STUDY ON THE VARIATION IN THE CONFIGURATION OF SUBTROPICAL ANTICYCLONE AND ITS MECHANISM DURING SEASONAL TRANSITION—PART III: THERMODYNAMIC DIAGNOSES*

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ABSTRACT

The mechanisms for the variation in the configuration of subtropical anticyclone during seasonal transition are explored from energy budget using the NCEP/NCAR reanalysis data. Based on the seasonal variations of temperature and heating fields, it is found that the significant diabatic heating associated with spring precipitation over southern China has impacts on subsequent Asian seasonal transition. The reversal of meridional temperature gradient in the vicinity of the WEB (westerly-easterly boundary) in the middle and upper troposphere also depends on the latitudinal position where temperature ridge locates. The northward shift of the warm temperature ridge results from the fact that the local temperature increase to the north of the WEB is more than that in its vicinity. The diagnostic results through thermodynamic equation show that physical mechanism responsible for seasonal transition is different from area to area over the Asian monsoon region. The dominant factors responsible for northward shift of the Bay of Bengal warm ridge are the meridional temperature in initial stages of the onset and the descending motion after the onset. The factors for causing the northward jump of the South China Sea warm ridge involve the zonal temperature advection, meridional temperature advection, and diabatic heating associated with the southern China spring rainfall. The subsidence is the factor leading to the northward migration of the South Asia warm ridge.

Key words: seasonal transition, meridional temperature gradient, thermodynamic equation, mechanism

I. INTRODUCTION

The preceding two parts (Mao et al. 2003; 2004) of this series study have substantiated that climatologically, the Asian summer monsoon onset consists of three consequential stages. The first onset occurs over eastern Bay of Bengal (BOB) in the first two pentads of May, the second over the South China Sea (SCS) in the fourth pentad of May, and the third over the India and Arabian Sea (IDO) in the first two pentads of June. The monsoon onset over each of these three areas is closely related to the break of ridgelines of subtropical high and the establishment of local seasonal transition axis (STA). Why is the summer monsoon onset earlier over Southeast Asia than over India?

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What are the mechanisms responsible for the Asian monsoon onset? Many efforts have been made to explore these questions. Some researchers (e. g. Ye et al. 1957; Flohn 1957; Murakami and Ding 1982; Yanai et al. 1992; Yanai and Li 1994; Li and Yanai 1996) have emphasized the thermal effects of the Tibetan Plateau as an elevated heat source on the Asian monsoon onset. Lau and Yang (1996) reported that the larger-scale forcing (such as thermal contrasts between Eurasian Continent and adjacent oceans) causes the monsoon onset over Southeast Asia. Wu and Zhang (1998) pointed out that the thermal and mechanical forcing of the Tibetan Plateau determines the timing and location of monsoon onset. Ueda and Yasunari (1998) found that the air temperature over Tibetan Plateau rises in the form of low-frequency oscillation with a 15-day period, and the thermal contrast between the Tibetan Plateau and its neighboring oceans may induce the acceleration and eastward extension of the low-level westerlies, leading to the Southeast Asian monsoon onset. Liu (2000) reported that the Tibetan Plateau heating expedites not only the local seasonal variation but also the seasonal transition in the Northern Hemispheric circulation. Song et al. (2000) accentuated the compositive impacts of the Tibetan Plateau and the Indochina heating source on the Southeast Asian monsoon onset.

He et al. (1987) showed that in 1979 the Asian monsoon circulation underwent two abrupt stages, which corresponded to the establishment of summer monsoon over Southeast Asia and India, respectively. These two abrupt changes are closely related to the successive warming in the mid-upper troposphere east of 85°E (eastern Tibetan Plateau and China mainland) and west of 85°E (western Tibetan Plateau and Iran) respectively, while Lau and Yang (1996) and Krishnakumar and Lau (1998) mentioned that symmetric instability of basic flow may trigger the monsoon onset. Undoubtedly, the Tibetan Plateau as an elevated heat source can warm local air column during seasonal transition. Since the monsoon onset depends on the reversal of meridional temperature gradient (MTG) near the ridge-surface of subtropical high (defined as WEB in Part I), how and to what extent does the Tibetan Plateau heating influence on the temperature in the vicinity of the WEB? In this part, the thermal processes in the mid-upper troposphere are investigated quantitatively based on thermodynamical equation, and the physical mechanisms associated with seasonal abrupt change are explored from the perspective of energy.

II. DATA AND METHODOLOGY

1. *Data*

The daily data used are same as in the preceding two parts (Mao et al. 2003; 2004), which are the NCEP/NCAR reanalysis products including the wind, geopotential height, and temperature fields from 1980 to 1998. Also employed are the monthly mean heating rate data at 28 sigma surfaces with a Gauss grid of 1.875° latitude $\times$1.875° longitude from 1958 to 1998, which involve deep convection, shallow convection, large-scale condensation, longwave radiation, solar short radiation, and vertical diffusion.

2. *Thermodynamical Equation*

$$c_p \frac{dT}{dt} - \frac{1}{\rho}\frac{dp}{dt} = Q, \tag{1}$$

$$\frac{\partial T}{\partial t} + V_n \cdot \nabla T + \omega \frac{\partial T}{\partial p} - \frac{\omega}{\rho c_p} = \frac{Q}{c_p}. \tag{2}$$

Equation (2) can be rewritten as

$$\frac{\partial T}{\partial t} = - V_h \cdot \nabla_h T - \left(\frac{P}{P_0}\right)^{\kappa} \omega \frac{\partial \theta}{\partial p} + \frac{Q}{c_p}, \tag{3}$$

where θ is potential temperature, $\kappa = \frac{R}{c_p}$, and the subscript h denotes the horizontal component. The other symbols in equations are commonly used in meteorology. The diabatic heating Q/c_p can be retrieved using Eq. (3) (Nitta 1972; Yanai et al. 1973).

In this paper, the relative importance regarding zonal temperature advection, meridional temperature advection, vertical temperature advection, and diabatic heating in seasonal transition processes is examined quantitatively. The physical mechanism associated with monsoon onset can be well understood based on the energy budget.

III. SEASONAL VARIATION IN TEMPERATURE FIELDS

During monsoon onset, in addition to a reversal of wind direction, the atmospheric meridional temperature gradient (MTG) in the vicinity of the WEB changes from negative to positive, which is the reason why we proposed to use the MTG to define the monsoon onset.

Figure 1 shows the climatoliogical mid-upper tropospheric (200 − 500 hPa) temperature fields. In April, a warm temperature ridge (thick dashed line) is close to the equator, and the isothermal distributions indicate that the direction of the MTG is from the polar region to the equator, winter monsoon therefore remains prevailing. In May, a warmer center over the Indochina Peninsula occurs south of the Tibetan Plateau. Because the warm temperature ridge migrates northward and reaches 20°N, in the north side of the warm temperature ridge, the MTG remains negative, while in the south side of the warm ridge, the MTG becomes positive, which corresponds to the break of ridgelines at low-levels over the BOB extending to the SCS (see Fig. 2c in Mao et al. 2003). But the warm temperature ridge over the IDO keeps to south of 10°N, with no monsoon bursting here. By June, monsoon bursts over the India and Arabian Sea as well as east of Philippines when the entire warm ridge jumps over 20°N. Note also that the thermal contrast between the Asian Continent and the Pacific Ocean becomes stronger compared to that in May.

The latitudinal location and configuration of the warm temperature ridge show that during the northward migration of the warm ridge, it does not move parallelly as a whole, there are large differences between longitudes in which the warm ridge over the Indochina Peninsula migrates fastest. The break of the ridgelines is associated with the reversal of the MTG over this region in May. Since the start date of the seasonal transition is the day when the warm temperature ridge moves to the same latitude where the WEB locates, as long as causes and its regional differences are made clear, we will realize the physical mechanism for the seasonal transition.

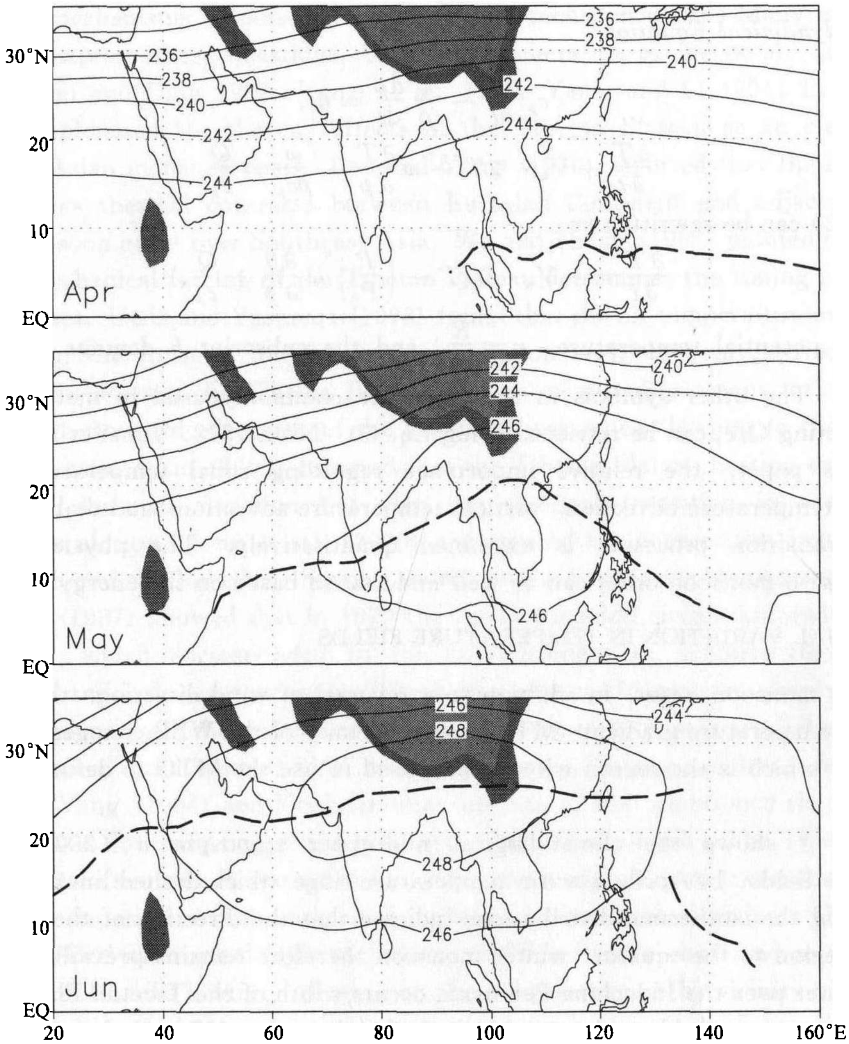

Fig. 1. Climatological mid-upper tropospheric (200 — 500 hPa) temperature fields (K) for April to June. Thick dashed lines denote the warm temperature ridgeline. Shaded areas indicate the terrain above 1500 m.

IV. HORIZONTAL DISTRIBUTIONS OF ENERGY BUDGET AND THEIR SEASONAL VARIATION

The energy driving the atmosphere motion comes from the solar radiation. Most of the radiation is first accepted by the earth's surface, and then supplies to the atmosphere in different forms. The various weather and climate phenomena that take place on the earth are related to the budget and transform of the energy. The atmospheric heat source and sink play the determining roles in large-scale monsoon circulation driven by thermal forcing. Besides, the heating fields also influence the variation of the temperature fields. Therefore, the distributions and seasonal variations of the heating fields are necessary to be analyzed.

The distributions of vertically integrated sensible heating (not shown) indicate that in

March, positive heating exists over large continents south of 40°N in which the heating exceeds 50 W/m^2 over the two peninsulas (Indochina and Indian Peninsulas). In April, sensible heating increases significantly over the Tibetan Plateau, Indian Peninsula, and Arabian Peninsula, while the heating decreases over the Indochina Peninsula and eastern China. In May, sensible heating over the Eurasia in middle and high latitudes increases as compared to that in the last month, the heating magnitude reaches 100 W/m^2 over western Tibetan Plateau and Mongolia region. In June, the distribution of the heating is similar to that in May, except for evident reduction over the southeastern Indian Peninsula. From spring to summer, sensible heating tends to decrease over the Indochina and Indian Peninsulas, but it tends to increase over the Tibetan Plateau.

Figure 2 shows that the spatial and temporal variations of vertically integrated latent heating tend to be opposite to those of sensible heating. The area with large latent heating values migrates northward over the oceanic surface from March to June, while the latent heating increases gradually over the Indochina Peninsula, eastern and southern Tibetan Plateau. In May, the larger latent heating exceeds 100 W/m^2 over the Southeast Asia due to the BOB and SCS summer monsoon onset. After the Indian summer monsoon bursts in June, the higher heating center with 300 W/m^2 occurs over the western Indian Peninsula.

In particular, the latent heating reaches 100 W/m^2 over the southern China in March, and increases to 200 W/m^2 in April which results from the spring rainfall over this region. Tian (1998) investigated the climatological characteristics and formation mechanism of the spring rainfall, but did not state its influences on the subsequent summer monsoon activities. We consider that this unique heating source has impact on the Asian seasonal transition, and may be an important factor that results in the regional differences of the Asian summer monsoon onset.

The distributions of total diabatic heating (Fig. 3) show that in March, negative heating less than 100 W/m^2 exists only over three Bays, namely, the SCS, BOB, and Arabian Sea because of smaller sensible and latent heating and lager radiative cooling. Positive heating over western Pacific and south equatorial Indian Ocean reflects the contribution from the latent heating. In April, as compared to Fig. 2, dominant component of positive total diabatic heating is the condensation latent heating over the southern China and Indochina Peninsula, although the diabatic heating exhibits positive over the Indian Peninsula, its magnitude is not comparable to that over the former. In May, the diabatic heating is stronger over the southeastern Tibetan Plateau than over the SCS. In June, the distributions of the diabatic heating are similar to those of latent heating. It is found from Fig. 3 that although the earth-atmosphere system accepts the same solar radiation at the same latitude for every month, the energy that the atmosphere obtains is different from the different longitudes due to different physical properties of the earth's surface. Since positive diabatic heating can warm air temperature, both the significant latent heating resulting from precipitation condensation over the southern China and Indochina Peninsula in April and positive diabatic heating over the Tibetan Plateau in May are favorable for the formation of high temperature center over the southeastern Tibetan Plateau.

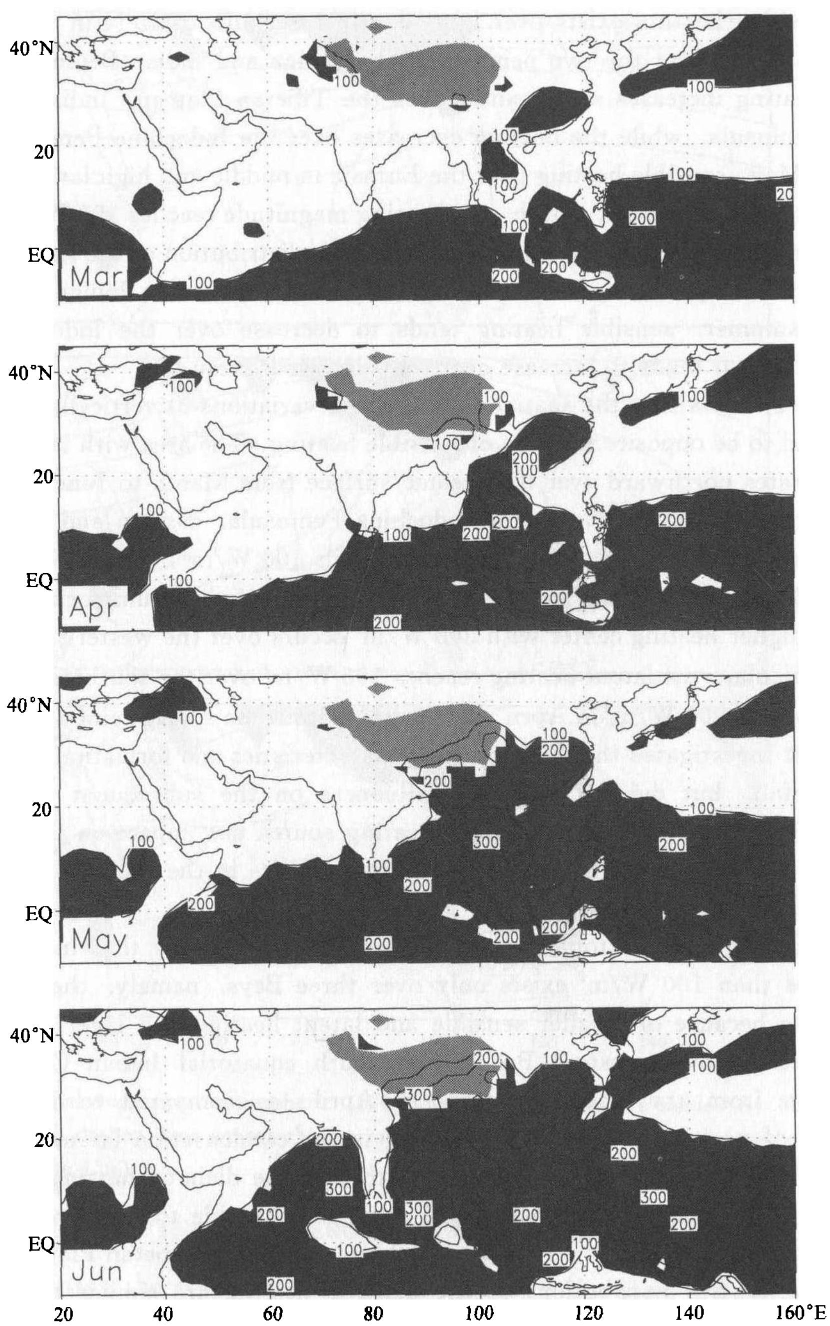

Fig. 2. Distributions of condensation latent heating integrated vertically (W/m^2) for March to June. Heavy shaded areas denote the Tibetan Plateau above 3000 m.

V. EVOLUTIONS OF CONVECTION AND TEMPERATURE FIELDS DURING SEASONAL VARIATION

Monthly mean temperature fields (Fig. 1) show that the warm temperature ridge locates the northernmost latitude over the Indochina Peninsula in May. In this section, the

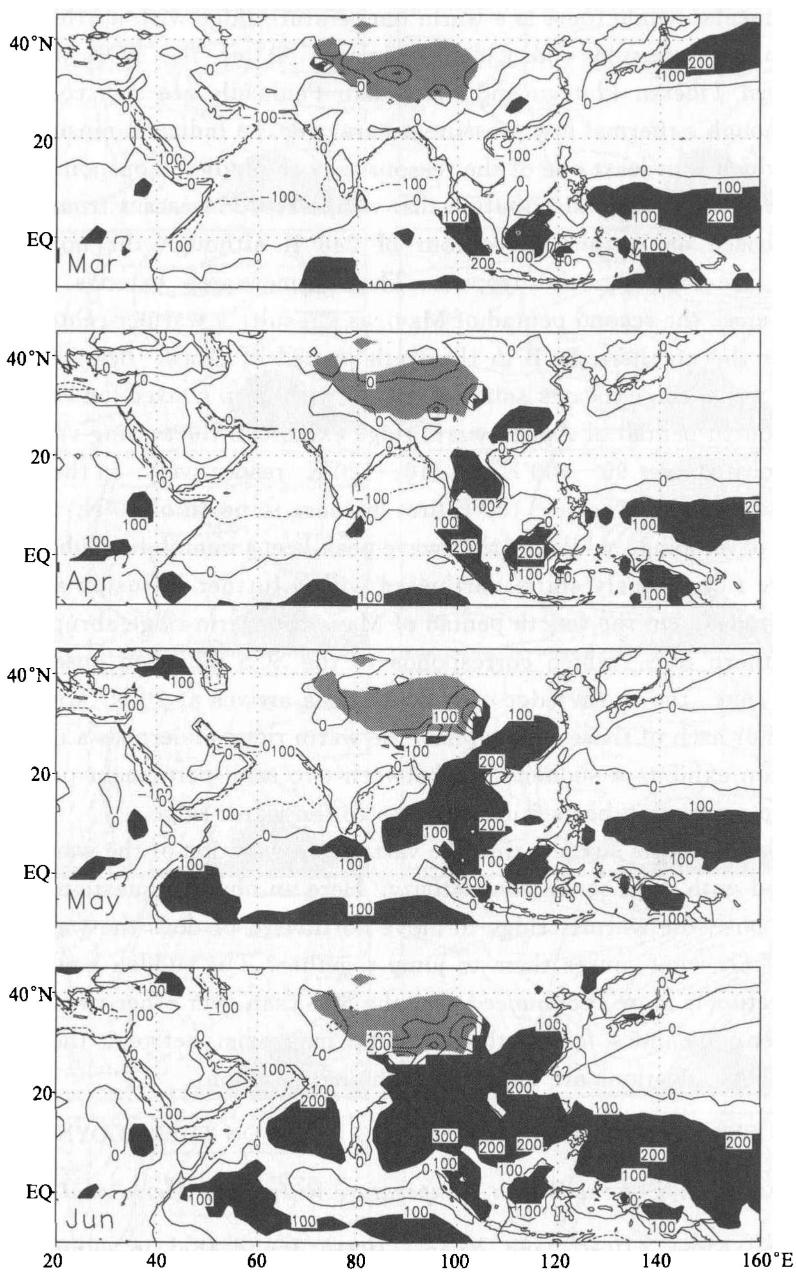

Fig. 3. As in Fig. 2, except for the total diabatic heating (negative values are denoted by dashed lines).

examination is focused on the differences and similarities among the BOB, SCS, and IDO based on pentad mean data.

The evolutions of mid-upper tropospheric (200—500 hPa) temperature fields (from the third pentad of April to the second pentad of June, not shown) indicate that prior to the first two pentads, a large cool temperature trough paralleling with the orographic fringe exists over the southwestern Tibetan Plateau extending to the east coast of the

Indian Peninsula, while there is a warm temperature ridge with north-south direction over the Indochina Peninsula and China mainland. After the BOB monsoon onset, the southwestern Tibetan Plateau and the Indian Peninsula are still controlled by the cool trough although a thermal low pressure occurs over the Indian Peninsula in low-levels (not shown), which is at least one of the reasons why the Indian monsoon bursts later than the BOB monsoon does. The distribution that temperature increases from east to west can be identified based on isothermal contour of 246 K although the horizontal temperature differences are small in the tropics. The warmer area extends westward and then northward since the second pentad of May, as a result, a warmer center higher than 248 K occurs over the northern BOB in the sixth pentad of April. Before the third pentad of April, the warm ridge locates south of 10°N, with deep convection existing east of 80°E. Since the fourth pentad of April, warm ridge exhibits a fluctuating variation, and the two peaks are located over 90—100°E and 110—120°E, respectively. In the first two pentads of May, the wave peak over 90—100°E first intrudes to north of 10°N, accompanied by deep convection developing, while another wave peak keeps immobile. Subsequently, the warm temperature ridge largely shifts northward with a further intrusion along 100°E than the other longitudes. On the fourth pentad of May, the warm ridge abruptly shifts from 10°N to the northern SCS, which corresponds to the SCS monsoon onset. By the first two pentads of June, the warm ridge over South Asia arrives at 20°N, with the IDO monsoon bursting. For each of these three areas, the warm ridge undergoes a northward jump, and this variation exhibits a sudden shift between two semi-permanent positions of the warm ridge. Deep convection has similar features to the warm ridge.

The above results suggest that the variation in position of the warm temperature ridge is associated with deep convective activity. Here an obvious question arouses: Does deep convection push the warmer ridge to move northward or does the warmer ridge pull deep convection? Or what causes them to jump together? The sudden jump of warm ridge and deep convection is more pronounced over the SCS than over other two areas, which implies that different dynamics for northward jump may exist between the different longitude regions. These questions are explored in the next section.

VI. DIAGNOSES FOR REVERSAL OF THE MTG BASED ON THERMODYNAMICAL EQUATION

1. *Northward Shift of the Warm Temperature Ridge and Meridional Circulation*

Lin and Miao (1981) and Yuan (1981) found that in summer, the northward migration of the subtropical high is related to the evolution of the meridional circulation. Since the pressure (geopotential height) field depends on the temperature, the structure and variability of the temperature field are certainly associated with the evolution of the meridional circulation. The roles of meridional and vertical temperature advections played in local temperature tendency can be revealed using the meridional circulation.

Figure 4 shows the pressure-latitude cross sections of the zonal wind and meridional circulation (left panels) as well as local temperature tendency and warm ridge during the BOB monsoon onset. The striking features seen from Fig. 4 (right panels) are that the almost vertical warm ridge (200—500 hPa) suddenly jumps from south side to north side

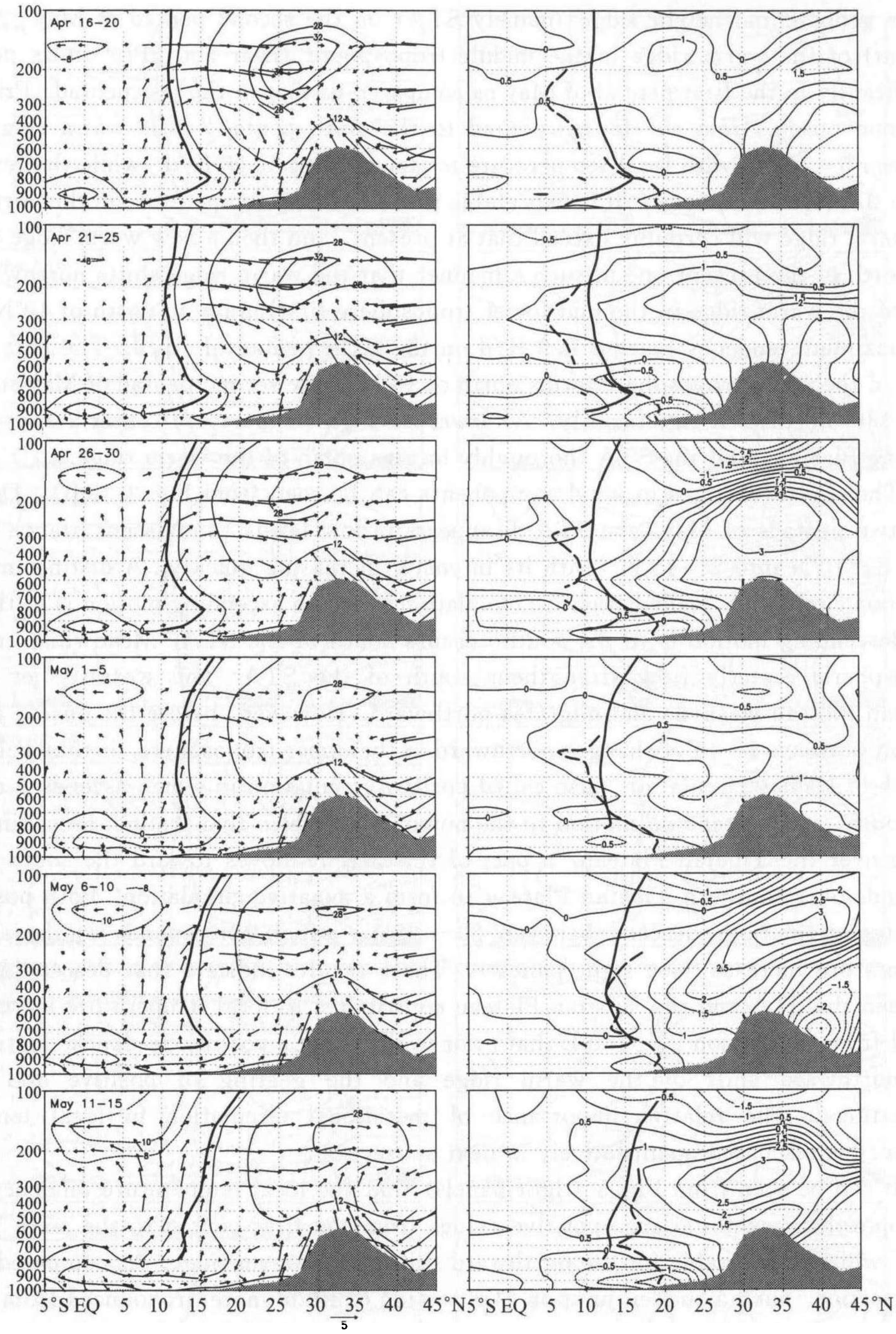

Fig. 4. Pressure-latitude cross sections (90—100°E) of the zonal wind (contour, m/s) and meridional circulation (left panels) as well as local temperature tendency (0.1 K/d) and warm temperature ridge (right panels) during the BOB monsoon onset (the fourth pentad of April to the third pentad of May). Thick solid lines denote geopotential height ridge. Thick dashed lines denote warm temperature ridge. Shading indicates orographic altitude. Vectors denote meridional circulation (m/s, —0.02 Pa/s).

of the geopotential height ridge (namely STA) on the second pentad of May. Actually, the part of the warm ridge in the middle troposphere (near 500 hPa) shifts northward significantly in the first pentad of May as compared to that in the last pentad. Prior to the monsoon onset (from the fourth pentad to the sixth pentad), the warm ridge locates between 5—10°N, with local temperature tendency being 0.05 K/d, while the tendency is above 0.1 K/d to its north. It is expectable that the actual temperature in the north side of the warm ridge will certainly exceed that at present, and then a new warm ridge occurs to the north of the present one in such a manner that the warm ridge shifts northward. The part of the warm ridge in the mid-lower troposphere first jumps to north of 10°N because the maximum tendency reaches 0.3 K/d on the sixth pentad of April. The part of warm ridge in the upper troposphere forms north of 18°N in the second pentad of May so that the STA tilts northward significantly. Afterwards, high temperature center stands over the Tibetan Plateau, and the STA thoroughly locates north of the warm ridge.

The abrupt changes in wind components can be seen from Fig. 4 (left). During the first two pentads of May, the STA disappears in low-levels, and ascent branch broadens from 5—10°N into 5—15°N, with its intensity increasing quickly. A distinct meridional monsoon circulation (called negative circulation in which ascending motion is to the north, and descending motion is to the south) occurs south of the STA. Meanwhile, the upper tropospheric easterly flow strengthens south of the STA, and westerly jet over the Tibetan Plateau weakens and migrates northward. However, before the onset, ascending motion between 5—10°N changes northward in the upper troposphere, it thus belongs to a branch of Hadley circulation (also called positive circulation in which ascending motion to the south, and descending motion to the north). Note also that distinct ascending motion exists over the Tibetan Plateau, a part of this ascent moves toward the south and then descends over southern Tibetan Plateau to form a negative circulation. Both positive and negative circulations descend between 20—25°N, which corresponds to the significant temperature increase (see right panels). These results indicate that descending motion between the STA and the Tibetan Plateau contributes to local temperature increase. It is found from the section 90—100°E that prior to the onset a positive feedback exists between the northward shift of the warm ridge and the gearing of positive and negative circulations. The relative importance of meridional circulation in local temperature tendency is discussed quantitatively in next subsection.

It can be seen from Fig. 5 (right panels) that the local temperature tendency in most of troposphere exhibits large positive values from the first pentad to the sixth pentad of May, which is favorable for the northward shift of the warm ridge. As compared with the BOB section, such a sudden jump in SCS section exhibits more pronounced from the third pentad to the fourth pentad of May. The feature associated with meridional circulation (left panels) is that the ascending branch abruptly occurs between 10—20°N to form a negative circulation (monsoon circulation), along with the increase of both upper tropospheric easterlies and lower tropospheric westerlies to the south of the STA. This monsoon circulation establishes under the circumstance that the vertical shear between the upper- and lower-troposphere zonal winds is not yet well developed. Lau and Yang (1996) and Krishnakumar and Lau (1998) suggested that this abruptness might be caused

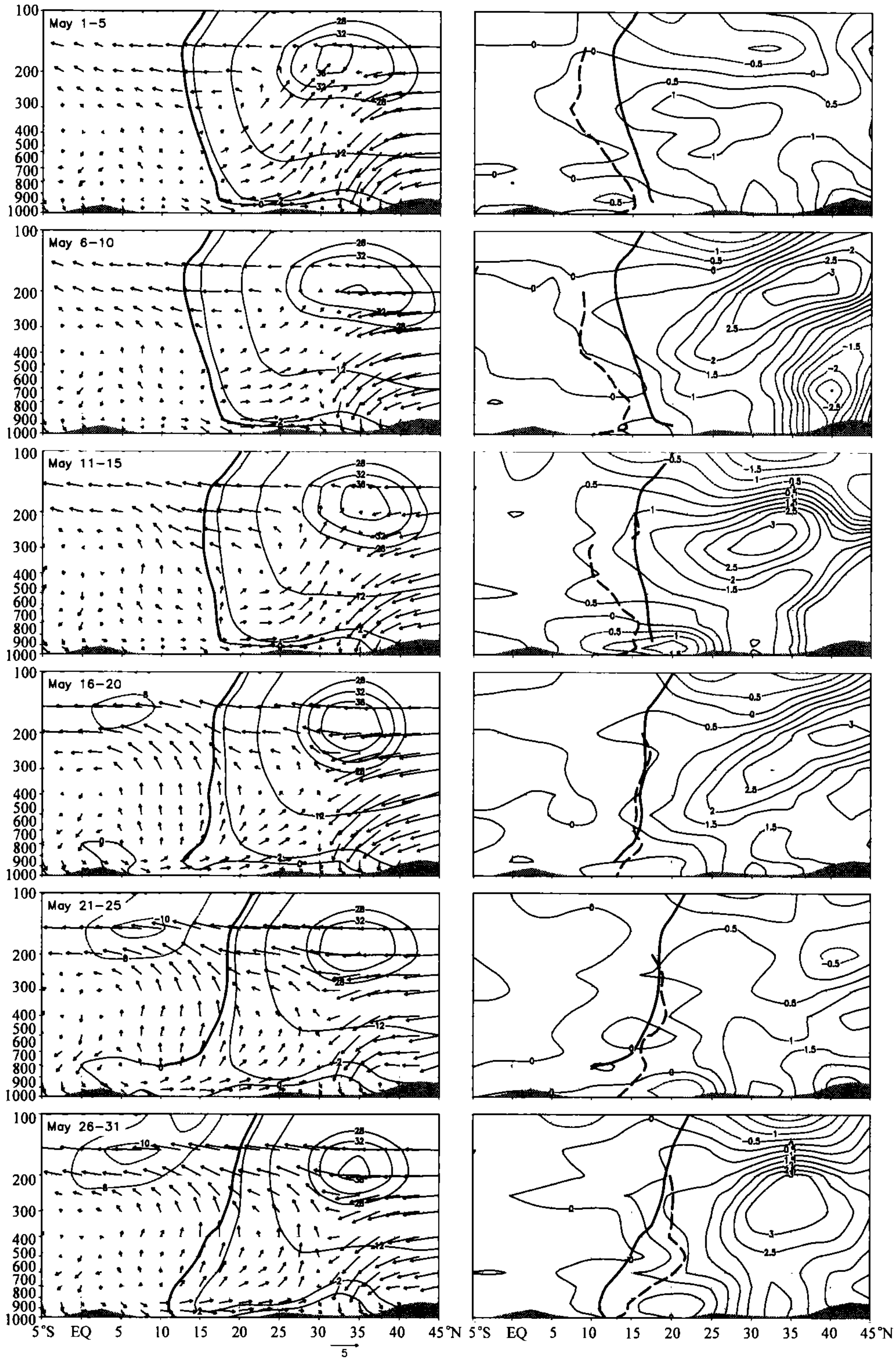

Fig. 5. As in Fig. 4, except for cross section (110—120°E) and the SCC monsoon onset (the first pentad of May to the sixth pentad of May).

by the symmetric instability in the basic flow. Note that the strongest ascending motion is located between 20—30°N, which corresponds to positive temperature tendency (see right panels). It is conceivable that local temperature increase is a positive correlation with the enhanced southerly flows, and these southerlies transport northward a lot of sensible heating, leading to temperature rise over the north side of the STA. The latitude-time cross sections (110—120°E) of precipitation, meridional wind, and southwesterlies given by Jiang (2001) can well illustrate the above features (see her Figs. 2. 10—2. 14). The evolution of the rainfall starting from the second dekad of March is well consistent with that of strong southerlies and southwesterlies. Of course, there are other processes such as zonal temperature advection and condensation latent heating (see Fig. 7) that cause the temperature increase over the north of the STA.

The situation for the South Asia section (60—85°E, not shown) indicates that a temperature increase center exists over 300 hPa and 30—35°N during monsoon onset. The ascending motion shifts from 5°N to 10—15°N, with the STA migrating northward. It is affirmed that the positive temperature tendency mainly results from the descending motion over the north of the warm ridge.

Thus it can be seen that the position variation of warm temperature ridge depends on the local temperature tendency, and the northward shift results from the temperature increase rate more over the north of the warm ridge than over its vicinity. The contributions of meridional circulations to the local temperature increase over the north of the STA are different from area to area. The gearing between positive and negative circulations is one of the factors that cause temperature increase over the northern BOB. The meridional warm advection caused by the southerlies between 20—30°N is favorable for temperature increase over the northern SCS. The subsidence of positive circulation is a dominant factor for temperature increase over the northern IDO.

2. *Relative Importance of Factors Influencing Local Temperature Tendency*

The above discussions show that the northward shift of the warm ridge depends on the local temperature tendency. The change of temperature field is determined by thermodynamical equation:

$$\frac{\partial T}{\partial t} = -u\frac{\partial T}{\partial x} - v\frac{\partial T}{\partial y} - \left(\frac{p}{p_0}\right)^{\kappa}\omega\frac{\partial \theta}{\partial p} + \frac{Q}{c_p}. \tag{4}$$

Equation (4) indicates that the factors influencing the local temperature tendency include horizontal, vertical temperature advections, and diabatic heating.

To identify the relative importance of four terms in Eq. (4), we perform vertical average for Eq. (4) over the 200—500 hPa layer (He et al. 1987):

$$\overline{\frac{\partial T}{\partial t}} = -\overline{u\frac{\partial T}{\partial x}} - \overline{v\frac{\partial T}{\partial y}} - \overline{\left(\frac{p}{p_0}\right)^{\kappa}\omega\frac{\partial \theta}{\partial p}} + \overline{\frac{Q}{c_p}}, \tag{5}$$

where

$$\overline{(\ \ )} = \frac{1}{500-200}\int_{500}^{200}\overline{(\ \ )}\mathrm{d}p. \tag{6}$$

Thus, Eq. (5) represents layer-averaged local temperature change and its influencing factors in the form of heating rate. Again, the temporal integrals of Eq. (5) can clearly reflect the relative importance of the four terms.

$$T_L = T_0 + \int_0^L \overline{\frac{\partial T}{\partial t}} dt, \tag{7}$$

where let $T_0 = 0$, T_L then represents the cumulative temperature increase.

Similarly,

$$T_u = -\int_0^L \overline{u \frac{\partial T}{\partial x}} dt, \tag{8}$$

$$T_v = -\int_0^L \overline{v \frac{\partial T}{\partial y}} dt, \tag{9}$$

$$T_\omega = -\int_0^L \overline{\left(\frac{p}{p_0}\right) \omega \frac{\partial \theta}{\partial p}} dt, \tag{10}$$

$$T_Q = \int_0^L \frac{\overline{Q}}{c_p} dt, \tag{11}$$

where T_u, T_v, T_ω and T_Q are the cumulative heating due to zonal, meridional, vertical temperature advections, and diabatic heating, respectively.

Figure 6a shows that the warm temperature ridge initially (1 April) locates around 5°N, and reaches 10°N on 28 April because of lager cumulative temperature increase there. Subsequently, the cumulative temperature increase over the north of the warm ridge remains unchangeable (about 0. 9 K), the warm ridge then keeps immobile. As noted in previous section, the warm ridge jumps to the north of 15°N on 12 May. Now it is focused on the period 1 — 10 May, the cumulative heating from zonal temperature advection (Fig. 6b) is negative, and the cumulative heating from meridional advection (Fig. 6c) occupies 1 — 5 K. Adiabatic cooling derived from vertical advection (Fig. 6d) almost counteracts the diabatic heating (Fig. 6e). The temperature increase caused by the total horizontal advection is about 1—2 K, which just corresponds to the magnitude of local temperature rise. Therefore, the meridional temperature advection is the dominant factor responsible for the northward shift of the warm ridge, which is in agreement with the qualitative analyses regarding the meridional circulation in previous subsection.

For the SCS, the warm ridge still situates at 8°N before 11 May. During the sudden jump, the cumulative temperature increase near 5°N is 0. 8 K, while around 20°N it exceeds 3 K. Compared with the cumulative rate, both zonal and meridional advections have contribution to temperature rise near 20°N (Fig. 7f). The maximum of zonal temperature advection near 25°N reflects the influence of the Tibetan Plateau heating source. Also noteworthy is that the diabatic heating rate exhibits the larger positive values between 25—30°N (Fig. 7e), which indicates quantitatively the role of the southern China spring rainfall in the Asian seasonal transition.

Figure 8a shows that during the period 1 — 10 June, the cumulative temperature increase at 10°N keeps 1. 5 K, while around 20°N it reaches 5. 5 K. It can be seen that the descending motion is the main factor to cause the temperature rise over northern IDO (Fig. 8d).

In summary, the dominant factors responsible for northward shift of the BOB warm ridge are the meridional temperature in initial stages of the onset and descending motion after the onset. The factors for causing the northward jump of the SCS warm ridge involve

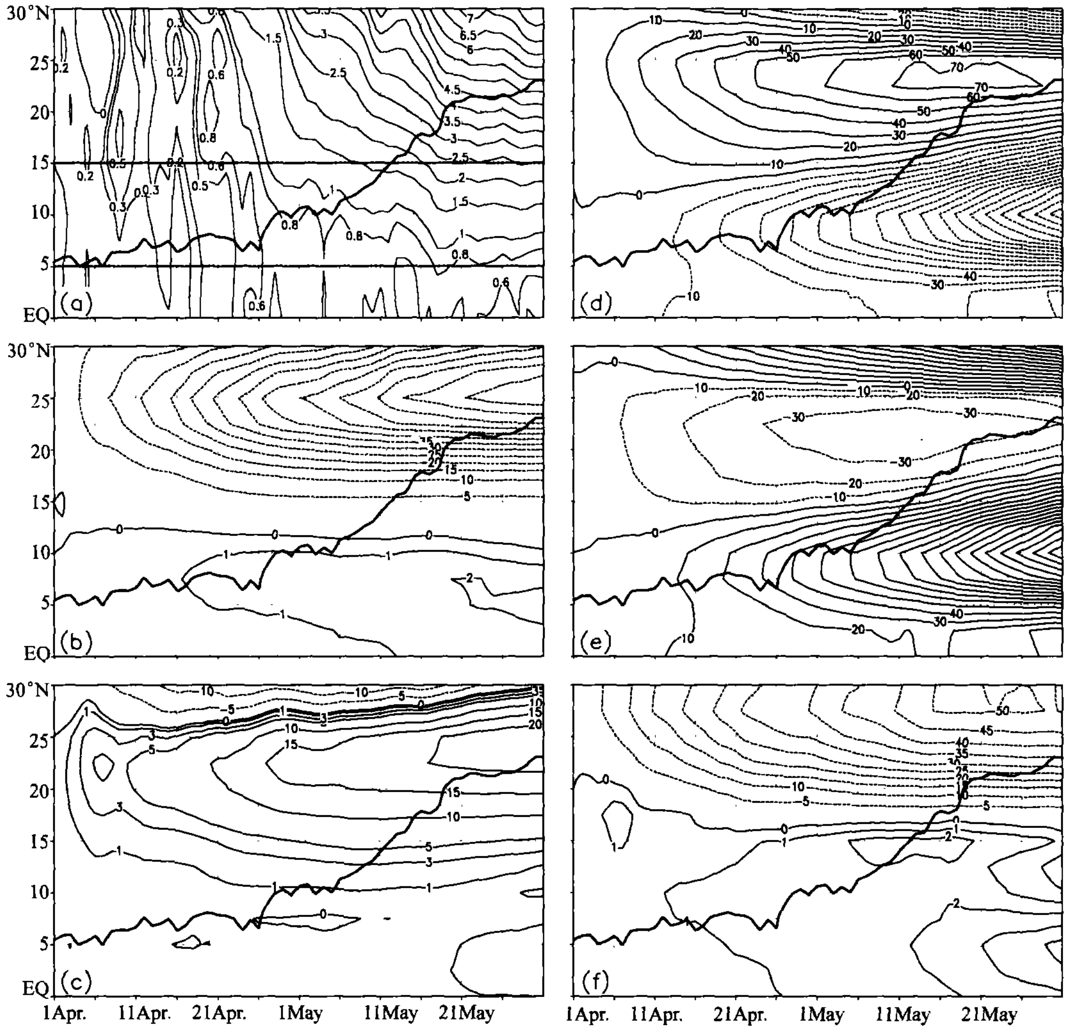

Fig. 6. Latitude-time cross sections (90 — 100°E) of mid-upper tropospheric (200 — 500 hPa) cumulative temperature increase (K) and cumulative heating (K). Thick solid lines denote the warm temperature ridgeline. (a) T_L; (b) T_u; (c) T_v; (d) T_ω; (e) T_Q; and (f) T_u+T_v.

the zonal temperature advection, meridional temperature advection, and diabatic heating. The former is related to the Tibetan Plateau heating source, and the latter two are associated with the southern China spring rainfall. The subsidence is the factor leading to the northward migration of the IDO warm ridge.

VII. CONCLUSIONS

(1) Seasonal variations of monthly mean temperature fields indicate that in May, the latitudinal positions where the warm temperature ridge locates are different from the different longitudes in which the warm temperature ridge reaches the northernmost latitude over the longitudes of the Indochina Peninsula, and is close to 20°N. The start date of the seasonal transition is the day when the warm temperature moves to the same

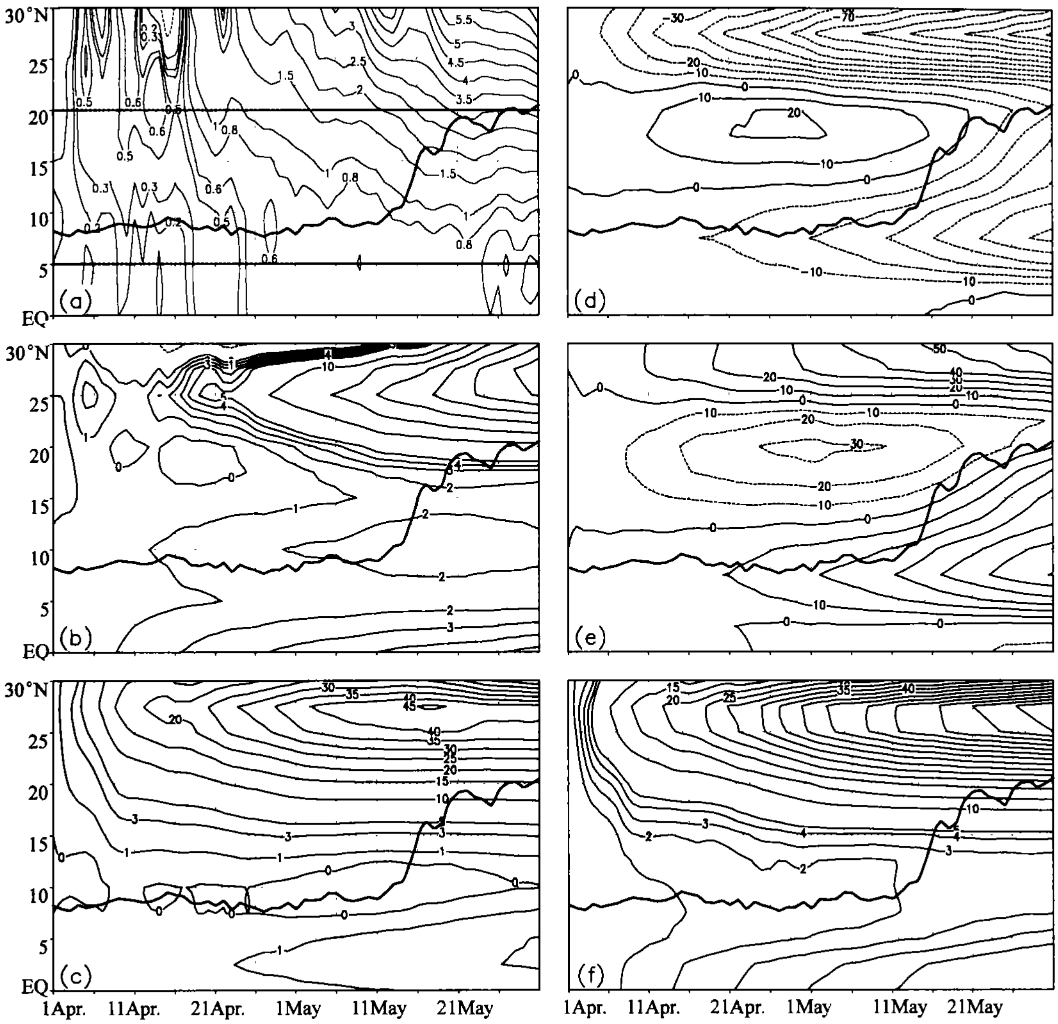

Fig. 7. As in Fig. 6, except for the SCS section (110—120°E).

latitude where the WEB locates. The reversal of the MTG in the vicinity of the WEB depends on the variation in the position of the warm ridge.

(2) Based on the distributions of monthly mean energy budget, it is found that significant diabatic heating source exists over the southern China, which results from the spring rainfall over this region in March and April. This unique heating source has impact on the Asian seasonal transition. The asymmetric energy distribution due to different physical properties of the earth's surface is one of the factors that result in the regional differences of the Asian summer monsoon onset.

(3) The variation in the position of the mid-upper tropospheric warm temperature is associated with deep convective activity. For each of these three areas (BOB, SCS, and IDO), the warm ridge undergoes a northward jump, and this variation exhibits a sudden shift between two semi-permanent positions of the warm ridge. Deep convection has similar features to the warm ridge. The northward shift of the warm temperature ridge results from the temperature increase rate more over the north of the warm ridge than over

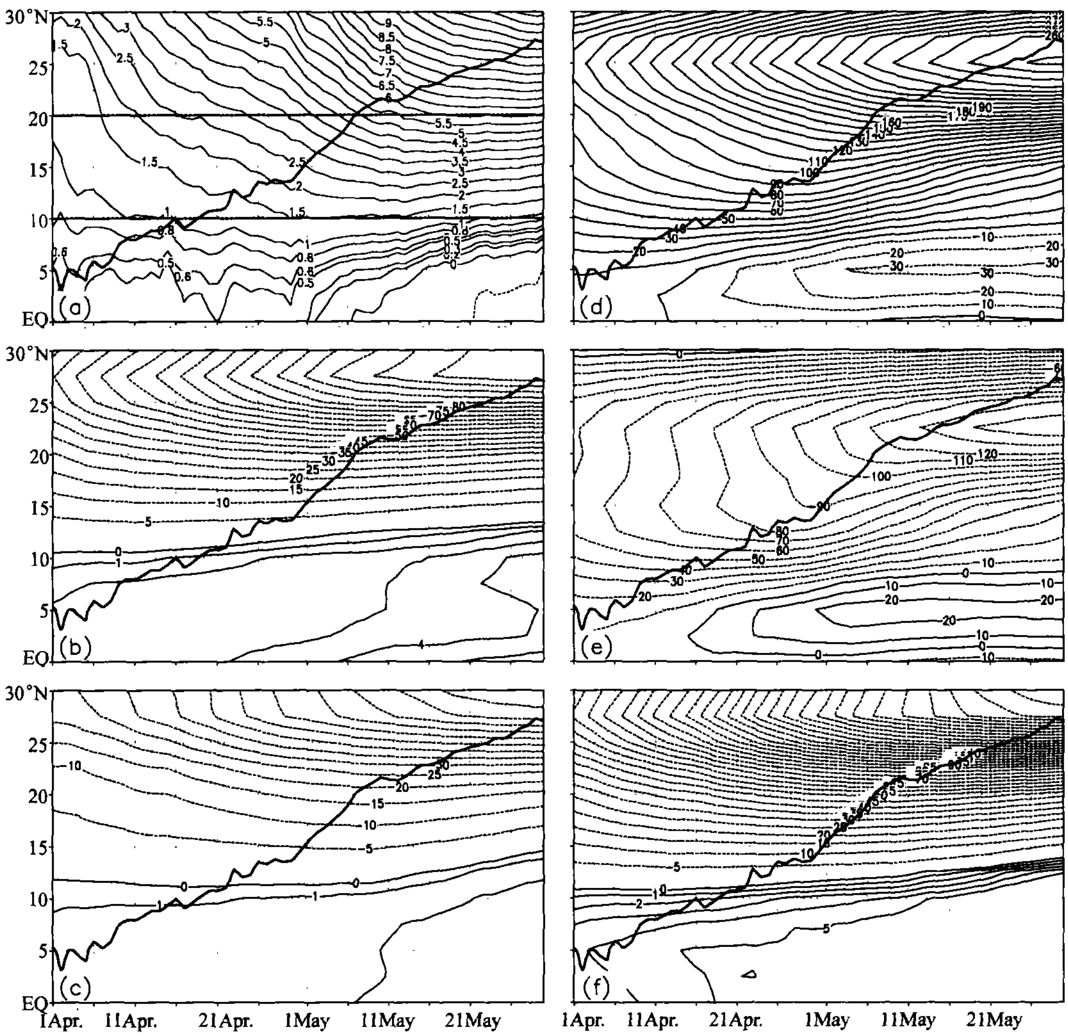

Fig. 8. As in Fig. 6, except for the IDO section (60—85°E).

its vicinity. The structure and variation of the temperature field are related to the evolution of the meridional circulation.

(4) Thermodynamical diagnoses indicate that the physical mechanisms responsible for the seasonal transition are different from area to area. The dominant factors responsible for northward shift of the BOB warm ridge are the meridional temperature in initial stages of the onset and descending motion after the onset. The factors for causing the northward jump of the SCS warm ridge involve the zonal temperature advection, meridional temperature advection, and diabatic heating associated with the southern China spring rainfall. The subsidence is the factor leading to the northward migration of the IDO warm ridge.

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