ADVANCES IN ATMOSPHERIC SCIENCES, VOL. 22, NO. 5, 2005, 665–676

South China Sea Summer Monsoon Onset in Relation to the Off-Equatorial ITCZ

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(Received 9 December 2004; revised 18 March 2005)

ABSTRACT

Observations of the South China Sea summer monsoon (SCSSM) demonstrate the different features between the early and late onsets of the monsoon. The determining factor related to the onset and the resultant monsoon rainfall might be the off-equatorial ITCZ besides the land-sea thermal contrast. The northward-propagating cumulus convection over the northern Indian Ocean could enhance the monsoon trough so that the effect of the horizontal advection of moisture and heat is substantially increased, thus westerlies can eventually penetrate and prevail over the South China Sea (SCS) region.

Key words: SCSSM onset, off-equatorial ITCZ, land-sea thermal contrast

1. Introduction

The onset of the East Asian Summer Monsoon (EASM) is often signified by an abrupt increase of precipitation (Tao and Chen, 1987) or convection (Hirasawa et al., 1995) around the middle of May. Tao and Chen (1987) suggested that such a switch first occurs over the SCS, which was accepted later by many meteorologists. During the South China Sea monsoon experiment (SCSMEX), the onset of the South China Sea summer monsoon (SCSSM) was investigated in many studies (Ding and Li, 1999; Li and Wu, 2000; Chen et al., 2000; Xu et al., 2001), and the onset criterion and the atmospheric circulation evolution associated with the onset are basically now understood. However, some studies have also shown that the summer monsoon over Asia does not begin over the South China Sea (SCS), but rather over the south of the Bay of Bengal (BOB) or the southwest of the Indochina Peninsula (Li and Qu, 1999; Lau et al. 2000; Liu et al., 2002; Zhang et al., 2004).

Up to now, the origin of the monsoon is still arguable as to whether it is due to the land-sea thermal contrast or the off-equatorial ITCZ. Some works have also questioned the role of the Tibetan Plateau in the onset of the SCSSM (e.g., Chen and Dell'osso, 1986). The controversial points are that the maximum heating center is not over the Tibetan Plateau, but located near Burma, so the systematic cumulus convection near India/Burma would be more important. And from the thermal wind point of view, the South Asian high that formes over the Tibetan Plateau requires the heating source to be south of the Tibetan Plateau because the heating source should be at the intersection between the westerly and the easterly (westerly to its north and easterly to its south) (Mao and Chan, 2004).

Thus, a study of the SCSSM from the circulation and land-sea thermal contrast points of view is necessary. This would enable an overall examination of these physical processes, give insight into the evolution of the monsoon period, and identify the relative contributions from each factor. The present paper presents an observational investigation of the early or late monsoon in the SCS domain. In the present work, the data used are mainly from the National Centers for Environmental Prediction (NCEP) reanalysis. These include the daily mean values of geopotential height,

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wind speed, air temperature, pressure, and specific humidity during 1955–2001. Daily outgoing longwave radiation (OLR) data from the National Oceanic and Atmospheric Administration for the 25 years of 1979–2001 are also used. The horizontal resolution of these data is $2.5° \times 2.5°$. The South China Sea summer monsoon onset (SCSSM) and its strength are defined in section 2. And in section 3, large-scale characteristics of the early and late SCSSM onset years are examined. A possible mechanism of SCSSM onset in relation to the land-sea thermal contrast is then argued in section 4. A summary and discussion are presented in section 5.

2. Consistency of monsoon onset and its strength

A consensus has yet to be reached concerning the onset of the EASM (5°–45°N, 105°–140°E) (e.g., Xie et al., 1998; Ding and Li, 1999). A general conclusion is that the SCSSM onset is associated with a switch of the zonal winds over the SCS from easterly to westerly, accompanied by an increase in convective activity.

Some meteorological parameters such as precipitation, prevailing wind, outgoing longwave radiation (OLR) and cloud-top temperature have been used to define the monsoon onset (Tao and Chen, 1987; Xie et al., 1998; Chan et al., 2000; Ding and Liu, 2001). The OLR data are commonly used as a proxy for deep convection since they have a strong relationship with tropical rainfall (Wu and Zhang, 1998), which is a direct reflection of the summer monsoon. Quite a few studies have used OLR values to define the onset and development of the summer monsoon (e.g., Wu and Zhang, 1998; Xu and Chan, 2001). Arkin and Ardanuy (1989) suggested that tropical rainfall may be estimated with OLR < 235 W m^{-2}. On the other hand, some authors (e.g., Wang and Wu, 1997) have preferred to use both the 850-hPa zonal wind and OLR to identify the SCSSM onset.

To discuss the interannual variability of the Asian summer monsoon, a measure of the monsoon intensity must be defined. Generally, three independent data sources can be used to define the intensity of the monsoon: the intensity of the monsoonal circulation, satellite-measured OLR field, and precipitation amounts over a representative area. Different monsoon indices have been proposed (e.g., Webster and Yang, 1992; Lu and Chan, 1999) to depict the evolution of the summer monsoon in a certain year. Webster and Yang (1992) proposed another strength index—the zonal wind shear defined by the zonal wind difference between 850 and 200 hPa over South Asia. Recently, Lu and Chan (1999) suggested a unified monsoon index for both summer and winter monsoons over South China. In both winter and summer, the meridional winds at upper and lower levels over the SCS are always in opposite directions, which makes it possible to use the meridional wind of this area as a unified index for both monsoons. It has been observed that the strength of the meridional wind anomaly at 1000 hPa over the SCS relates well to the summer rainfall (Lu and Chan, 1999). Li and Zeng (2003) described another unified dynamical index emphasizing the seasonal cycle. Huang (2004) suggested an index measuring the interannual variation of the EASM. So far, the Asian monsoon index is defined from the thermodynamic (e.g., rainfall, OLR) and dynamic (e.g., zonal wind, meridional wind) perspectives. The index based on rainfall elements might be easily influenced by local thermodynamic conditions, and the other index defined by zonal wind might work only over the South Asian domain (Huang, 2004).

In our study, the summer monsoon onset date (MOD) over the SCS domain is defined by using OLR and 850-hPa zonal wind together. Because the OLR data are only available from 1975, the onset dates of the years before 1975 cannot be defined using the OLR. In order to extend the number of years studied, it is necessary to find another index with a longer record, the value of which can give an onset pentad consistent with that given by the OLR data. A good candidate is the 850 hPa zonal winds over the central and southern part of the SCS (5°–15°N, 110°–120°E), which reflects the migration of the subtropical high in and out of the SCS. The onset pentad is therefore defined as the first pentad after mid-April when the 5-day mean value of the 850-hPa zonal winds is westerly over the central and southern part of the SCS and lasts for more than two pentads.

The mean onset pentad of 44 years is the 28th pentad (P28), and the 44 years (1958–2001) are stratified into three groups by using one pentad before and one after the 28th pentad (P28) as the threshold: 13 early, 21 normal and 10 late (Table 1), with the average onset pentad in each category being in the 25th pentad, 28th pentad and 31st pentad respectively. Thus the normal onset range is within P27–P29, and the early and late onset years are outside this range (early <P27, late >P29). The 13 early onset years are: 1960, 1966, 1971, 1972, 1974, 1976, 1984, 1985, 1994, 1996, 1999, 2000, and 2001, while the 10 late onset years are: 1959, 1963, 1968, 1970, 1973, 1975, 1982, 1987, 1991 and 1993. In the following sections, the large-scale features associated with the early and late onsets are analyzed.

Table 1. Classification of MOD. Early, Normal, and Late represent the onset occurrences of prior to the 27th pentad, within the 27th–29th pentads, and later than the 29th pentad, respectively.

Classification of MOD	Year
Early (<P27)	1960, 1966, 1971, 1972, 1974, 1976, 1984, 1985, 1994, 1996, 1990, 2000, 2001
Normal (P27–P29)	1958, 1961, 1962, 1964, 1965, 1967, 1969, 1977, 1978, 1979, 1980, 1981, 1983, 1986, 1988, 1989, 1990, 1992, 1995, 1997, 1998
Late (>P29)	1959, 1963, 1968, 1970, 1973, 1975, 1982, 1987, 1991, 1993

As for the SCSSM strength index (MSI), the May–June monthly zonal wind difference between 850 hPa and 200 hPa over the SCS domain is used, because the meridional component of the SCSSM period is not as obvious as that of winter. Figure 1 also shows the normalized MSI time series, Using one standard deviation as the threshold, 7 strong (1960, 1961, 1965, 1975, 1976, 1986, 2001) and 10 weak (1959, 1968, 1970, 1980, 1983, 1987, 1991, 1992, 1993, 1998) years can be identified.

We found a large interannual variation in the MOD, from late April to early June, and in the MSI. Even though the SCSSM onset is an abrupt transition, its strength variations seem to follow the early or late onsets simultaneously over a large latitudinal extent with large interannual variability. Note that early onset generally corresponds to strong SCSSM, such as 1960, 1976, 2001, and late onset with weak SCSSM (1959, 1968, 1970, 1987, 1991, 1993). An early onset matching a strong monsoon or a late onset corresponding to a weak monsoon is also reflected by a negative correlation (−0.53) between MOD and MSI, which is significant above the 99% level. The similar variations in both the SCSSM intensity and MOD can also be observed (Fig. 1). For early onset years, the MSI appears to be weak in both year −2, and year −1, but strong at year 0 and year 1. The reverse occurs for the late onset years, where the MSI seems to be strong or normal-to-strong in both year −2 and year −1, but less strong or weak at year 0 and year 1. This result suggests that if the SCSSM reaches onset early (late), its strength is likely to be strong (weak), and the adjustment cycle of SCSSM from strong to weak is generally related to the MOD cycle from early onset to late onset, and vice versa. In other words, an extremely early or late onset of the SCSSM probably brings abundant or sparse rainfall, so abnormal (above or below normal) rainfall usually does not result from a normal onset.

Though a one-to-one correspondence cannot be found between the strong (weak) monsoon and a certain early (late) onset year, some certain years follow this rule, and this agrees well with the result of Shi and Zhu (1997). Thus such a consistent tendency of both MOD and MSI is very obvious, such that the strength of SCSSM tends to be abnormal when the monsoon reaches onset extremely early or late. On the other hand, the strength of the SCSSM is neither strong nor weak if there is a normal monsoon onset.

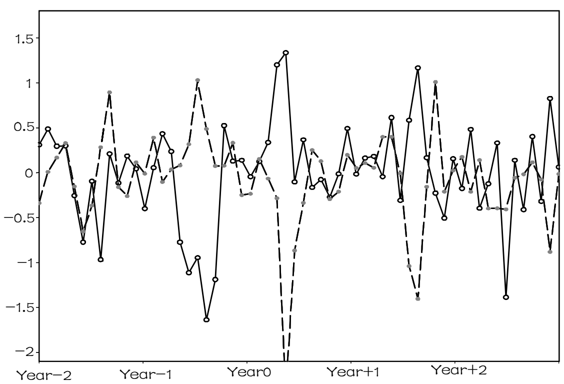

Fig. 1. The variation of 850-hPa zonal wind anomalies (unit: m s^{-1}) over the southern part of the SCS (5–15°N, 110–120°E) for early onset years (solid line) and late onset years (dashed line). Year 0 refers to the monsoon onset year, Year −1 (+1) to one year before (after), etc.

3. Large-scale characteristics of the early and late SCSSM onset years

3.1 *Zonal winds*

Generally, the SCSSM onset is associated with an abrupt increase in cloudiness and rainfall (Wang and Wu, 1997). This occurs because once the Western Pacific Subtropical High (WPSH) moves to the east of the Philippines, the SCS region is susceptible to convection (e.g., Ding and Liu, 2001; Chan et al., 2002). However, no consensus has yet been reached on the mechanism responsible for the release of the stored convective available potential energy.

It is worthwhile to note that over the SCS, the wind shift (from easterly to westerly) in the lower and mid-troposphere levels is due to the eastward retreat of the

* Year 0 is defined as the year in which the MOD of a particular type is considered. Year ($-n$) refers to n year(s) before, and year (n) n year(s) after.

WPSH and development of the monsoon trough, suggesting that the establishment of the monsoon trough over the SCS be regarded as a sign of the SCSSM onset (Pan, 2004), while that (from westerly to easterly) in the upper levels is linked to the establishment of the South Asian high. The burst in the westerlies in the lower and midtroposphere levels and easterlies in the upper levels can be seen further from time cross-sections of the zonal winds (Fig. 2). At 850 hPa, the switch to westerlies occurs at the end of the 24th(32nd) pentad in the early (late) onset case (Fig. 2). Note that in the late case, easterlies prevail throughout the domain during the onset of the early cases (Fig. 2). In other words, the low-level westerlies in the tropical region (5°–15°N) begin over the Bay of Bengal, then extend to the SCS, and reach a maximum around the SCSSM onset.

When the 850-hPa westerlies appear over the SCS in the early cases (24th–25th pentads) of Fig. 2, the shift to easterlies also occurs in the 200-hPa easterlies (Fig. 2c). However, around this time, the westerlies continue to dominate the SCS at 200 hPa in the late cases until the 32nd–33rd pentads (Fig. 2d). For both the early and late cases, the transition at 200 hPa from westerlies to easterlies and at 850 hPa from easterlies to westerlies occur nearly in the same pentad (24th–25th pentads in the early case, 32nd–33rd pentads in the late cases), which is consistent with the result of Xu and Chan (2001) in the study of the 1998 Asian summer monsoon.

This implies that the establishment of the typical monsoon circulation at the upper levels coincides with that at the lower levels. To some extent, the characteristics of the SCSSM onset are that the easterlies (westerlies) control the upper (lower) levels of the SCS. The significant easterlies at 200 hPa are associated with the South Asian high, while the strong westerlies at 850 hPa emanate from the Arabian Sea and the Southern Hemisphere. The westerlies are apparently responsible for providing the moisture for the convection, as has been discussed by Chen et al. (1988) and Chan et al. (2002).

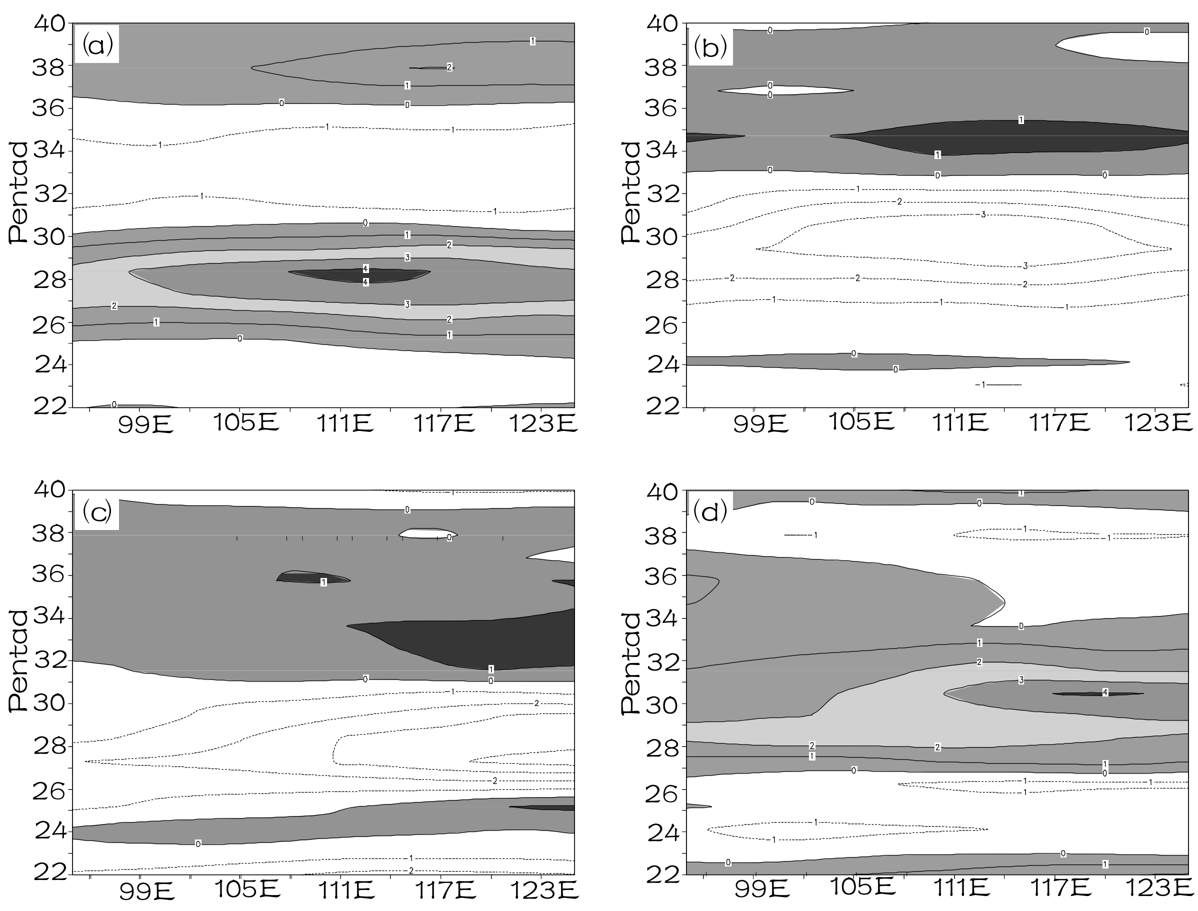

Fig. 2. Hovmüller diagrams of zonal wind anomaly (units: m s^{-1}) averaged within 5°–15°N at 850 and 200 hPa for the early (a, c) and late cases (b, d). Shading indicates westerlies.

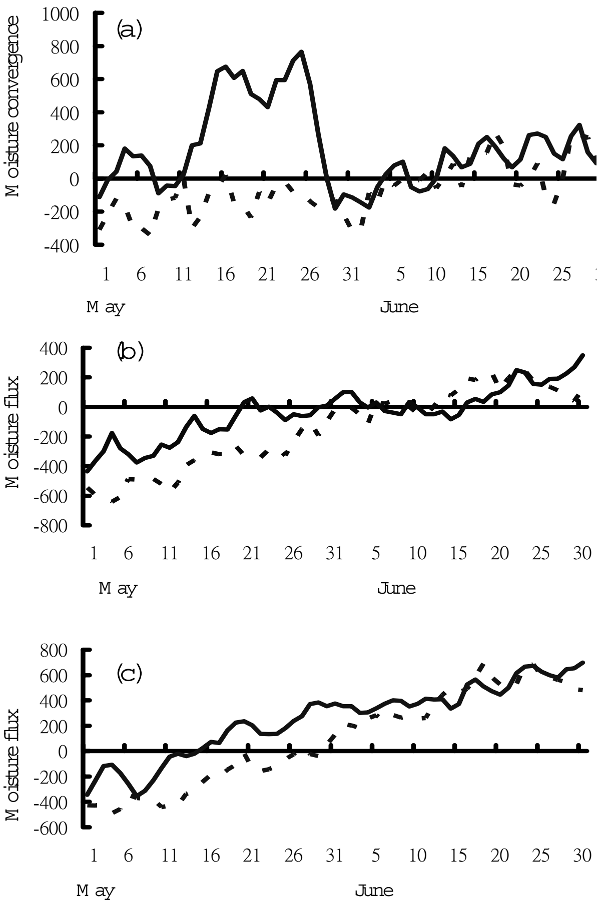

Fig. 3. (a) Moisture convergence over the SCS (5°–20°N, 110°–120°E), moisture flux at (b) east flank and (c)west flank during May–June for the early case (solid) and late case (dashed). (Units: g m kg^{-1} s^{-1}).

3.2 *Moisture transport*

Previous studies (e.g., Chen et al., 1988) have shown that water vapor is transported eastward from the Indian Ocean to the SCS and northeastward by the western branch of the subtropical high. The moisture convergence over the SCS should therefore have a direct contribution to the SCSSM onset. This aspect is evaluated by computing the moisture flux integrated from 1000 to 200 hPa within the SCS.

The first moisture convergence occurs in early May in the early onset cases (Fig. 3a). It abruptly increases to a large value in mid May, and then decreases in late May. In contrast, moisture convergence over the SCS only occurs after mid-June for the late case. The magnitude is also not as large as that in the early onset case. To determine the relative contributions to this moisture increase by the westerlies and the flow from the WPSH, the moisture fluxes along the eastern and western flanks can be examined. Along the eastern flank, moisture flows generally into the SCS from the east in May and out in mid and late June (Fig. 3b), which is consistent with the location of the WPSH. Before the SCSSM onset, the WPSH dominates over the SCS so that a southeasterly flow brings moisture into the SCS. After the onset, the WPSH retreats out of the SCS so that westerlies prevail over the SCS. Notice from Fig. 4b that the difference between the two cases is not too significant. Further, after the onset in the early case, the moisture flux is around zero, indicating that very little contribution to the moisture flux comes from the east. Along the western flank, the moisture flux changes to positive in mid May and keeps increasing for the early case (Fig. 3c). For the late case, the flux does not become positive until the end of May. This difference in moisture flux variations reflects the strength of the westerlies from the BOB. Therefore, it appears that the westerlies are more important in moistening the atmosphere over the SCS as far as the onset is concerned.

4. SCSSM onset in relation to the land-sea thermal contrast

In this section, the global distribution of Q_1 (apparent heat source) and Q_2 (apparent moisture sink) will be determined over a 22-year period (from 1980–2001) as basic variables for the description of the seasonal variability of the land-sea thermal contrast:

$$Q_1 = c_p \left[\frac{\partial T}{\partial t} + \boldsymbol{V} \cdot \nabla T + \left(\frac{p}{p_0} \right)^{R/c_p} \cdot \omega \cdot \frac{\partial \theta}{\partial p} \right] \quad (1)$$

$$Q_2 = -L \left[\frac{\partial q}{\partial t} + \boldsymbol{V} \cdot \nabla q + \omega \cdot \frac{\partial q}{\partial p} \right] , \quad (2)$$

where T is the air temperature, θ the potential temperature, R and c_p the gas constant and the specific heat at constant pressure of dry air respectively, $\boldsymbol{V}$ the zonal wind, ω vertical p-velocity, q specific humidity, L the latent heat of condensation, p the pressure, and p_0 the surface pressure. A positive (negative) Q_1 indicates the presence of apparent heating (cooling). In contrast, a positive (negative) Q_2 corresponds to regions of apparent moisture sinks (sources). In other words, positive (negative) Q_2 implies an apparent condensation (evaporation) region.

Most of the work in the literature has focused on the Tibetan Plateau's role in the establishment and maintenance of the Asian summer monsoon as an elevated heat source (e.g., Xu and Chan 2001). While the diabatic heating and warm horizontal advection play a primary role in the temperature increase over

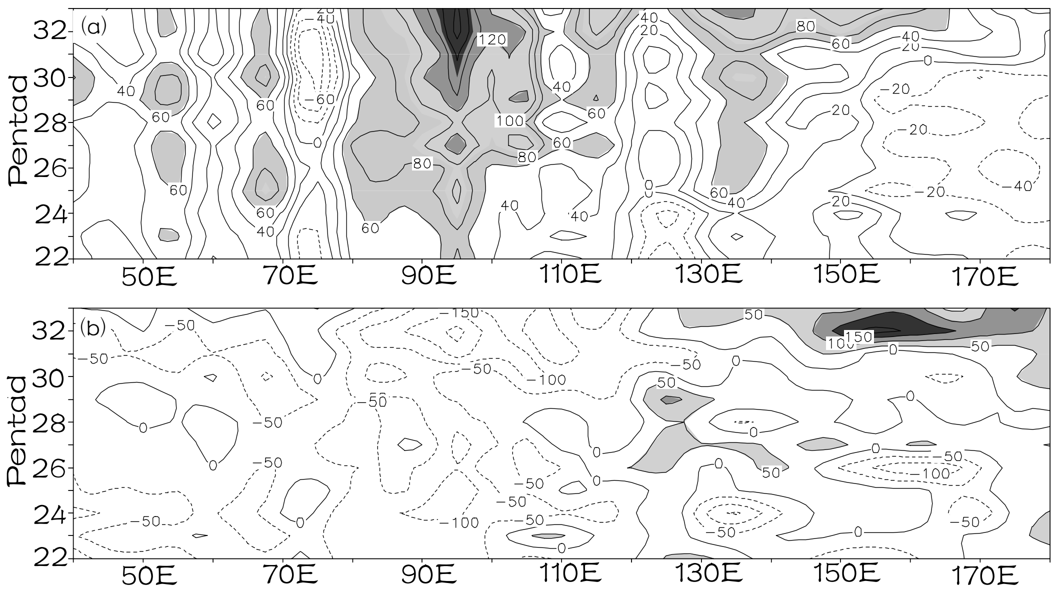

Fig. 4. (a) Composites of Q_1 (integrated from the surface to 100 hPa) (units: W m^{-2}) averaged for 22 years and (b) difference in Q_1 (integrated from the surface to 100 hPa) between early and late onset years along the Tibet region (27.5°–37.5°N) for a total of 12 pentads around the SCSSM onset period. Shading indicates values larger than (a) 60 W m^{-2}, (b) 50 W m^{-2} .

the eastern plateau (Yanai et al., 1992), it is the sensible heating over the plateau region in spring that leads to the reversal of the meridional temperature gradient. The SCSSM onset in middle May is then characterized by the eastward intrusion of low-level southwesterlies to the west of 120°E. Such an abrupt change is a response to the differential heating between the Asian continental landmass and the adjacent oceans.

However, while the Tibetan Plateau is indeed a major heating source, the maximum occurs over the plateau after the SCSSM onset (Fig. 4a). Comparing the early and late onset years (Fig. 4b), the difference at the eastern flank of the Tibetan Plateau is not obvious (the area with values larger than 50 W m^{-2} is small), but it varies a lot over the plateau (the area with values smaller than -50 W m^{-2} is not small), which may indicate that the role of the Tibetan Plateau in SCSSM onset is still important. However, some arguments about the over-emphasis of the role of the Tibetan Plateau in the onset of SCSSM (e.g., Chen and Dell'osso, 1986; Zhang et al., 2004) suggest that other factors might come into play as well, such as the off-equatorial ITCZ. Numerical experiments have further indicated that some "wet" process such as low-level deep cumulus convection is crucial for the onset and maintenance of the monsoonal circulation, even though the uneven heating between land and sea and the Earth's rotation may be the basic reasons for the genesis of the monsoon (e.g., Chen and Dell'osso, 1986). While the topographic and thermodynamic forcings of the Tibetan Plateau on the development of the SCSSM should not be ignored, some other forcings may be important in leading to the establishment of the SCSSM.

Chen et al. (1988) suggested that the main source of the SCSSM is the water vapor transported eastward from the Indian Ocean to the SCS, and the cumulus convection over the Indian Ocean becoming stronger since April. Thus it is better to examine the Q_1 and Q_2 averaged over 40°–80°E. Similar distributions of Q_1 and Q_2 would suggest that heating might result from the cumulus convection and rainfall. For the early case, the cumulus convection over the southern Indian Ocean becomes obvious in late April in association with a northward propagation of large values of Q_1 and Q_2 before the SCSSM onset (Figs. 5a and 6a). The latent heating release propagates northward as well, and reaches a peak at a similar time (Fig. 5a) when the Q_2 also becomes the strongest in the northern Indian Ocean (Fig. 6a). After the onset, the timing of the peaks in Q_1 and Q_2 correspond well to each other indicating contributions from the released latent heat of condensation. In other words, before onset can take place, it must wait for the cumulus convection to be strong enough and plenty of moisture and external

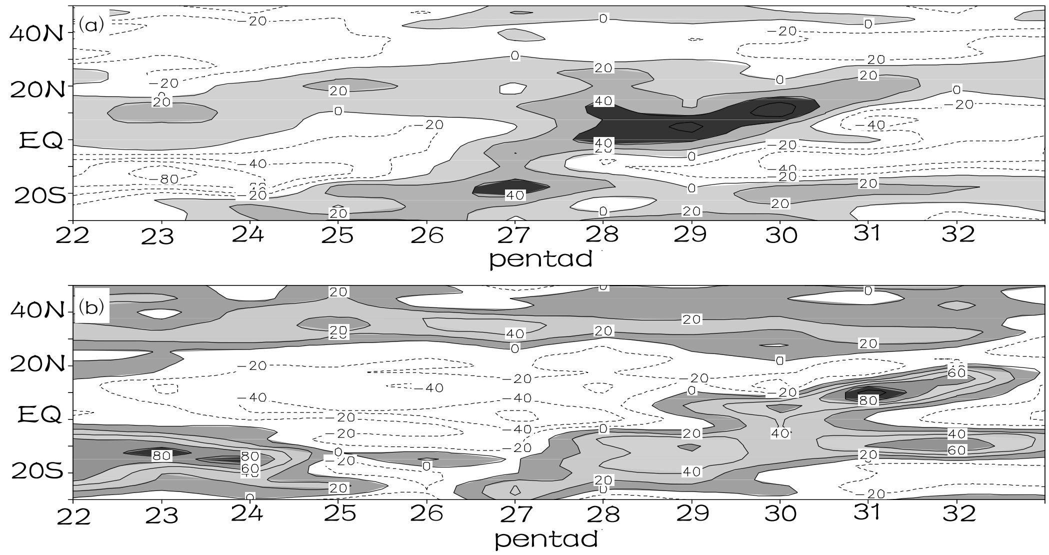

Fig. 5. Composites of Q_1 anomaly (integrated from the surface to 100 hPa) (units: W m^{-2}) for (a) early onset years and (b) late onset years averaged over (40°–80°E) starting for a total of 12 pentads. Shading indicates positive values.

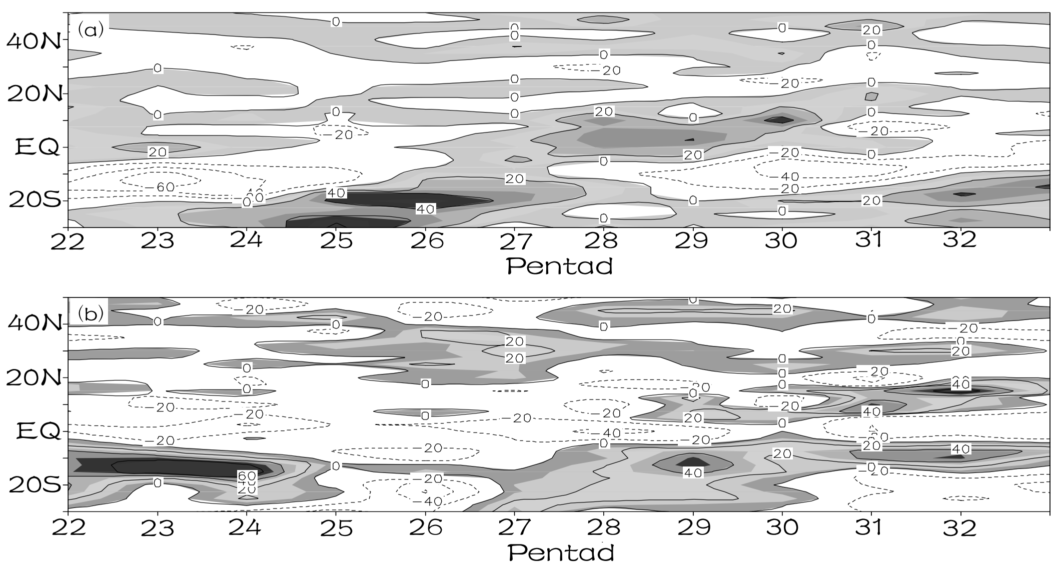

Fig. 6. Composites of Q_2 anomaly (integrated from the surface to 100 hPa) (units: W m^{-2}) for (a) early onset years and (b) late onset years averaged over (40°–80°E) for a total of 12 pentads. Shading indicates positive values.

forcing to be in such a configuration that the cross-equatorial westerly jet flows are prevalent from the Indian Ocean through the Bay of Bengal then to the SCS domain.

Note the cumulus convection in this area is suppressed in the late case from April to May (Figs. 5b and 6b). The values of Q_1 and Q_2 are mostly negative, which suggests evaporation exceeding precipitation, and are not active until mid May in the southern Indian Ocean and early June in the northern Indian Ocean. In particular, we offer the hypothesis that "prior to the onset, the buildup of planetary-scale and land-sea temperature gradients under the topographic and thermodynamic forcings of the Tibetan Plateau reach a critical stage; with the trigger of a northward-propagating off-equatorial ITCZ or cumulus convection, the troposphere is in a state of readiness for the onset of the SCSSM". This hypothesis could give an explanation for why the SCSSM onset is early or late in certain years, but still needs to be tested in future work.

5. Summary and discussion

5.1 *Summary*

This paper examines some observational variations of the SCSSM during the period 1958–2000, with the objective of identifying the large-scale features associated with the early and late onset years. The years are first classified into early, normal and late onset years based on the onset pentad. It is found that onset has occurred as early as the 25th pentad and as late as the 31st or 32nd pentad, which suggests a large interannual variability in the onset date. Likewise, a strong or weak monsoon is generally consistent with an early or late onset, though a one-to-one correspondence is not found.

The main difference in the 850-hPa flows between early and late onset cases is the timing of the westerly outbreak over the SCS domain. So the feature of the SCSSM onset is the easterlies (westerlies) controlling the upper (lower) levels of the SCS. Besides the topographic and thermodynamic forcings of the Tibetan Plateau, the northward-propagating cumulus convection over the northern Indian Ocean might enhance the monsoon trough. Plenty of moisture is then brought to the SCS. Together with the flow from the western flank of the WPSH, the effect of the horizontal advection of moisture and heat is substantially increased, thus destabilizing the atmosphere and weakening the subtropical ridge there. Westerlies can eventually penetrate and prevail over the SCS region (which is the SCSSM onset).

5.2 *Discussion*

The observations demonstrate the different features between the early and late onsets, but whether the origin of the monsoon is from the land-ocean thermal contrast or from the ITCZ is still unclear, and the determining factors related to the onset and the resultant monsoon rainfall are yet to be identified. In many respects, they are still poorly understood.

The importance of the Tibetan Plateau as an elevated heat source during the northern summer has been emphasized by many authors (e.g., Yeh, 1981; Luo and Yanai, 1983; Ding and Hu, 1988; Ding and Wang, 1988; Wu and Zhang, 1998). Yeh (1981) pointed out that the Tibetan Plateau is situated in the belt of the subtropics, and the heat source over it will greatly affect the structure of the Hadley circulation. Luo and Yanai (1983) also found that the Tibetan Plateau exerts profound topographic and thermal influence upon the low-level wind field. He et al. (1987) identified two stages existing in the upper-troposphere warming over the eastern Tibetan Plateau in May and over the Iran Plateau in June, while Li and Yanai (1996) also mentioned that the diabatic heating over the Tibetan Plateau in May and June results from sensible heating. Yanai et al. (1992) considered that the diabatic heating over the eastern Tibetan Plateau results in the tropospheric warming and leads to the reversal of the meridional gradient of temperature and the prevalence of low-level southwesterlies over the Bay of Bengal. Wu and Zhang (1998) documented that the Tibetan Plateau acts as a huge stove to release sensible heat, and it also contributes to the formation of the India-Burma trough, which allows the westerly jet to move through the Bay of Bengal region into the SCS region. They also suggested that the downwind advection of heating over the Tibetan Plateau is responsible for the onset of the summer monsoon over the SCS region being earlier than over India. Ueda and Yasunari (1998) indicated that the reversal of the meridional gradient of temperature in the troposphere over the Tibetan Plateau is the key factor for the outbreak of the Asian monsoon. Generally in boreal summer, a Hadley circulation is replaced by another regional circulation system over the Tibetan Plateau and its adjacent area. The large-scale mean circulation of the Asian monsoon is characterized by strong cross-equatorial flows from the Southern to the Northern Hemisphere in the lower troposphere, near the longitudes of the east coast of Africa and the maritime continent of Borneo and Indonesia, as well as a strong westerly flow over South India and a southwesterly flow over China. This circulation is superposed at 200 hPa by an intense anticyclone centered over the Tibetan Plateau and strong easterly winds near 10°N stretching from Indochina to West Africa.

In addendum, a further view of the land-sea thermal contrast can be had by looking at the variations of the Q_1 and Q_2 anomalies over the 90 days prior to the onset day of the SCSSM as shown in Figs. 7 and 8 for the location of the Tibetan Plateau (27.5°–37.5°N, 80°–100°E). Yet, as has also been mentioned in previous studies, there have been a number of cases of topographic and thermodynamic forcings of the Tibetan Plateau coherent in the establishment of the SCSSM that have been observed or numerically tested (e.g., Yeh, 1981; Luo and Yanai, 1983; Wu and Zhang, 1998). The region we are focusing upon in this section is the off-equatorial ITCZ, within the Tropics, and within the longitude range of about 40°E to the dateline. In the previous section, Figs. 7 and 8 suggest that the burst of the SCSSM is highly associated with the

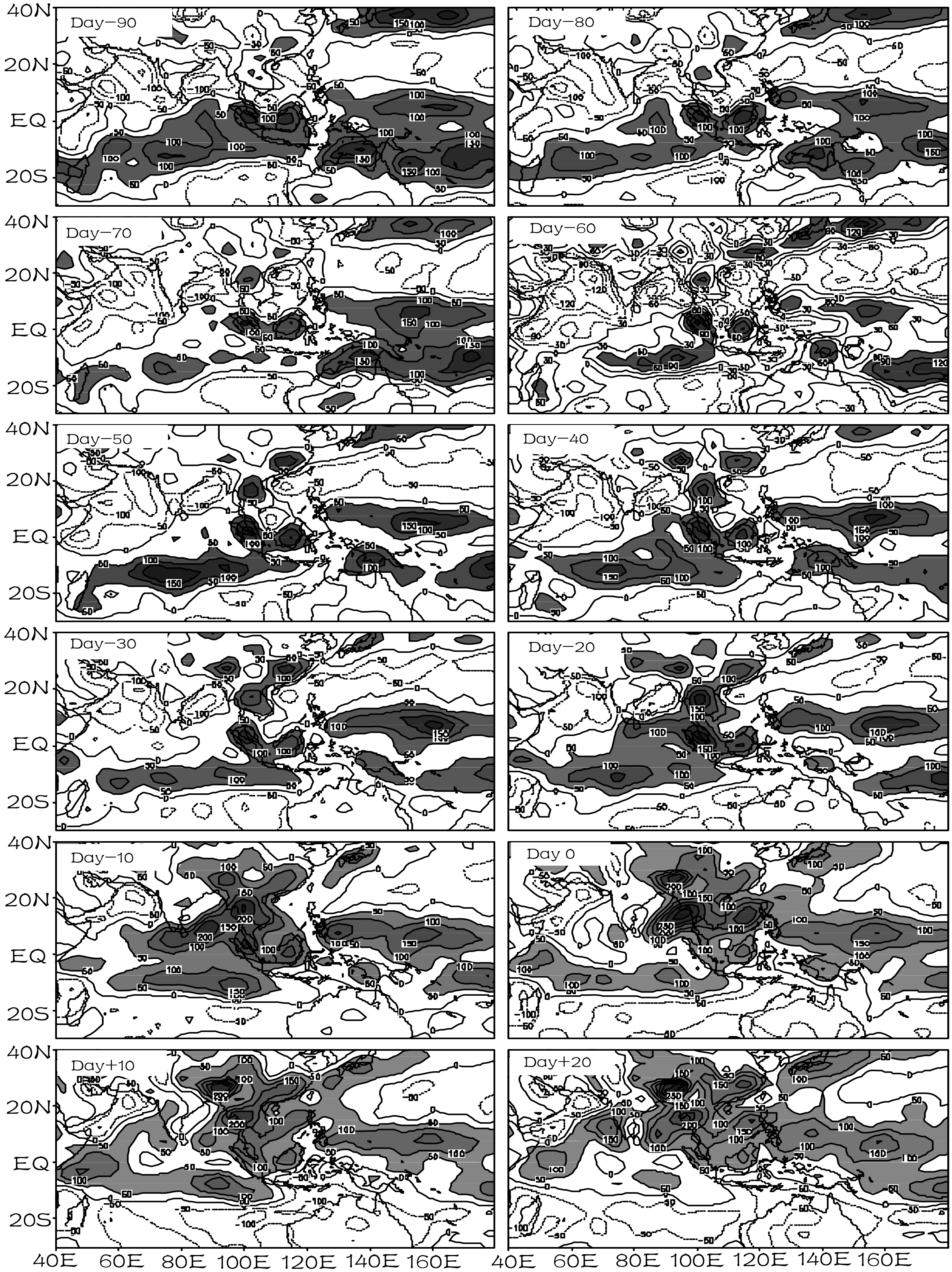

Fig. 7. Composite Q_1 anomaly (units: W m^{-2}) prior to the onset date of the SCSSM. Shading indicates values larger than 50 W m^{-2}.

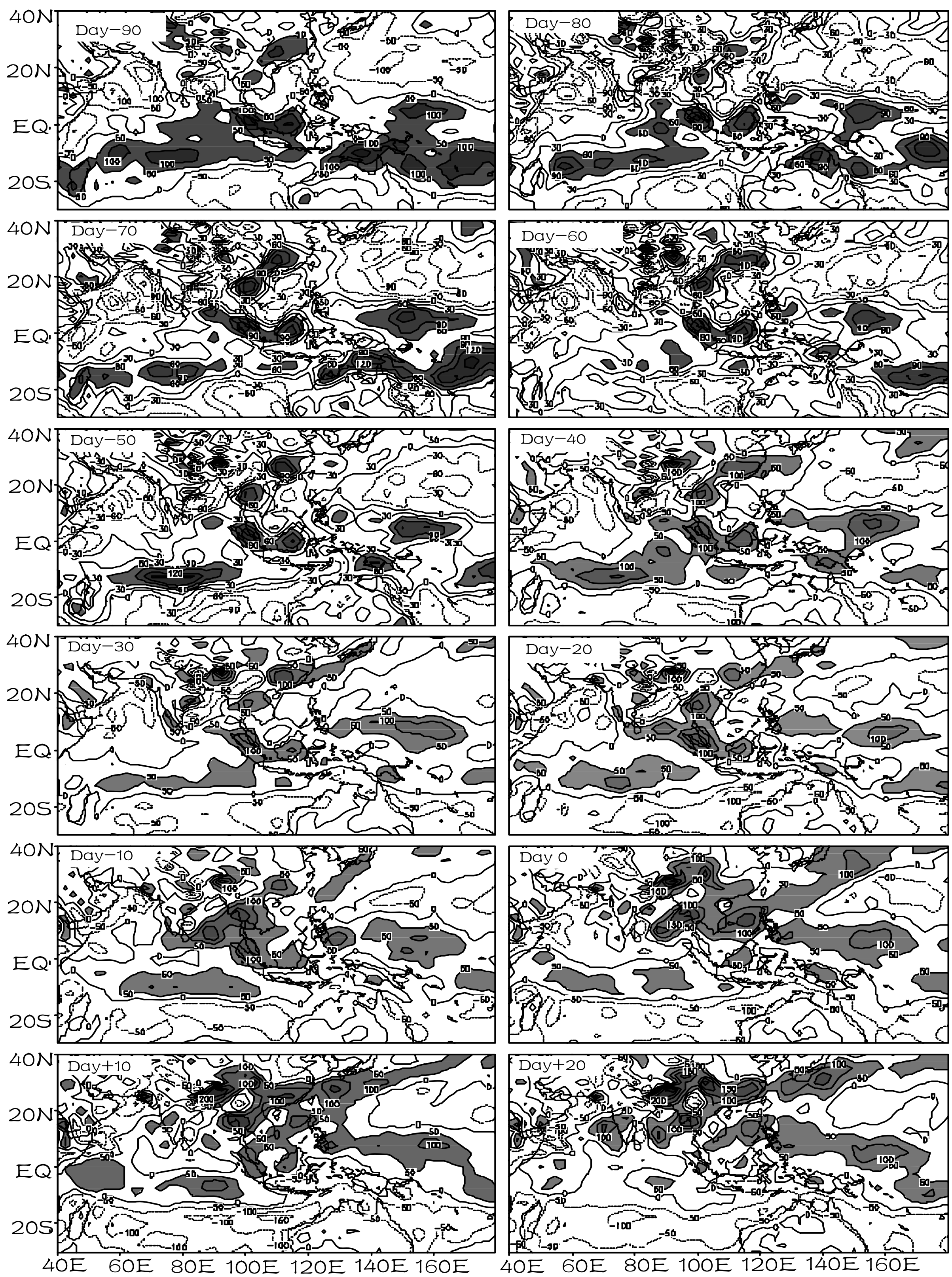

Fig. 8. Composite Q_2 anomaly (units: W m^{-2}) prior to the onset date of the SCSSM. Shading indicates values larger than 50 W m^{-2}.

northward-propagating cumulus convections from the southern Indian Ocean or the Southern Hemisphere to the SCS. This is further supported by Figs. 7–8 showing a systematic northward movement of the maximum Q_1 and Q_2, which are not over the Tibetan Plateau, but over the equatorial Indian Ocean and western Pacific. In other words, at day -90 (90 days prior to the onset), two obvious bands of convection (the maximum Q_1 and Q_2 indicates strong latent heating, and obvious moisture condensation) are located at the southern Indian Ocean, the equatorial western Pacific to the north of the Australian continent. These two bands shrink with a slight northward movement from day -80 to day -60, but enlarge at day -50 and -40 accompanied by a slow northward propagation, and shrink again at day -30, and later enhanced again at day -20 and day -10. Eventually these two bands, one from the southern Indian Ocean with an obvious northeastward propagation, and the other from the equatorial western Pacific with an obvious northwestward propagation, merge together over the SCS domain, which indicates the outbreak of the SCSSM. The notable feature of the northward propagation along the land bridge in Southeast Asia was also suggested by Lau and Yang (1996) and Hsu et al. (1999). These characteristics are indeed suggestive of an influence on the moisture source of the SCSSM. No doubt, this is partly a reflection of the propagation fact that the robust cumulus convections emanate from the equatorial Indian Ocean and the equatorial western Pacific on the intraseasonal scale, sometimes active or breaking in midcourse. Fu and Wang (2004) also found such systematically stronger northward-propagating intraseasonal oscillations in the coupled model.

Therefore, though the Tibetan Plateau's thermodynamic forcing is key in the monsoon onset, the notion that the northward-propagating off-equatorial ITCZ or cumulus convections from the Southern Hemisphere are one of the factors responsible for the monsoon outbreak is somewhat reasonable.

Acknowledgments. This research was partially supported by the City University of Hong Kong Research under Grant No. 7001336 and by the Chinese Academy of Sciences (KZCX3-SW-226).

REFERENCES

Arkin, P. A., and P. E. Ardanuy, 1989: Estimating climatic-scale precipitation from space: A Review. *J. Climate*, **2**, 1229–1238.

Chan, J. C. L., W. Ai, and J. Xu, 2002: Mechanism responsible for the maintenance of the 1998 South China Sea summer monsoon. *J. Meteor. Soc. Japan*, **80**, 1103–1113.

Chen Longxun, Li Wei, Zhao Ping and Tao Shiyan, 2001: On the process of summer monsoon onset over East Asia. *Climatic and Environmental Research*, **5**, 345–355. (in Chinese)

Chen, T. C., M. C. Yen, and M. Murakami, 1988: The water vapor transport associated with the 30–50 day oscillation over the Asian monsoon regions during 1979 summer. *Mon. Wea. Rev.*, **116**, 1983–2002.

Chen, S. J., and L. Dell'Osso, 1986: Numerical experiments on the sensitivity of the monsoon circulation to differential heating. *Quart. J. Roy. Meteor. Soc.*, **112**, 93–112.

Ding Yihui, and Hu Jian, 1988: The variation of the heat sources in East Asia in the early summer of 1984 and their effects on the large-scale circulation in East Asia. *Adv. Atmos. Sci.*, **6**, 171–180.

Ding Yihui, and Wang Xiaofang, 1988: An analysis of the distribution of apparent heat sources and sinks over the middle reaches of Yangtze River during the Meiyu season in 1983. *Journal of Tropical Meteorology*, **4**, 134–145. (in Chinese)

Ding Yihui, and Li Chongyin, 1999: *Onset and Evolution of the South China Sea Monsoon and Its Interaction with the Ocean*. China Meteorological Press, Beijing, 350pp.

Ding, Y. H., and Y. Liu, 2001: Onset and evolution of the summer monsoon over the South China Sea during SCSMEX field experiment in 1998. *J. Meteor. Soc. Japan*, **79**, 255–276.

Fu, X. H., and B. Wang, 2004: Differences of boreal summer intraseasonal oscillations simulated in an atmosphere-ocean coupled model and an atmosphere-only model. *J. Climate*, **17**, 1263–1271.

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

Hirasawa, H., K. Kato, and T. Takeda, 1995: Abrupt change in the characteristics of the cloud zone in subtropical East Asia around the middle of May. *J. Meteor. Soc. Japan*, **73**, 221–239.

Hsu, H. -H., C. -T. Teng, and C. -T. Chen, 1999: Evolution of large-scale circulation and heating during the first transition of Asian Summer Monsoon. *J. Climate*. **12**, 793–810.

Huang Gang, 2004: An index measuring the interannual variation of the East Asian Summer Monsoon–The EAP index. *Adv. Atmos. Sci.*, **21**, 41–52.

Lau, K. M., K. M. Kim, and S. Yang, 2000: Dynamical and boundary forcing characteristics of regional components of the Asian Summer Monsoon. *J. Climate*, **13**, 2461–2482.

Lau, K. M., and S. Yang, 1996: The Asian monsoon and predictability of the tropical ocean-atmosphere system. *Quart. J. Roy. Meteor. Soc.*, **122**, 945–957.

Li, C., and M. Yanai, 1996: The onset and interannual variability of the Asian summer monsoon in relation to land-sea thermal contrast. *J. Climate*, **9**, 358–375.

Li Chongyin, and Qu Xin, 1999: Atmospheric circulation evolutions associated with summer monsoon onset in the South China Sea. *Chinese J. Atmos. Sci.*, **23**, 311–325. (in Chinese)

Li Chongyin, and Wu Jingbo, 2000: On the onset of the South China Sea summer monsoon in 1998. *Adv. Atmos. Sci.*, **17**, 193–204.

Li Jianping, and Zeng Qingcun, 2003: A new monsoon index and the geographical distribution of the global monsoons. *Adv. Atmos. Sci.*, **20**, 299–302.

Liu, Y. M., J. C. L. Chan, J. Y. Mao, and G. X. Wu, 2002: The role of Bay of Bengal convection in the onset of the 1998 South China Sea summer monsoon. *Mon. Wea. Rev.*, **130**, 2731–2744.

Lu, E., and J. C. L. Chan, 1999: A unified monsoon index for South China. *J. Climate*, **12**, 2375–2385.

Luo, H., and M. Yanai, 1984: The large-scale circulation and heat sources over the Tibetan Plateau and surrounding areas during the early summer of 1979. Part II: Heat and moisture budgets. *Mon. Wea. Rev.*, **112**, 966–989.

Mao, J. Y., and J. C. L. Chan, 2005: Intraseasonal variability of South China Sea summer monsoon. *J. Climate*, **18**, 2388–2402.

Pan Jing, 2004: The activities of the SCSSM trough and its impact on the climate of the North Hemisphere. PhD Dissertation, Institute of Atmospheric Physics/Chinese Acaderny of Sclences, 109pp. (in Chinese)

Shi, N., and Q. G. Zhu, 1996: An abrupt change in the intensity of the East Asia summer monsoon index and its relationship with temperature and precipitation over East China. *Int. J. Climatol.*, **16**, 757–764.

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

Ueda, H., and T. Yasunari, 1998: Role of warming over the Tibetan Plateau in early onset of the summer monsoon over the Bay of Bengal and the South China Sea. *J. Meteor. Soc, Japan*, **76**, 1–12.

Wang Bin, and Wu Renguang, 1997: Peculiar temporal structure of the South China Sea summer monsoon. *Adv. Atmos. Sci.*, **14**, 177–194.

Webster, P. S., and S. Yang, 1992: Monsoon and ENSO: Selectively interactive systems. *Quart. J. Roy. Meteor. Soc.*, **118**, 887–926.

Wu, G., and Y. Zhang, 1998: Tibetan Plateau forcing and the timing of the monsoon onset over South Asia and the South China Sea. *Mon. Wea. Rev.*, **126**, 913–927.

Xie, A., Y.-S. Chung, X. Liu, and Q. Ye, 1998: The interannual variations of the summer monsoon onset over the South China Sea. *Theoretical* & *Applied Climatology*, **59**, 201–213.

Xu Haiming, He Jinhai, and Zhou Bing, 2001: Composite analysis of summer monsoon onset process over South China Sea. *Journal of Tropical Meteorology*, **17**, 10–22. (in Chinese)

Xu, J., and J. C. L. Chan, 2001: First transition of the Asian summer monsoon in 1998 and the effect of the Tibet-tropical ocean thermal contrast. *J. Meteor. Soc. Japan*, **79**, 241–253.

Yanai, M., Li. C. Y., and Z. Song, 1992: Seasonal heating of the Tibetan Plateau and its effects on the evolution of the Asian summer monsoon. *J. Meteor. Soc. Japan*, **70**, 319–351.

Yeh, T. C., 1981: Some characteristics of the summer circulation over the Qinghai-Xizang (Tibet) Plateau and its neighborhood. *Bull. Amer. Meteor. Soc.*, **62**, 14–19.

Zhang, Z., J. C. L. Chan, and Y.-H Ding, 2004: Characteristics, evolution and mechanisms of the summer monsoon onset over Southeast Asia. *Int. J. Climatol.*, **24**, 1461–1482.