A Climatological Monsoon Break in Rainfall over Indochina—A Singularity in the Seasonal March of the Asian Summer Monsoon

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ABSTRACT

This study investigated the climatological pentad mean annual cycle of rainfall in Thailand and the associated atmospheric circulation fields. The data used included two different data of rainfall: rain gauge data for Thailand from the Thai Meteorological Department and satellite-derived rainfall data from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP).

Climatological mean pentad values of rainfall taken over 50 yr clearly show a distinct climatological monsoon break (CMB) occurring over Thailand in late June. The occurrence of the CMB coincides with a drastic change of large-scale monsoon circulation in the seasonal march. The CMB is a significant singularity in the seasonal march of the Southeast Asia monsoon, which divides the rainy season into the early monsoon and the later monsoon over the Indochina Peninsula.

A quasi-stationary ridge dynamically induced by the north–south-oriented mountain range in Indochina is likely to cause the CMB. The formation of the strong ridge over the mountain ranges of Indochina is preceded by a sudden enhancement (northward expansion) of the upstream monsoon westerlies along a latitudinal band between 15° and 20°N in late June. The CMB also has an impact downstream. The orographically induced stationary Rossby waves enhance the cyclonic circulation to the lee of Indochina, and over the South China Sea. The enhancement of cyclonic circulation may be responsible for the summer monsoon rains peaking in late June over the South China Sea and the western North Pacific, and in the baiu front.

1. Introduction

The last decade has seen a surge in studies of the monsoon over the Indochina Peninsula (IP). The summer monsoon over the IP has many peculiar features that distinguish it from the adjacent Indian and western North Pacific [WNP; including the South China Sea (SCS)] monsoons (Matsumoto 1997; Wang and LinHo 2002).

Matsumoto (1997) noted that there are many significant rain events over Thailand, referred to as premonsoon rains, before the monsoon onset in mid-May. Rainfall increases gradually from the premonsoon rains to the summer monsoon onset. The monsoon onset over Thailand is a more gradual process than the abrupt change from very dry conditions at onset over India or the SCS. Wang and LinHo (2002) analyzed rainy season characteristics and noted that the monsoon over the IP differs in many essential aspects from the adjacent Indian and WNP summer monsoons. They suggested that the IP might serve as a transition zone, or boundary, between the two tropical monsoon subsystems: the Indian monsoon and the WNP monsoon. Interesting features in monsoon rainfall over the IP include an early onset, late peak, and withdrawal.

Matsumoto (1997) found that the rainy season in Bangkok starts in mid-May. Zhang et al. (2002) found that rainy season onset over all Thailand is around 9 May. Wang and LinHo (2002) mapped the onset of the rainy season for the entire Asian–Pacific monsoon and showed that the summer rainy season starts in the IP in early May (P26–27: "pentad" will be referred to as "P" in this study), which is the earliest onset over land in

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Asia. The rainy season peaks over the IP in September (P52–56), which is the latest peak over land in Asia (Wang and LinHo 2002).

Another unusual feature of the IP monsoon is its significant bimodality in the seasonal march. There is a strong peak in rainfall in September and a weaker peak in late May to early June (Fig. 2). A notable break in monsoon rainfall occurs over Thailand (Fig. 2) in late June. This midsummer break is called the climatological monsoon break (CMB). The cause of the CMB over Thailand has not been studied. The primary purpose of this study is to document circulation changes before, during, and after the CMB and to uncover the fundamental processes responsible for the CMB.

It has long been recognized that the annual evolution of the rainy season and summer monsoon over Asia is not smooth. Stepwise evolution, subseasons, and sudden changes have been studied extensively for many years (e.g., Yoshino 1965; Kato 1985; Ninomiya and Muraki 1986; He et al. 1987; Tao and Chen 1987; Lau et al. 1988; Matsumoto 1992; Nakazawa 1992; Murakami and Matsumoto 1994; Ueda et al. 1995; Ueda and Yasunari 1996; Wang and Xu 1997; Kawamura and Murakami 1998; Kang et al. 1999; Wu and Wang 2000, 2001; LinHo and Wang 2002). Wang and Xu (1997) used the term climatological intraseasonal oscillations (CISOs) to refer to intraseasonal components of outgoing longwave radiation (OLR) in climatological time series. They identified four subseasonal cycles in the Asian–Pacific summer monsoon. Each CISO event was statistically significant, had a coherent dynamic structure, and propagated in a systematic way; consequently, they nearly qualified as a physical mode on climatological maps. They showed that extremely wet and dry phases of the CISO represent monsoon singularities, that is, wet and dry events that occur with regularity during the fixed pentads. LinHo and Wang (2002) found rapid variation (on an intraseasonal time scale) in the climatological time series and named this the fast annual cycle (FAC). The FAC over the IP is much weaker than over the WNP or South Asia (however, Fig. 2 shows that the CMB time scale is intraseasonal). Exploring how the CMB is related to intraseasonal variation (ISV) is warranted.

Monsoon breaks over India occur when the monsoon trough is close to the foot of the Himalayas. At this time, there is a striking decrease in rainfall over most of India, although rainfall increases along the Himalayas and over southern India (Rao 1976). Many observational studies during the past two decades have sought to explain active/break monsoon spells from the perspective of ISV in the summer monsoon, especially over south Asia. Yasunari (1979, 1980, 1981) found cloud bands to propagate northward over India on a time scale of 30–50 days and related the bands to monsoon active/break spells over India. Krishnamurti and Ardanuy (1980) related monsoon breaks over India to high pressure anomalies in a westward-propagating mode with a period of 10 to 20 days. Monsoon breaks over Asia are now thought to be equivalent to break spells in the ISV. However, this dose not explain the CMB over Thailand. Wang and Xu (1997) and LinHo and Wang (2002) showed that climatological ISV components of OLR over the IP is much weaker than the other monsoon regions. However the CMB over Thailand is prominent (Fig. 2). Therefore, this study addresses why the transient ISV over the IP is much weaker than over India and the WNP, but the CMB over Thailand is obvious. Understanding the causes of monsoon singularities in other regions of the Asian–Pacific monsoon will be enhanced as causes of the CMB are revealed in this study.

Section 2 documents the data used in this study. Section 3 describes the characteristics of the annual rainfall over Thailand. Spatial structures in rainfall and atmospheric circulation fields are investigated in section 4. Possible dynamics behind the CMB are discussed in section 5, and conclusions are given in section 6.

2. Data

Daily rainfall data were taken from 32 stations in Thailand (Fig. 1) and compiled by the Thai Meteorological Department (TMD). Data used are from 1951 to 2000. In Fig. 1, the stations are almost uniformly distributed in Thailand. These stations have 50-yr daily rainfall records. Missing data from each station are limited to less than 1%.

Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997) data were used to analyze the large-scale rainfall distribution. CMAP is a global dataset on a 2.5×2.5 grid from 1979 to 2000.

The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996) yielded zonal and meridional winds (u, v) and geopotential heights (z) at 850 and 700 hPa on a 2.5×2.5 global grid. Also, to calculate vertically integrated moisture flux, horizontal winds, geopotential heights, relative humidity, and air temperature from the surface to 500 hPa as well as surface pressure were used.

Pentad data calculated from daily data for each field describe the climatological subseasonal variability.

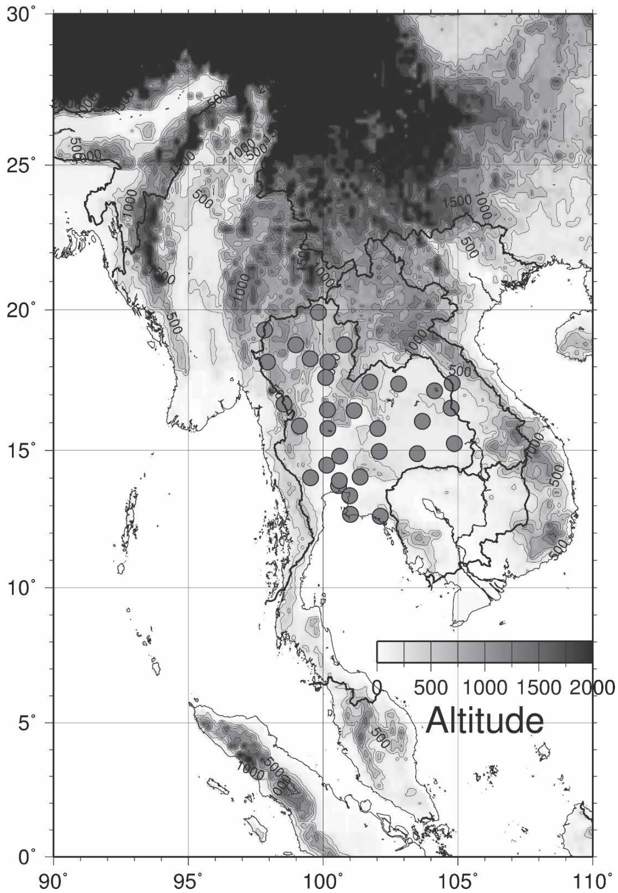

FIG. 1. Location of the 32 stations in Thailand for which daily rainfall records are available for the period 1951–2000.

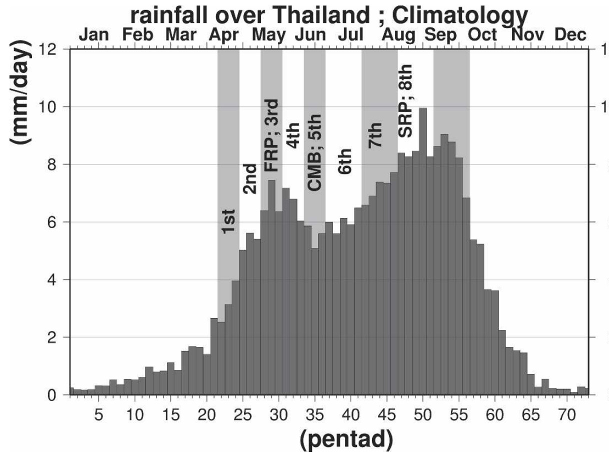

FIG. 2. Climatological pentad mean time series of rainfall index for 50 yr in units of mm day^{-1}. The index is averaged rain gauge observations at 32 stations. To illustrate atmospheric circulation associated with this rainfall index, the rainy season was divided into eight periods.

3. A singularity in Thailand rainfall

Matsumoto (1997) described two rainfall peaks in the summer monsoon in Thailand. However, his rainfall data were limited to Bangkok, Thailand, and he focused on the onset and withdrawal of the summer monsoon in Thailand. This study uses area-averaged pentad mean rainfall for 50 yr (1951–2000) for 32 stations over Thailand, as shown in Fig. 2, to examine the characteristics of the climatological seasonal march of rainfall over Thailand.

The summer monsoon (rainy season) over the IP occurs from May to October, while the winter monsoon (dry season) is from November to April. Seasonal rainfall variation shows two significant peaks in the climatological annual cycle over Thailand: the first rainy peak (FRP) occurs from P28 to P30; and the second rainy peak (SRP) occurs from P49 to P51. A monsoon break occurs in P34 to P36 between the FRP and SRP, and this break is defined as the CMB. The rainfall time series is averaged over 50 yr at 32 stations, but the CMB is still obvious. Climatological seasonal variation in the rainfall at each station shows that the CMB occurs at almost all stations during the same period (not shown). The CMB occurs over all Thailand at nearly the same time each year.

The CMB is statistically meaningful for climatology because it has statistical significance over 50 yr. We confirmed that the CMB is a true monsoon singularity rather than an artifact owing to extreme events in some years by examining the occurrence frequency of each pentad such that each pentad's rainfall was more than a threshold value over 50 yr. The threshold value was the mean rainfall during the rainy season, defined as the average from P26 to P60 based on previous studies (Matsumoto 1997; Zhang et al. 2002). The rainy seasonal mean was 6.7 mm day^{-1}. Figure 3 shows the frequency of occurrence of precipitation more than the threshold value. Low frequencies occur during the dry season and around the CMB, which indicates that rainfall during the CMB is less than the rainy season mean in most years. A small number of extreme events do not make up the CMB. Instead, the CMB is a real monsoon singularity that is fixed to the seasonal march. A period of high frequency also appears during the SRP. The SRP is an interesting phenomenon because the Indian monsoon and WNP monsoon are weak in this period, when monsoon rainfall over Thailand is highest. By contrast, monsoon precipitation is high in all three regions during the FRP. Monsoon rains over Thailand have unique characteristics that distinguish them from the Indian and WNP monsoons. The CMB divides the monsoon season over Thailand into the FRP and SRP; surrounding large-scale atmospheric circulation fea-

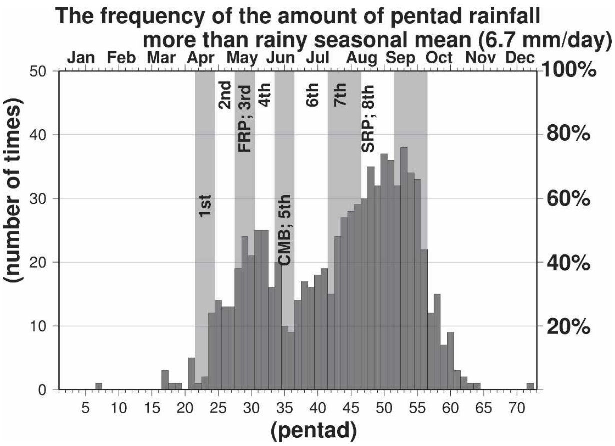

FIG. 3. The occurrence frequency for each pentad when each pentad rainfall is more than a threshold value. The threshold value was determined as the rainy seasonal mean. The rainy seasonal mean averaged was at 6.7 mm day^{-1}.

tures before and after the CMB also differ. The CMB may not be a local phenomenon, but rather a significant monsoon singularity associated with the seasonal march of the Asian–Pacific monsoon system.

This study defines eight periods in the summer monsoon season to facilitate an examination of atmospheric circulations associated with seasonal variation in rainfall over Thailand. The definitions are shown in Fig. 2. The first and second periods occur during times in the annual cycle after the dry season when rainfall totals are increasing. The third period is the FRP. During the fourth period, rainfall totals are at a peak and start to decline. The fifth period is the CMB. During the sixth and seventh periods, rainfall totals again increase until the eighth period, which is the SRP.

4. Atmospheric circulation associated with rainfall over Thailand

The previous section presented the characteristics of the CMB, namely, a monsoon singularity with a considerable drop in rainfall. Figure 4 shows a sequence of total CMAP rainfall distribution to illustrate the large-scale seasonal march of rainfall. Figure 5 shows the associated atmospheric circulations, namely the 850-hPa geopotential height and wind fields. The eight periods shown in Figs. 4 and 5 are as defined in section 3. Figures 6 and 7 show difference fields between one period and the previous period for rainfall and atmospheric circulation, respectively. The focus is on the CMB as an important singularity in the seasonal march of the Asian–Pacific monsoon.

a. The first increasing phase (periods 1 to 4)

In period 1 (P22–24), the climatological three-pentad mean of the 850-hPa horizontal wind (Fig. 5a) shows no monsoon westerlies over the south and Southeast Asia monsoon regions. The largest rainfall is centered at 5°N, 90°E over the eastern Indian Ocean (Fig. 4a).

During period 2 (P25–27), a trough appears over the Bay of Bengal (BOB). At the same time, weak monsoon southwesterlies appear over the eastern Indian Ocean and reach the IP (Fig. 5b). Simultaneously, high precipitation is observed over the eastern Indian Ocean and southern IP (Fig. 4b).

During period 3 (P28–30), a large amount of rainfall is observed over the IP (Fig. 4c). The strong monsoon rainfall over the BOB and IP precedes the other tropical monsoon regions. The monsoon southwesterlies are established over the BOB, as the trough deepens over the BOB (Fig. 5c). Over the IP, the wind direction is southwesterly. The difference fields in rainfall show that the increase of rainfall is observed over the BOB, IP, northern SCS, and the south of Japan (Fig. 6a). The monsoon westerlies are enhanced along the 10°N latitude (Fig. 7a), which is consistent with the increase in rainfall. Thus, the southwesterlies from the trough over the BOB is likely to induce the increase in rainfall over the IP.

The monsoon westerlies over south Asia are drastically stronger during period 4 (P31–33), which corresponds with the monsoon onset in India (Fig. 5d). The intensified monsoon is consistent with the abrupt increase in rainfall over India (Fig. 4d).

b. During the CMB

During the CMB (P34–36), the strongest band of monsoon rainfall is oriented northwest–southeast from north India to the WNP (Fig. 4e). An area of low rainfall amount is locally observed over the western IP, which is consistent with area-averaged rain gauge rainfall. The monsoon rainband migrates northward over the Arabian Sea, Indian subcontinent, and BOB (Fig. 5e). Over the IP, the monsoon wind changes direction from southwesterly to westerly as the trough over the BOB weakens. A strong ridge is formed over western Indochina, which is consistent with the area of low rainfall amount there. Simultaneously, a trough over the SCS, where the monsoon westerlies meet trade winds, deepens, as rainfall increases, and the monsoon westerlies enlarge eastward to the Philippines. At the same time, the subtropical high over the south of Japan intensifies, which strengthens moisture transport that

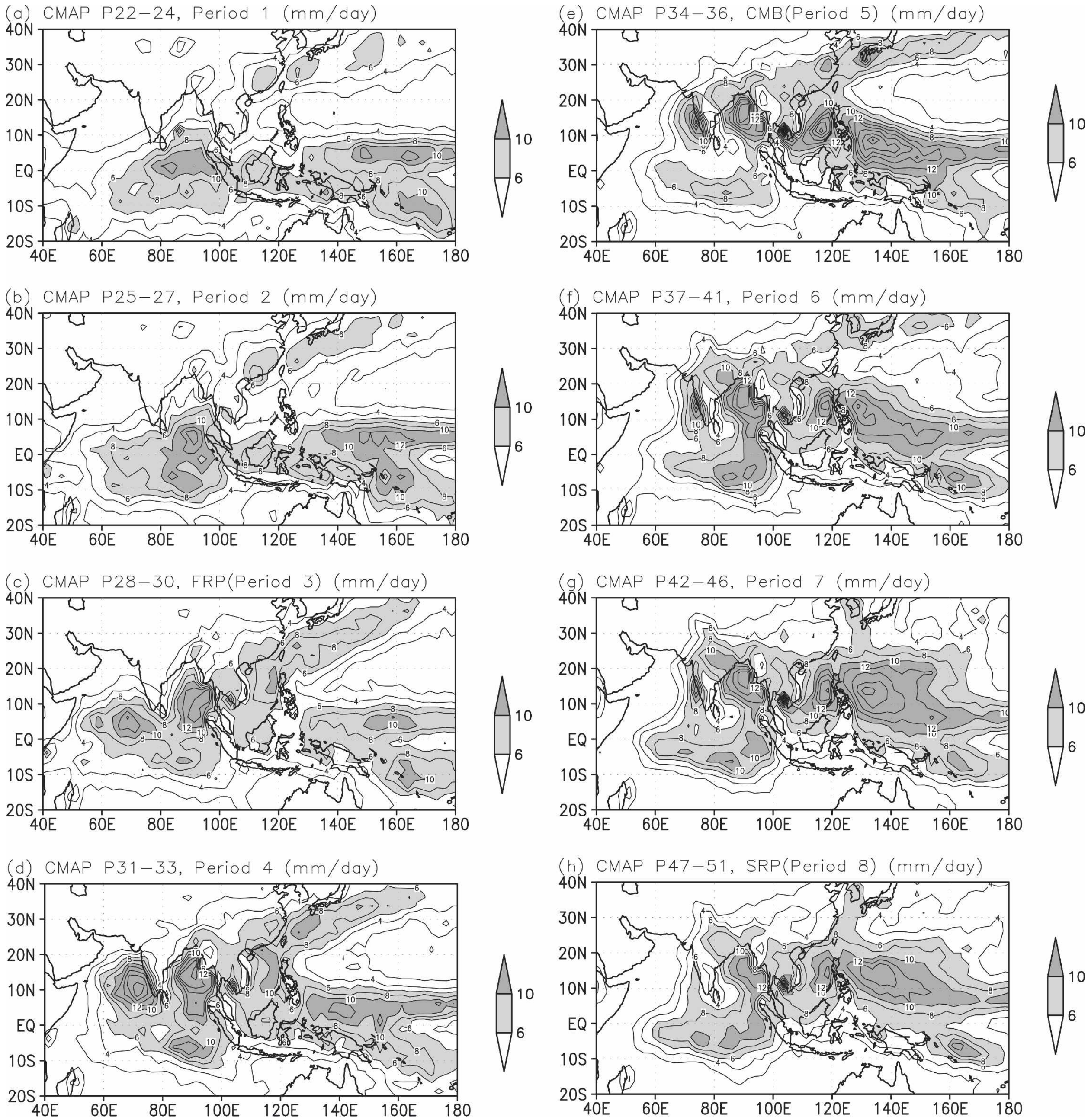

FIG. 4. Spatial distributions of rainfall by CMAP in each period based on rainfall index. Light (dark) shaded region denotes the rainfall in excess of 6 mm day^{-1} (10 mm day^{-1}).

supports increased rainfall over Japan during the onset of the baiu season there.

The differences between periods 4 and 5 show that decreased rainfall is observed over the western and central IP (Fig. 6b). The area of decrease in rainfall is confined to Thailand, while the monsoon rainfall increases over most of the Asian monsoon regions, such as the northern Indian subcontinent, SCS, WNP, and the baiu region, except for western IP. Over India and the BOB, the northward expansion of the stronger monsoon westerlies is noticeable, which is consistent with the northward migration of rainfall (Figs. 6b and 7b). When the low-level monsoon westerlies strengthen between 15° and 20°N upstream from the monsoon (over the BOB), a strong ridge appears over the western IP, which is clearly anchored at the Indochina

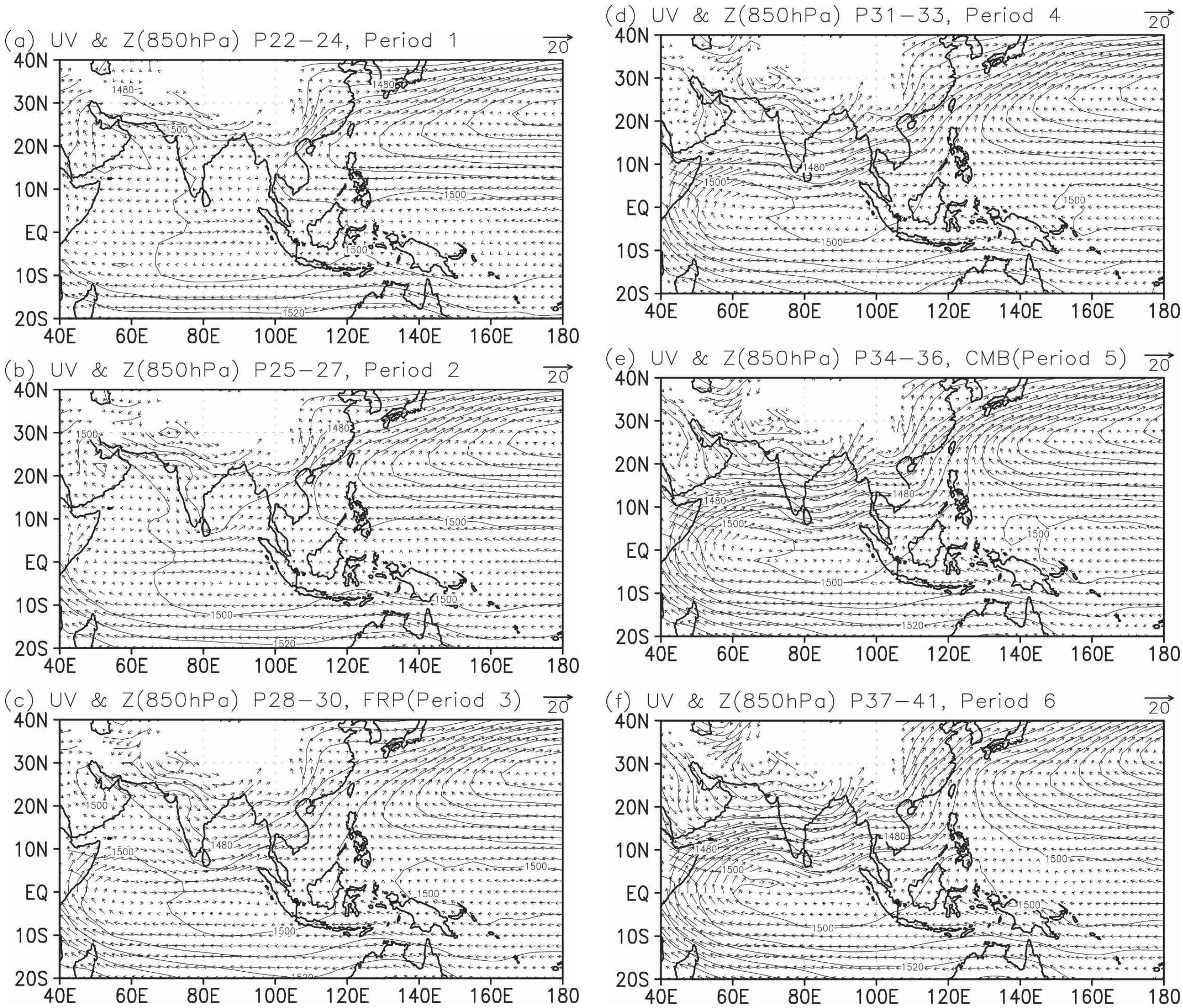

FIG. 5. Each period mean distribution of horizontal wind vector and geopotential height at 850 hPa. Unit vector is 20 m s^{-1} as denoted at upper right. Contour interval of geopotential height is 10 gpm.

mountain ranges (the Arakan, Dawna, and Tenasserim Mountains). This ridge formation is likely to be responsible for the reduction of rainfall to form the CMB over the IP. The formation dynamics of the strong ridge will be discussed in the next section. Note also that a clear cyclonic circulation appears downstream from the monsoon westerlies in association with the deepening of the trough over the eastern IP and SCS. Another cyclonic circulation also appears over the WNP. Both the cyclonic circulations are consistent with the rainfall increase there. The processes of rainfall increase over the SCS and WNP are discussed in the next section. In addition, the subtropical high centered at 20°N, 135°E over the WNP intensified, and this may be a response of the greater amounts of rainfall and convection over the SCS and WNP (Nitta 1987).

c. The second phase of increasing rainfall (periods 6 to 8)

During period 6 (P37–41) to period 7 (P42–46), the amount of rainfall is gradually decreased over the SCS and WNP regions. Rainfall also decreases over the Japanese Archipelago, which corresponds with the end of baiu (Figs. 4f,g). Simultaneously, the subtropical high retreats northward over the WNP. The reduction in rainfall over the Japanese Archipelago and SCS is associated with weakening of moisture transport along the periphery of the subtropical high, which is caused by the retreat of the subtropical high.

During periods 7 and 8 (P42–46 and P47–51), the monsoon westerlies reach the WNP. The trough deepens over the WNP. During the SRP, the IP has a high

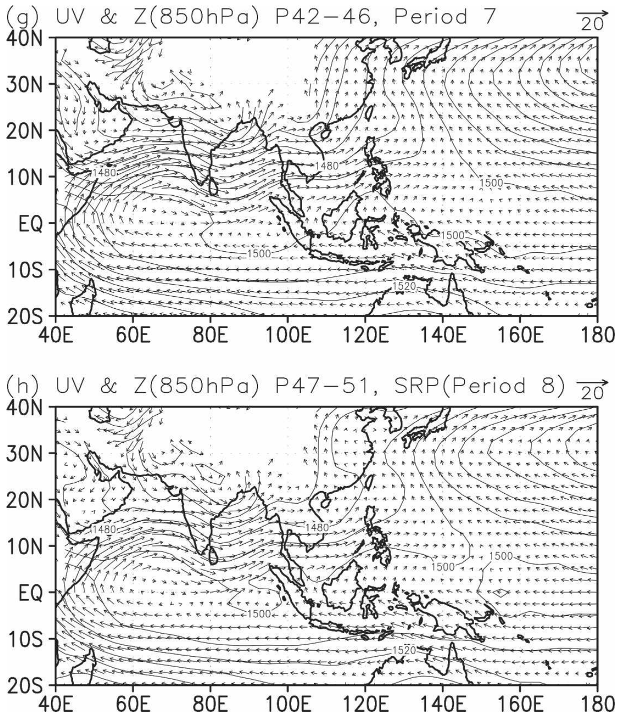

FIG. 5. (*Continued*)

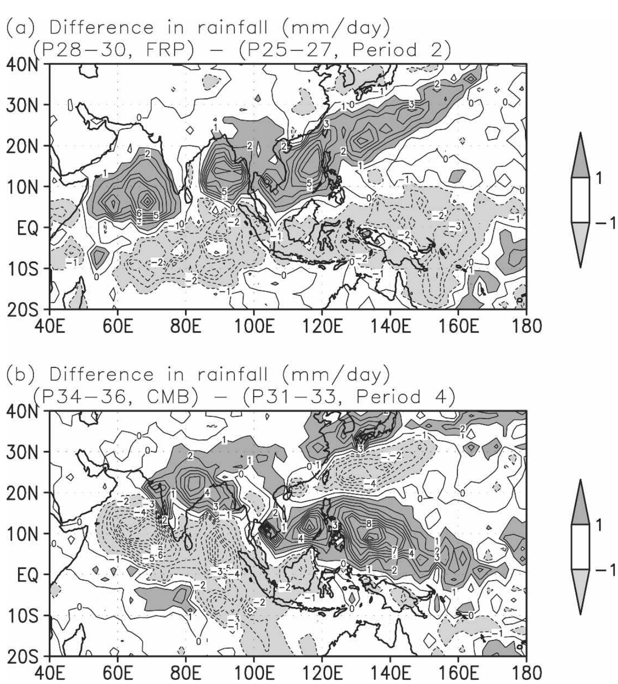

FIG. 6. Difference in rainfall (CMAP) between one period and the previous period, based on the rainfall index. Light (dark) shaded denotes rainfall increased (decreased). The periods are (a) periods 3 − 2 and (b) periods 5 − 4.

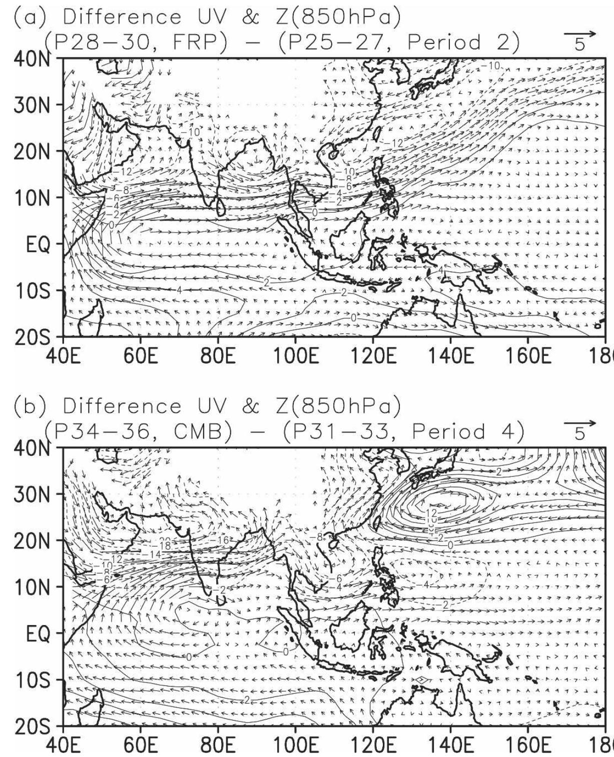

FIG. 7. As in Fig. 6 except for in geopotential height and wind at 850 hPa.

amount of rainfall, which is consistent with the rain gauge rainfall data. The SRP is the late rainy peak over the whole tropical Asian monsoon regions. The strong trough is deepened over the SCS and the WNP, which is clearly consistent with the frequent tracks of tropical cyclones in this period.

To examine the enhancement of cyclonic activity over the Asian monsoon regions, a Hovmöller diagram of vorticity at 700 hPa along the latitude band between 10° and 20°N is produced as shown in Fig. 8. Positive vorticity over the SCS and WNP at this latitude band increases with time between the CMB and SRP, then decreases after the SRP. The cyclonic vorticity is strongest from August to mid-September, expanding westward to the IP. The cyclonic vorticity likely induces moisture convergence and heavy rainfall over the IP during the SRP. In addition, each wave packet of positive vorticity shows systematic westward phase propagation after the CMB, starting from the SCS and WNP, which corresponds to the frequent tracks of tropical cyclones.

d. Different rainfall systems

In this subsection, the two different rainfall systems before and after the CMB are discussed. Figure 9 shows related series of atmospheric circulation changes.

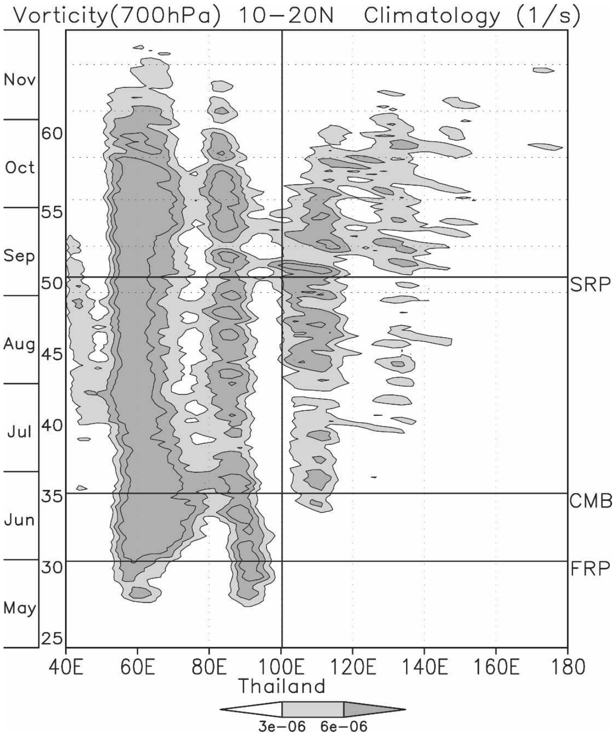

FIG. 8. Hovmöller diagram of daily-averaged vorticity in the lower troposphere (700 hPa) along the latitudinal band averaged within 10°–20°N.

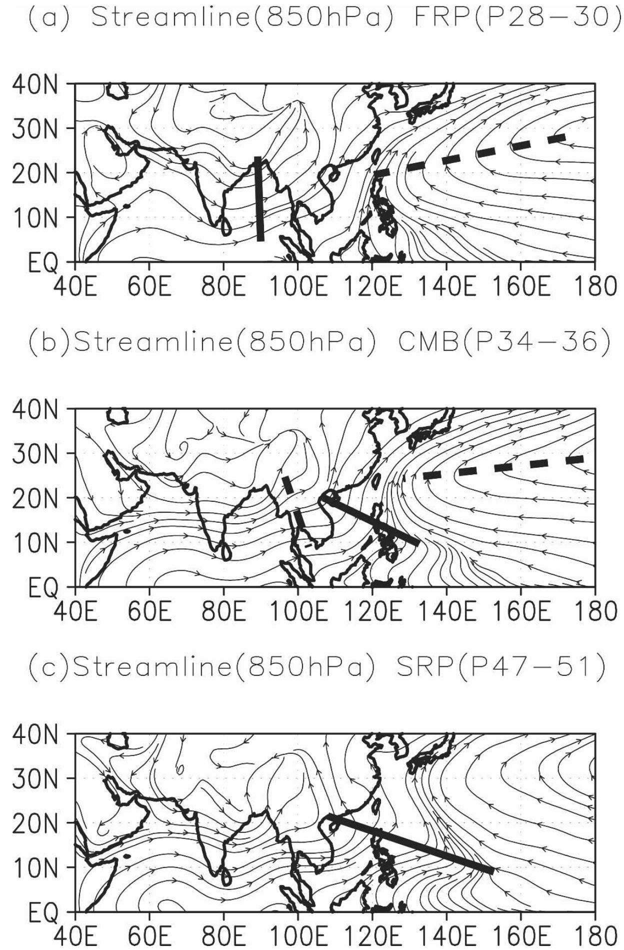

FIG. 9. Atmospheric change associated with rainfall over Thailand during (a) the former increasing rainfall phase (FRP; period 3), (b) the CMB (period 5), and (c) the latter increasing phase (SRP; period 8). The solid (dashed) lines denote ridges (troughs).

The upper panel shows a trough over the BOB when monsoon westerlies are enhanced along a latitudinal band within 5°–10°N. The monsoon southwesterlies from the trough over the BOB are associated with increased rainfall over the IP during the FRP.

During the CMB, a trough appears over the SCS and WNP. During this period, the rainfall center shifts eastward from south Asia to the WNP, which is consistent with vorticity fields in Fig. 8.

In the bottom panel, the trough is formed over the SCS and WNP, which has intensified since the CMB. After the trough has developed over the SCS and WNP, transient waves with positive vorticity (cyclones) display systematic westward phase propagation (Fig. 8). Harr and Elsberry (1991, 1995) showed that variability in the low-level zonal wind is related to the motion of tropical depressions over the SCS and WNP. Periods 7 and 8 contain atmospheric structures that favor westward-propagating tropical depressions. During the SRP, these tropical depressions are associated with the increased rainfall over the IP.

The CMB is a transition period to change rainfall systems from the monsoon southwesterlies to tropical depressions. Thus, the dynamics that cause rain during the FRP differ from those of the SRP.

5. Discussions

a. Topographic effects

A strong ridge anchored at the Indochina mountain ranges is intensified during the CMB, which is associated with the decrease in rainfall, when the monsoon westerlies are enhanced along 15°–20°N. This suggests an enhanced interaction between the monsoon westerlies and topography. This subsection investigates a mechanism of the formation of the ridge over Indochina, in terms of topographic effects.

Gadgil (1977) explained the observed trough in the southern BOB and the ridge to the east over Burma as stationary Rossby waves generated downstream from the Western Ghats Mountains on the Indian subcontinent. The IP is quite similar topographically to the Indian subcontinent. The latitudinal mountain ranges on the IP, the Arakan, Dawna, and Tenasserim Mountains, all have an average height exceeding 1 km, which is higher than the Western Ghats. The strong monsoon westerlies extend up to 600 hPa (not shown) and the strongest westerlies are around 850 and 700 hPa. Monsoon westerlies are thicker than the mountain ranges over Indochina, so stationary Rossby wave theory can be applied to the IP. Gadgil (1977) discussed features on the seasonal time scale. Figure 7b clearly shows evidence of topographic effects during three pentads. In terms of the conservation of potential vorticity, therefore, a strong ridge forms over the Indochina mountain ranges and a cyclonic circulation forms downstream of the IP as a stationary Rossby wave. In this case, the wavelength of the stationary Rossby wave λ is given by

$$\lambda = 2\pi\sqrt{\frac{U}{\beta}}, \tag{1}$$

where β is the planetary vorticity gradient and U is the average mean flow. When $\beta = 2.2 \times 10^{-11}$ at 15°N, and when $U = 10\ \mathrm{m\,s^{-1}}$ over the BOB at 850 hPa, the wavelength is about 4200 km. The half-wavelength (2100 km) is nearly equivalent to the zonal extent of the cyclonic circulation over the SCS (from the eastern IP to the Philippines).

b. The mechanism of increase in rainfall over the SCS and WNP during the CMB

The CMB over the IP appears simultaneously with an increase in rainfall over the SCS and WNP, which is consistent with a strong ridge over the IP and a deep trough over the SCS and WNP (Figs. 6b, 7b). In this subsection, the processes of rainfall increase over the SCS and WNP during the CMB are discussed.

The previous studies noticed a drastic increase in convections and rainfall over the SCS, WNP, and baiu region around late June (e.g., Wang and LinHo 2002). Wang and Xu (1997) noted that the wet event of CISO in mid-June corresponds with the drastic increase of convections over the WNP as a monsoon singularity. Wu and Wang (2001) pointed out the importance for a dramatic increase in rainfall of the slow seasonal evolution of large-scale circulation as well as the CISO over the SCS and WNP. LinHo and Wang (2002) found that the negative OLR anomalies on an intraseasonal time scale propagate eastward rapidly from south Asia to the SCS and WNP around the same time. Rainfall shows a simultaneous increase over the SCS, WNP, and baiu region during the CMB of the IP. Ueda et al. (1995) suggested that SST warming is not a sufficient condition, although it is certainly one important ingredient, to allow such a dramatic increase in convections over the WNP. Warm SSTs support unstable conditions over the SCS and WNP. An additional condition, such as a rapid atmospheric trigger, is necessary for a drastic increase in rainfall. The deep trough over the SCS appears, which is associated with the stationary Rossby waves generated over Indochina mountain ranges in the strong monsoon westerlies. The appearance of the trough over the SCS could trigger the increase in rainfall over the SCS.

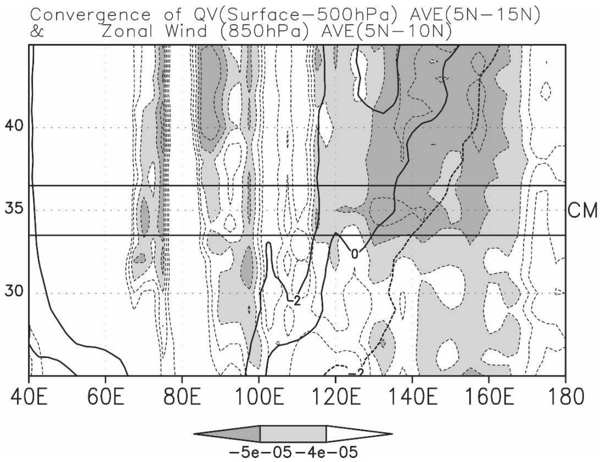

FIG. 10. Hovmöller diagram of pentad-averaged moisture convergence along the latitudinal band averaged within 5°–15°N, which is integrated from the surface to 500 hPa (shaded and thin contours). The averaged latitude is consistent with the rainfall increase over the WNP. The thick contours denote the latitudinal (5°–10°N) averaged zonal wind at 850 hPa from -2 to $2\ \mathrm{m\,s^{-1}}$. The contour interval is $2\ \mathrm{m\,s^{-1}}$.

Figures 6b and 7b shows that rainfall also increase over the WNP, and another cyclonic circulation is located there.

To reveal the processes of rainfall increase over the WNP, the Hovmöllor diagram of convergence of water vapor flux integrated from the surface to 500 hPa along a latitudinal band between 5° and 15°N was produced (Fig. 10). In this latitude band, the rainfall increases greatly during the CMB. To confirm the relationship the rainfall increases and the expansion of the monsoon westerlies, the zonal wind along 5°–15° was added in the figure. This latitude is the southern part of the cyclonic circulation over the WNP where the westerlies are enhanced (Fig. 7b). In this figure, the moisture convergence over the WNP is drastically intensified during the CMB. The intensification of moisture convergence seems to be a trigger for causing the drastic increase in rainfall over the WNP in the CMB. At P34, in addition, the monsoon westerlies rapidly expand eastward and reach the WNP, which corresponds with the intensification of moisture convergence over the WNP. This suggests that the enhancement of monsoon westerlies,

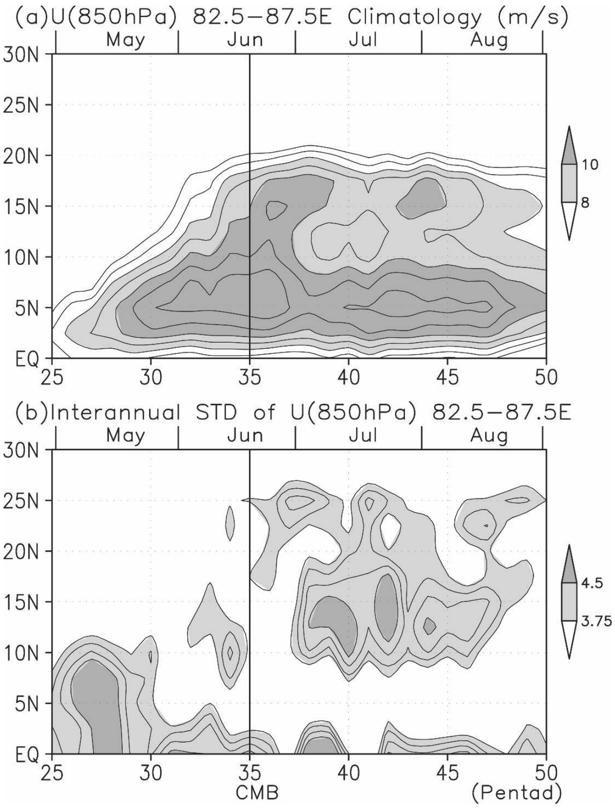

FIG. 11. The time–latitude cross section of climatological zonal wind (a) averaged over 82.5°–87.5°E and (b) its year-to-year std dev.

which is associated with the CMB, presumably makes the moisture convergence intensify over the WNP.

c. *Seasonal phase lock of the CMB*

This section discusses mechanisms behind the strong tendency of the CMB to occur at nearly the same time each year. The northward expansion of monsoon westerlies over the Indian monsoon region is likely responsible for the CMB, and how this drives the CMB at nearly the same time each year will be addressed. Figure 11 shows a time–latitude cross section of the climatological mean zonal wind over 50 yr (Fig. 11a) and its year-to-year standard deviation (Fig. 11b) along 85°E (the BOB longitude). In early and mid-June, strong zonal winds exist between the equator and 10°N. In late June, the area of the strong zonal wind expands dramatically northward. The standard deviation in early June between 5° and 10°N and in late June between 15° and 20°N is smaller than at any other period and latitude in Fig. 11b. Enhanced monsoon westerlies appear over the BOB at about the same time each year, in early June between 5° and 10°N and in late June between 15° and 20°N. This determines CMB timing. However, the origins and speed of the northward expansion of the monsoon westerlies may differ from year to year. The timing of the northward expansion of the monsoon westerly is consistent with the first seasonal phase lock of ISV of the monsoon westerlies (Nakazawa 1992). In addition, the area of expanded monsoon westerlies matches the area of increased rain, which indicates that northward migration of the active rainfall area accompanies the expansion of the monsoon westerlies. The northward migration of rainfall is likely to be associated with the northward propagation of ISV (Yasunari 1979, 1980, 1981). The first northward propagation of ISV over south and Southeast Asia tends to occur at nearly the same time almost every year, so the first northward propagation of the ISV over south Asia in June could be referred to as a climatological ISV over the South Asian monsoon region. Examination of the 850-hPa zonal winds for each year shows that the northward propagation occurs in June, although the origins and northward speeds of the ISV vary. Therefore, the CMB is associated with climatological ISV over the BOB. The ISV of OLR is weak (Wang and Xu 1997; LinHo and Wang 2002) over the IP, whereas the CMB in rainfall over Thailand is obvious. Because the decrease in rainfall during the CMB is spatially limited and complicated, OLR could not show the decrease in rainfall.

What physical mechanism causes the seasonal phase locking of the CMB to the annual cycle? The timing of the CMB may be significant because the summer solstice is coincident with the timing of the northward expansion of the monsoon westerlies and the heavy rains in the Indian monsoon, and this expansion is associated with the CMB. Jiang et al. (2004) noted that easterly vertical shear is important for the northward propagation of the ISV over the BOB. Easterly vertical shear is characterized by low-level monsoon westerlies and upper-level monsoon easterlies that are part of the Tibetan high. Intensification of monsoon easterlies in the upper troposphere as part of the Tibetan high is essential for the northward-propagating ISV. The Tibetan high is intensified by heating over and around the Tibetan Plateau during the onset of the Indian summer monsoon, so the timing of the peak in solar radiation, around the summer solstice, is closely related to the northward expansion of the low-level monsoon westerlies. In late July, low-level monsoon westerlies reintensify (Fig. 11a). However, the standard deviation in late July exceeds that in late June (Fig. 11b). The Asian monsoon may fluctuate because of internal dynamics rather than pure external forcing after the CMB.

6. Conclusions

This study investigated the climatological pentad mean annual cycle of rainfall in Thailand and the associated atmospheric circulation fields. The data used included rain gauge data for Thailand from the Thai Meteorological Department, satellite-derived rainfall data from CMAP, and atmospheric fields from the NCEP–NCAR reanalysis.

Climatological mean pentad values of rainfall taken over 50 yr clearly show a distinct CMB occurring over Thailand in late June. The atmospheric circulation changes drastically during the CMB not only over the IP but also over the Asian–Pacific monsoon region. Thus the CMB is a singularity in the Southeast Asian monsoon.

The decrease in rainfall is associated with the formation of a strong ridge over the mountain ranges of Indochina. The strong ridge anchored at the mountain ranges forms when monsoon westerlies strengthen along the latitudinal band of 15°–20°N. Enhancement of the monsoon westerlies over the BOB accompanies the northward expansion of the monsoon westerlies over south Asia as a part of the climatological ISV. The strong ridge in the lower troposphere over the mountain ranges of Indochina associated with the CMB is part of a stationary Rossby wave train that is generated by the strong monsoon westerlies and north–south-oriented topography. Therefore, small displacements of the enhanced monsoon westerlies have a significant nonlinear effect because of the effects of the mountain ranges of Indochina. The CMB also has an impact downstream. The stationary Rossby waves propagate downwind and a trough is enhanced over the SCS. This trough likely triggers a drastic increase in rainfall over the SCS. The trough over the SCS was deepened, which makes the monsoon westerlies reach the Philippines. The eastward-expanded monsoon westerlies intensify the moisture convergence over the WNP, which induces rainfall increase there.

Two types of rainfall systems were identified over Indochina. Rainfall associated with the first type peaks in late May to early June and is caused by an intensification of the monsoon southwesterlies along 10°N. Rainfall associated with the second type peaks in early September and is related to active tropical depressions originating in the SCS and WNP. The CMB occurs between the two peaks.

Thus, the CMB, namely the formation of the wave train on the strong monsoon westerlies, is a fundamental component the seasonal march of the Southeast Asian monsoon. It serves as a bridge that links monsoon rainfall from south Asia to the SCS and WNP. The CMB is, therefore, not solely a local phenomenon, but is an important singularity in the seasonal march of the monsoon over the Asian–Pacific region. Further study using regional climate models may help verify CMB forcing mechanisms.

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REFERENCES

Gadgil, S., 1977: Orographic effects on the southwest monsoon: A review. *Pure Appl. Geophys.,* **115,** 1413–1430.

Harr, P. A., and R. L. Elsberry, 1991: Tropical cyclone track characteristics as a function of large-scale circulation anomalies. *Mon. Wea. Rev.,* **119,** 1448–1468.

——, and ——, 1995: Large-scale circulation variation over the tropical western North Pacific. Part I: Spatial patterns and tropical cyclone characteristics. *Mon. Wea. Rev.,* **123,** 1225–1246.

He, H., W. McGinnis, Z. Song, and M. Yanai, 1987: Onset of the Asian summer monsoon in 1979 and the effect of the Tibetan Plateau. *Mon. Wea. Rev.,* **115,** 1966–1995.

Jiang, X., T. Li, and B. Wang, 2004: Structures and mechanisms of the northward propagating boreal summer intraseasonal oscillation. *J. Climate,* **17,** 1022–1039.

Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.,* **77,** 437–471.

Kang, I.-S., C.-H. Ho, Y.-K. Lim, and K.-M. Lau, 1999: Principal modes of climatological, seasonal, and intraseasonal variations of the Asian summer monsoon. *Mon. Wea. Rev.,* **127,** 322–340.

Kato, K., 1985: On the abrupt change in the structure of the baiu front over the China Continent in late May of 1979. *J. Meteor. Soc. Japan,* **63,** 737–750.

Kawamura, R., and T. Murakami, 1998: Baiu near Japan and its relation to summer monsoons over Southeast Asia and the western North Pacific. *J. Meteor. Soc. Japan,* **76,** 619–639.

Krishnamurti, T. N., and P. Ardanuy, 1980: The 10 to 20-day westward propagation mode and "breaks in the monsoons." *Tellus,* **32,** 15–26.

Lau, K.-M., G. J. Yang, and S. H. Shen, 1988: Seasonal and intraseasonal climatology of summer monsoon rainfall over East China. *Mon. Wea. Rev.,* **116,** 18–37.

LinHo, and B. Wang, 2002: The time–space structure of the Asian–Pacific summer monsoon: A fast annual cycle view. *J. Climate,* **15,** 2001–2019.

Matsumoto, J., 1992: The seasonal changes in Asian and Australian monsoon regions. *J. Meteor. Soc. Japan,* **70,** 257–273.

——, 1997: Seasonal transition of summer rainy season over Indochina and adjacent monsoon region. *Adv. Atmos. Sci.,* **14,** 231–245.

Murakami, T., and J. Matsumoto, 1994: Summer monsoon over the Asian Continent and Western North Pacific. *J. Meteor. Soc. Japan,* **72,** 719–745.

Nakazawa, T., 1992: Seasonal phase lock of intraseasonal variation during the Asian summer monsoon. *J. Meteor. Soc. Japan,* **70,** 597–611.

Ninomiya, K., and H. Muraki, 1986: Large-scale circulation over East Asia during Baiu period 1979. *J. Meteor. Soc. Japan,* **64,** 409–429.

Nitta, T., 1987: Convective activities in the tropical western Pacific and their impacts on the Northern Hemisphere summer circulation. *J. Meteor. Soc. Japan,* **65,** 373–390.

Rao, Y., 1976: *Southwest Monsoons. Meteor. Monogr.,* No. 1, Indian Meteorological Department, 367 pp.

Tao, S., and L. Chen, 1987: A review of recent research on the East Asian summer monsoon in China. *Monsoon Meteorology,* C.-P. Chan and T. N. Krishnamurti, Eds., Oxford University Press, 60–92.

Ueda, H., and T. Yasunari, 1996: Maturing process of the summer monsoon over the Western North Pacific—A coupled ocean/atmosphere system. *J. Meteor. Soc. Japan,* **74,** 493–508.

——, ——, and R. Kawamura, 1995: Abrupt seasonal change of large-scale convection activity over the western Pacific in northern summer. *J. Meteor. Soc. Japan,* **73,** 795–809.

Wang, B., and X. Xu, 1997: Northern Hemisphere summer monsoon singularities and climatological intraseasonal oscillation. *J. Climate,* **10,** 1071–1085.

——, and LinHo, 2002: Rainy season of the Asian–Pacific summer monsoon. *J. Climate,* **15,** 386–398.

Wu, R., and B. Wang, 2000: Interannual variability of summer monsoon onset over the western North Pacific and the underlying processes. *J. Climate,* **13,** 2483–2501.

——, and ——, 2001: Multi-stage onset of the summer monsoon over the western North Pacific. *Climate Dyn.,* **17,** 277–289.

Xie, P., and P. Arkin, 1997: Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. *Bull. Amer. Meteor. Soc.,* **78,** 2539–2558.

Yasunari, T., 1979: Cloudiness fluctuations associated with the northern hemisphere summer monsoon. *J. Meteor. Soc. Japan,* **57,** 227–242.

——, 1980: A quasi-stationary appearance of 30- to 40-day period in the cloudiness fluctuations during the summer monsoon over India. *J. Meteor. Soc. Japan,* **58,** 225–229.

——, 1981: Structure of the Indian monsoon system with around 40-day period. *J. Meteor. Soc. Japan,* **59,** 336–354.

Yoshino, M., 1965: Four stages of the rainy season in early summer over East Asia (part 1). *J. Meteor. Soc. Japan,* **43,** 231–245.

Zhang, Y., T. Li, B. Wang, and G. Wu, 2002: Onset of the summer monsoon over the Indochina Peninsula: Climatology and interannual variations. *J. Climate,* **15,** 3206–3221.