Interaction between the Summer Monsoons in East Asia and the South China Sea: Intraseasonal Monsoon Modes

TSING-CHANG CHEN, MING-CHENG YEN,* AND SHU-PING WENG

Atmospheric Science Program, Department of Geological and Atmospheric Sciences, Iowa State University, Ames, Iowa

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ABSTRACT

The summer monsoons in East and Southeast Asia are characterized, respectively, by the Mei-yu (in eastern China)–Baiu (in Japan) front (MBF) and by the monsoon trough stretching from northern Indochina to the Philippine Sea. These two major monsoon elements are separated by the North Pacific anticyclone. As indicated by the 850-mb zonal wind and cumulus convection over some key areas, a distinct opposite-phase intraseasonal variation exists between the two monsoon elements. Two approaches are adopted to explore the cause of this opposite-phase variation (which reflects the coupling between the two monsoon components): 1) the correlation coefficient patterns between the 850-mb zonal-wind monsoon index and the 850-mb streamfunction field and 2) the composite 850-mb streamline charts and the 120°E zonal-wind cross sections. It is shown that the opposite-phase variation between the two monsoon elements is caused by the anomalous circulation associated with the northward-migrating 30–60-day monsoon trough/ridge from the equator to 20° N and with the westward-propagating 12–24-day monsoon low–high along the latitude of ~15°–20° N. Results obtained in this study are used to address two often discussed phenomena of the East Asian monsoon: 1) the rapid northward shift of the MBF across the Yangtze River basin during the Mei-yu onset is related to the north–south meridional oscillation of the MBF; and 2) the three longitudinally oriented location zones of extremely heavy rain events in eastern China are formed by the alternation of deep cumulus convection zones associated with the intraseasonal monsoon vortices (centered in the northern part of the South China Sea) between extreme monsoon conditions.

1. Introduction

The East Asian monsoon is characterized by the Mei-yu–Baiu front (MBF), which is climatologically one of the major convergence zones within the global atmospheric circulation. During the summer monsoon season, the MBF extends from southern China to Japan and the Aleutian Islands, and exhibits a definite moisture gradient (but no temperature gradient). Water vapor originating from the South China Sea (SCS)–western tropical Pacific (WTP) region is advected by the monsoon southwesterlies along the MBF (Murakami 1959). Moisture and precipitation associated with the front are maintained by water vapor flux convergence toward this front (Akiyama 1973). This hydrological process is driven by the large-scale divergent circulation (Chen et al. 1988a). Following the seasonal march, the MBF migrates northward from northern Indochina during May to the Yangtze River basin during June, triggering the onset of the East Asian summer monsoon. However, this northward progression is not uniform, but undergoes a north–south intraseasonal oscillation induced by the eastward propagation of the intraseasonal global divergent circulation (Chen and Murakami 1988). Conceivably, the seasonal evolution of the East Asian monsoon is modulated by this meridional oscillation.

Analyzing temporal evolution of the East Asian summer monsoon rainfall, Lau et al. (1988) identified two onsets of major monsoon rain: the first (known as the Mei-Yu) occurs in central China during the first half of June, and the second occurs over northeastern China during late July. The multiple onsets of the East Asian summer monsoon rain transpire as major rainbands undergo rapid transitions that progress wavelike from south to north between three quasi-stationary locations: southern China (premonsoon rain), central China (the Mei-yu front), and northeastern China (the polar front). As suggested by Lau et al.’s empirical orthogonal function analysis of the East Asian summer monsoon rainfall, the rapid transitions of these major rainbands may result from the phase lock between the 30–60-day and 20-day intraseasonal monsoon modes.
The water vapor in the SCS–WTP region is supplied by the convergence of water vapor flux transported by the monsoon westerlies and by the trade easterlies (e.g., Chen et al. 1988a). It was observed that the intensity of south Asian monsoon westerlies (an indicator of the monsoon life cycle) is modulated by the northward-propagating 30–60-day monsoon trough/ridge (Krishnamurti and Subrahmanyam 1982; Krishnamurti et al. 1985) and that the trade easterlies associated with the North Pacific anticyclone are oscillated by the eastward-propagating, global 30–60-day mode (Chen 1987). For these reasons, water vapor and water vapor flux in the SCS–WTP region fluctuate intraseasonally (Chen et al. 1988b). Since rainfall is maintained by the convergence of water vapor flux, it is expected that the SCS–WTP monsoon rain undergoes an intraseasonal fluctuation. Actually, this inference is substantiated by a dipole structure of tropical convection with an intraseasonal time scale identified by Lau and Chan’s (1986) EOF analysis of the outgoing longwave radiation (OLR) between the Indian Ocean and the SCS–WTP regions.

Examining the time variation of the SCS summer monsoon, Chen and Chen (1995) showed that the 30–60-day monsoon trough–ridge in the SCS–WTP region is coupled with the 30–60-day global divergent circulation in a way similar to the type of coupling that occurs over the Indian monsoon region. The life cycle of the SCS summer monsoon is regulated by the 30–60-day monsoon trough–ridge in such a manner that the monsoon circulation reaches its maximum (minimum) intensity when this trough (ridge) moves to approximately 15°–20°N. The SCS summer monsoon break occurs simultaneously with a phase lock in the north SCS between the northward-propagating 30–60-day monsoon ridge and the westward-propagating 12–24-day closed high. The most northward extension of the anomalous circulation associated with these two intraseasonal monsoon modes may reach the Yangtze River basin.

As shown in previous studies, intraseasonal oscillations exist in both the summer monsoons in East Asia and the SCS–WTP region. For East Asia, the MBF undergoes a north–south intraseasonal oscillation, and the monsoon rainfall, subsequently, exhibits an intraseasonal fluctuation. In the SCS–WTP region, the summer monsoon life cycle, cumulus convection, and hydrological processes are regulated by the intraseasonal monsoon mode. In view of the intraseasonal oscillations of the monsoons over the two regions, a question is raised: Is there a physical and systematic link between intraseasonal oscillations of the summer monsoon in East Asia and the SCS–WTP region? In their examination of the evolution of the upper-level circulation during the Mei-Yu onset in eastern China, Yeh et al. (1959) presented a series of 3-km streamline charts. Synoptic development shown in these charts suggest that the Mei-Yu onset is related in some way to the northward migration of the monsoon trough, which extends longitudinally from the Indian subcontinent to the SCS.

Based on the $y-t$ diagram of a convective index along 140°E, Chen and Murakami (1988) pointed out that the cumulus convection associated with the ITCZ in the WTP region is enhanced (suppressed) when the MBF migrates southward (northward). Conceivably, there should be a physical link between the intraseasonal oscillations of the summers monsoons in East Asia and the SCS–WTP region.

So far, the mechanism responsible for the physical link between the monsoons in these two regions has not been systematically examined. In view of the research findings by Yeh et al. (1959) and Chen and Murakami (1988), the possible coupling of the two monsoon components may be explained through the coherent intraseasonal oscillations of the MBF and the life cycle of the SCS summer monsoon. A research task along this line is undertaken with two major data sources: the National Centers for Environmental Prediction–National Center for Atmospheric Research NCEP–NCAR reanalysis data (Kalnay et al. 1996) for the period 1979–93 and the equivalent blackbody temperature ($T_{BB}$) observed by the Japanese Geostationary Meteorological Satellite (GMS) for the period 1980–93. Results of this study are presented in the following arrangement. The possible coupling of the two monsoon components is demonstrated by the coherent intraseasonal oscillations of monsoon westerlies and cumulus convection in section 2. The coupling mechanism of the two monsoon components is illustrated by correlation coefficient patterns, composite synoptic charts, and zonal-wind cross sections in section 3. Section 4 offers a discussion on the possible effects of the coupling of the two monsoons on two often discussed phenomena of the East Asian monsoon: 1) the rapid shifts of upper westerlies and the MBF during the Mei-yu onset and 2) the formation of three east–west-oriented zones of extremely heavy rain events in eastern China. Finally, section 5 is devoted to concluding remarks.

2. Indication of possible coupling

a. Time series of monsoon indices

To indicate the temporal variation of monsoon intensity and life cycle, rainfall and low-level wind are commonly adopted as monsoon indices (e.g., Murakami 1972; Krishnamurti 1985). For this reason, an inference of the possible coupling of the summer monsoons in East and Southeast Asia may be made with time series of these indices at some key climatological locations associated with the MBF and with the SCS–WTP monsoon trough.

Since there are no precipitation data available over the open seas, some precipitation proxy generated from satellite data will be used. Arkin and Ardanuy (1989) proposed that tropical rainfall may be estimated with OLR values below the threshold of 235 W m$^{-2}$. Nitta and Sekine (1994) inferred that deep cumulus convec-
tion was associated with cloud tops above 400 mb with $T_{bb}$ values below 250 K. Instead of this $T_{bb}$ threshold, Chen and Chen (1995) used the $T_{bb}$ value of 270 K as a threshold to include low convective clouds. For the rainfall and cumulus convection proxies, we define the following two convective indices:

$$\Delta \text{OLR} = 235 \text{ W m}^{-2} - \text{OLR} \quad (=0 \text{ if } \Delta \text{OLR} \leq 0)$$

and

$$\Delta T_{bb} = 270 \text{ K} - T_{bb} \quad (=0 \text{ if } \Delta T_{bb} \leq 0).$$

In addition to the rainfall and cumulus convection proxies, the 1-yr rainfall estimations ($P$) for 1979 generated from the IR data by Susskind and Pfaffertner (1989) are also incorporated into our analysis.

Shown in Fig. 1 are histograms of $P$, $\Delta \text{OLR}$, $\Delta T_{bb}$ and time series of $\bar{u}(850 \text{ mb})$ for the summers of 1979 (left) and for 1989 (right) at locations of maximum rms (root-mean-square) value centers for these variables over the SCS and eastern China. The selection of these two summers will be explained later. Affected by the interannual variation of the summer circulation in the East Asia–western Pacific region (e.g., Nitta 1987; Chen and Weng 1997), locations of the rms centers may not always be the same in every summer. In Fig. 1, the upper histograms and time series are located at or south of the Yangtze River basin, and the lower ones are at or south of the SCS–WTP summer monsoon trough. For comparison, the upper histograms and time series are plotted upside down. Variations of the summer monsoons in East Asia and the SCS–WTP region are well indicated by these monsoon indices. Previous studies cited in the introduction report that variations of the two monsoon components are modulated by the 30–60-day and 12–24-day monsoon modes. To illustrate the roles of these two intraseasonal modes in the variations of the two monsoon components, a combination of the seasonal mean value, the 30–60-day, and the 12–24-day bandpass filtered monsoon indices [solid lines in Fig. 1, denoted by ( ) for later discussion] are superimposed on histograms and time series. The second-order Butterworth bandpass filter (Murakami 1979) is used to isolate these two temporal regimes. Salient features of these indices include the following.

1) The MBF undergoes a north–south intraseasonal oscillation, and the East Asian monsoon rain exhibits an intraseasonal fluctuation that is caused by the northward migration of intraseasonal monsoon modes (Lau et al. 1988). These temporal variations in the East Asian monsoon are reflected by the histograms of $\bar{P}$, $\Delta T_{bb}$, $\Delta \text{OLR}$, and the time series of $\bar{u}(850 \text{ mb})$.

2) The life cycle of the SCS–WTP summer monsoon is regulated basically by the northward-migrating 30–60-day monsoon trough–ridge, and the onset (break) of this monsoon is often the result of a phase lock between the 30–60-day monsoon trough (ridge) and the 12–24-day monsoon low (high) [Chen and Chen (1995); Chen and Weng 1997]—however, to make the paper more self-contained, the propagation properties of the two intraseasonal monsoon modes are provided in the appendix. Since cumulus convection in the SCS–WTP region undergoes a pronounced intraseasonal fluctuation (e.g., Lau and Chan 1986), it is not surprising to see that a clear intraseasonal fluctuation emerges from the monsoon indices in the SCS–WTP region.

3) The spectral analysis of all monsoon indices reveals that both the 30–60-day and 12–24-day signals are distinct but may not always exist simultaneously in both monsoon components. One intraseasonal mode may be more significant than the other during one monsoon season, while the reverse situation may be true in another monsoon season. The contrast between the bandpass filtered monsoon indices of the two monsoons in the summer shows clearly that, in the summer of 1979, the 30–60-day mode is more pronounced than the 12–24-day mode, while the 12–24-day mode is the dominant one in the summer of 1989. Due to this, we selected these two summers to illustrate the possible coupling of the two monsoons by these two intraseasonal monsoon modes.

4) The most interesting and important feature emerging from Fig. 1 is the opposite-phase variation of the same monsoon indices between the two monsoons. The correlation coefficients of all corresponding bandpass-filtered monsoon indices during the two monsoons are above 0.7 in the summer of 1989. For the summer of 1979, the correlation coefficients of $\bar{u}(850 \text{ mb})$, $\bar{p}$, and OLR between the two monsoons are $-0.7$, $-0.65$, and slightly below $-0.5$, respectively. The opposite-phase variations are by no means accidental, but indicate strongly a physical coupling of the two monsoons through some mechanism. In view of contributions of the 30–60-day and 12–24-day modes to the temporal variations of all monsoon indices, it is conceivable that they play a vital role in the coupling of the two monsoons.

Presented in Fig. 1 are monsoon indices of only two monsoon seasons. One may question whether the opposite-phase variations of monsoon indices between the two monsoons exist in other summers. Showing histograms and time series of monsoon indices for all seasons may not be a practical way to answer this question. Thus, we show correlation coefficients of the bandpass filtered monsoon indices between the two for all summer seasons (Fig. 2). For most monsoon seasons, correlation coefficients of $\bar{u}(850 \text{ mb})$, $\Delta \text{OLR}$, and $\Delta T_{bb}$ are over $-0.7$, $-0.6$, and $-0.5$, respectively. Evidently, the opposite-phase relationship between the temporal evolu-

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1 The power spectra of monsoon indices averaged over 15 summers (1979–93) are given in the appendix.
Fig. 1. Indices of the monsoon westerlies and cumulus convection (or precipitation) for East Asia (upper part of each) and the SCS (lower part of each) during the summers of 1979 (left) and 1989 (right) are displayed. (left) A daily histogram of Goddard precipitation estimation (top), \( \Delta \text{OLR} \) \( (=235 \text{ W m}^{-2} - \text{OLR}) \) (middle), and the time series of daily \( u(850 \text{ mb}) \) (bottom) for the summer of 1979; (right) daily histograms of \( \Delta \text{OLR} \) (top) and of \( \Delta T_{\text{mb}} \) \( (=270 \text{ K} - T_{\text{mb}}) \) (middle), and the time series of daily \( u(850 \text{ mb}) \) (bottom) for the summer of 1989. The location of each of the monsoon indices is shown. Superimposed on both the histogram and the time series is the combined 30–60-day and 12–24-day filtered time series of this variable (thick solid line).

The intensity variation of the monsoon circulation is reflected by the low-level monsoon westerlies. Thus, the temporal evolutions of the 1979 and 1989 SCS summer monsoons are revealed clearly from both the \( u(850 \text{ mb}) \) \( y-t \) diagram south of 20°N and the \( \hat{u}(850 \text{ mb}) \) indices of these two monsoon seasons. For the East Asian summer monsoon, let us use the latitudinal location of the maximum \( u(850 \text{ mb}) \) to indicate approximately the core of the monsoon southwesterlies associated with the MBF. With this approximation, the north–south oscillation of the MBF latitudinal location is reflected by the north–south oscillation of the maximum \( u(850 \text{ mb}) \) location [revealed from the \( u(850 \text{ mb}) \) \( y-t \) diagram north of 20°N].

The question posed above may be answered by a simple contrast between the SCS \( \hat{u}(850 \text{ mb}) \) index and the filtered MBF location index.

1) The amplitude of the MBF’s location index is about 20° latitude, which is comparable to the north–south extent covered by the northward migration of the 30–60-day monsoon trough/ridge in the SCS region. For the 12–24-day monsoon mode, its latitudinal extent has not yet been presented (but will be illustrated later). Actually, the circulation pattern and size of this intraseasonal mode in the SCS region are similar to those associated with the 30–60-day monsoon trough–ridge.

2) For both the 1979 and 1989 summer monsoon seasons, the MBF location index oscillated in the north–south direction coherently with the intraseasonal variation of the SCS \( \hat{u}(850 \text{ mb}) \) index. The correlation coefficients between these two time series are 0.91 and 0.95, respectively. The coherent relationship between the two indices, shown in Fig. 3, is present in every summer (not shown) analyzed in this study.

The coherent intraseasonal variations in the monsoon indices of the two monsoon components presented in this section and in section 2a lead us to a basic question: What is the physical mechanism responsible for the coherent northward migration (southward retreat) of the MBF in East Asia and the intraseasonal variation of the summer monsoon intensity in the SCS–WTP region? In other words, what is responsible for the coupling of the monsoons over these two regions?

3. Coupling mechanism

Analyzing the lower-tropospheric streamline charts during the second half of May and early June 1956, Yeh et al. (1959) pointed out that the Mei-Yu onset (triggered...
by the northward movement of a shear line) is linked to the northward migration and eastward extension of the monsoon trough from the Indian subcontinent to the SCS region. With a careful examination of Yeh et al.’s streamline charts (their Fig. 16), one can also see that the lower-tropospheric high retreats from the SCS region into the western North Pacific when the monsoon trough moves northward. As observed by previous studies (e.g., Chen and Chen 1995), the northward-migrating 30–60-day monsoon trough/ridge over the SCS–WTP region is associated with an anomalous cyclonic (anticyclonic) circulation, and the westward-propagating 12–24-day monsoon mode (which plays an important role in the SCS monsoon onset and break) possesses a double-celled structure with its northern cell moving along 15°–20°N and its southern cell along the equator. The opposite-phase intraseasonal variations of monsoon indices between the two monsoons may be established through their coupling by the anomalous circulation cells associated with these two intraseasonal monsoon modes. This argument will be substantiated by using two approaches in the following two sections: 1) correlation coefficient patterns between the \( \bar{u}(850 \text{ mb}) \) monsoon index and the lower-tropospheric streamfunction and (2) composite synoptic charts of lower-tropospheric streamline and zonal-wind cross sections.

**a. Correlation coefficient pattern**

The spatial structure of summer monsoon disturbances associated with the intraseasonal fluctuation of a monsoon index may be inferred from the correlation

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**Fig. 3.** The latitude–time \((y-t)\) diagrams of \(u(850 \text{ mb})\) \((115^\circ E)\) for (a) the summer of 1979 and (b) the summer of 1989. The thick solid line is a combination of the 30–60-day and 12–24-day bandpass filtered \(\bar{u}(850 \text{ mb})\) time series at the center of the root-mean-square value of \(\Delta u(850 \text{ mb}) := u(850 \text{ mb}) - \bar{u}(850 \text{ mb})\) (\(\bar{u} = \text{ summer mean(\(u)\)}\) in the SCS region. The thick dashed line is the combined bandpass filtered 30–60-day and 12–24-day maximum \(\Delta u(850 \text{ mb})\) location associated with the MBF. The contour interval of \(u(850 \text{ mb})\) is 2.5 m \(\text{s}^{-1}\). Values of 2.5 m \(\text{s}^{-1}\) ≤ \(u(850 \text{ mb})\) < 5 m \(\text{s}^{-1}\) and 5 m \(\text{s}^{-1}\) ≤ \(u(850 \text{ mb})\) are lightly and heavily stippled, respectively.
Coefficient pattern between this index and streamfunction. Therefore, we construct the correlation coefficient patterns between the monsoon index of the 850-mb monsoon westerlies and the 850-mb streamfunction fields for three temporal regimes: 1) the departure from the seasonal cycle, $\Delta(t)$; 2) the 30–60-day mode, $(\sim)$; and 3) the 12–24-day mode, $(\hat{\sim})$. The seasonal cycle is simply removed by a least-squares-fit approach, while the two intraseasonal modes are isolated by the first-order Butterworth filter (Murakami 1979). For convenience of later discussion, we denote the correlation coefficients of the three temporal regimes as $\sigma_{\Delta,\hat{\sim}}$, $\sigma_{\hat{\sim},\hat{\sim}}$, and $\sigma_{\hat{\sim},\hat{\sim}}$.

The procedure for computing the three aforementioned correlation coefficients is outlined as follows.

1) The monsoon westerly indices $[\Delta\bar{u}(850 \text{ mb}), \bar{u}(850 \text{ mb}), \text{and } \hat{u}(850 \text{ mb})]$ are formed at rms $[\Delta\bar{u}(850 \text{ mb})]$ centers over the SCS and eastern China for each summer monsoon season.

2) We select in each monsoon season the periods when $[\Delta\bar{u}(850 \text{ mb})] \geq 0.8\text{SD}$, and $[\bar{u}(850 \text{ mb})] \geq 0.8\text{SD}_2$, $[\text{SD}_1$ and $\text{SD}_2$, are standard deviations of $\bar{u}(850 \text{ mb})$ and $\hat{u}(850 \text{ mb})$, respectively, over the entire monsoon season] for constructing the correlation coefficient patterns.

3) The filtered monsoon westerly index and streamfunction of selected periods for all 15 monsoon seasons in each temporal regime are assembled sequentially (following the order of years) to form a new set of data. Correlation coefficient patterns, $\sigma_{\Delta,\hat{\sim}}$, $\sigma_{\hat{\sim},\hat{\sim}}$, and $\sigma_{\hat{\sim},\hat{\sim}}$, of three temporal regimes are constructed with the three newly formed time series of the filtered monsoon westerly index and the corresponding streamfunction anomalies.

Correlation coefficient patterns of the three temporal regimes with the SCS (East Asia) monsoon westerly indices are displayed in Figs. 4a–c (Figs. 4d–f). Salient features of these patterns are as follows.

1) A double-celled structure dominates all three types of correlation coefficient patterns. As revealed from Fig. 1, the most distinctive intraseasonal oscillations in the $\Delta\bar{u}(850 \text{ mb})$ index are the 30–60-day and 12–24-day monsoon modes. The resemblance of the spatial pattern between $\sigma_{\Delta,\hat{\sim}}$ and the other two (i.e., $\sigma_{\hat{\sim},\hat{\sim}}$ and $\sigma_{\hat{\sim},\hat{\sim}}$) indicates that the intraseasonal variation of $\Delta\bar{u}(850 \text{ mb})$ fields is dominated primarily by these two intraseasonal monsoon modes.

2) The $\Delta\hat{u}(850 \text{ mb})$ index in the SCS reaches its maximum (minimum) value when the 30–60-day monsoon trough (ridge) arrives at 15°N–20°N and a new 30–60-day monsoon ridge (trough) appears near the equator. The double-celled structure of $\sigma_{\hat{\sim},\hat{\sim}}$ manifests when the anomalous cyclonic (anticyclonic) circulation centered in the north SCS attains its maximum intensity as the 30–60-day monsoon trough (ridge) arrives there. At this stage, the anomalous circulation at the equator should be opposite to its northern counterpart.

3) Because of the spatial structure of the 12–24 day monsoon mode, the double-celled pattern of $\sigma_{\hat{\sim},\hat{\sim}}$ is not a surprise. In view of the differences of propagation properties and life cycle between both this intraseasonal mode and the 30–60-day monsoon mode, the similarity in the spatial structures of $\sigma_{\hat{\sim},\hat{\sim}}$ and $\sigma_{\hat{\sim},\hat{\sim}}$ is surprising and may reflect two possibilities: (a) the relationship between the $\hat{u}(850 \text{ mb})$ monsoon index and the anomalous circulation associated with the 12–24-day monsoon mode behaves in a way similar to the 30–60-day monsoon mode, and (b) the spatial structure of the anomalous circulation of the 12–24-day monsoon mode during its mature stage over the SCS resembles that of the 30–60-day monsoon mode.

4) If opposite-phase intraseasonal variations of the monsoon indices in the two monsoons are induced by the same monsoon disturbance, the following may be expected: the spatial patterns of $\sigma_{\hat{\sim},\hat{\sim}}$ and $\sigma_{\hat{\sim},\hat{\sim}}$ constructed with the 850-mb zonal-wind indices of both the SCS and East Asian monsoons should resemble each other but have opposite spatial structures. This expectation is confirmed by the contrast of corresponding correlation coefficient patterns in Figs. 4a–c and Figs. 4d–f.

In summary, feature 4 of the correlation coefficient patterns shown in Fig. 4 implies that the intraseasonal oscillations of the two monsoons are coupled by the anomalous circulations associated with the northward-migrating 30–60-day monsoon trough–ridge and with the westward-propagating 12–24-day monsoon low–high. However, a more quantitative assessment of contributions of the two intraseasonal modes to the intraseasonal variations of monsoon indices cannot be attained directly from the correlation coefficient patterns. Instead, this assessment may be made through the use of composite streamline charts and zonal-wind cross sections.

b. Composite charts

To illustrate the possible coupling of the two monsoon components, composite charts are prepared for three temporal regimes: total, 30–60-day, and 12–24-day. The composite procedure starts with the case selection using the following criteria.

1) The SCS zonal-wind monsoon index $[\bar{u}(850 \text{ mb})$ or $\hat{u}(850 \text{ mb})]$ exceeds 0.8 of its standard deviation (SD$_1$ or SD$_2$) over an entire monsoon season.

2) A distinct closed cyclonic (anticyclonic) cell is centered at the northern SCS when the zonal-wind index $\bar{u}(850 \text{ mb})$ or $\hat{u}(850 \text{ mb})$) meets criterion 1.

3) For the 30–60-day monsoon mode, a clear trough (ridge) appears in the northern SCS when the $\hat{u}(850 \text{ mb})$ index satisfies criterion 1.

All cases selected (see Tables 1 and 2) are averaged
Fig. 4. Listed in Table 1(2) are the dates when the maximum and minimum \( \bar{u}_{(850 \text{ mb})} \) in the SCS are larger than 0.8 s and smaller than 0.8 s, respectively. Here, \( \sigma \) is the standard deviation of the \( \bar{u}_{(850 \text{ mb})} \) or \( \bar{u}_{(850 \text{ mb})} \) time series over a summer (May–Aug). The selected cycles of \( \bar{u}_{(850 \text{ mb})} \) [or \( \bar{u}_{(850 \text{ mb})} \)] time series are combined chronologically to form a new time series. The correlation coefficients between the newly formed time series and the corresponding \( \bar{u}_{(850 \text{ mb})} \) [or \( \bar{u}_{(850 \text{ mb})} \)] fields are shown in (b) [(c)]. The SCS \( \Delta \bar{u}_{(850 \text{ mb})} \) time series covered by the selected \( \bar{u}_{(850 \text{ mb})} \) and \( \bar{u}_{(850 \text{ mb})} \) cycles are lined up to form a new \( \Delta \bar{u}_{(850 \text{ mb})} \) time series that is correlated with the corresponding \( \Delta u_{(850 \text{ mb})} \) fields. The correlation coefficients generated from the selected \( \Delta \bar{u}_{(850 \text{ mb})} \) time series and \( \Delta u_{(850 \text{ mb})} \) fields are displayed in (a). Shown in (d)–(f) are the correlation coefficient patterns corresponding to those in (a)–(c) with the zonal-wind index over East Asia and the 850-mb filtered streamfunction.

over a 3-day window with its center date coinciding with the maximum/minimum \( \bar{u}_{(850 \text{ mb})} \) or \( \bar{u}_{(850 \text{ mb})} \) index. After this composite procedure is complete, two types of composite charts are constructed:

1) the 850-mb streamline charts superimposed with the \( T_{\text{bb}} \) index and
2) the zonal-wind cross sections at 120°E superimposed with the latitudinal distribution of the \( T_{\text{bb}} \) index along this longitude.

1) The 30–60-day mode

(i) The 850-mb streamline charts

In order to save space, shown in Fig. 5 are only the streamline and \( T_{\text{bb}} \) anomaly charts of the two extreme monsoon conditions when the SCS–\( \bar{u}_{(850 \text{ mb})} \) index reaches its maximum (minimum) value and the differences between them.

1) Total 850-mb winds. Because of the northwestward extension of the North Pacific anticyclone’s ridge line into northern China during the active monsoon phase (Fig. 5a), the MBF convection zone retreats northeastward to cover only the region stretching from the Yangtze Delta to the northwest Pacific. The monsoon trough (south of the North Pacific anticyclone) is deepened and stretched from northern Indochina to the Philippine Sea. The deepening of this trough enhances cumulus convection and also strengthens the monsoon westerlies associated with
During the break monsoon phase, the ridge line of the North Pacific anticyclone (Fig. 5b) intrudes southward into the northern part of the SCS. The MBF convection zone is extended southwestward to cover southern China in such a way that the SCS convection zone is moved equatorward and the monsoon westerlies are significantly weakened.

The opposite-phase intraseasonal variation of the monsoon indices in the SCS and East Asian region can be inferred directly from the comparison of the monsoon circulation between the two extreme monsoon conditions. This comparison is accomplished easily by the difference between them (Fig. 5c). Like the $\sigma_{anomaly}$ pattern (Fig. 4a), the difference streamline chart of $\Delta V(850 \text{ mb})$ between the active and break monsoon phases exhibits a double-celled structure. A meridional juxtaposition of the east–west-elongated positive and negative $\Delta T_{\text{anomaly}}$ zones can be seen clearly, although major convective zones are associated with the SCS monsoon trough and the MBF. A reversal of the $\Delta V(850 \text{ mb})$ monsoon flow pattern and the meridional juxtaposition of convection zones are expected when the contrast of the two extreme monsoon conditions is reversed. It becomes evident that the northward (southward) shifts of the MBF and the convection zone associated with the MBF in eastern China follow the strengthening (weakening) of the SCS monsoon westerlies and the enhancement (suppression) of cumulus convection along the monsoon trough.

2) The 30–60-day filtered winds. The role played by the 30–60-day monsoon mode is illustrated by the composite charts of the 30–60-day filtered streamline, and $\Delta T_{\text{anomaly}}$ anomalies corresponding to Figs. 4a–c are displayed in Figs. 5d–f. The double-celled structure of the $\Delta V(850 \text{ mb})$ streamline and the meridional juxtaposition of $\Delta T_{\text{anomaly}}$ in Fig. 5c appear again in the $\Delta V(850 \text{ mb})$, $\Delta T_{\text{anomaly}}$ chart of the active monsoon phase (Fig. 5d). In contrast, the directions of the $\Delta V(850 \text{ mb})$ flow pattern and of the meridional $\Delta T_{\text{anomaly}}$ juxtaposition are reversed during the break monsoon

<table>
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<th>Year</th>
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China Sea and the Philippine Sea. Thus, the monsoon trough in the SCS becomes oriented southeastward (across the sea to reach Borneo) in such a way that the SCS convection zone is moved equatorward and the monsoon westerlies are significantly weakened.

The opposite-phase intraseasonal variation of the monsoon indices in the SCS and East Asian region can be inferred directly from the comparison of the monsoon circulation between the two extreme monsoon conditions. This comparison is accomplished easily by the difference between them (Fig. 5c). Like the $\sigma_{anomaly}$ pattern (Fig. 4a), the difference streamline chart of $\Delta V(850 \text{ mb})$ between the active and break monsoon phases exhibits a double-celled structure. A meridional juxtaposition of the east–west-elongated positive and negative $\Delta T_{\text{anomaly}}$ zones can be seen clearly, although major convective zones are associated with the SCS monsoon trough and the MBF. A reversal of the $\Delta V(850 \text{ mb})$ monsoon flow pattern and the meridional juxtaposition of convection zones are expected when the contrast of the two extreme monsoon conditions is reversed. It becomes evident that the northward (southward) shifts of the MBF and the convection zone associated with the MBF in eastern China follow the strengthening (weakening) of the SCS monsoon westerlies and the enhancement (suppression) of cumulus convection along the monsoon trough.

2) The 30–60-day filtered winds. The role played by the 30–60-day monsoon mode is illustrated by the composite charts of the 30–60-day filtered streamline, and $\Delta T_{\text{anomaly}}$ anomalies corresponding to Figs. 4a–c are displayed in Figs. 5d–f. The double-celled structure of the $\Delta V(850 \text{ mb})$ streamline and the meridional juxtaposition of $\Delta T_{\text{anomaly}}$ in Fig. 5c appear again in the $\Delta V(850 \text{ mb})$, $\Delta T_{\text{anomaly}}$ chart of the active monsoon phase (Fig. 5d). In contrast, the directions of the $\Delta V(850 \text{ mb})$ flow pattern and of the meridional $\Delta T_{\text{anomaly}}$ juxtaposition are reversed during the break monsoon

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Fig. 5. Composite streamline charts superimposed with $\Delta T_{\text{BB}} [=270 \text{K} - T_{\text{BB}} \geq 0$ or $=0$ if $270 \text{K} - T_{\text{BB}} < 0]$ for (a) active phase, (b) break phase, and (c) the difference between (a) and (b). The corresponding 30–60-day filtered streamline and $\Delta T_{\text{BB}}$ charts are displayed in (d)–(f). Any 30–60-day cycle of the SCS $\tilde{u}(850 \text{mb})$ index $\pm 0.8\sigma$ [$\sigma$ the standard deviation of the SCS $\tilde{u}(850 \text{mb})$ index] during any given summer is selected for composite. The composite procedure includes a 3-day window with the centered date corresponding to date of the maximum or minimum $\tilde{u}(850 \text{mb})$ index. Values of $0 \leq \Delta T_{\text{BB}} \leq 15 \text{K}$ and $0 \leq \tilde{u}(850 \text{mb}) \leq 5^\circ$ are lightly stippled, while those of $15 \text{K} \leq \Delta T_{\text{BB}} \leq 5 \text{K}$ are heavily stippled.

phase (Fig. 5e). Recall that the SCS summer monsoon attains its maximum (minimum) intensity when a 30–60-day monsoon trough (ridge) arrives at 20$^\circ$N with a cyclonic (anticyclonic) cell and a maximum (minimum) convection zone (Chen and Chen 1995). The reversal of the $\Delta \tilde{V}(850 \text{mb})$ circulation pattern and of the meridional $\Delta \tilde{T}_{\text{BB}}$ juxtaposition between the two extreme monsoon conditions (Figs. 5a,b) re-
sults from the alternative northward migration of the 30–60-day monsoon trough and ridge from the equator to near 20°N. To strengthen this argument further, displayed in Fig. 5f is the difference in [V(850 mb), ΔTmm(120°E)] between the two extreme monsoon conditions. The close resemblance of circulation and ΔTmm anomalies between Figs. 5c and 5f indicates that the 30–60-day monsoon mode is a major agent responsible for the change in synoptic conditions between the active and break monsoon phases and, in turn, for the opposite-phase variation in the monsoon indices of the SCS and East Asian monsoons.

(ii) Latitude–height cross sections

The latitude–height cross sections of u(120°E) and Δu(120°E) superimposed with the corresponding Tmm anomalies are displayed in Fig. 6 for the following reasons: 1) to understand the three-dimensional structures of both the monsoon circulation and the 30–60-day monsoon mode and 2) to attain a more quantitative measurement of the 30–60-day monsoon mode’s contribution to the intraseasonal variation of monsoon indices.

1) Total zonal wind. The comparison between Figs. 6a and 6b shows that the intraseasonal changes of the u(120°E) cross section in the extratropics and Tropics differ from each other. For the former region, the lower-tropospheric monsoon flow undergoes a distinct and quasiperiodic oscillation between the two extreme monsoon conditions, which are the MBF monsoon westerlies centered at 40°E (25°N) in the active (break) monsoon. This intraseasonal oscillation went undetected by previous studies (Yeh et al. 1959; Lau and Yang 1996), which stressed the rapid northward shifts of both the upper-level westerlies
and the location of the MBF during the Mei-Yu onset in eastern China. Concerning the Tropics, a sharp contrast between the lower SCS monsoon westerlies and the upper tropical easterly jet during the active monsoon is present. Although during the break monsoon, easterly flow dominates the entire troposphere. This flow regime change is consistent with that of the Indian monsoon (Chen and Chen 1995; Chen and Weng 1997). Because grates northward from near the equator to about 20°N, and the location of the MBF during the Mei-Yu onset in eastern China. Concerning the Tropics, a sharp contrast between the lower SCS monsoon westerlies and the upper tropical easterly jet during the active monsoon is present. Although during the break monsoon, easterly flow dominates the entire troposphere. This flow regime change is consistent with that of the Indian monsoon (Chen and Chen 1995; Chen and Weng 1997). Because of its link with the monsoon life cycle, the 30–60-day monsoon trough–ridge exhibits a quasiperiodical northward migration. Although the occurrence of the 12–24-day monsoon mode may not be quasiperiodic, this intraseasonal mode plays a substantial role in the onset and break of the SCS summer monsoon. Conceivably, the intraseasonal variations of monsoon indices (Figs. 1 and 2) may be partially caused by the 12–24-day monsoon mode. Despite the similar structure of the $\sigma_{\Delta T_{\text{sh}}, \text{sh}}$ and $\sigma_{\Delta T_{\text{sh}}, \text{sh}}$ patterns (Fig. 4), it may not be legitimate to claim that the 12–24-day monsoon mode plays little or no part compared to the 30–60-day monsoon mode. This concern is clarified by composite charts of $V(850 \text{ mb})$ streamlines and $\Delta T_{\text{sh}}$ anomalies following the same procedure as in the 30–60-day monsoon mode shown in section 3b(1).

(i) The 850-mb streamline charts

The monsoon trough in the northern part of the SCS deepens during the maximum SCS $u(850 \text{ mb})$ index. At this time, the western part of the ridge line of the North Pacific anticyclone extends into northern China, and a ridge line appears near the equator (Fig. 7a). In contrast, during the minimum $u(850 \text{ mb})$ index phase, the ridge line of the North Pacific anticyclone intrudes southwestward into the northern SCS region, and then the monsoon trough becomes NW–SE oriented across the SCS (Fig. 7b). The synoptic change of the monsoon flow between the two extreme $u(850 \text{ mb})$ indices is similar to that between the two extreme monsoon conditions shown in Figs. 5a and 5b. The convection zones associated with the monsoon trough and the MBF undergo a similar meridional oscillation as these two monsoon elements do between the active and break SCS summer monsoon. The possible structure of disturbances causing the variation of $u(850 \text{ mb})$ and $\Delta T_{\text{sh}}$ indices may be inferred from the $\sigma_{\Delta T_{\text{sh}}, \text{sh}}$ patterns (Figs. 4c,f), but these disturbances can be depicted well by the $\Delta T_{\text{sh}}$ chart between the two extreme $u(850 \text{ mb})$ indices in Fig. 7c. A well-defined double-celled structure shows up in this difference streamline chart, and a meridional juxtaposition of positive and negative $\Delta T_{\text{sh}}$ zones is also discernible. The direction alternation in the northern cell of the difference streamline chart and in the associated convection zones results in the opposite-phase variations of the zonal wind and $T_{\text{sh}}$ indices between the two monsoons with the timescale of 12–24 days.

For the 12–24-day monsoon mode (Figs. 7d–f), a clear double-celled structure, like the $\sigma_{\Delta T_{\text{sh}}, \text{sh}}$ pattern, appears in the $\Delta V(850 \text{ mb})$ streamline during both the maximum and minimum $u(850 \text{ mb})$ indices, with the exception of a reversal of anomalous circulation direction. Just as in $\Delta T_{\text{sh}}$ anomalies shown in Figs. 5d and 5f, the $\Delta T_{\text{sh}}$ anomalies exhibit a north–south juxtaposition. The appearance of the cyclonic (anticyclonic) $\Delta V(850 \text{ mb})$ northern cell facilitates the northeastward (southwestward) retreat (inversion) of the ridge line out of (into) the northern SCS

2) THE 12–24-DAY MODE

The characteristics of the 12–24-day monsoon mode differ in several ways from those of the 30–60-day monsoon mode. The 30–60-day monsoon trough–ridge migrates northward from near the equator to about 20°N, while the 12–24-day monsoon low–high propagates westward along two tracks (the equator and 15°–20°N) (Chen and Chen 1995; Chen and Weng 1997). Because of its link with the monsoon life cycle, the 30–60-day monsoon trough–ridge exhibits a quasiperiodical northward migration. Although the occurrence of the 12–24-day monsoon mode may not be quasiperiodic, this intraseasonal mode plays a substantial role in the onset and break of the SCS summer monsoon. Conceivably, the intraseasonal variations of monsoon indices (Figs. 1 and 2) may be partially caused by the 12–24-day monsoon mode. Despite the similar structure of the $\sigma_{\Delta T_{\text{sh}}, \text{sh}}$ and $\sigma_{\Delta T_{\text{sh}}, \text{sh}}$ patterns (Fig. 4), it may not be legitimate to claim that the 12–24-day monsoon mode plays little or no part compared to the 30–60-day monsoon mode. This concern is clarified by composite charts of $V(850 \text{ mb})$ streamlines and $\Delta T_{\text{sh}}$ anomalies following the same procedure as in the 30–60-day monsoon mode shown in section 3b(1).

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For the 12–24-day monsoon mode (Figs. 7d–f), a clear double-celled structure, like the $\sigma_{\Delta T_{\text{sh}}, \text{sh}}$ pattern, appears in the $\Delta V(850 \text{ mb})$ streamline during both the maximum and minimum $u(850 \text{ mb})$ indices, with the exception of a reversal of anomalous circulation direction. Just as in $\Delta T_{\text{sh}}$ anomalies shown in Figs. 5d and 5f, the $\Delta T_{\text{sh}}$ anomalies exhibit a north–south juxtaposition. The appearance of the cyclonic (anticyclonic) $\Delta V(850 \text{ mb})$ northern cell facilitates the northeastward (southwestward) retreat (inversion) of the ridge line out of (into) the northern SCS...
and the deepening (filling) of the monsoon trough. The effect of the 12–24-day monsoon mode on the change of monsoon flow is further substantiated by the structure resemblance between $\Delta \left[ V(850 \text{ mb}), T_{BB} \right]$ (Fig. 7c) and $\Delta \left[ \hat{V}(850 \text{ mb}), \hat{T}_{BB} \right]$ (Fig. 7f). Conceivably, the 12–24-day intraseasonal variations in the monsoon westerlies and cumulus convection of the two monsoon components can be induced by the westward-propagating 12–24-day monsoon mode.

(ii) Latitude–height cross section

During the maximum $\hat{u}(850 \text{ mb})$ index, the $u(120^\circ \text{E})$ structure (Fig. 8a) is characterized by the following sa-
lient features: the tropical monsoon westerlies, the tropical easterly jet, and the midlatitude westerlies coupled with the MBF centered at about 40°N. This structure is similar to that of the active monsoon phase (Fig. 6a). In contrast, during the minimum $\bar{u}(850$ mb) index phase (Fig. 8b) the monsoon westerlies south of 20°N disappear, and the midlatitude westerlies extend down to couple with the MBF, just as the $\bar{u}(120°E)$ cross section behaves during the break monsoon phase. The cumulus convection is active in the maximum $\bar{u}(850$ mb) index phase at about 20° and 35°N but is somewhat opposite and less organized during the minimum $\bar{u}(850$ mb) index phase.

Shown by our previous study (Chen and Chen 1995), the vertical extent of the 12–24-day monsoon mode can reach to a level near 300–400 mb. A well-organized $\bar{u}(120°E)$ vortex with this vertical extent (Fig. 8c) is centered at about 20°N. The amplitudes of $\bar{u}(120°E)$ and $\Delta T_{\text{BB}}(120°E)$ are about two-thirds those of $\Delta u(120°E)$ and $\Delta T_{\text{BB}}(120°E)$. The resemblances between cross sections of both $\Delta u(120°E)$ and $\Delta \bar{u}(120°E)$ and $\Delta T_{\text{BB}}(120°E)$ and $\Delta T_{\text{BB}}(120°E)$ reveal that the intraseasonal changes of the monsoon zonal flow and cumulus convection are caused by the westward propagation of the 12–24-day monsoon mode.

Recall that $\Delta[\bar{u}, \Delta T_{\text{BB}}](120°E)$ in Fig. 6f may reach amplitudes of about one-half that of $\Delta[u, \Delta T_{\text{BB}}](120°E)$ (Fig. 6c). In spite of the differences in the propagation property and the occurrence frequency between the 30–60-day and 12–24-day monsoon modes, there is a phase lock between these two intraseasonal monsoon modes during the SCS monsoon onset and break. Our analysis of 1979–93 in this study (not shown) reveals that this phase lock occurs in more than 75% of the SCS monsoon onsets and breaks. Compared to the amplitudes of $u$ and $T_{\text{BB}}$ anomalies, the 30–60-day and 12–24-day monsoon modes contribute the most to the intraseasonal variations in the $\Delta u$ and $\Delta T_{\text{BB}}$ indices of the two monsoons during the active and break monsoon phases.
4. Discussion

The East Asian summer monsoon is a complex circulation system that contains a number of interesting phenomena during its evolution (e.g., Lau and Li 1984; Tao and Chen 1987; Murakami 1987). The following two phenomena are often discussed in literature.

1) During the Mei-Yu onset, the large-scale summer circulation over East Asia undergoes some abrupt changes (e.g., Yeh et al. 1959; Tao and Ding 1981). These changes consist of several stepwise northward shifts of the upper-level westerlies over East Asia and the northward shift of the lower-level MBF from a position south of the Yangtze River to north of the river.

2) As shown in Fig. 9, three east–west-oriented location zones of extremely heavy rain events emerge over the eastern half of China (Tao and Ding 1981). Regardless of the mechanism generating the heavy rain, Tao and Ding suggested that these three longitudinal zones are related, during the Mei-yu onset, to the stepwise northward shifts of the upper-level westerlies over East Asia and to the northward shift of MBF across the Yangtze River.

Concerning the first phenomenon, the northward shift of the upper westerlies (covering only a limited time period of the entire monsoon season) is a part of the seasonal march of the East Asian circulation, which does not reverse its course until late summer. On the other hand, the MBF oscillates meridionally in a coherent manner with the SCS summer monsoon. Since the MBF should couple in a certain way with the upper westerlies, it would be of interest to explore how the northward shift of the MBF across the Yangtze River (in concert with the tropical northward shifts of upper westerlies) during the Mei-Yu onset occurs within the context of the MBF’s north–south intraseasonal oscillation. For this purpose, let us use the coherent intraseasonal oscillations of the filtered MBF location (thick dashed line) and the SCS $\bar{u}(850 \text{ mb})$ (thick solid line) indices in Fig. 3 to illustrate the role played by the intraseasonal monsoon modes in the northward shifts of the MBF location across the Yangtze River basin. The climatological onset date of the Mei-yu season in the Yangtze River basin is approximately 10–30 June (Tao and Chen 1987). As indicated by the SCS $\bar{u}(850 \text{ mb})$ index (Fig. 3a), the SCS monsoon westerlies re-intensify in the second life cycle of this monsoon after 10 June. According to observations at this stage by Chen and Chen (1995) a 30–60-day monsoon ridge reaches the northern SCS at about $20^\circ$N, and a newly originated 30–60-day monsoon trough emerges near the equator. The location index shows that the MBF is located south of the Yangtze River. About a week to 10 days later, the MBF shifted across the Yangtze River. Accompanying the MBF’s northward shift, the strong zonal wind associated with this front (particularly below 500 mb) also moves northward [indicated by the $u(120^\circ \text{E})$ cross section in Fig. 6b] over the Yangtze River. Let us next compare the $u(120^\circ \text{E})$ and $\bar{u}(120^\circ \text{E})$ cross sections in Fig. 6. During the active (break) monsoon phase, it is inferred from the coincidence of negative (positive) $\bar{u}$ anomalies around $20^\circ$–$35^\circ$N with the weak (strong) MBF westerlies that the anomalous circulation associated with the 30–60-day monsoon trough (ridge) plays a significant role in the northward shift of the MBF across the Yangtze River basin. This argument is further supported by the resemblance of the $\Delta u(120^\circ \text{E})$ (Fig. 6c) and $\Delta \bar{u}(120^\circ \text{E})$ (Fig. 6f) cross sections. For the 1989 summer (Fig. 3b), the zonal-wind monsoon index indicated that the intensity of the 12–24-day monsoon mode (revealed from the zonal-wind monsoon index) was stronger than that of the 30–60-day monsoon mode. As shown by the location index, the MBF moved rapidly across $30^\circ$N in early June primarily due to the 12–24-day monsoon mode [also inferred from the comparison between the $u(120^\circ \text{E})$ and $\bar{u}(120^\circ \text{E})$ cross sections in Fig. 8].

Our discussion so far focuses only on the lower half of the troposphere. Nevertheless, the northward shift of upper westerlies observed by Yeh et al. (1959) for the 1956 Mei-yu onset is not easily perceived from Figs. 6 and 8. This northward shift of the upper westerlies associated with the Mei-yu onset may not occur at the exact same latitudinal location and cover the same north–south extent in every summer monsoon season. It is likely that the northward shifts of upper westerlies with the Mei-yu onset of 1979–93 are obscured by the
composite procedure. Yeh et al. (1959) stressed in their study only the rapid northward shifts of the upper westerlies during the Mei-ju onset. In contrast, we have shown in this paper that the two intraseasonal monsoon modes are vital to the coupling, that is, the opposite-phase oscillation between the lower-tropospheric circulations of the SCS and East Asian summer monsoons, which was neglected previously.

Recently, Lau and Yang (1996) suggested that during the East Asian monsoon onset, the rapid northward shift of the upper-level westerlies over East Asia is linked to the abrupt northward migration of the ascending branch of the local Hadley circulation over the SCS region. This migration may be induced by the symmetric instability of the basic flow in May. Actually, the northward migration of the ascending branch of the local Hadley circulation can be coupled with that of the two intraseasonal monsoon modes (Chen and Chen 1995). Whether the northward migration of the 30–60-day monsoon trough–ridge and the westward propagation of the 12–24-day monsoon mode are related to the symmetric instability of the monsoon flow is beyond the scope of this study. Nevertheless, the discussion above supports the argument that the two intraseasonal monsoon modes are indispensable components of the East Asian monsoon to the northward shifts of the MBF location.

For the second phenomenon, the rainstorms in eastern China during the summer monsoon season are steered by the upper westerlies to move along the MBF. Perhaps because of this, Tao and Ding (1981) suggested a possible link between the three east–west-oriented location zones (around 20°, 30°, and 40°N) and the stepwise northward shift of the upper westerlies over eastern China; however, it was not made clear in their study how these three zones of heavy rain events are linked to the northward shifts of the MBF and the upper westerlies during the Mei-ju onset. Actually, these three location zones presented by Tao and Ding’s heavy rain events match well with Lau et al.’s (1988) quasi-stationary locations of three rainbands formed by multiple onsets of the East Asian summer monsoon. Lau et al.’s analysis strongly indicates that the three location zones of heavy rain events in eastern China may not be linked only to the Mei-ju onset. Therefore, some questions are raised from the difference between the observations of these two studies. Why should there be three location zones for heavy rain events? What is the dynamical mechanism responsible for the formation of these three zones? Answers to these two questions may be derived from results presented in sections 3 and 4.

Recall that during the active SCS monsoon phase the center and northern periphery of the intraseasonal monsoon cyclonic vortex are located at about 20° and 35°N (Fig. 6d), respectively. As inferred from \( \Delta T_{\text{MBF}}(120°\text{E}) \) (Fig. 6a) or \( \Delta T_{\text{MBF}}(120°\text{E}) \) (Fig. 6d), deep cumulus convection associated with the 30–60-day monsoon trough and along the MBF exists around the two aforementioned latitudinal locations. In contrast, the intraseasonal monsoon vortex becomes anticyclonic during the SCS break monsoon phase (Fig. 6e). Although the anticyclonic vortex center is still located at about 20°N, the maximum 850-mb zonal wind of the MBF is shifted southward to about 25°N (Fig. 6b), and the southern periphery of this vortex reaches the equator (Fig. 6e). At this stage, the deep cumulus convection positioned along the MBF and associated with the 30–60-day monsoon trough appears near the maximum 850-mb zonal wind and the southern rim of the vortex, respectively.

As far as the 12–24-day monsoon mode is concerned, the convection zones associated with the vortex of this monsoon mode (shown in Fig. 8) for the two extreme \( \bar{u}(850\text{mb}) \) indices alternate in the same manner as the 30–60-day monsoon mode.

The locations of maximum \( \Delta T_{\text{MBF}}(120°\text{E}) \) and \( \Delta T_{\text{MBF}}(120°\text{E}) \) [or \( \Delta T_{\text{MBF}}(120°\text{E}) \)] during the SCS active monsoon phase [or the maximum SCS \( \bar{u}(850\text{mb}) \) phase] coincide with the latitudinal locations of Tao and Ding’s (1981) heavy rain zones around 20°N and near 40°N. Contrarily, the activation of the MBF during the break SCS monsoon phase [or the minimum SCS \( \bar{u}(850\text{mb}) \) phase] enables heavy rain events to occur around the Yangtze River (approximately 30°N). We cannot deny that heavy rain events in eastern China during the Mei-ju onset are related to the northward shifts of the MBF and upper westerlies. Since these shifts occur over only a limited time period during the East Asian monsoon season, the three zones of heavy rain events over eastern China are unlikely to be the results of only these stepwise northward shifts. By engaging the following comparisons:

1. the zonal wind and \( \Delta T_{\text{MBF}} \) cross sections at 120°E in Figs. 6 and 8,
2. the coherent intraseasonal oscillation of the SCS \( \bar{u}(850\text{mb}) \) and the MBF location indices in Fig. 3, and
3. the coherent intraseasonal oscillations of \( T_{\text{MBF}}(140°\text{E}) \) between the ITCZ and the MBF observed by Chen and Murakami (1988, their Fig. 1),

we are able to connect the zoning of the eastern China heavy rain locations to the maximum convection zone associated with the MBF and the monsoon trough during extreme SCS monsoon conditions indicated by the \( \bar{u}(850\text{mb}) \) index. In other words, the three heavy rain zones shown by Tao and Ding are a result of the intraseasonal north–south oscillations of the MBF coupled with the northward migration of the intraseasonal monsoon trough–ridge in the SCS region. This argument sheds some light on the question posed above.

5. Concluding remarks

Although the establishment of the summer monsoon life cycle by the 30–60-day and 12–24-day monsoon modes in the SCS–WTP region has been documented in previous studies, the possible links of this monsoon
component to the monsoon over East Asia and to any likely effect of the former monsoon component on the latter monsoon component were not systematically explored in the past. However, findings from several previous studies lead us to hypothesize a link between the monsoons in these two regions.

1) The isoline of monsoon onset dates in East Asia, compiled by Tao and Chen (1987, their Fig. 3.9), shows a clear northward advancement of the East Asian monsoon from southern to eastern China following the northward seasonal march of the sun.

2) The $T_{bb}(140^\circ E)$ $y$–$t$ diagram of Chen and Murakami (1988) revealed that a coherent intraseasonal north–south oscillation exists between the MBF and the ITCZ in the western North Pacific.

3) The Indian monsoon onset is triggered by the arrival of the northward-migrating 30–60-day monsoon trough at 15°–20°N (Krishnamurti and Subrahmanyan 1982; Chen et al. 1988b). This migrating intraseasonal monsoon trough extends eastward to the western tropical Pacific, causing the second life cycle of the SCS summer monsoon to synchronize with the first life cycle of the Indian monsoon during the active monsoon phase (Chen and Chen 1995). Since the Mei-yu onset follows the rapid northward shift of the upper westerlies across the Tibetan Plateau, there should be some connection between the monsoons in Southeast and East Asia.

4) A number of previous studies cited in the introduction have observed the existence of 12–24-day and 30–60-day intraseasonal oscillations in the East Asian and SCS summer monsoons. The former intraseasonal oscillation was found to be more regional, while the latter intraseasonal oscillation may be coupled with the eastward-propagating Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972). In light of the coupling between the regional 30–60-day monsoon mode and the global-scale MJO, a coherent intraseasonal oscillation may exist between the summer monsoons in East Asia and the SCS.

The NCEP–NCAR reanalysis data, OLR, and GMS $T_{bb}$ of Japan for the period 1979–93 were analyzed in this study to seek the possible link between the summer monsoons in East Asia and the SCS. The major findings of this study are as follows:

1) an opposite-phase intraseasonal oscillation of a) the 850-mb monsoon zonal wind index and b) the cumulus convection proxy (i.e., $\Delta T_{bb}$ and $\Delta$OLR) of the two monsoons components, and

2) in-phase intraseasonal oscillations of a) the MBF and the ITCZ and b) the MBF latitudinal location and the intensity of the SCS monsoon westerlies.

Regardless of the separation of the summer monsoon components in East Asia and the SCS by the North Pacific anticyclone, these coherent intraseasonal oscillations between the two monsoon components are caused essentially by the intraseasonal flip-flop oscillation between the anomalous cyclonic and anticyclonic vortices associated with (a) the northward-migrating 30–60-day monsoon trough and ridge, respectively, and (b) the westward propagating 12–24-day monsoon low and high, respectively. Evidently, intraseasonal oscillation of summer monsoons in East Asia and the SCS is coupled through these monsoon vortices. Observations made in the South China Sea Monsoon Experiment during the summer of 1998 (Lau et al. 1998) provided high quality data over the Southeast–East Asian region and offer a unique opportunity for us to test further the suggested link between the intraseasonal oscillations of the summer monsoons in East and Southeast Asia.

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APPENDIX

Propagation Properties of the Two Intraseasonal Monsoon Modes and Averaged Power Spectra of Monsoon Indices

a. Propagation properties of the two intraseasonal monsoon modes

The northward propagation of the 30–60-day monsoon mode and the westward propagation of the 12–24-day monsoon mode were depicted in terms of lagged correlation coefficient patterns (between the filtered 850-mb zonal-wind monsoon index and streamfunction) by Chen and Chen (1995, their Figs. 9 and 15) for the summer of 1979 and by Chen and Weng (1997) for 15 summers during the period 1979–93. The lagged correlation coefficient pattern certainly provides a statistical summary for propagation properties and for the structure of intraseasonal monsoon modes. However, the lagged correlation coefficient patterns are only on expansion of the correlation coefficient patterns (shown in Fig. 4) in time. It may be informative to illustrate the propagation properties of the intraseasonal monsoon modes by some direct approach.

Using the 30–60-day filtered 850-mb wind fields, Chen and Chen (1995) portrayed the 30–60-day monsoon trough/ridge over the SCS–WTP region. The same approach was adopted to determine latitudinal locations of these monsoon troughs/ridges in this study. A $y$–$t$ diagram of latitudinal locations of the 30–60-day monsoon troughs (solid stars) and ridges (open circles) and
$\Delta T_{BB}$ anomalies averaged between 115° and 120°E are shown for all SCS monsoon life cycles selected over 15 summers (1979–93) by Chen and Weng (1997) in Fig. A1. Locations of the 30–60-day monsoon troughs and ridges are not the same every summer season. However, as revealed from Fig. A1, the quasiperiodic alteration between the 30–60-day monsoon trough and ridge is obvious. The copresence of the 30–60-day monsoon trough and ridge is also true when one of them arrives at 15°–20°N.

The occurrence of the 12–24-day monsoon mode in the SCS–WTP region is identified by the streamline charts of the 12–24-day filtered 850-mb wind when a closed low coincides with a $\Delta T_{BB}$ center. The 12–24-day monsoon mode possesses a double-cell structure with the same polarity. A clear view of this monsoon mode’s propagation track can be gained from the occurrence frequency chart in Fig. A2. Low centers of the 12–24-day monsoon mode identified every day are marked with an open circle (solid triangle) for the northern (southern) track. The two-track propagation of the 12–24-day monsoon mode is apparent, but the SCS monsoon is primarily affected by the northern track. This intraseasonal mode propagates westward and rarely appears east of 150°E.

b. Averaged power spectra of monsoon indices

The $u(850\text{ mb})$, $T_{BB}$, and OLR monsoon indices associated with the SCS monsoon and the MBF were subjected to spectral analysis using a scheme proposed by Madden and Julian (1971, 1972). The power spectra (of these monsoon indices) averaged over 15 summers (1979–93) are displayed in Fig. A3. A predominant signal of the 30–60-day mode emerges with a confidence level close to or higher than 99% (smooth solid curve). In addition to this intraseasonal mode, there is still a noticeable signal within the period between 12 and 24 days. To make this signal more discernible in the power spectra of monsoon indices, we applied simple harmonic analysis [as Chen and Chen (1995) suggested] to exclude signals with periods longer than 30 days from the
monsoon indices. The 12–24-day signal with a confidence level of 99% (smooth dashed curve) stands out in the power spectra (dashed lines) of the residual time series.

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