated and the vigorous convection and torrential rain are triggered when an in-phase ISO moves into the region.

5. The interannual variability of the monsoon onset over Indochina

As shown in Table 1, the monsoon onset dates differ remarkably from one year to another. We first examine the relationship between variations of the onset date and the total May–June rainfall amount in Indochina (Fig. 13). The climatological mean rainfall amount is 360 mm and the variance is 62 mm. For the 11 wettest years with anomalous rainfall exceeding 50 mm (0.8 std dev), there are 8 early onset years (the monsoon onsets before 9 May). For the 10 driest years with rainfall anomaly less than −50 mm, there are 9 late onset years (the monsoon onsets after 9 May). Thus, during the developing stage (May–June) of the summer monsoon, a wet year is usually concurrent with an early onset while a dry year is with a late onset. In other words, the timing of the monsoon onset tends to be in

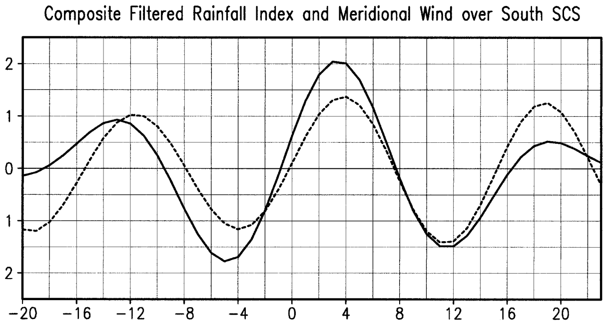

FIG. 10. Daily evolution of composite filtered rainfall index (solid line, in unit of mm day^{-1}), and meridional wind over 5°–10°N, 102.5°–107.5°E (dashed line, in unit of 0.1 m s^{-1}) by a 12–25-day bandpass filter by centering the onset day.

phase with the strength of the monsoon rainfall in May–June.

The relationship between the early/late onset and Niño-3 SSTA can be readily seen from Table 3, which lists the bimonthly Niño-3 SSTA (5°S–5°N, 90°–150°W), in January–February (JF−) preceding the monsoon onset, April–May (AM0) during monsoon onset, and November–December (ND+) after monsoon onset for both early and late onset cases. The left panel of the table shows the 14 early onset cases and the right shows the 15 late onset cases. The SSTA has been normalized by its standard deviation. The numbers of the years with positive and negative SSTA on each column are shown in the bottom rows, respectively. For the early onset years, 10, 12, and 11 cold years are found in JF−, AM0, and ND+, respectively. This implies that for the early onset years the cold SSTA tends to persist from the

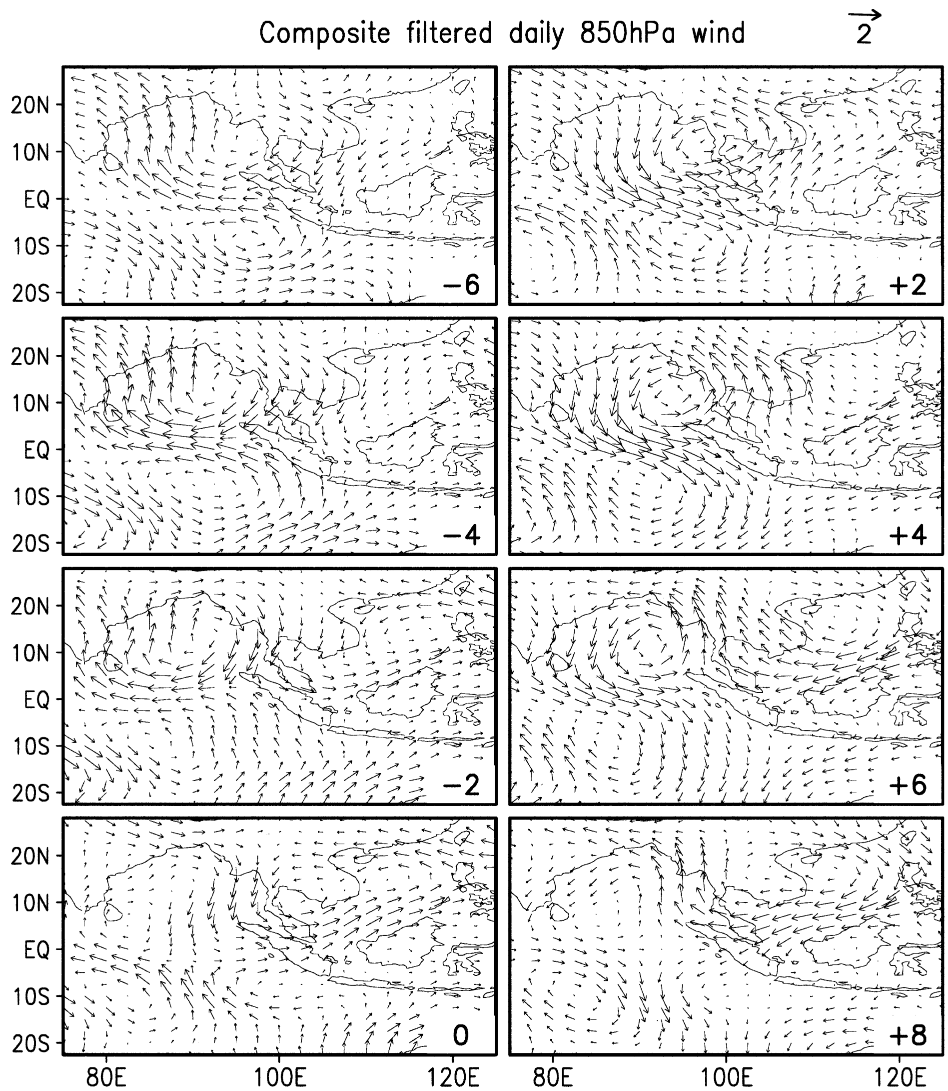

FIG. 11. Daily evolution of the composite filtered 850-hPa wind (m s^{-1}) for the phase-locking years by centering the onset days.

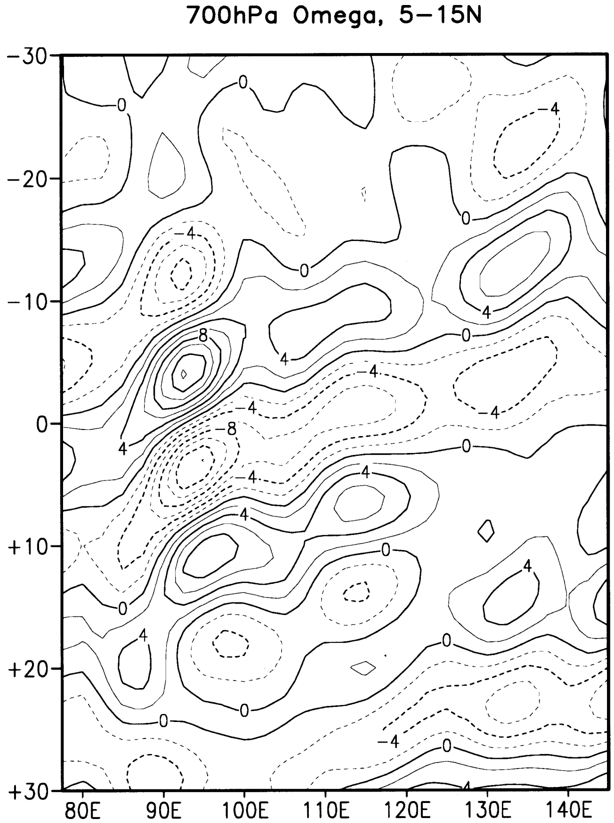

FIG. 12. The longitude–time cross section of the composite filtered omega at 700 hPa for the phase-locking years along 5°–15°N in units of 0.001 Pa s^{-1}. The interval is 0.002 (Pa s^{-1}).

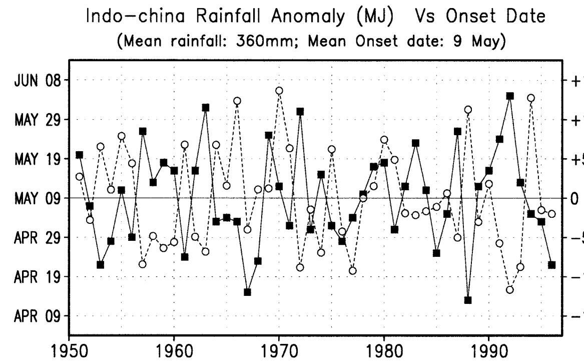

FIG. 13. The monsoon onset date and interannual anomaly of rainfall over Indochina during May–Jun in units of mm.

preceding winter through the following winter. For the late onset years, 13 and 11 warm SSTA years are found in late spring (MA0) and following winter (ND+), while only 8 warm years are found in the preceding winter (JF−). Thus, the late monsoon onset is often associated with the positive Niño-3 SSTA in spring and the following winter.

The relationship between the western Pacific (10°–20°N, 135°–165°E) SSTA and the monsoon onset date is shown in Table 4. For the 14 early onset cases, there are 11, 10, and 10 years in which the warm SSTA appears in JF−, AM0, and ND+, respectively. This implies that for the early onset years the western Pacific SSTA tends to be warm and persists through the whole year. For the late onset years, 10 cold years are identified in the preceding winter while only 8 cold years are found in both the late spring and following winter, suggesting that the relationship between the late monsoon onset

TABLE 3. The bimonthly Niño-3 SST anomalies (normalized by std dev) for the early (left columns) and late (right columns) onset years, with early/late more than 1 week compared to the mean onset date of 9 May in the (JF−) preceding winter, (AM0) during monsoon onset, and (ND+) following winter. The years marked with "*" are the years that are used in the composite study. Each year's anomalous onset date is also given with "+" and "−" denoting early/late days compared with 9 May. In the bottom two rows, the sums of the SST with positive/negative signals in each column are presented. Bold denotes negative SST for the early onset and positive SST for the late onset.

Year early	Date	JF−	AM0	ND+	Year late	Date	JF−	AM0	ND+
1988*	−26	0.71	**−0.82**	**−1.29**	1992*	+26	**1.64**	**2.26**	**0.10**
1967*	−24	**−0.40**	**−0.74**	**−0.83**	1963	+23	−0.69	**0.20**	**0.90**
1996*	−17	**−0.82**	**−0.78**	**−0.61**	1972*	+22	−0.42	**0.98**	**2.22**
1953	−17	0.46	1.14	0.43	1957*	+17	−0.51	**0.94**	**1.42**
1968*	−16	**−1.75**	**−1.06**	0.53	1987*	+17	**1.21**	**1.78**	**1.33**
1961	−15	**−0.24**	0.16	**−0.37**	1969*	+16	**0.80**	**1.31**	**0.98**
1985*	−14	**−0.84**	**−0.81**	**−0.35**	1991*	+15	**0.43**	**0.92**	**1.14**
1976*	−11	**−1.76**	**−0.43**	0.83	1983*	+14	**3.24**	**2.83**	−0.43
1954*	−11	0.16	**−1.45**	**−0.87**	1951*	+11	−0.23	**0.28**	**0.96**
1956*	−10	**−1.16**	**−0.66**	**−0.81**	1959*	+9	**0.05**	**0.28**	−0.12
1981*	−8	**−0.80**	**−0.45**	0.15	1980*	+9	**0.43**	**0.23**	**0.34**
1973*	−8	1.80	**−0.87**	**−1.41**	1979*	+8	−0.18	**0.41**	**0.38**
1975*	−7	**−0.70**	**−0.92**	**−1.39**	1960	+7	−0.30	−0.20	−0.53
1971*	−7	**−1.71**	**−0.98**	**−1.04**	1962	+7	−0.34	−1.07	−0.80
					1990*	+7	**0.16**	**0.63**	**0.28**
+ SST		4	2	3	+ SST		8	13	11
− SST		10	12	11	− SST		7	2	4

TABLE 4. Same as Table 3, but for the SSTAs in the western Pacific (10°–20°N, 135°–165°E).

Year early	Date	JF−	AM0	ND+	Year late	Date	JF−	AM0	ND+
1988*	−26	**0.45**	**1.21**	**1.42**	1992*	+26	**−1.49**	**−1.66**	**−1.44**
1967*	−24	**0.64**	**0.18**	−0.38	1963	+23	1.34	1.07	1.63
1996*	−17	**0.96**	**1.36**	−0.16	1972*	+22	**−0.50**	**−1.78**	**−1.86**
1953	−17	**0.90**	**0.11**	−0.34	1957*	+17	0.09	0.38	
1968*	−16	**0.28**	−0.15	**0.08**	1987*	+17	**−0.55**	0.11	0.67
1961	−15	**1.23**	**0.95**	**0.12**	1969*	+16	**−0.73**	0.09	0.95
1985*	−14	**0.66**	**0.19**	**0.44**	1991*	+15	**−1.05**	**−0.08**	**−1.94**
1976*	−11	−0.20	−1.24	−1.12	1983*	+14	**−1.73**	**−2.07**	1.50
1954*	−11	**0.29**	**0.73**	**0.74**	1951*	+11	0.78	**−0.36**	**−0.31**
1956*	−10	**0.22**	−0.17	**0.72**	1959*	+9	**−1.37**	**−1.58**	**−0.31**
1981*	−8	**0.91**	**1.65**	**1.31**	1980*	+9	**−0.57**	**−0.29**	0.66
1973*	−8	−1.08	−0.25	**1.14**	1979*	+8	**−0.47**	**−0.80**	**−0.05**
1975*	−7	−0.45	**0.08**	−0.48	1960	+7	0.37	0.25	0.95
1971*	−7	**1.16**	**0.79**	**0.23**	1962	+7	1.34	1.75	1.24
					1990*	+7	0.17	0.29	**−0.89**
+ SST		11	10	10	+ SST		5	7	7
− SST		3	4	4	− SST		10	8	8

and the western Pacific SSTA is not robust except in the preceding winter.

It has been suggested that the relationship between the monsoon onset and the SSTA in the central-eastern Pacific may depend on the phase of ENSO (Ju and Slingo 1995). To examine this, we compared the early/late monsoon onset cases with the El Niño (La Niña) events defined by Kiladis and Diaz (1989). It is found that in the 14 early onset years, there are 7 La Niña events. Five of them are in the developing phase and two in the decaying phase. In the 15 late onset years, there are 8 El Niño events. Five of them are in the developing phase and three in the decaying phase. This suggests that although the monsoon–ENSO relationship is not completely dependent of the phases of ENSO, there are more tendencies locking to the ENSO developing phase.

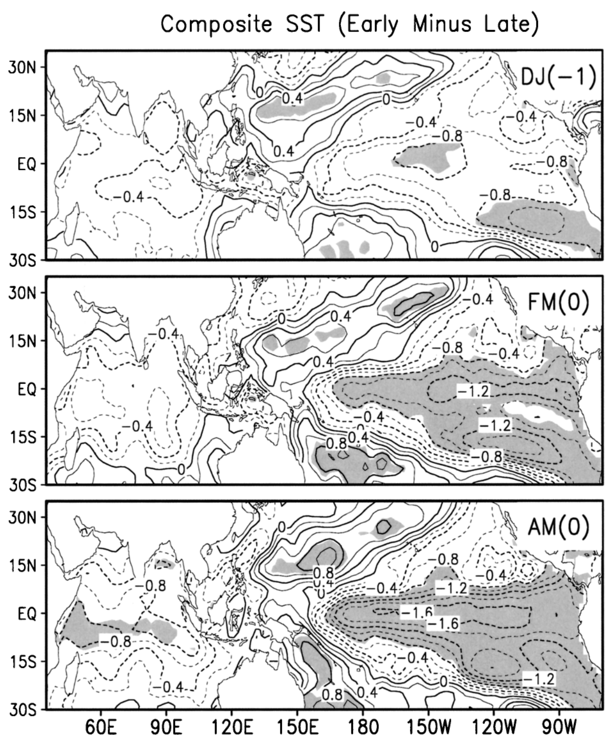

FIG. 14. The bimonthly early-minus-late composite of SST from preceding winter to spring. The interval is 0.1°C. The ''0'' denotes the year of the reference monsoon year. Shading indicates that the *t* test passes the 5% significance level.

Since a robust relationship between the early/late onset and SSTA in the central-eastern Pacific occurs in spring, in the following, 12 early onset years with the cold Niño-3 SSTA and 12 late onset years with the warm SSTA are selected to compose the circulation and SST patterns. These years are marked with an * in Table 3.

Figure 14 presents the bimonthly early-minus-late SST composites from the preceding December–January through April–May. The SST anomalies exhibit a clear El Niño pattern in the tropical Pacific Ocean. In the preceding winter, cold SST anomalies are identified in the central Pacific while the warm SST anomalies occur in the western Pacific. During spring, the cold SST anomalies increase in the central-eastern Pacific while the warm SST anomalies persist in the western Pacific. Cold SST anomalies appear in the equatorial Indian Ocean during the preceding winter and early spring, but the strong signals reaching the 5% significance level occur only in late spring [AM(0)].

Figure 15 shows the early-minus-late composites of the surface wind. Corresponding to the SST anomalies in the Pacific Ocean, the anomalous easterlies appear across the equatorial Pacific Ocean and persist from the preceding winter to spring. In the Indian Ocean, the anomalous southwesterlies occur in the northeastern Indian Ocean in the preceding winter [DJ(−1) and FM(0)], and extend westward to the western coast of the Indochina Peninsula in the late spring [AM(0)]. The appearance of the anomalous westerlies in the Indian Ocean and easterlies in the equatorial Pacific exhibits

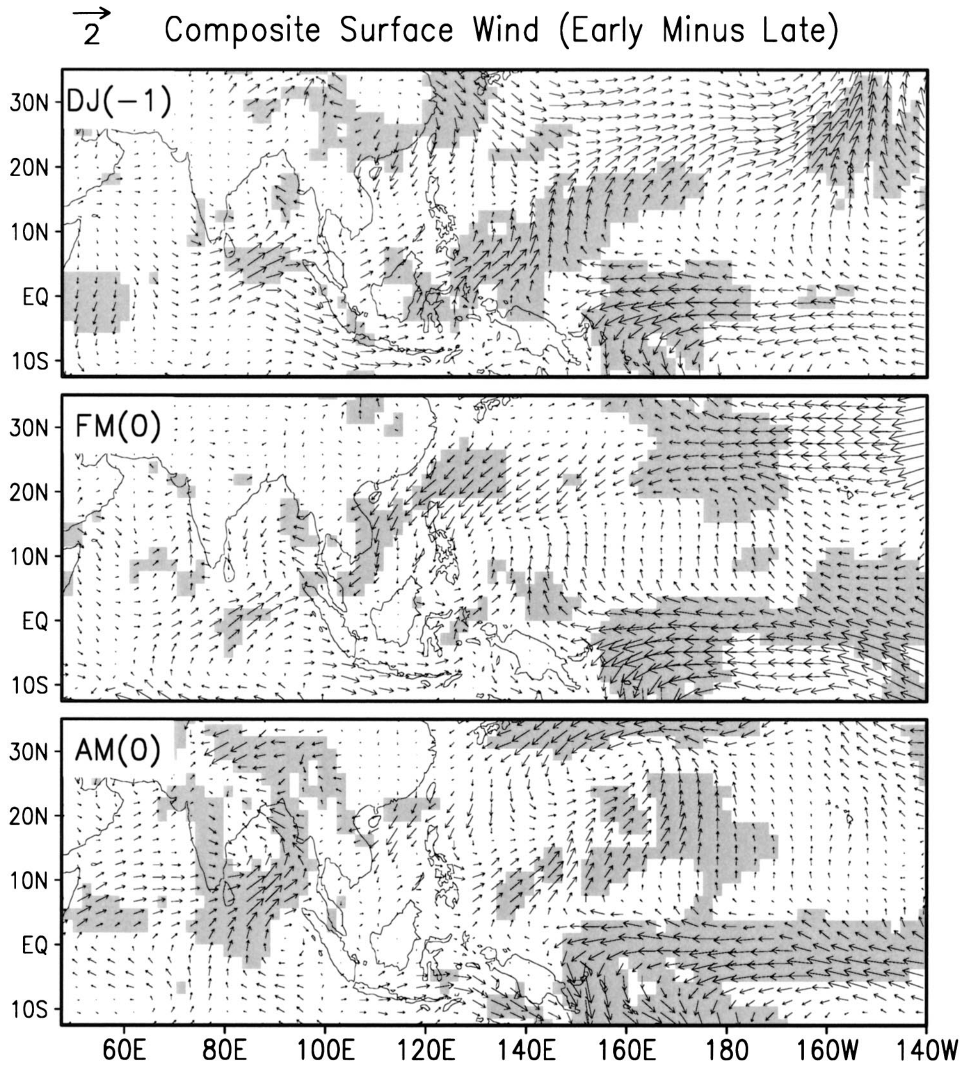

FIG. 15. Same as Fig. 14, but for the surface wind (m s^{-1}). Shading indicates the *t* test of zonal wind or meridional wind, and both exceed the 5% significance level.

an overturning of the Walker circulation associated with ENSO. Substantial northeasterly wind anomalies can also be identified in the northern SCS. These anomalous northeasterlies develop during the preceding winter and persist through spring as part of the anomalous cyclone over the Philippine Sea. Along with the southwesterly anomalies over the northern Indian Ocean, they provide anomalous low-level convergence over the southern Indochina, Sumatra, and the southern SCS regions.

The convective activities represented by the OLR are shown in Fig. 16. A minimum OLR center, as the proxy for the increase of tropical convective activities, appears over the southern SCS and the southern Indochina Peninsula during the preceding winter and spring. Meanwhile, a maximum OLR center is located in the central Pacific. Consistent with the anomalous wind fields, these two anomalous OLR centers reflect the overturning of the Walker circulation. The southern Indian Ocean is under the control of the positive OLR anomalies with a maximum center over the southeastern region (5°–20°S, 100°–120°E) in the late spring [AM(0)]. The anomalous southwesterly flows in the equatorial Indian Ocean and equatorially asymmetric distribution of the OLR anomalies suggest that the local Hadley circulation across the SCS–maritime continent–Indian Ocean is possibly one of the mechanisms regulating the early/late monsoon onset. The atmospheric vertical motions relevant to the changes of the Walker circulation and the local Hadley circulation, revealed by the velocity potential composite at 200 hPa (figure omitted), are consistent with the OLR and surface wind fields.

The western Pacific subtropical high (WPSH) is an important system regulating the weather and climate regimes over Southeast Asia (Chang et al. 2000a,b). Its variations in the early/late onset years are presented in Fig. 17, which outlines the 5860- and 5865-m contours

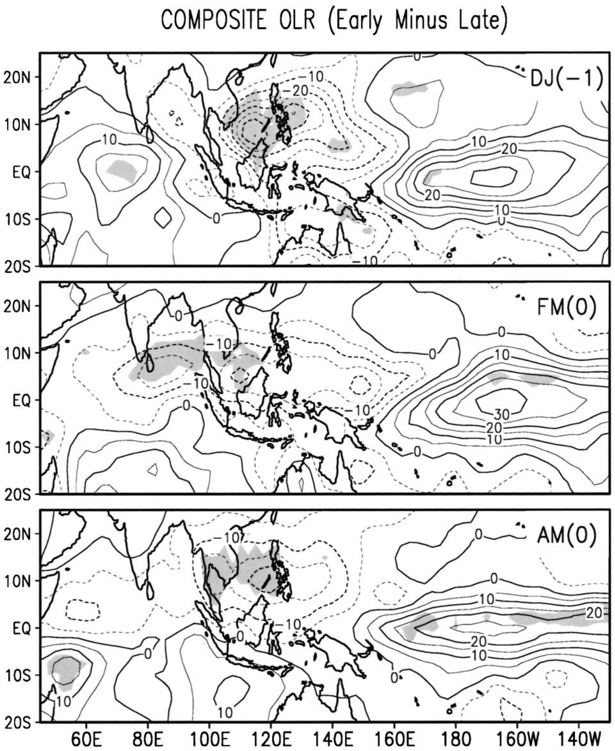

FIG. 16. Same as Fig. 14, but for the OLR. The value of the interval is 5 W m^{-2}. Shading indicates that the t test passes the 5% significance level.

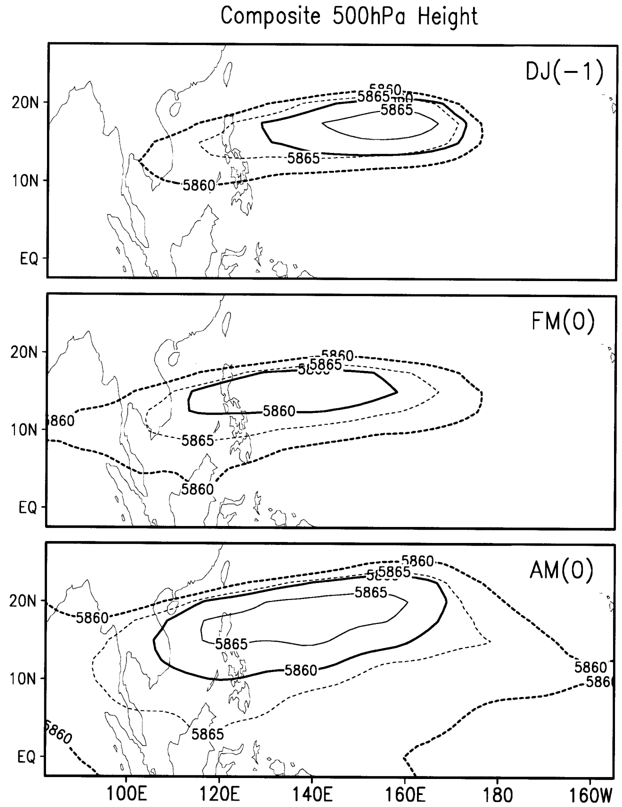

FIG. 17. The composite bimonthly 500-hPa geopotential height, indicated by 5865- (thin) and 5860-m (thick) contours for early (solid) and late (dashed) onset years.

of 500-hPa height, respectively. These two contours are chosen because they describe the characteristics of the subtropical ridge. The strength and shape of WPSH between the early and late monsoon onset composites differ remarkably. The early onset is preceded by a weakening and eastward withdrawal of the subtropical ridge. For the late onset composite, the subtropical ridge controls the entire SCS and the southern part of the Indochina Peninsula. The weakening and eastward shift of the WPSH is favorable for the enhanced rising motion and strong convection over the Indochina Peninsula, while the strong WPSH leads to the suppressed convective activities.

6. Discussion

As shown above, substantial precursory signals for the early and late monsoon onset are found not only in the SST fields, but also in the surface wind, OLR, and height fields. How are these signals linked to each other and what processes determine the interannual variation of the monsoon onset?

For a large-scale point of view, the year-to-year variations of the monsoon onset over Indochina are primarily regulated by the SST anomalies in the western and central-eastern Pacific during the preceding winter and spring (Fig. 14). Associated with the SST anomalies, the change of the Walker circulation is manifested by the shift of the strong convective activities (Fig. 16) and the anomalous easterlies in the equatorial Pacific (Fig. 15). The change of the local Hadley circulation is characterized by contrast of the enhanced convective activities over the SCS/western Pacific and the suppressed convective activities over the southeastern Indian Ocean, and by the cross equatorial southwesterly flow over the eastern Indian Ocean. These results agree with that of Ju and Slingo (1995), who considered the monsoon onset in term of the changes of the large-scale flows over the Arabian Sea and the Indian subcontinent.

Locally, the anomalies of the convective activity over the SCS and southern Indochina during winter and spring are directly responsible for the year-to-year variations of the monsoon onset in Indochina. For an early onset year, the enhanced anomalous convection is associated with the low-level convergence between the anomalous northeasterly over the northern SCS and the southwesterly over the northeastern Indian Ocean. What are the possible processes that cause and maintain the two branches of anomalous flows?

It is apparent that the anomalous westerly in the equatorial Indian Ocean is a result of the enhanced Walker circulation associated with the cold SST anomalies in the central-eastern Pacific and warm SST anomalies in the western Pacific (La Niña). Since the prevailing wind

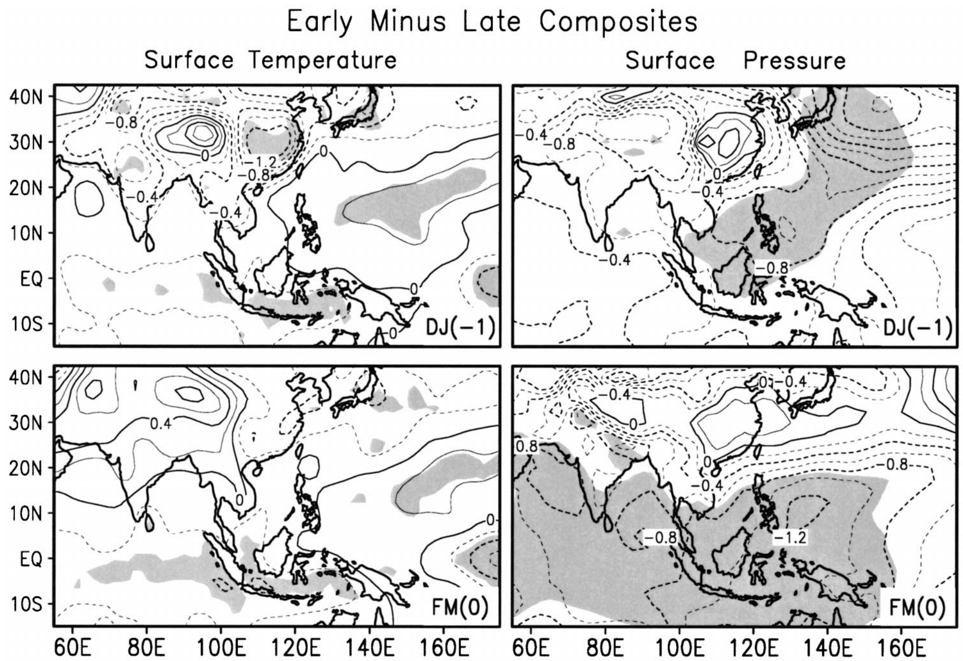

FIG. 18. The bimonthly early-minus-late composites of (left) the surface temperature and (right) surface pressure in the preceding Dec–Jan and Feb–Mar. Shading indicates that the t test passes the 5% significant level. The interval for the surface temperature and surface pressure are 0.2°C and 20 Pa, respectively.

flow is westerly over the equatorial Indian Ocean during spring, this westerly anomaly leads a cold local SST anomaly due to the enhanced surface evaporation. The cooling in the equatorial Indian Ocean, along with the anomalous warming over the south Asian continent (left panels of Fig. 18), results in an enhanced land–ocean thermal contrast, and thus favors the development of the anomalous southerly flow in the northern Indian Ocean. As a result, the anomalous southwesterlies prevail in the northern Indian Ocean in spring.

The anomalous northeasterly flow over the northern SCS and along the coast of southeastern China in the preceding winter is associated with an enhanced east Asian winter monsoon. As shown in Fig. 18, an anomalous high pressure center (top-right panel) is overlapped with anomalous cold surface temperature (top-left panel) over east Asia in the preceding winter. Meanwhile, an anomalous low pressure center locates over the warm SSTA region in the western Pacific. This enhanced surface pressure gradient occurs across the coast of southeastern China between the two anomalous centers. As a result, the anomalous northeasterly flow is aroused along the coast of southeastern China. During February–March (bottom panels), the northeasterly flow persists even though the land–sea thermal contrast reduces. The anomalous northeasterly flow is maintained by the cyclonic circulation associated with the anomalous Philippine Sea cyclone, which is induced due to the enhanced convective heating related to La Niña. Wang et al. (2000) proposed that the anomalous Philippine Sea cyclone is maintained by a local positive air–sea feedback mechanism. On the one hand, the total wind speed to the east of the anomalous cyclone is decreased because the direction of the induced cyclonic flow is opposite to the prevailing trade wind, and thus leads to the SST warming in this region due to the decrease of the surface evaporation. On the other hand, the warm SST anomaly, in turn, favors amplification of the anomalous cyclone by exciting ascending Rossby waves to its west. Due to such a positive feedback, the anomalous Philippine Sea cyclone is maintained through spring. An opposite behavior is noted for a late onset year with the El Niño conditions.

7. Conclusions

In this paper, by using the domain-averaged daily rainfall index from 30 rain gauge stations over the central Indochina Peninsula for 1951–96, the general onset date of the summer monsoon in this region is defined based on the strength and persistence of the rainfall during late spring to early summer. The spatial and temporal characteristics of atmospheric circulation and tropical convection, as well as the ISO activities associated with monsoon onset are investigated. A composite study is conducted according to the early/late onset categories in order to identify the precursory signals and their underlying processes affecting the year-to-year variations of the monsoon onset. The following are the conclusions derived from this observational study:

1) The climatological monsoon onset over Indochina is

on 9 May. For individual years, the date differs remarkably from one year to another, ranging from mid-April to early June. The climatological monsoon outbreak is characterized by the northward extension of intense convection over Sumatra, the arrival of the tropical southwesterly over the western coastal region of Indochina, and the intensification of the India–Burma trough in the 500-hPa height field. Accompanying them, the western Pacific subtropical high retreats eastward and the subtropical westerly flow weakens. The study indicates that the tropical influence plays a fundamental role in triggering the monsoon onset over the Indochina Peninsula.

2) The intraseasonal oscillation (ISO) with a dominant timescale of 12–25 days is closely related to the timing of monsoon onset. It is originated from the SCS and the western Pacific and propagates westward into the Indochina Peninsula.
3) There is a close relationship between the early/late monsoon onset and the cold/warm SSTA in the western and central-eastern Pacific during boreal spring. Robust precursory signals for the early onset include the cold SSTA in the central-eastern Pacific and the warm SSTA in the western Pacific in the preceding winter. No close relationship exists between the late onset and the preceding winter SSTA in the central-eastern Pacific. However, the cold SSTA in the western Pacific is a precursory signal for the late onset.
4) The composite SST field for the early-minus-late monsoon onset exhibits a typical ENSO pattern in the preceding winter and spring. For an early onset year, strong convective activity occurs over southern Indochina and the southern SCS during the preceding winter and spring, and is closely associated with the changes of the Walker circulation and the local Hadley circulation. The strong convective activity is maintained by an anomalous low-level convergence between the anomalous northeasterlies over the northern SCS and the southwesterlies in the northeastern Indian Ocean. The enhancement of the southwesterly flow south of Indochina is possibly associated with the enhanced Walker circulation related to ENSO and the land–sea thermal contrast between the warm southern Asian continent and the relatively cold Indian Ocean. The northeasterly anomalies originated in winter owing to a thermal contrast between cold east Asia and the warm tropical western Pacific, and are further maintained by a positive feedback between the anomalous Philippine Sea cyclone and warm SST anomalies in the western Pacific, a process described by Wang et al. (2000). The situation is reversed for a late monsoon onset case.

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