

Tibetan Plateau Forcing and the Timing of the Monsoon Onset over South Asia and the South China Sea

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ABSTRACT

Observations were employed to study the thermal characteristics of the Tibetan Plateau and its neighboring regions, and their impacts on the onset of the Asian monsoon in 1989. Special attention was paid to the diagnosis of the temporal and spatial distributions of surface sensible and latent heat fluxes. Results show that the whole procedure of the outbreak of the Asian monsoon onset is composed of three consequential stages. The first is the monsoon onset over the eastern coast of the Bay of Bengal (BOB) in early May. It is followed by the onset of the East Asian monsoon over the South China Sea (SCS) by 20 May, then the onset of the South Asian monsoon over India by 10 June. It was shown that the onset of the BOB monsoon is directly linked to the thermal as well as mechanical forcing of the Tibetan Plateau. It then generates a favorable environment for the SCS monsoon onset. Afterward, as the whole flow pattern in tropical Asia shifts westward, the onset of the South Asian monsoon occurs.

Finally, the timing of the onset of the Asian monsoon in 1989 was explored. It was shown that the onset of the Asian monsoon occurs when the warm or rising phase of different low-frequency oscillations reach the "East Asian monsoon area" (EAMA) concurrently. These include the warm phase of the eastward propagating two- to three-week oscillation (TTO) of the upper-layer temperature in middle latitudes, the rising phase of the northward propagating Madden-Julian oscillation of the southern tropical divergence, and the rising phase of the westward propagating TTO of the western Pacific divergence. It was concluded that the timing of the Asian monsoon onset is determined when the favorable phases of different low-frequency oscillations are locked over the EAMA.

1. Introduction

The seasonal transition from winter to summer in the Far East and South Asia is characterized by the abrupt changes in general circulation and weather patterns in the region (Ye et al. 1959; Matsumoto 1992; Murakami and Matsumoto 1994). This usually occurs in May and June in association with the onset of the Asian monsoon (Krishnamuti et al. 1985; Hirasawa et al. 1995). It has long been recognized that climate mean outbreak of the Asian summer monsoon starts first in the South China Sea (SCS) region in early and middle May, then propagates westward gradually. In early and middle June, it reaches the South Asian subcontinent region. The onset of the South Asian monsoon then occurs (refer to Tao and Chen 1987; Chang and Chen 1995).

Both the onsets of the East Asian and South Asian monsoons are consequences of the atmospheric response to the changes in the contrast of thermal heating between land and ocean (Murakami and Ding 1982; Johnson et

al. 1987; Luo and Yanai 1983, 1984; He et al. 1987; Chen et al. 1991). Particularly, the elevated heating of the Tibetan Plateau to the atmosphere plays a fundamental role in the formation and maintenance of the summer circulation at least over Asia (Staff Members of Academia Sinica 1957; Flohn 1957, 1969; Ye et al. 1979; Ye 1981, 1982; Luo and Yanai 1983, 1984; Chen et al. 1985; He et al. 1987; Li and Yanai 1996). This has been further elucidated and confirmed by a series of numerical experiments (Hahn and Manabe 1975). In the case in which orography is excluded, an Asian monsoon does not appear, and the convergence and rainy belt over the Indian Ocean is located near the equator just like the intertropical convergence zone (ITCZ) over the Pacific. Their experiments remind us that for the formation of the Asian monsoon, the differential heating between land and ocean is only a part of the story, and the important mechanism must lie in the influences of the Tibetan Plateau.

It has long been puzzled why the monsoon onset occurs earlier in the SCS region than over the Indian region, and how this East Asian monsoon onset is linked to the forcing of the Plateau. Chang and Chen (1995) briefly reviewed several hypotheses that had been proposed to explain the SCS monsoon onset considering

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the plateau forcing, and suggested that the SCS monsoon onset is triggered by the approach of a midlatitude trough–front system. Since in late spring and early summer, most of such trough systems, which intrude the SCS region, are associated with the India–Burma trough, then, how the Tibetan Plateau affects the formation of the India–Burma trough requires further investigation.

Another important subject in the literature of monsoon study that needs to be clarified further concerns the timing of monsoon onset. Chen and Chang (1980) and Krishnamurti et al. (1981) found that the onset of the East Asian monsoon is associated with the generation, development, and movement of the so-called onset vortex or “monsoon low” in the lower troposphere. The low-frequency oscillation (LFO) of either the Madden and Julian (1971, 1972) mode (MJO) or two- to three-week mode (TTO) has been considered as an important mechanism that modulates the monsoon activities (Tao et al. 1963; Murakami 1976; Krishnamurti and Bhalme 1976; Yasunari 1979, 1980; Krishnamurti and Subrahmanyam 1982; Krishnamurti et al. 1985; Lorenc 1984; Murakami et al. 1986; Chen 1987; Nakazawa 1992; Chen and Chen 1995). Their studies remind us that when the timing of monsoon onset is studied, in addition to the plateau impacts, attention should also be paid to the atmospheric motions with different frequencies.

In this paper we employ observational study for the year 1989, and try to get some new insights into the problems concerning how the plateau forcing is associated with the Asian monsoon onset and what the causes are of the location and timing of the early monsoon onset over Asia. The data used include the European Centre for Medium-Range Weather Forecasts (ECMWF) analyzed data, the ECMWF Tropical Ocean Global Atmosphere (TOGA) complementary data, the National Centers for Environmental Prediction (NCEP)–National Meteorological Center (NMC) outgoing longwave radiation (OLR) data, and the Chinese rainfall data collected from 336 observation stations and archived at the Data Center of the Institute of Atmospheric Physics. The OLR data is resolved at a network of 5° longitude by 5° latitude. The ECMWF data is a twice-daily (0000 and 1200 UTC) objective analyzed grid data with a resolution of 2.5° longitude by 2.5° latitude. The ECMWF TOGA complementary data are contended in the ECMWF extended FGGE level-III dataset, which is also twice-daily grid data but with a finer resolution of 1.125° latitude by 1.125° longitude. It includes surface wind stress, latent heat flux, sensible heat flux, net radiation, etc. The data were used as initial values of the ECMWF assimilation system for routine medium-range weather forecasts. It compensates the sparse data coverage over the plateau area and over the oceans. From the performance of the ECMWF forecasts and from the comparison of its analyses with those of Ye et al. (1979) by using the station-based observation,

we found the dataset, in general, is good in fidelity and is appropriate for the present research.

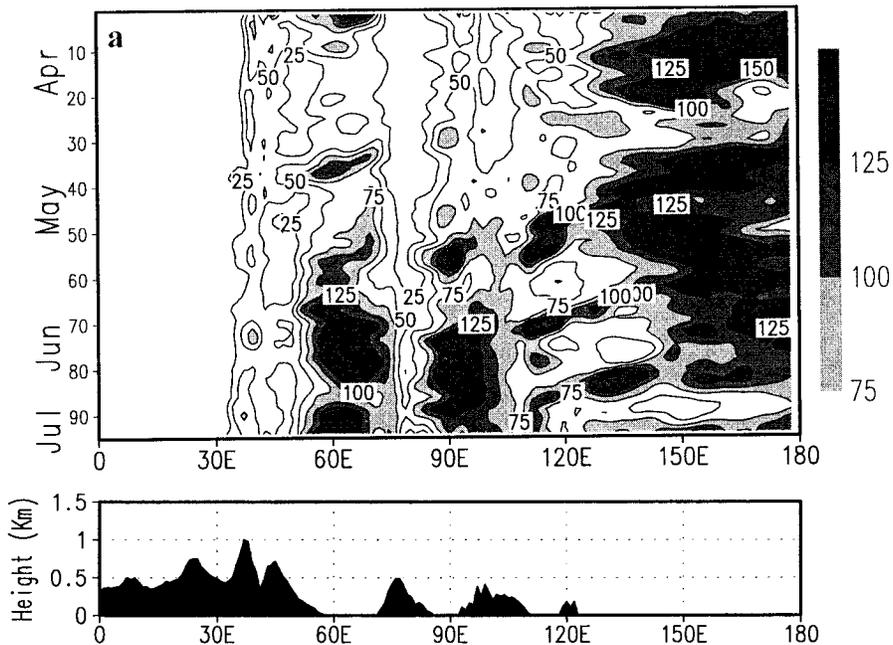
In sections 2 and 3, the seasonal variations of differential heating between land and ocean in the Asian monsoon region are examined. The significance of the surface sensible heat flux of the Tibetan Plateau in the seasonal transition of the general circulation over the monsoon area is also considered. The focus in section 4 is on how the huge sensible heating of the Tibetan Plateau in late spring and early summer in 1989 leads to the onset of the Asian monsoon in the SCS region. The importance of both the thermal and mechanical forcing of the plateau is investigated. In section 5, efforts are made to understand the timing of the Asian monsoon onsets at various locations by considering the interactions of different low-frequency oscillations during the seasonal transition period. Some conclusions and discussions are given in section 6.

2. The thermal features in the boreal tropical and subtropical regions

During the seasonal transition from winter to summer in 1989, the evolutions of surface latent heat flux in different latitude zones are shown in Fig. 1. At the bottom of each panel is shown the distribution of the mean orography meridionally averaged over the latitude domain. In the Tropics (Fig. 1a), the latent heat flux from ocean surface is much larger than that from land surface. Over the ocean surface, the intensity of most of the maximum centers exceeds 200 W m^{-2} . On the contrary, over the African continent it is basically below 50 W m^{-2} . From late May to late June, there are three maximum perturbation centers of more than 200 W m^{-2} appearing over the western Pacific and propagating westward. The first and second centers correspond to typhoons 8903 and 8905, respectively, and the third one appeared already in late June.

The evolution of surface latent heat flux in the subtropical zone (27.5° – 37.5°N), in which the Tibetan Plateau is located, is shown in Fig. 1b. In the area west of 80°E , the land surface latent heat flux is basically below 50 W m^{-2} , similar to what occurs in the Tropics. Whereas in the area east of 80°E , its seasonal transition is remarkable. Before early May, as the offshore cyclone disturbances develop near the east China coast and propagate eastward one after another, periodically intensified latent heat fluxes are observed over the western Pacific. At the same time the flux over land surface is weaker. After the middle of May, the aforementioned cyclogenesis disappears, and the surface latent heat flux over the western Pacific becomes very weak. The maximum surface latent heat flux is now observed over the plateau. The contrast in surface latent heat flux between land surface and sea surface is reversed. This of course will exert considerable impacts on the general circulation of the atmosphere. However, since the substantial expansion and increase of surface latent heat flux over the

LH (Wm^{-2}) 10-20°N 1989



LH (Wm^{-2}) 27.5-37.5°N 1989

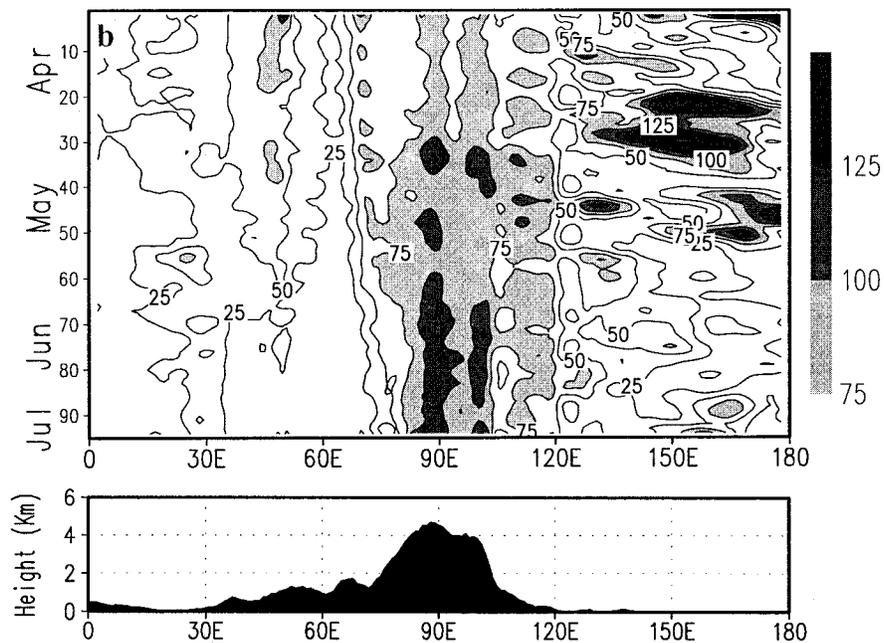


FIG. 1. The evolution of longitudinal distribution of surface latent heat flux during April to July 1989, averaged over 10°–20°N (a), and 27.5°–37.5°N (b). Interval is 25 $W m^{-2}$. Stippling indicates the area of more than 75 $W m^{-2}$. The numbers along the vertical coordinate denote the days counted from 1 April. At the bottom of each panel is the meridional mean orographic height (km) averaged over the corresponding latitude zone.

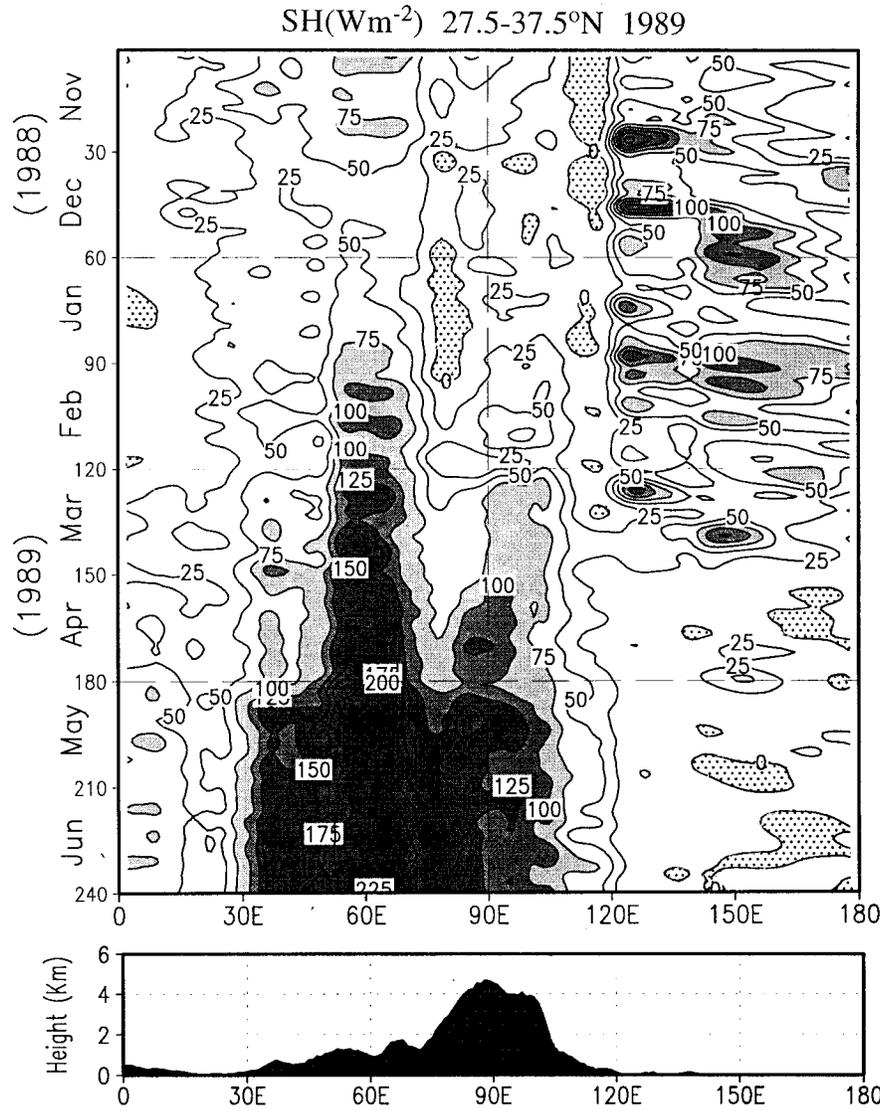


FIG. 2. The evolution of longitudinal distribution of surface sensible heat flux during November 1988 to June 1989, meridionally averaged between 27.5° and 37.5°N. Interval is 25 W m⁻². Light stippling indicates the area of more than 75 W m⁻², and dotted stippling, negative values. The numbers along the vertical coordinate denote the days counted from 1 November. The bottom panel shows the meridional mean orographic height (km) averaged between 27.5° and 37.5°N.

plateau and over east China occur in May onward, in accordance with the rainy season there, when we are studying those persistent processes that have occurred over the plateau for some period and contributed to the Asian monsoon onset, we may therefore disregard the influence of surface latent heat flux.

The evolution of surface sensible heat flux from November 1988 to June 1989 in the same subtropical zone is shown in Fig. 2. Over the western Pacific and before the spring equinox, the offshore cyclone disturbances accompanied by large surface sensible heat flux develop one after another and propagate eastward. After the equinox, as the surface air temperature along the east China coastal region is getting warmer and closer to the

nearby sea surface temperature, those cyclone disturbances observed in Fig. 1b do not generate large surface sensible heat flux when they move from continent area to the offshore ocean area after the equinox. The western Pacific then becomes tranquil. Some regions are even controlled by downward flux. Although the sensible heat flux to the west of 30°E remains small, prominent changes occur over the massive Asian continental area during the period. In the winter months, downward flux are observed over the western part of the Tibetan Plateau and along the coast area of eastern China. From late winter onward, upward flux of more than 100 W m⁻² appears at first in Iran, Afghanistan, and Pamirs, and is intensified in April. At the beginning of May, this strong

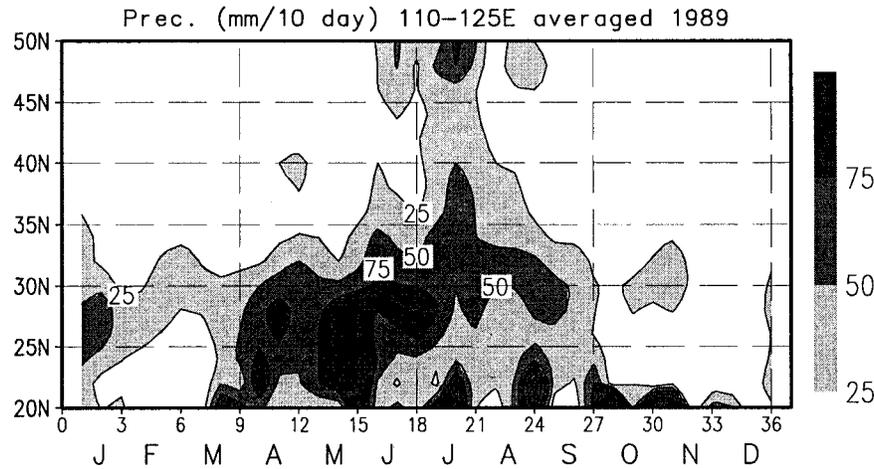


FIG. 3. Meridional movement of the 10-day total rainfall in 1989 averaged over the longitude domain 110°–125°E. Interval is 25 mm. Shading varies from very heavy, heavy, to light, indicating respectively, regions of more than 75 mm, between 75 and 50 mm, and between 50 and 25 mm.

positive sensible heat flux has already covered the vast areas from the Mediterranean Sea to the Tibetan Plateau, with maximum intensity of more than 200 W m^{-2} located over the western part of the plateau. Along the eastern coast of the Asian continent, the reversal of the gradient of surface sensible heat flux occurs by the spring equinox. After the equinox, the thermal contrast between land and sea has changed completely. It is important that, unlike the seasonal change of latent heat flux, the strong surface sensible heating over the plateau and middle-east Asian region appears much earlier and persists afterward. Therefore, the surface sensible heating of the plateau must have played some roles in the seasonal transition at least over the surrounding regions and in the onset of the Asian monsoon. This has been discussed by many other authors and will be further investigated below.

3. Sensible heating of the Tibetan Plateau

During the onset of the East Asian monsoon, heavy rainfall occurs in the SCS and extends to the coast area

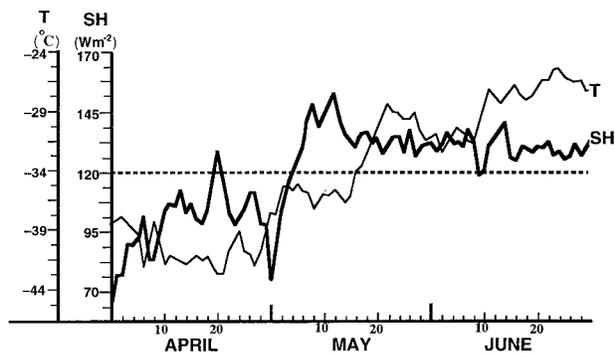


FIG. 4. The evolutions during April–June of surface sensible heat flux SH (W m^{-2}), and 300-hPa temperature T ($^{\circ}\text{C}$), averaged over the Tibetan Plateau area (27.5° – 37.5°N , 80° – 100°E).

of southern China. Figure 3 shows the latitude–time distribution of 10-day total rainfall in 1989, which is averaged over the 110° – 125°E domain. It can be seen from the figure that after the middle of May, heavy rain of more than 75 mm occurs in the area south of 30°N . In addition, according to daily observations, low OLR of less than 220 W m^{-2} and strong southerlies at 850 hPa in the SCS region of (10° – 20°N , 110° – 120°E) appears by 20 May. Then 20 May can be chosen as the onset day of the SCS monsoon in 1989. This choice is close to that of 15–20 May by Chen et al. (1996) based upon the analysis of temperature of blackbody at cloud top (T_{BB}), and that of 10–15 May by Q. Ye (1995, personal communication) based upon the minimum OLR data analysis. On 10 June, the heavy rainband jumps northward to the latitude domain 27° – 30°N , and dry period starts in southern China. In the following sections, we will show this is the time when the South Asian monsoon onset occurs.

To understand the thermal aspects of the plateau in association with the Asian summer monsoon onset, the evolutions of the daily mean sensible heat flux at the surface (SH) and the temperature at 300 hPa (T) averaged over the Tibetan Plateau area (27.5° – 37.5°N , 80° – 100°E) in the transition period are shown in Fig. 4. The temperature at 300 hPa is representative of the average temperature of the tropospheric column over the plateau. Three spells of abrupt increase in T occur at the ends of late April, middle May, and early June, respectively. The amounts of temperature increase during the abrupt warming spells are 6° , 7° , and 4°C over 5, 6, and 3 days, respectively. Enhanced sensible heating appears about 10 days before the first and second spells. The structure of the three-spell abrupt warming is an important characteristic phenomenon. We will discuss this further when we investigate the monsoon onset and low-frequency oscillations of the year in the following context.

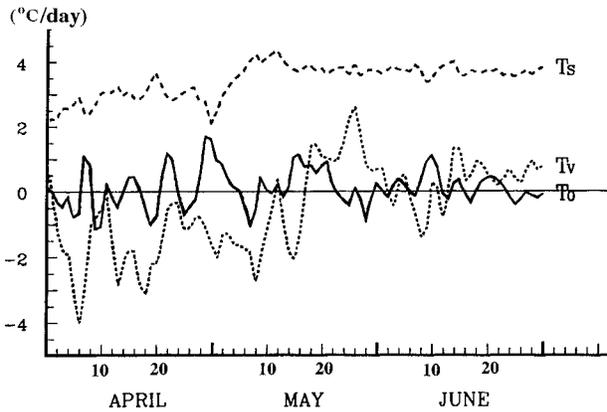


FIG. 5. The evolutions during April–June 1989 of the 200–500-hPa mean temperature advection (T_v) and local change (T_o), and the surface sensible heating averaged over the Tibetan Plateau (27.5° – 37.5° N, 80° – 100° E) and measured in degrees Celsius per day.

To further elucidate the importance of surface sensible heating in the increase of column temperature, let T_o be the local change of temperature of the unit column between 200 and 500 hPa; T_v , the heating rate due to temperature advection; and T_s , the heating rate due to surface sensible heat flux SH. Then we can employ observation data and the thermal dynamic equation to estimate the relative importance of surface sensible heating and horizontal advection. The results for the plateau area (27.5° – 37.5° N, 80° – 100° E) are shown in Fig. 5. It becomes obvious that during the seasonal transition, the surface-sensible heating over the plateau prevails over advection in the local temperature change. From the beginning of April to the end of June and over the plateau, the increase of column temperature due to surface sensible heating gradually increases from 2° to 4° C per day, while that due to advection is secondary and even negative. It is interesting to notice that before the onset of the SCS monsoon (20 May), advection over the plateau plays a role in cooling the atmosphere. This implies that the outflow over the eastern flank of the plateau is warmer than the inflow over its western flank, and the warm temperature ridge must be located to the east of the plateau. This is indeed the case of observed temperature distribution (as will be shown in Fig. 6b). Astonishingly, just 3 days before the onset, the advection T_v changes its sign. In conjunction with sensible heating, it contributes to the strong atmospheric abrupt warming over the plateau as identified in Fig. 4. This important feature reminds us that not only the surface sensible heating of the plateau, but also the atmospheric circulation pattern are important factors in contributing to the onset of the Asian monsoon.

The surface sensible heat flux over the plateau warms the atmosphere at an elevation of about 5 km, while the surrounding air is far above the boundary underneath. Such elevated heating of the plateau has profound impacts on the seasonal transition in Asian regions. As a matter of fact, from the beginning of May onward, the

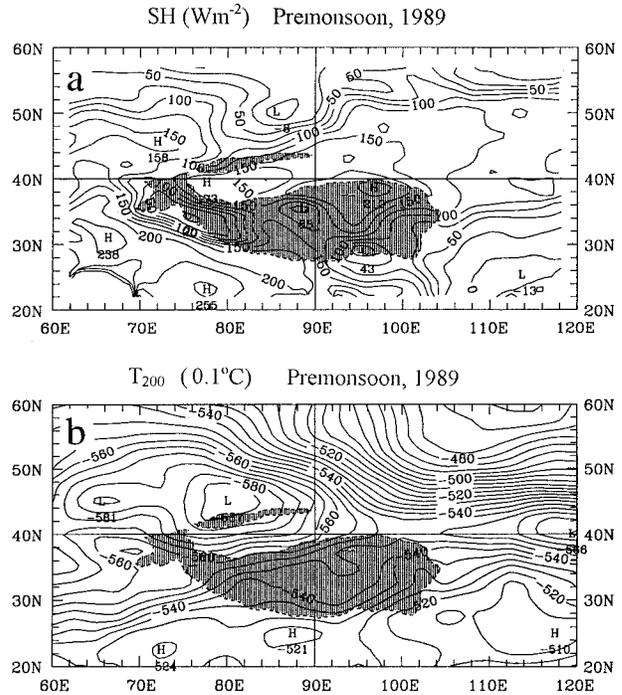


FIG. 6. Distributions in the BOB monsoon onset period (5–11 May) of 1989 of (a) the mean surface sensible heat flux (interval in 25 W m^{-2}), and (b) the mean 200-hPa temperature (unit in 0.1°C , interval in 0.5°C). Stippling indicates where the elevation of the Tibetan Plateau is higher than 3 km.

difference in the meridional mean 200-hPa temperature between the subtropics from 27.5° to 37.5° N and the Tropics from the equator to 10° N, becomes positive in the monsoon region while it remains negative in other longitude regions (figures not shown). The general feature is the same as what was reported by Li and Yanai (1996), and in favor of the development and maintenance of the South Asian high over the plateau and the Asian monsoon onset.

4. Thermal and mechanical forcing of the Tibetan Plateau and the Asia monsoon onset

One of the essential questions concerning the Asian monsoon onset is why the monsoon onset occurs earlier in the SCS region than in the Indian region? In this section we try to explore the possible mechanism linking the Asian monsoon onset to the plateau forcing.

As shown in Fig. 4, the second abrupt warming of the atmosphere over the plateau starts in the middle of May. To present the basic thermal and dynamical characteristics of the circulation before the SCS onset, we specified the period May 5–11 as the “pre-SCS onset” period since during this period the column temperature over the plateau and its local change are representative and remain relatively stable (Figs. 4 and 5). The distributions of surface sensible heat flux and 200-hPa temperature averaged over this period are shown in Fig. 6.

The sensible heat flux is below 100 W m^{-2} only over the Pamirs and Kunlun Shan mountain ranges, and over the southeast flank of the plateau. From the remainder of the plateau, large SH flux of more than 100 W m^{-2} is pumped to the atmosphere. Particularly over the central and northeastern parts of the plateau the sensible heat flux exceeds 150 W m^{-2} . Compared with the western Pacific where the surface sensible heat flux is less than 25 W m^{-2} (refer to Fig. 2), the plateau does look like a huge stove. However, the warmest column is by no means over the plateau. Due to the strong advection, the warm temperature ridge in each layer of the upper troposphere is shifted to the downwind side of the sensible heat source over the plateau. As shown in Fig. 6b, the apparent east–west gradient of temperature at 200 hPa is just over the plateau. Along 30°N , a cold trough of -56°C is at 60°E , while a warm center of -52°C is near 115°E , just to the north of the SCS. Over the plateau region, the column over its eastern part is warmer than that over its western part. Since the warm column over the plateau in summer acts as a huge chimney that sucks the lower-tropospheric air from the surroundings (Wu et al. 1996), and since the air column over the eastern plateau is warmer than that over its western part, the lower-tropospheric inflow along the eastern boundary of the plateau should be stronger than the inflow along its western boundary, as will be shown in Fig. 8a. This means that convection should be easier to develop along the eastern coast of the Bay of Bengal (BOB), in favor of the earlier monsoon onset in the eastern region.

Now let us turn to consider the mechanical forcing of the plateau. The westerly flows at 850 or 700 hPa are much below the crest of the Tibetan Plateau. When they impinge upon the plateau at its southwestern corner before the monsoon onset, according to the research of Wu (1984), the deflecting effect of the orography must prevail over the climbing effect, and airflow should go around the plateau. An anticyclonic pattern over the Arabian Sea and the India subcontinent and cyclonic pattern over the northern part of the BOB will be formed. The latter in weather practice is termed the India–Burma trough, which appears long before the Indian monsoon onset (Yin 1949). In a set of numerical experiments based on a dynamical circulation model and designed by Zhu (refer to Wu et al. 1996), it was shown that this India–Burma trough exists only when the Tibetan Plateau is presented. This supports the aforementioned postulation. Due to such mechanical forcing, the lower-tropospheric southerly inflow must be intensified in the northeastern coast of the Bay of Bengal. Figure 7 is the Hovmöller diagram of both the pentad mean wind vector at 850 hPa and the OLR averaged over 10° – 20°N . After the middle of April, persistent southerly inflow is observed between 100° and 110°E . The India–Burma trough is already discernible at the beginning of May and develops until late May (Fig. 8). During this period, northerlies and northeasterlies in the eastern front of an anticyclone that is centered over the southern

Arabian Peninsula (Fig. 8a) develop over the Arabian Sea and Indian subcontinent. By using daily observations we found that by 10 May, convergence is intensified in the eastern front of the India–Burma trough. From Fig. 7 we see that deep convection with $\text{OLR} < 220 \text{ W m}^{-2}$ occurs first along the western coast of Burma. About one week later, enhanced southerlies and deep convection start to appear in the SCS, and the SCS monsoon onset occurs. As to the Indian region, it is only when the aforementioned northerlies and northeasterlies are replaced by strong southerlies that the onset of the South Asian monsoon occurs, which is already in early June. In Fig. 7, the top branch of the “cactus” shape of the 220 W m^{-2} OLR isoline represents the deep convection over the eastern BOB region. The second branch is to its east and represents the onset of the SCS monsoon, while the third is to its west and represents the onset of the Indian monsoon. These mean that the earliest development of systematic deep convection and southerlies is over the eastern coast of the Bay of Bengal and before the onset of the SCS monsoon. It can therefore be defined as the BOB monsoon onset.

Our further concern is, what is the role played in the SCS monsoon onset of the deep convection over Burma after the BOB monsoon onset. In Figs. 8 and 9, the fields of streamlines at 850 and 200 hPa before and after the SCS monsoon onset are presented. During the BOB onset period (5–11 May) at 850 hPa (Fig. 8a) a deep trough extends from Burma to the Bay of Bengal. A prominent anticyclone is centered over Saudi Arabia and controls the Arabian Sea and the Indian subcontinent. From the Indian Ocean to the western Pacific Ocean, the cross-equator flow is very weak, limited only to the region from 90° to 120°E . Southerlies are confined to the longitude domain just next to the east of the Plateau. At 200 hPa (Fig. 9a), the pattern of streamline curvature above the North Indian Ocean is opposite to that at 850 hPa: a deep trough is over the Arabian Sea and Indian subcontinent, while an anticyclone is located to the east, over the Bay of Bengal and the Indochina Peninsula, and centered near Rangoon. The southerlies at 850 hPa and the anticyclone at 200 hPa appearing in the BOB monsoon onset period are in correspondence with the strong development of deep convection identified by $\text{OLR} < 220 \text{ W m}^{-2}$ as shown in Fig. 7 by the top branch of the OLR cactus. The development of the 850-hPa Burma trough and 200-hPa anticyclone is significant in triggering the SCS monsoon onset. First, it causes strong upper-layer divergence just over the SCS region (Fig. 9a). Second, by pumping the intensified crossing-equator northeasterlies in upper layers and southwesterlies in lower layers in the longitude domain from 90° to 120°E , it switches on a forced secondary meridional circulation with rising branch in the SCS region. Such structure develops further (figures not shown) after the BOB monsoon onset. By about 20 May the 850-hPa crossing-equator southwesterly flow has become very strong and ranged from the Indian Ocean to the SCS

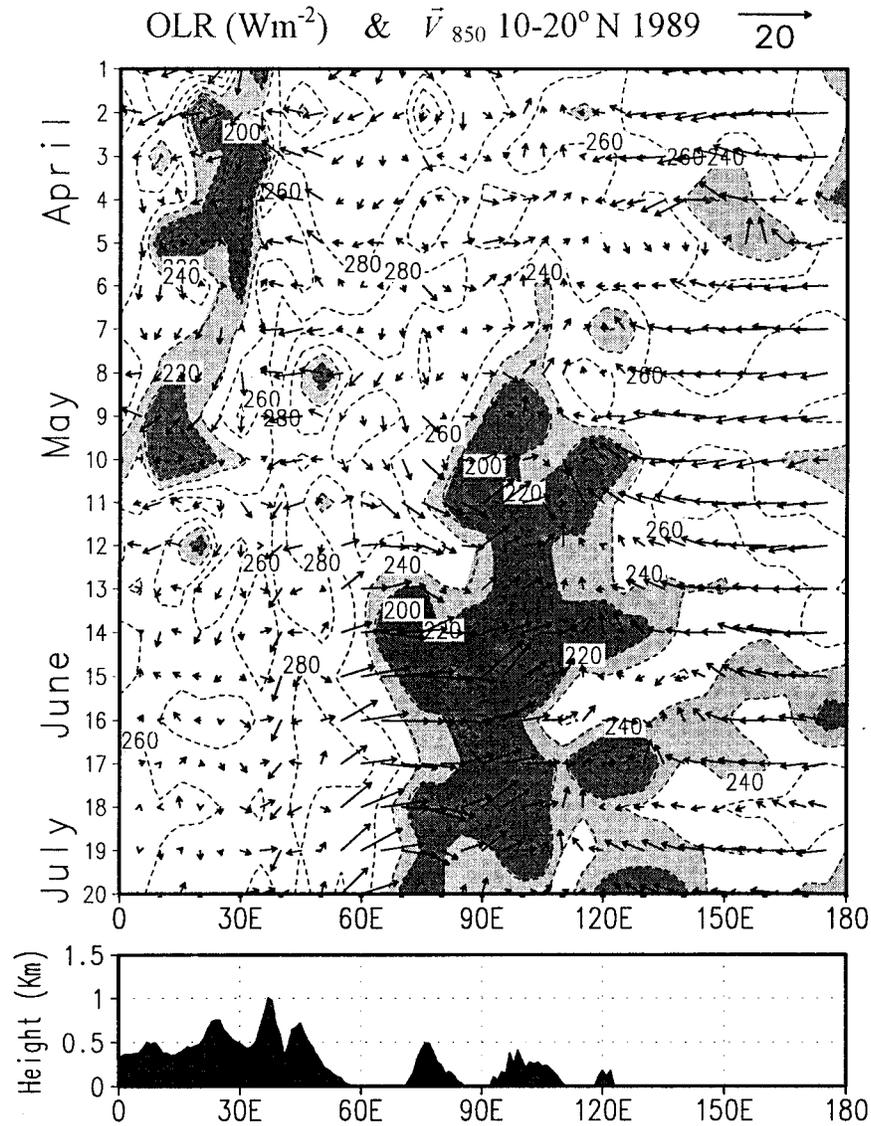


FIG. 7. The evolutions during April–mid-July 1989 of the longitudinal distribution of 850-hPa wind and outgoing longwave radiation (OLR, in contour) meridionally averaged from 10° to 20°N. Contour interval is 20 W m⁻². The numbers along the vertical coordinate denote the pentad order counted from 1 April. The bottom panel shows the meridional mean orographic height (km) averaged between 10° and 20°N

(Fig. 8b). Much intensified convergence is located just over the SCS region, and the onset of the SCS monsoon occurs. At the same time, a small-size 850-hPa cyclone and a similar size 200-hPa anticyclone (Fig. 9b) appear over the north SCS, in correspondence with the deep convection shown in Fig. 7 by the second branch of the small value (<220 W m⁻²) OLR cactus.

Furthermore, during and after the onset of the SCS monsoon the circulation patterns over the monsoon region shifts westward. By late May at 850 hPa (Fig. 8b), the Burma trough has moved to Bihar, and the anticyclone over the Arabian Sea and India has been weakened. At 200 hPa (Fig. 9b), the trough originally over

the Arabian Sea has moved to North Africa, while the anticyclone that was over Burma has now settled over the south of the Tibetan Plateau. To its south, upper-layer easterlies are intensified, and divergence prevails over the northern Indian Ocean. All these are then mature for the onset of the Indian monsoon, which occurred in early June of 1989 as signified by the third branch of the small-value cactus shown in Fig. 7, about 20 days later than the monsoon onset in the SCS region. The general features shown in Figs. 7–9 can be seen from the evolutions of the 10-yr mean 850-hPa flow for the period 1980–89 calculated by Nakazawa (1992, Fig. 2a). In which the strong impinging 850-hPa flow toward

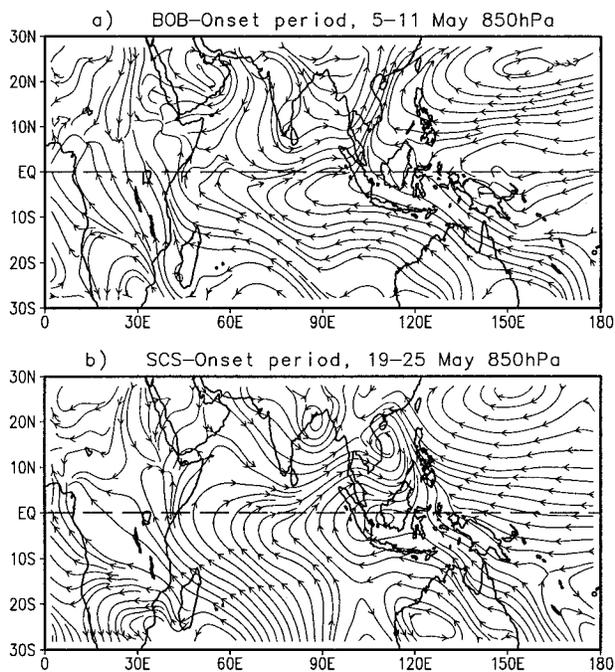


FIG. 8. Distributions of the 850-hPa streamline in 1989 during the BOB monsoon onset period 5–11 May (a), and the SCS monsoon onset period 19–25 May (b).

the Tibetan Plateau in May is observed to the south-western corner of the plateau. The early development of southerlies along the western coast of the Indochina Peninsula and the existence of anticyclonic flow over the Arabian Sea in May, and the replacement of northerlies by southerlies along the western coast of India in early June are all in agreement with those shown in Figs. 7–9.

The three-stage characteristics of the Asian monsoon outbreak revealed in the study is also in good agreement with the annual march of the rainy season in the Asian region as identified from OLR diagnosis. Using the 12-yr OLR data from 1975 to 87 (except for 1978), Nakazawa (1992) analyzed the climate mean marching of OLR distribution from 1 May to 20 June. The results (refer to his Fig. 2b) show that deep convection signified by OLR being less than 240 W m^{-2} appears first over the western coast of Burma, then extend to the SCS region by 21 May and southwestern India in early June. Using the pentad mean OLR data of the Climate Analysis Center (CAC) NMC, which are on a $2.5^\circ \times 2.5^\circ$ grid and over the period from June 1974 to December 1989, Wang (1994) found that over southern Asia the rainy season, which is defined as the pentad mean OLR being lower than or equal to 230 W m^{-2} , begins at the northern tip of Sumatra in early April and reaches the southeastern Bay of Bengal in early May, followed by northeastward and northwestward propagation. He also reported that the high reflective clouds (HRC) atlas of Garcia (1985) shows that, “from April to May an HRC

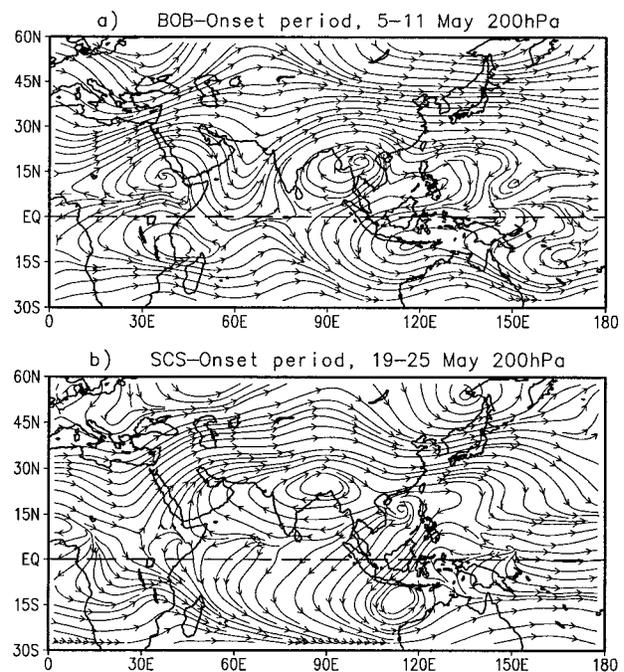


FIG. 9. The same as Fig. 8 except for the 200-hPa streamline.

maximum in the equatorial eastern Indian Ocean (2°N , 95°E) shifts to Northern Sumatra and the southern Andaman Sea (10°N , 95°E). Therefore, the HRC maximum moves steadily northwestward along the west coast of Burma until August when it reaches its northern most position in the head of the Bay of Bengal (90°E , 18°N). In addition, the marching of such a low OLR center is accompanied by the intensification of either southwesterlies or southerlies, as shown in Fig. 7. All these support our postulation that the BOB monsoon onset is the earliest monsoon onset in Asia during the spring seasonal transition.

5. Phase locking of low-frequency oscillations and the timing of the Asian monsoon onset

Undoubtedly, the thermal and mechanical forcings of the Tibetan Plateau have important impacts on the Asian monsoon climate, including its onset, lull, revival, and withdrawal. However, significant sensible heating of the plateau to the atmosphere usually begins before the spring equinox, much earlier than the monsoon onset. The intrinsic mechanism linking the Plateau heating directly to the timing of monsoon onset does not seem to be obvious. To find the mechanism that is associated with the timing of the Asian monsoon onset, in conducting this study we have analyzed different frequency oscillations in different latitude regions. Some significant oscillations associated with monsoon onset have been obtained and are presented below.

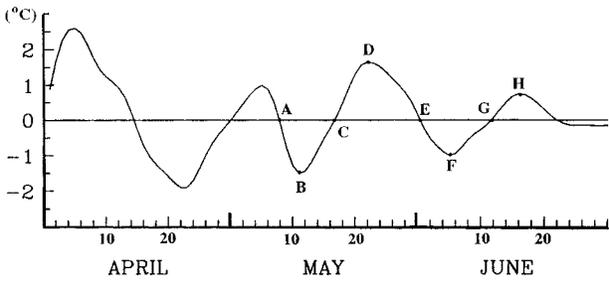


FIG. 10. The evolutions in the seasonal transition period (1 April–30 June) 1989 of the temperature departure from the corresponding period mean at 200 hPa and averaged over the Tibetan Plateau area (27.5°–37.5°N, 80°–100°E). The data used have been treated by using a 15–25-day bandpass filter. The ordinate denotes the departure of temperature (°C). The letters A–H denote different phases of the low-frequency oscillation.

a. Two- to three-week oscillation of the extratropical temperature

As discussed earlier, the elevated heating of the plateau warms the air column above and sucks lower-tropospheric air from surroundings. This becomes one of the key topics of monsoon dynamics. However, the temperature rise in the troposphere is by no means uniform.

From Fig. 4 we see that there are three spells of rapid increase: from the end of April to the beginning of May, 14–21 May, and around 10 June. These three spells correspond to the onsets of the BOB, SCS, and South Asian monsoons, respectively. The period between two successive spells is about two to three weeks, with 20 days on average. Take a 15–25-day bandpass filter and act on the 200-hPa temperature evolution. The filtered curve is shown in Fig. 10, which is generally in phase with the original evolution. The letters A–H indicate the key phases of the evolution. To see the features of temperature evolution at each of the key phases, we employ the same bandpass filter to the original ECMWF 200-hPa temperature series for each grid point and for the period January to July 1989, and extract the filtered temperature fields at the times corresponding to the eight key phases as shown in Fig. 10, then present these fields in Fig. 11. From this figure the following becomes apparent.

1) Significant TTO of the 200-hPa temperature with amplitude greater than 1.0°C occurs mainly in the extratropics. It can be detected in the Tropics only over the western Pacific (at phase C) and over the SCS region when the East Asian monsoon onset appears (at phase D). Whether the former is the extension of the propa-

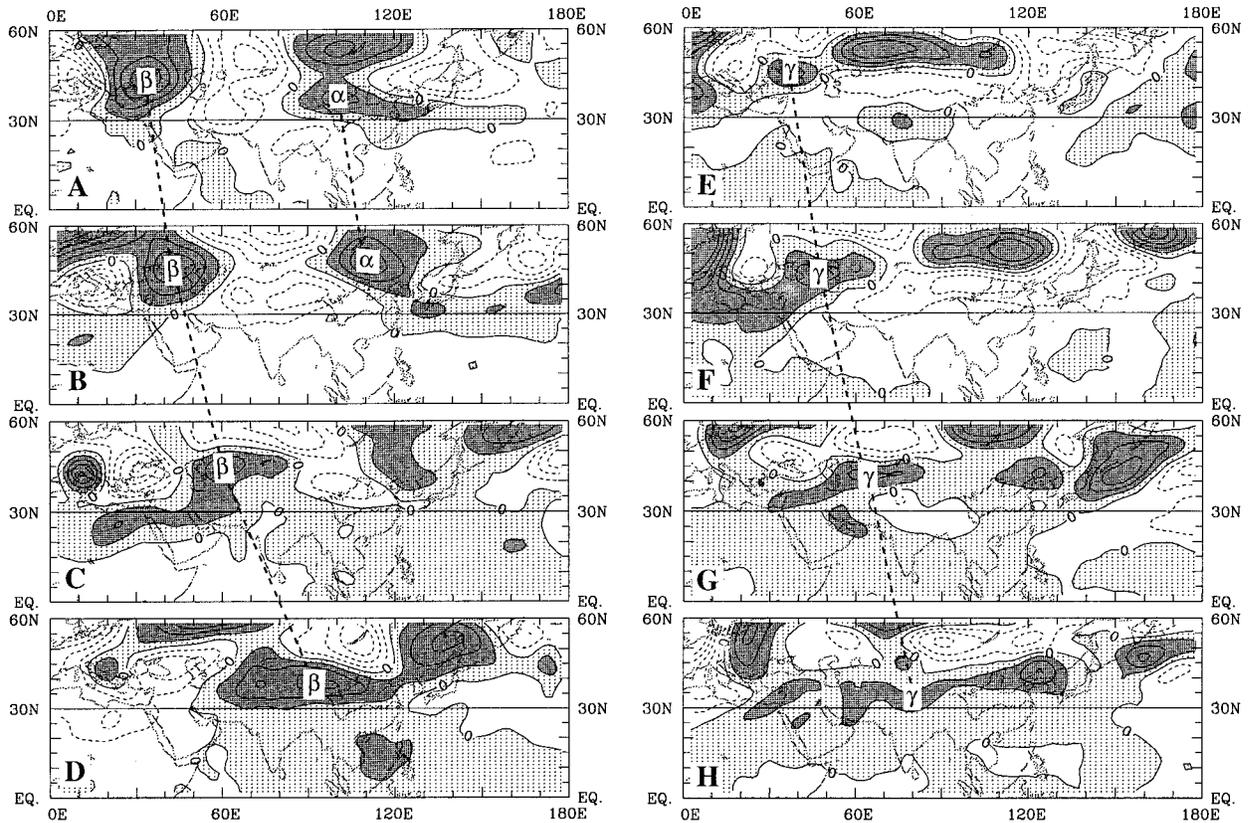


FIG. 11. Distributions of the 200-hPa filtered temperature in 1989 at different phases of the low-frequency oscillation as shown in Fig. 10. The data used have been processed by a 15–25-day bandpass filter. The interval is 1°C. Stippling denotes warm anomaly; and heavy stippling, warmer than 1°C: (a) 7 May, (b) 14 May, (c) 17 May, (d) 22 May, (e) 30 May, (f) 4 June, (g) 11 June, and (h) 15 June.

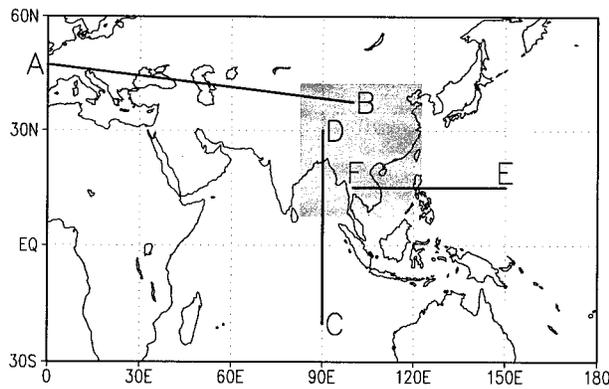


FIG. 12. The locations of the East Asian monsoon area (EAMA) as defined by the block, and the cross sections *AB*, *CD*, and *EF* for presenting the propagation of different low-frequency oscillations as shown in Figs. 13 and 14.

gation of the warm TTO α requires further work and will not be discussed here. The latter may result partly from the arrival of the warm phase of the TTO, and partly from the large latent heat release due to condensation since it appears at the time typhoon 8903 intruded into the SCS and the SCS monsoon onset occurred, and since such kind of δ -function-type warming may be decomposed into different harmonics.

2) Before the East Asian monsoon onset, the temperature variation in the Tropics east of the meridian 90°E is in phase with that over the Tibetan Plateau. When the TTO of the 200-hPa (or 300-hPa, or 500-hPa, figures not shown) temperature over the plateau evolves from its valley (B) to its crest (D), the upper-tropospheric temperature over the tropical region east of 90°E also increases.

3) The warm ridge that controls the area over the Tibetan Plateau and northern China during the monsoon onset (phase D) can be traced backward to the European origin. At phase A, while the previous warm TTO α departs the Plateau, another warm center, β is just over the Mediterranean Sea. This warm ridge moves gradually east-southeastward with a speed of about 4° of longitude per day. From phase B onward, the air over the plateau and the western Pacific is getting warmer. By phase D when this warm ridge starts to invade the plateau area, the onset of East Asian monsoon occurs.

4) The warm ridge γ over the plateau at phase H, which is associated with the Indian monsoon onset, can also be traced backward to the European origin. The warm ridge γ over the Mediterranean Sea at phase E follows system β and propagates in a similar manner and at a similar speed.

In summary, during the spring transition period the low-frequency oscillation TTO of the upper-tropospheric temperature dominates the extratropical regions. In early May, when the warm ridge of the TTO mode develops over Europe and moves east-southeastward into

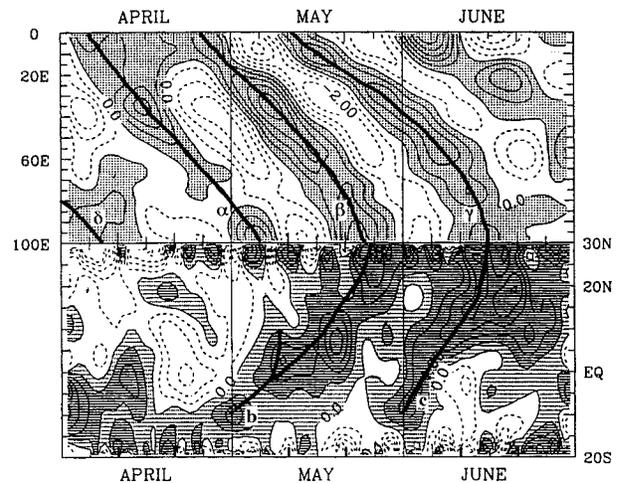


FIG. 13. The evolutions in the seasonal transition period of 1989 and along the *AB* cross section shown in Fig. 12 of the 200-hPa temperature, which has been proceeded by a 15–25-day bandpass filter (upper panel, interval in 1.0°C), and along the *CD* cross section of the 200-hPa divergence (lower panel, interval in $2.0 \times 10^{-6} \text{ s}^{-1}$).

the Tibetan Plateau area, favorable backgrounds for East Asian monsoon onset is then created.

b. 30- to 60-day oscillations of the tropical divergence

To see how the low-frequency oscillations affect the onset of the Asian monsoon, we choose a rectangular block bounded by the meridians 85° and 120°E and the latitude circles 10° and 40°N as shown in Fig. 12. The eastern part of the Tibetan Plateau, and the main parts of the SCS and the Bay of Bengal are all within the block. The onsets of the BOB monsoon and SCS monsoon both occur within this block as well. It can be regarded as the most sensitive region for the East Asian monsoon onset, and defined as the “East Asian monsoon area” (EAMA). The cross sections *AB*, *CD*, and *EF* are made to present how the low-frequency oscillations affect the region from its north, south, and east directions, respectively. No obvious low-frequency oscillations from the west were observed to enter the region. This is possibly because the northern part of the western boundary is occupied by the plateau, and its southern part is on the downwind side of the surface anticyclone that exists over the Arabian Sea before the onset of the Southeast Asian monsoon.

In the lower panel of Fig. 13, we present the Hovmöller diagrams of the horizontal divergence at 200 hPa that is averaged over the longitude range from 85° to 95°E and along the *C* (20°S , 90°E) to *D* (30°N , 90°E) cross section shown in Fig. 12. From April to June, there are three main divergence centers that develop in the southern Tropics, respectively, in early April, May, and June, with a period of about one month. The latter two centers are coupled with lower-tropospheric con-

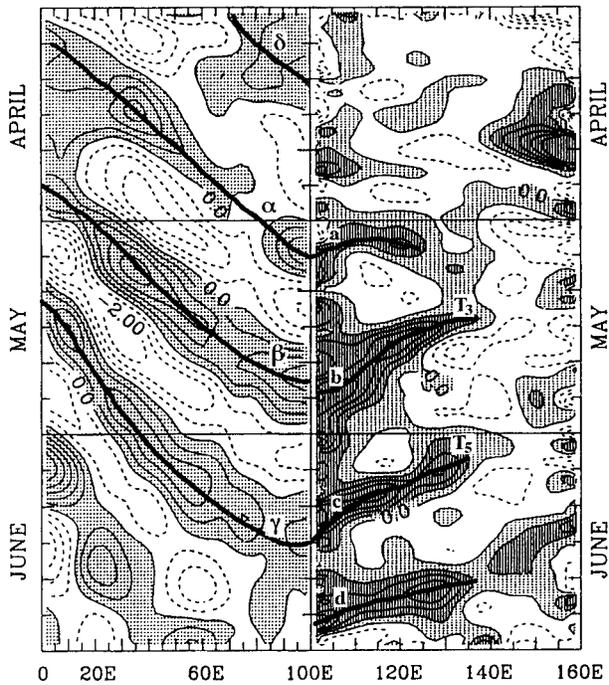


FIG. 14. (Left) Same as the upper panel of Fig. 18. (Right) The evolution in the seasonal transition period of 1989 and along the *EF* cross section shown in Fig. 12 of the 200-hPa divergence averaged meridionally from 10° to 20°N . Interval is $2.0 \times 10^{-6} \text{ s}^{-1}$. Stippling indicated divergence. T3 and T5 denote typhoon 8903 and typhoon 8905, respectively.

vergence (figures not shown). These centers propagate northward toward the Northern Hemisphere at a speed of about 1.2° – 1.5° of latitude per day. Since the MJO is characterized by lower-layer convergence and upper-layer divergence, and since their arrival provides favorable conditions for the development of strong convection and heavy rainfall, it can be defined as the rising phase of the MJO. The arrival of the main rising phase of the MJO in the Andaman Sea in early May is in correspondence with the BOB monsoon onset. As discussed in the previous sections, the BOB monsoon onset then generates a favorable environment for the SCS monsoon onset. As the MJO reaches the southern boundary at 10°N of the East Asian monsoon area, the SCS monsoon onset then occurs. From Fig. 13 it becomes prominent that the onset of the East Asian monsoon in the middle of May and the onset of the South Asian monsoon in early June are both associated with the arrival of the MJO at the latitude belt 10° – 20°N .

c. Two- to three-week oscillations of the tropical divergence

In the right panel of Fig. 14, is shown the evolution of the divergence at 200 hPa averaged between 10° and 20°N and along the *E*(160°E) to *F*(100°E) cross section as indicated in Fig. 12. The corresponding evolution at

850 hPa is similar but with opposite signs, and not shown here. From May to June, there are four strong divergence perturbations (a–d) that appear over the western Pacific and then propagate westward, appearing as another kind of TTO. Since these strong upper-layer divergence centers are coupled with strong lower-layer convergence, the passage of such TTO always causes vigorous upward motion and torrential rain. It has been identified from weather charts that TTOb and TTOc correspond respectively to typhoon 8903 (T3) and 8905 (T5). TTOb reaches the SCS region and becomes intensified there by 20 May, the time of the East Asian monsoon onset. TTOc propagates at a speed of about 3.5° of longitude per day. It reaches the eastern coast of the Bay of Bengal by 10 June when the South Asian monsoon onset occurs.

d. Phase locking of low-frequency oscillations and monsoon onset

The upper panel of Fig. 13 and the left panel of Fig. 14 present the eastward propagation of the TTO mode of the upper-tropospheric temperature along the *A* (45°N , 0°) to *B* (37.5°N , 100°E) cross section as shown in Fig. 12. They summarize the propagation feature of the TTO shown in Fig. 11. From April to June, there are four warm surges of the TTO and each propagates from Europe and invades the plateau area. The first warm surge that reaches the plateau region in the first half of April does not match the northward propagating MJO and the westward propagating TTO of the tropical divergence, and monsoon onset does not appear. However, the second, third, and fourth warm surges over the plateau are in phase with the arrivals of the northward propagating MJO at the southern boundary (10°N) of the EAMA (Fig. 13, lower panel) and the westward propagating TTO at the eastern boundary of the EAMA (Fig. 14, right panel). The MJO and TTO have different oscillation periods by nature. To understand why they can be phase locked in the EAMA, consider the divergence at 200 hPa of more than $2 \times 10^{-6} \text{ s}^{-1}$ as a presentation of strong rising phase. Then from Fig. 13 we see that the lifetime of the strong rising phase of this MJO is short in the Southern Hemisphere but rather sustained while propagating toward the EAMA. At its southern border, the strong rising phase of the MJO persists from early May to 22 May, and that of the MJOc persists for the whole of June. The persistency of the strong rising phase of this MJO thus provides the possibility for the phase-locking of it with the two TTOs. In other words, all the favorable phases of these low-frequency oscillations arrive in the EAMA at about the same time. They act together in the region and cause vigorous development of strong convection and torrential rain over large areas. The BOB, SCS, and South Asian monsoon onsets thus occur correspondingly. It can therefore be concluded that the timing of monsoon onset is determined by the phase locking within the

EAMA of different kinds of low-frequency oscillations from different directions.

6. Conclusions and discussion

It has been confirmed in this case study that during the seasonal transition of 1989 there exists complete reversal in thermal contrast between land and sea surface in the Asian subtropical region. Large surface latent and sensible heat fluxes are observed over the western Pacific during winter months but over the Tibetan Plateau in summer months. The changes in latent heat flux are more or less accompanied with the onset of the Asian monsoon. However, the maximum center of surface sensible heat flux over the western and central Tibetan Plateau is established in early spring. It warms up the atmospheric column aloft by several degrees Celsius per day. Lower-layer air in the surrounding area of the plateau are then sucked toward the plateau. The north to south gradient of the upper-tropospheric temperature to the south of the plateau is gradually reversed, in favor of the development of tropical easterlies. All these provide suitable conditions for the seasonal transition of the circulation in East Asia and the onset of a monsoon.

The onset of the SCS monsoon in 1989 was by 20 May, and the onset of the South Asian monsoon was about 20 days later. The reasons that caused the monsoon onset in East Asia leading that in South Asia were explored. Both the thermal forcing and mechanical forcing were proven to be important factors in causing the earlier occurrence of the SCS monsoon. Thermally, the strong surface sensible heating of western and central Tibet in collaboration with the westerly advection of temperature makes the air temperature over the eastern plateau warmer than that over the western plateau. Strong southerly inflow toward the plateau in the spring season thus occurs first in the eastern part of the Bay of Bengal. Mechanically, the orographically deflected westerly forms low-layer anticyclonic flow over the Arabian Sea and South Asia subcontinent and cyclonic flow over the Bay of Bengal. Deep convection and strong southerlies thus develop vigorously in the eastern front of the Burma trough, and a lower-layer low and an upper-layer high are observed near Rangoon. A low value OLR center appears first in this region in the latitude belt from 10° to 20° N. This is defined as the onset of the BOB monsoon. After the BOB monsoon onset, the upper-layer anticyclone, which is centered near Rangoon, drives strong northerlies toward the Southern Hemisphere over the SCS region. Strong lower-layer return flows of southerlies and southwesterlies then develop in the region. The onset of the SCS monsoon thus occurs. After this onset, the warm South Asian high develops over the Tibetan Plateau, and the monsoonal meridional circulation with southerlies in the lower layer and northerlies aloft shifts westward to the Arabian Sea. The surface anticyclone circulation that occupied the Arabian Sea during the onset of the East

Asian monsoon has retreated to the north of the Arabian Peninsula. Cyclonic circulation then prevails over the Arabian Sea, and the onset of the South Asian monsoon occurs. The overall structure discussed in this paragraph has been confirmed by examining the weather data of the last 14 yr and will be published in a parallel paper.

Based upon these results, we suggest that the whole procedure of the outbreak of the Asian monsoon is better divided into three stages. The first is the BOB monsoon onset, followed by the SCS monsoon onset, and the consequent Indian monsoon onset. Such division provides linkage of the Asian monsoon onset to the Tibetan Plateau forcing more closely. It is due to the thermal as well as mechanical forcing of the Tibetan Plateau that the BOB monsoon onset occurs, and the SCS monsoon onset is then induced.

Another important point revealed in this study is about the timing of the monsoon onset. It is proved that monsoon onset occurs when different kinds of the low-frequency oscillations are in phase in the EAMA region. When the warm phase of the TTO of midlatitude upper-layer temperature, which is originally over Europe, arrives in the EAMA concurrently with the arrival of the rising phase of the northward propagating MJO of the tropical divergence, which originates in the Southern Hemisphere, favorable background for the development of atmospheric overturning over a large area is then created. When the tropical TTO of strong convection that originates over the western tropical Pacific and propagates westward reaches this EAMA region, vigorous convection and torrential rain are triggered, and the onsets of the Asian monsoon occur. In other words, the onsets of the Asian monsoon occur when the favorable phases of different kinds of the low-frequency oscillations are locked over the EAMA region.

All the calculations performed in this study have been repeated by using the data for 1988 rather than 1989 and very similar results have been obtained. These will be published in a separate paper. However, since the results obtained here are based on a case study, they should be considered preliminary, and further research is required. Nevertheless, it is still not clear how the MJO of the divergence develop in the Southern Hemisphere. Is it merely a tropical phenomenon, or generated in the southern middle latitudes? Tao et al. (1983) analyzed the low-frequency characteristics of the upper-layer zonal wind component at middle latitudes in the Southern Hemisphere and the difference in tropospheric temperature between 25° and 45° S in the period from May to July of 1979. Besides the most developed TTO, the MJO was identified in the locations of 100° and 120° E. However, how this Southern Hemispheric midlatitude MJO is related to the Asian monsoon is unclear. Another unsolved question is associated with the mechanism that drives the MJO propagating northward to the monsoon region. Is it the lower-tropospheric return flow from the winter hemisphere to the summer hemisphere, or the thermal contrast between land and sea? In fact,

there must be other problems we have not thought of. In any circumstances, to further understand the monsoon dynamics, we need to consider the monsoon a spectacular phenomenon that occurs in a system composed of the atmosphere, hydrosphere, and lithosphere, and under the interaction among these subsystems.

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