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EVOLUTION OF TROPOSPHERIC TEMPERATURE FIELDS AND CORRESPONDING THERMAL MECHANISMS BEFORE/AFTER ONSET PERIODS OF ASIAN SUMMER MONSOON

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ABSTRACT: The evolution of the tropospheric temperature fields over Indian and South China Sea monsoon areas and their thermal mechanisms are compared and analyzed during the period from March to June, 1996. The results show that the onsets of the Indian and South China Sea summer monsoons are closely associated with the seasonal warming in the troposphere over the zonal belt of 10°N ~ 30°N in these areas, which leads to the inversion of meridional temperature gradient. During the pre-onset period, the warming over the South China Sea monsoon region is mainly due to the warm horizontal advection and diabatic (latent) heating processes. Meanwhile, the warming is suppressed by the vertical adiabatic process (cooling). In spring over the Indian monsoon region, the significant adiabatic heating due to the subsidence motion, which compensates the cooling due to the strong cold advection and diabatic cooling processes, results in a larger warming rate than over the South China Sea monsoon region. However, the meridional temperature gradient over the Indian monsoon region is so large during the late winter and early spring that it takes longer time to warm the troposphere to have the reversion of meridional temperature gradient than it does over the South China Sea monsoon region. It results in the phenomenon that the South China Sea summer monsoon generally breaks out earlier than the Indian summer monsoon.

Key words: Asian monsoon; temperature evolution; troposphere; thermal mechanism

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1 INTRODUCTION

The outbreak of the Asian summer monsoon is the response of the atmosphere to the seasonal variation of thermodynamic difference between land and sea (Murakami and Ding, 1982; Johnson, Yanai and Schoak, 1987; Luo and Yanai, 1983; Luo and Yanai, 1984; Chen, Zhu and Luo et al., 1991). As part of the essential constituents of the system, monsoons break out by quite a large difference in date over India and the South China Sea (SCS), with the former averaged in the second pentad of June and the latter the fourth pentad of May (He, Wen, Jian and Sui, 2000), a difference of about 4 pentads. It is closely related with the fact that the land-sea thermodynamic difference shows variably in seasonal characteristics for the two regions. Studying the evolution of the 1979 thermodynamic field, He (1998) suggests that diabatic cooling and cold advection may be attributing to a late set-up of the monsoon by lowering mean temperature and postponing warming in the troposphere in the region of the Indian Monsoon. In fact, more diagnostic study and

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discussions are needed regarding the issue that the difference in the date of establishment of monsoons in the SCS and Indian regions are directly associated with the cause of summer monsoon formation. Taking the year 1996 as example, the current work compares and studies the evolutions of the tropospheric temperature field and its thermal factors: horizontal temperature advection, vertical diabatic temperature variation, and adiabatic heating, before and after the summer monsoon establishment in the Indian and SCS regions, in an attempt to reveal more diagnostic foundations for the causes.

2 DIAGNOSTIC EQUATIONS AND DATA

The diagnostic equations are of thermodynamics and water moisture (Luo et al., 1984) as in

$$c_p\left[\frac{\partial T}{\partial t}+\vec{V}\cdot\nabla T+\left(\frac{p}{p_0}\right)^k w\frac{\partial q}{\partial p}\right]=Q_1 \tag{1}$$

$$-L\left[\frac{\partial q}{\partial t}+\vec{V}\cdot\nabla q+w\frac{\partial q}{\partial p}\right]=Q_2 \tag{2}$$

where T is the temperature, L the condensation latent heat, θ the potential temperature, q the specific humidity, ω the vertical p-velocity, p_0 is 1000 hPa, $k=R/c_p$, and $\vec{V}$ the horizontal wind vector.

Set $Q_{11}=C_p\frac{\partial T}{\partial t}, Q_{12}=C_p\vec{V}\cdot\nabla T, Q_{13}=C_p\left(\frac{p}{p_0}\right)^k w\frac{\partial q}{\partial p}$, then Eq.(1) can be rewritten as

$$Q_{11}=Q_1-Q_{12}-Q_{13} \tag{3}$$

$Q_{11}, -Q_{12}$ and $-Q_{13}$ are the local variation of temperature, horizontal advection and vertical adiabatic variation, respectively. Q_1 and Q_2 are the source of apparent heat and the sink of apparent moisture, respectively. The vertical integration is expressed by:

$$\langle Q_1\rangle=\frac{1}{g}\int_{p_0}^{p_s}Q_1\mathrm{d}p\approx LP+S+\langle Q_R\rangle \tag{4}$$

$$\langle Q_2\rangle=\frac{1}{g}\int_{p_0}^{p_s}Q_2\mathrm{d}p\approx L(P-E) \tag{5}$$

where P, S , and E are respectively the amount of precipitation, ground-surface sensible heat flux and ground-surface evaporation, and $\langle Q_R\rangle$ is the radiative heating term in the vertical integration.

The data used in the work are NCEP's objective analysis grid data of daily temperature, wind field and specific humidity for individual mandatory levels from March 7 to July 14, 1996. The resolution is 2.5° × 2.5° and there are 7 vertical levels at 1000, 850, 700, 300, 200, and 100 hPa, respectively. The equations above are used to compute the apparent heat sources and apparent moisture sinks. A 5-day average treatment is conducted for various elements within the studied sections of time to derive pentad-mean data series, such as the first pentad from March 7 to 11, totaling at 26 pentads.

To have the following discussion more conveniently, it is necessary to determine the time of outbreak of the summer monsoon in the South China Sea and India. As it is not the focus of the current work, the definition of monsoon outbreak will not be elaborated here. Relying on the seasonal transition between upper and lower wind fields and the changeover of north-south gradient of 850-hPa geopotential height, He et al. (2000) determine that in India and the SCS the summer monsoon breaks out by a difference of 5 pentads, in the second pentad of May and the first pentad of June in 1996, respectively. The difference is close to the climatological mean difference of 4 pentads. The summer monsoons end in late September (the 54th pentad) and late August through early September (47th pentad) respectively.

3 EVOLUTION OF TROPOSPHERIC TEMPERATURE

Fig.1 gives the annual variation of the north-south temperature contrast on vertical time sections over India and the region of the South China Sea. The abscissa denotes the pentad, which comprises 5 days for each unit and accounts to 73 pentads in a year. In the Indian Monsoon region, no obvious seasonal changes are found with the sign of the temperature contrast for layers higher than 200 hPa or so in the troposphere (Fig.1a) and a distribution of cold air in the south but warm

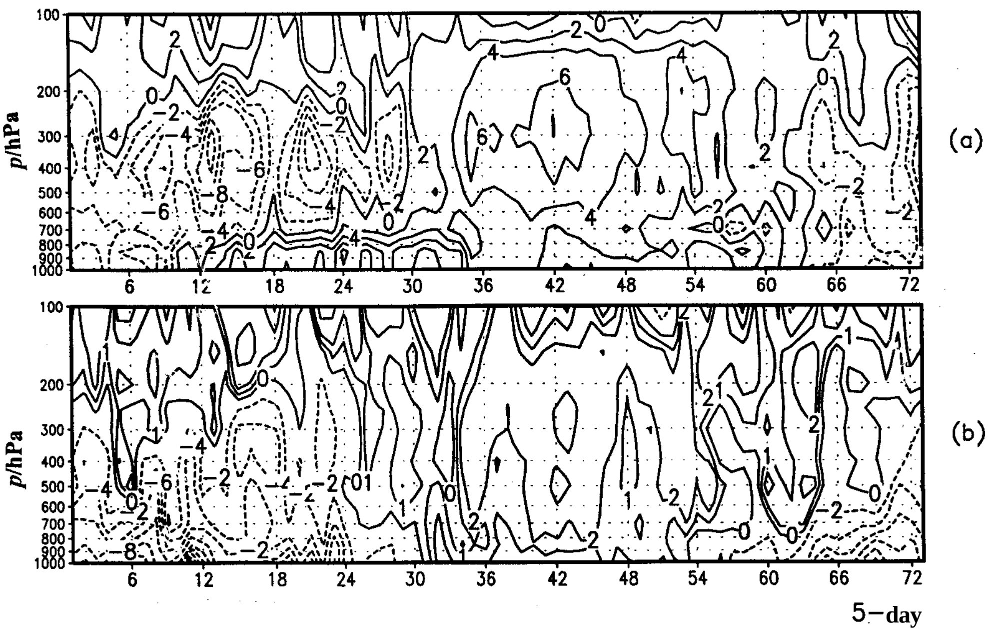

Fig.1 Height-time cross sections of north-south temperature contrast ($T_{22.5^\circ N} - T_{EQ}$) along 75°N ~ 80°N (a) and 110°E ~ 115°E (b).

air in the north is displayed throughout the year, with seasonal changes evident only in magnitude. Starting from early March (the 12th pentad), the north-south temperature contrast begins to reverse on layers below 850 hPa and trend of cold-south and warm-north rapidly enhances to increase the difference to 4 ~ 5°C. It is contributed by the fact that the Indian continent, which has received much less precipitation before the outbreak of the summer monsoon, has a warming rate on the ground surface that is larger than that on the ocean surface due to the solar radiation that is seasonally intensified. The result is a much faster warming rate in the near-surface layer. The strong temperature contrast has maintained till mid-June. Afterwards, the contrast drops as the Indian Monsoon sets up and brings more rain to lower the temperature of the underlying surface and eventually that of the near-surface layer. The most pronounced seasonal change happens in the mid- and upper- levels of the troposphere (600 hPa ~ 250 hPa), where the warm-south and cold-north pattern maintains till the end of May (30^{th} pentad) when it begins to reverse, which is very close to the outbreak timing of the Indian summer monsoon (31^{st} pentad). In addition, the temperature contrast is also significant in the mid- and upper- levels of the troposphere, oscillating between –8°C and 6°C. Beginning from late August and early September, the contrast of cold-south versus warm-north decreases substantially with the difference dropping near zero in the lower levels, corresponding to the end of the summer monsoon. For the whole troposphere, the second reversal of the temperature contrast does not complete until early November.

Before the South China Sea monsoon breaks out, the north-south temperature difference is largely on the negative side in layers beneath 250 hPa over the SCS monsoon region, with the minimum zone being within the near-surface layer (Fig.1b). Due to the affluence of cold air from the north in winter and early spring, the lower levels of the troposphere in the south of China and northern SCS are so cold that the lower-level temperature reversal comes late as compared to the Indian monsoon region or the mid-tropospheric atmosphere. High consistence is found between the timing of the monsoon outbreak in the SCS and the reversal time of the north-south temperature contrast in the middle troposphere (the 26^{th} pentad). When it comes to autumn, the temperature contrast, which is associated with the ending of the summer monsoon (the 54^{th} pentad), once again reverses, first in the near-surface layer and then in the middle and higher layers, being much similar to the situation in the Indian region. It is noted that the temperature contrast is generally weaker in the SCS monsoon region than in the Indian monsoon region.

It is known from the discussion that there is a close link between the summer monsoon outbreaks and reversal timing of temperature contrast in the middle and higher levels of the troposphere. Taking 500 hPa to represent the approximate mean state of the troposphere, Fig.2 shows the latitude-time evolution of 500-hPa temperature in the Indian and South China Sea monsoon regions. The abscissa is in the unit of pentad, setting March 7 ~ 11 as the first pentad. For the SCS monsoon region, warming begins to occur from the 13^{th} pentad (May 6 ~ 10) extensively and steadily between 10°N and 25°N with the warm ridge centering near 20°N (Fig.2b). The ridge moves northward over time and mostly stays between 20°N and 30°N from the middle of June (21^{st} pentad) to the middle of July (26^{th} pentad), though temporally weakening in the 20^{th} pentad (June 9 ~ 14). For the Indian monsoon region, extensive and steady warming takes place north of 15°N starting from the 19^{th} pentad while weak cooling appears in the equatorial region, increasing the north-south temperature contrast. From the 21^{st} pentad, the warm ridge stays between 20°N and 35°N. The study above suggests that the mid-tropospheric levels begin warming about a month earlier in the SCS monsoon region than in the Indian monsoon region, being generally close to the difference in the timing of monsoon set-up for the two localities.

The difference in the evolution of tropospheric temperature is apparently linked with the thermodynamic processes that may affect local variations of the temperature. Among them are horizontal advection of temperature, diabatic temperature changes due to vertical motion, and adiabatic processes, which are focused in the following sections of the paper.

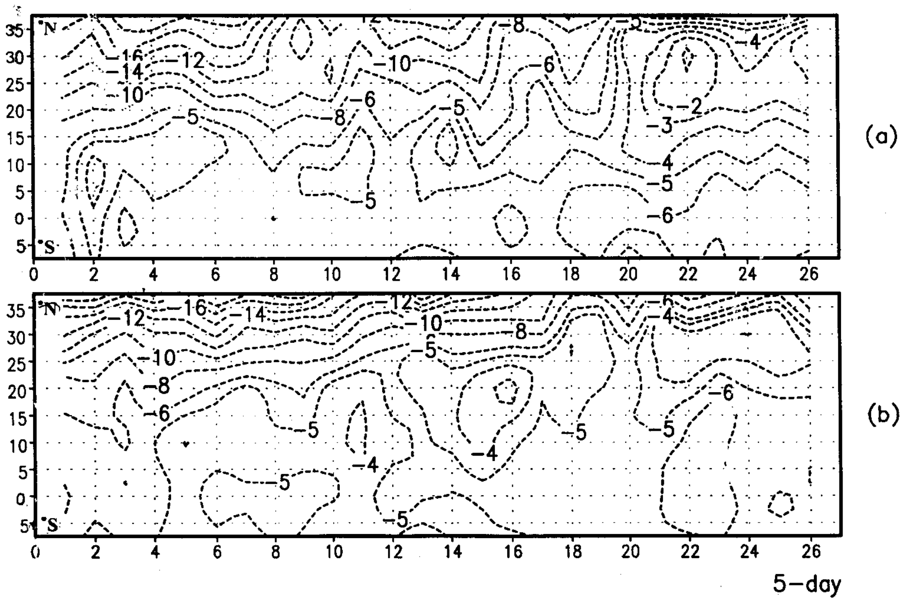

Fig.2 Time-latitude cross sections of temperature (°C) at 500 hPa along 80°E (a) and 112°E (b)

4 THERMODYNAMIC MECHANISM RESPONSIBLE FOR TEMPERATURE CHANGES IN MIDDLE AND UPPER TROPOSPHERE

4.1 *Horizontal advection of temperature*

Fig.3 gives the height-time cross section of the advection term $-Q_{12}$ for the horizontal tem-

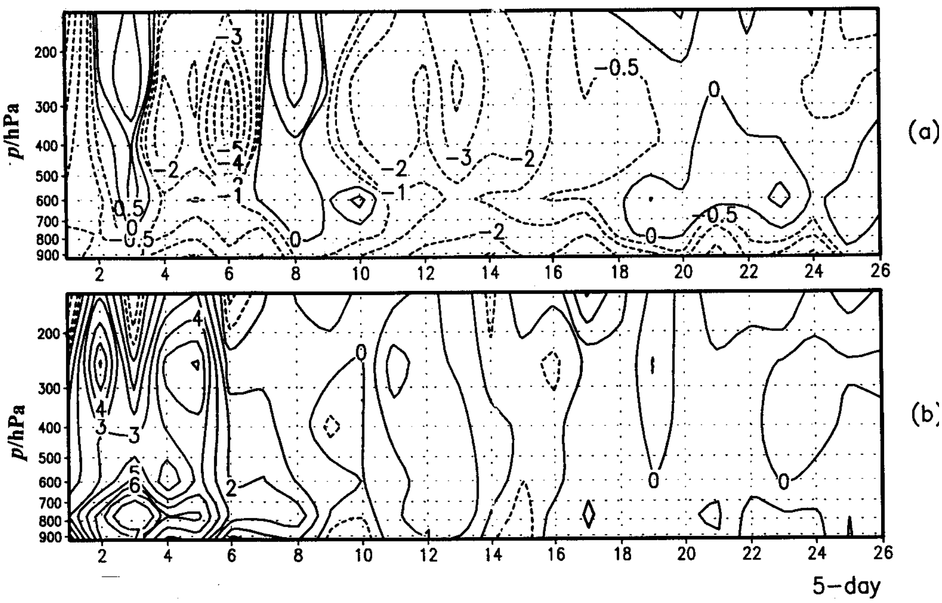

Fig.3 Height-time cross sections of horizontal advection of temperature ($-Q_{12}$) averaged over the regions of 75°E ~ 80°E and 22.5°N ~ 25°N (a) and 110°E ~ 115°E and 22.5°N ~ 25°N (b) in unit of °C/day

perature in north India and south of China. For the former, strong cold advection appears in the troposphere before the outbreak of the summer monsoon (the 18th pentad) except for the third and eighth pentad when temporary periods of warm advection are dominant. The center of the cold advection is located in the middle- and higher- levels of the troposphere, being unfavorable for the warming there. It is just the opposite for the SCS region. Before the outbreak, the troposphere is basically in the control of a strong warm advection, with the maximum centers being respectively in the lower and upper levels. After the outbreak, the warm advection obviously weakens. It is an indication that the warm advection is one of the important thermodynamic mechanisms in the warming of the troposphere over the south of China.

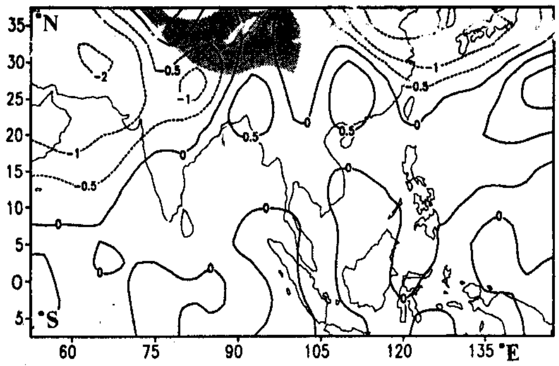

Fig.4 The 25-day mean values of $\langle -Q_{12} \rangle$ from April 11 to May 5 in unit of 100 $\mathrm{W/m^2}$.

Fig.4 exhibits the horizontal distribution of the horizontal advection $-Q_{12}$ that has been integrated vertically and averaged over 5 pentads before the outbreak of the SCS summer monsoon. Warm advection is found in the area east of 90°E and between 10°N and 30°N, with the centers in the Bay of Bengal and the continental south China near 110°E. For the Indian monsoon region, the cold advection is the main feature apart from southern Indian Peninsula where there is weak warm advection. Strong cold advection appears in the southwest flank of the Tibetan Plateau and north India. The out-of-phase feature shown by the advection of horizontal temperature for both regions is caused by opposite allocation of flow and temperature fields due to thermal and thermodynamic of the Tibetan Plateau.

4.2 *Temperature changes due to vertical adiabatic processes*

Being similar to the horizontal advection term, effects due to temperature changes by vertical adiabatic processes vary sharply over the two monsoon regions (Fig.5). In the north Indian, there is significant adiabatic warming in layers above 600 hPa, which is caused by vertical descent motion of the air (Fig.5a). The warming rate, on the average, is larger than the cooling rate that results from simultaneous cold advection. Moreover, the warming maximum due to descending adiabatic motion is located in middle- and higher- levels of the troposphere (400 hPa ~ 200 hPa). With the establishment of the summer monsoon, however, the descending motion over the Indian continent decreases substantially and changes into a ascending one and the adiabatic warming also weakens to a large extent, which corresponds to the northward shift of precipitating synoptic systems. For the south of China, the pre-onset period sees an adiabatic cooling in dominant status below the level of 200 hPa over all but the 10th pentad in which an obvious adiabatic warming occurs. It is taking place associating with the early rainy weather in the first rainy season of south China. It should also be noted that the most significant layer for the process of vertical adiabatic cooling is located in the middle and lower troposphere (Fig.6b). After the set-up, positive vertical adiabatic temperature changes occur alternatively with negative ones in the troposphere, possibly relating to the west-extending of the subtropical high in the west Pacific.

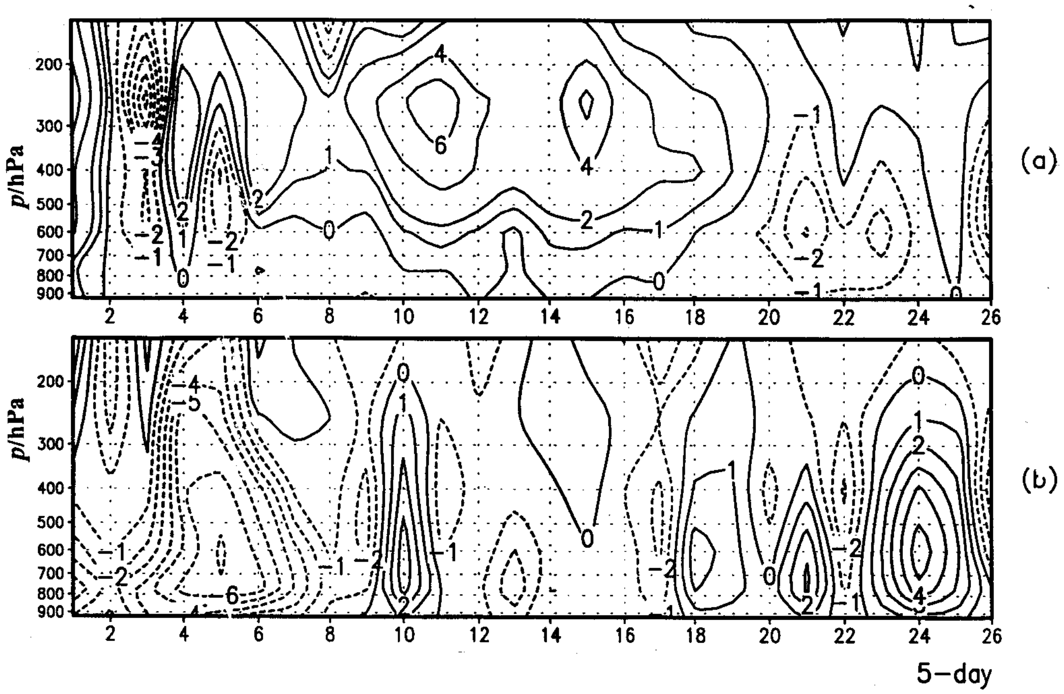

Fig.5 Height-time cross sections of temperature variation by adiabatic processes ($-Q_{13}$) averaged over the regions of 75°E ~ 80°E and 22.5°N ~ 25°N (a) and 110°E ~ 115°E and 22.5°N ~ 25°N (b) in unit of °C/day

4.3 *Diabatic heating – apparent heat source*

Fig.6 gives the vertical time cross section of the source of apparent heat that is averaged on the pentad basis for north India and south China. For the former area, the apparent heat source (Fig.6a), though with weaker intensity, evolves over time almost entirely out of phase with the term of adiabatic temperature variation (Fig.5a). From the 7th pentad to the 18th pentad (the outbreak of the Indian monsoon), tropospheric layers above 600 hPa are basically controlled by diabatic cooling with the center at the 300-hPa level. For the lower levels, the heat source ($Q_1 >$ 0) mainly consists of sensible heating in the near-surface layer. For the south China region, however, the troposphere is a source of heating from the 4th to the 9th pentad (Fig.6b) and the diabatic heating is primarily caused by latent heat by ascending motion of the air, which is indicated by the distribution of moisture sinks (figure omitted).

4.4 *Relative importance of local warming in middle and upper troposphere by all thermodynamic processes*

Local temperature variations depend on the net effect attributed by various thermodynamic processes. To have a qualitative account of the primary and secondary factors for warming before the monsoon outbreak in the middle and upper levels of the troposphere over the land surface of these regions, Tab.1 lists the mean values of various thermodynamic factors on layers 500 hPa ~ 250 hPa over Regions A and B from the 8th to the 12th pentad, i.e. for the 25 days leading up to the outbreak in the SCS.

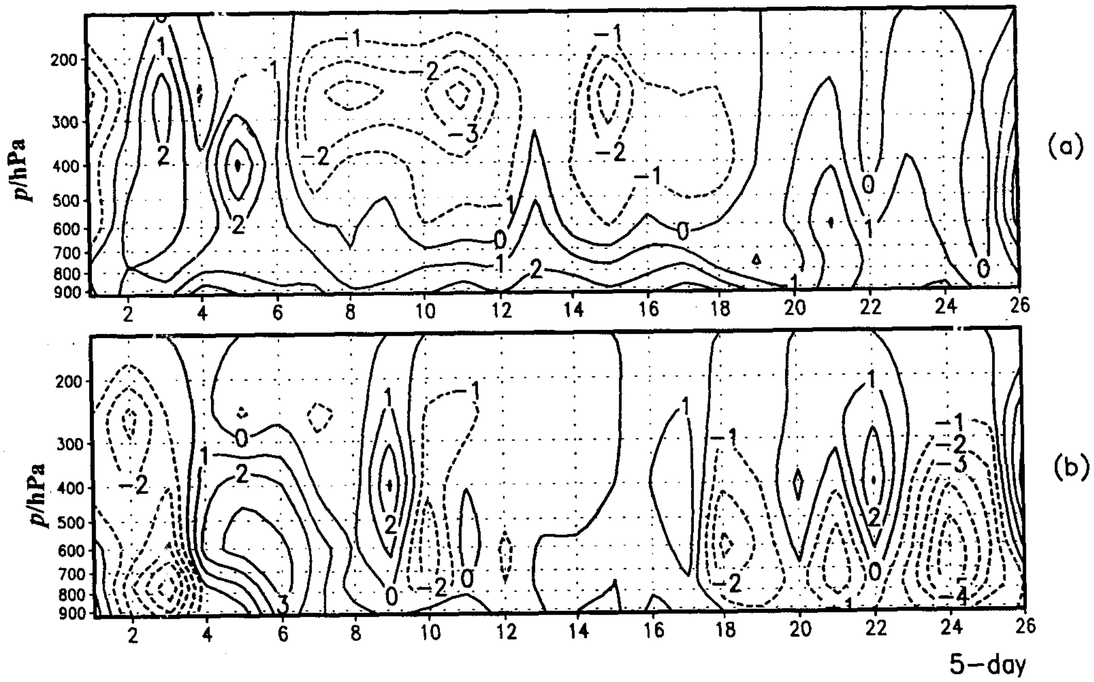

Fig.6 Height-time cross sections of the apparent heat source (Q_1) averaged over the regions of 75°E ~ 80°E and 22.5°N ~ 25°N (a) and 110°E ~ 115°E and 22.5°N ~ 25°N (b) in unit of °C/day

For East Asia (Region A), both the horizontal advection of temperature Q_{12} and the adiabatic heating term Q_1 are favorable for local warming with the former having larger contribution than the latter. With $\langle Q_1 \rangle$ in comparable magnitude with $\langle Q_2 \rangle$, we know that the adiabatic heating is primarily caused by heating from condensation during the precipitation (Luo and Yanai, 1984). The local cooling rate, which is resulted from the term of temperature variation by vertical adiabatic processes, is more significant than the heating rate caused by either the $-Q_{12}$ or the Q_1 term. In spite of it, temperature is rising because the joint action by strong warm advection and relatively weak adiabatic heating has been larger than the role of adiabatic cooling.

For the Indian Monsoon region (Region B), both the horizontal advection of temperature (cold advection) and the adiabatic heating term Q_1 are prohibiting the tropospheric warming with the

Tab. 1 25-day (April 11 ~ May 5) mean values of the thermodynamic factors averaged over layer 500 hPa ~ 250 hPa over Regions A and B in unit of °C/day

Region A: 102.5°E ~ 117.5°E, 20°N ~ 27.5°N					Region B: 70°E ~ 85°E (a) and 20°N ~ 27.5°N				
Q_{11}	$-Q_{12}$	$-Q_{13}$	Q_1	Q_2	Q_{11}	$-Q_{12}$	$-Q_{13}$	Q_1	Q_2
0.1	0.49	–0.5	0.11	0.12	0.23	–0.98	2.88	–1.67	–0.24

latter having a larger cooling rate than the former. The vertical adiabatic heating, which is caused by descending motion of the air, is the only essential thermodynamic mechanism for the warming of the mid- and higher- troposphere over the Indian monsoon region. It causes such dramatic warming effect that it retains considerable warming rate in the same levels of the troposphere for the Indian region after compensating for the cooling resulted from the cold advection and adiabatic cooling. Tab.1 shows another remarkable point: the warming rate in the middle and upper troposphere is much more significant in the Indian region than in the East Asia region.

5 CONCLUSIONS AND DISCUSSIONS

As what we have found with the case-diagnosing results above, summer monsoons in the Indian and South China Sea regions break out in close relation with the reversal of local north-south temperature contrast in the middle and upper levels of the troposphere, which is caused by seasonal warming at the same layers between 10°N and 30°N. The warming in the South China Sea monsoon region before the outbreak is jointly contributed by warm advection and adiabatic heating while the term of temperature changes due to vertical adiabatic processes is present to prohibit the warming of local tropospheric levels. The significant warming by descending adiabatic processes over the Indian monsoon region is a very important mechanism that warms the middle- and upper-tropospheric atmosphere at a rate much greater than that over the south China region in addition to offsetting the temperature fall due to strong cold advection and diabatic cooling. However, as the north-south temperature contrast (warm in the south versus cold in the north) is significant in the Indian monsoon region, it reverses on a later date in the middle and upper levels of the troposphere as compared to the South China Sea region. In contrast, the warming rate is mild over south China, but the temperature contrast is also less obvious in mid-troposphere in late winter and early spring and the reversal is thus on an early date. It is why the Indian monsoon breaks out later than the South China Sea monsoon.

In fact, differences in thermodynamic processes of seasonal warming in the tropospheric layers for both regions are the consequences of varied responses of the atmospheric circulation to seasonal changes of solar radiation heating due to differences between land and sea and the terrain itself in Asia. As pointed out by many meteorologists, the thermodynamic and mechanical forcing may be playing an essential role in it.

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