SUMMARY

The mean state and year-to-year variations of the tropospheric temperature fields and their relationship with the establishment of the summertime East Asian monsoon (EAM) and the Indian monsoon (INM) are studied using the NCEP reanalysis data of 15 years (1982-1996). The results show that the seasonal shift of the South Asian High in the upper troposphere and the establishment of the EAM and the INM are closely related to the seasonal warming which causes a reversal of the meridional gradient of upper tropospheric mean temperature over the monsoon règions. On the average of 15 years, the reversal time of the temperature gradient in the EAM region (INM region) is concurrent with (one pentad earlier than) the onset time of the summer monsoon. In most years of the 15-year period, the reversal of temperature gradient coincides or precedes the onset time of the summer monsoon in both the EAM region and the INM region. The results suggest an important role of thermal processes on the establishment of the Asian monsoon. The contributors to the upper tropospheric warming over the EAM region are the strong horizontal warm advection and the diabetic heating against the adiabatic cooling due to upward motion. In the INM region, strong adiabatic heating by subsidence and the diabetic heating are major warming processes against the strong horizontal cold advection related to the persistent northwestlies to the southwestern periphery of the Tibetan Plateau. It appears that the early or late establishment of the Asian summer monsoon is not directly related to the differential warming near the surface.

Keywords: Asian summer monsoon, monsoon index

1. INTRODUCTION

As a major component of the global climate system, Asian summer monsoon has been a worldwide research topic for decades. Since the late 1970's, Asian summer monsoon has been conceptually broadened and viewed as a combination of the Indian monsoon system and the East Asian monsoon (e.g., Krishinamurti and Bhalme 1976, Jin and Chen 1983, Shen et al. 1982, Chen et al. 1983, Tong et al. 1983, He et al. 1987). The establishment of the East Asian summer monsoon is generally earlier than that of the Indian monsoon (e.g., Lighthill and Pearce 1981, He et al. 1987, Tao and Chen 1987, Lau and Yang 1997). But the relevant mechanisms are still unclear.

The general circulation over Asia undergoes abrupt seasonal changes during late spring and early summer. These changes are closely related to the tropospheric seasonal warming over the Asian monsoon region (Yeh 1959, Murakami and Ding 1982, Krishnamurti 1985, He et al. 1987, Yanai et al. 1992). The importance of the Tibetan Plateau as an elevated heat source for the abrupt changes of the circulation has been noticed by many previous studies (e.g. Flohn 1957, 1960, Ye and Gao 1979, Lou and Yanai 1984, Murakami 1987).

In a study of the Asian summer monsoon in the period of late spring and earlier summer 1979, He et al. (1987) showed that the general circulation underwent two distinct stages of abrupt transitions resulting in the successive onsets of early summer rains over Southeast Asia and India. Their analysis revealed that the two transitions were closely related to a reversal of meridional gradient of mean temperature in the upper troposphere, as originally suggested by Flohn (1957). Recently, Li and Yanai (1996) showed that the onset of the Asian summer monsoon is, on the average over a 14-year period, concurrent with the reversal of meridional temperature gradient south of the Tibetan Plateau. However, it is not clear whether the above results hold for each individual year. The responsible physical mechanisms for the different timing of onset of the Indian monsoon and the East Asian summer monsoon also remain to be investigated.

In this paper, we tackle the above problems through an analysis of the evolution of the tropospheric temperature in relation to the onset of the Asian summer monsoons, and an analysis of the heat budget to investigate the responsible processes for the different timing of EAM and INM onsets. The data and the definition of an onset index of the East Asian summer monsoon and the Indian monsoon are introduced in section 2. The index is determined from the wind and geopotential height fields. The evolution of the South Asian High (SAH) and the tropospheric temperature fields are discussed in section 3. In section 4, the reversal times of the meridional temperature difference in upper troposphere over the EAM region and the INM region for the study period are identified. A comparison is then made between the reversal time and the monsoon onset time determined by the onset index. Section 5 gives the results of case studies to reveal the physical mechanism for the upper tropospheric warming over the land areas in the two monsoon regions. The final section consists of a summary of the main results of this paper and discussions of a major scientific question concerning the role of seasonal heating in the upper troposphere relative to the heating in the lower troposphere by land-sea thermal contrasts in the establishment of Asian summer monsoon.

geostrophic wind between the two points. A positive Ih implies that the South China Sea region is affected by westlies and is free of the control of the SCSH, and that the Arabian Sea is controlled by southwestlies and free of the effects of the AH.

The summer monsoon onset index used in this study is then defined as

$$Iwh = Iw \bullet Ih / |Ih|, \text{ when } Iw > 0 \quad (3)$$

Otherwise

$$Iwh = Iw \quad (4)$$

Figure 1 shows the variation of pentad mean Iwh averaged during the 15-year period (1982-1996) in the East Asian monsoon region, 5°N-20°N, 90° E-120°E (hereafter referred to as the EAM region or R1). The abscissa represents the series number of pentads (72 pentads year^{-1}). We can see that for the 15-year mean, the summer monsoon period (Iwh>0) over region R1 starts at the 28th pentad (16-20 May) and ends at the 56th pentad (6-10 October). For the Indian monsoon region, 5°N-20°N, 50°E-85°E (hereafter referred to as the INM region or R2), the summer monsoon lasts 18 pentads from the 32nd to 49th pentad (figure omitted). The onset time of the summer monsoon of the two regions verified against each year during the 15-year period can be found in Table 1 and 2 of this paper.

3. EVOLUTION OF SOUTH ASIAN HIGH AND TROPOSPHERIC TEMPERATURE

South Asian High (SAH) is a major component of the Asian summer monsoon system. The timing of monsoon development is directly related to the seasonal transition of the SAH. If thermodynamic processes play an important role in the evolution of the SAH that leads to the establishment and evolution of monsoon, the mean tropospheric temperature is then expected to be closely related to the seasonal transition of the SAH. To explore this relationship, we first show the seasonal migration of the center of the SAH.

(a) *Seasonal transition of South Asian High*

Figure 2 shows the annual evolution of the center of the SAH within the domain of interest (20°S-50°N, 40°-150°E) determined by the 200 hPa winds during the fifteen years. The abscissa is time and the ordinate is longitude and latitude. The annual mean latitude and longitude are 21.9°N and 101.2°E, respectively. In the figure, the center moves into the domain in mid-April, and moves steadily northwestward. It reaches northernmost (30°N) and westernmost (55°E) position in August (43rd pentad) and starts moving southeastward. As discussed in Section 2, the onset of south China monsoon and the Indian monsoon based on the fifteen-year average data are respectively the fourth pentad of May (28th pentad) and the second pentad of June (32nd pentad). The corresponding centers of the SAH are (17.5°N, 102°E) and (22.5°N, 96°E).

(b) *Evolution of the tropospheric temperature and South Asian High*

To examine the evolution of thermal and pressure fields, we show in Fig. 3 the mean upper tropospheric temperature (200 and 500 hPa) during the development period of the East Asian monsoon in pentad 27-29 (Fig. 3a, 3b, 3c) and the Indian monsoon in pentad 31-33 (Fig. 3d, 3e, 3f). The locations of zero zonal winds at 200 hPa are also shown in Fig. 3 by the dashed lines that in the Northern Hemisphere represent the ridge of the South Asian High. One of the salient features in the figure is the collocated and synchronized northward movement of the pressure ridge and the thermal ridge in the monsoon region. Within 50°E-120°E, they move from 15°N at 27th pentad to 22°N at 32nd pentad. Such a tight coupling monsoon period, the warm center is -28°C in the East Asia region (pentad 28-54) and -26°C in the Indian region (pentad 32-49). This is consistent with the fact that the center of the South Asian High is more northwestward and stronger during the prevailing Indian monsoon period. Third, the evolution of pressure ridge closely follows the evolution of the temperature ridge. Finally, the formation of a warm center near 30°N leads to a reversal of meridional temperature gradient in the region south of 30°N. The reversal of the temperature gradient occurs at 28th pentad in the East Asian monsoon region, and 31st pentad in the Indian monsoon region, consistent with the establishment of the corresponding monsoon system.

4. REVERSAL OF TEMPERATURE GRADIENT AND MONSOON ONSET

Figure 6 shows the difference of the mean upper tropospheric temperature between 25°N and 5°N in the East Asian monsoon region (90°E-120°E) and the Indian monsoon region (50°E-85°E) based on the 15-year average. In the East Asian monsoon region, the reversal as defined by the change of sign of the temperature difference occurs at 28th pentad, the same as the mean monsoon onset time determined by the above defined monsoon onset index *Iwh*. In the Indian monsoon region, the reversal occurs at 31st pentad, one pentad earlier than the mean monsoon onset time identified by the onset index *Iwh*. Clearly, the reversal of meridional temperature gradient agrees with the monsoon onset in the corresponding regions.

Using the same method discussed above, we calculated the reversal time of the meridional temperature gradient for each year of 1982-1996 as listed in Table 1 and Table 2. Also listed in the tables are the yearly onset time of the East Asian monsoon and the Indian monsoon calculated using the method discussed in Section 2. In the East Asian monsoon region (Table 1), the number of years that the reversal of temperature gradient occurs earlier, simultaneously, and later than the monsoon onset is respectively, 5, 8, and 2 years. The corresponding number of years in the Indian monsoon region is 9, 4, and 2. The mean monsoon onset time and temperature reversal time of the fifteen years shown in Tables 1 and 2 are slightly different from those estimated from the 15-year mean data. Table 1 shows that, in the East Asian monsoon region, the temperature gradient reverses about one pentad earlier than the monsoon onset. Table 2 shows that, in the Indian monsoon region, the temperature reversal time is about 1.5 pentads earlier than the monsoon onset time. This indicates that the reversal of mean upper-tropospheric temperature gradient is an important precursor of monsoon onset.

Table 1 Time of monsoon onset and the reversal of upper tropospheric temperature gradient in the East Asia monsoon region (pentad)

Year	82	83	84	85	86	87	88	89	90	91	92	93	94	95	96	mean
Onset	31	29	24	30	27	32	29	28	28	32	28	30	25	27	26	28.4
T. reversal	26	29	25	30	27	27	29	26	28	29	28	31	25	27	25	27.5

Table 2 Time of monsoon onset and the reversal of upper tropospheric temperature gradient in the Indian monsoon region (pentad)

Year	82	83	84	85	86	87	88	89	90	91	92	93	94	95	96	mean
Onset	32	34	31	32	32	31	31	31	29	32	35	32	31	33	31	32
T. reversal	31	34	30	29	32	31	28	28	27	32	34	32	32	27	30	30.5

A more careful examination of Fig. 8 shows that the absolute value of the meridional temperature gradient during the pre-onset period (before the reversal of the temperature gradient) over region R2 is larger than its counterpart over region R1. This may also be related to the later reversal of the meridional temperature gradient over region R2.

The underlying physical processes for the reversal of the meridional temperature gradient in the upper troposphere are closely related to the thermal processes. To gain a better understanding of the responsible processes for the rapid seasonal warming over the northern land areas, we examined the spatial and temporal variations of the warming (or cooling) rates caused by the horizontal temperature advection, the adiabatic heating related to the vertical motions and the diabetic heating over the two regions in 1994. Following Yanai et al (1973), the apparent heat source, Q1, and moisture sink, Q2, are defined as

$$Q_1 = c_p\left[\frac{\partial T}{\partial t} + \bar{V}\bullet\nabla T + \left(\frac{p}{p_0}\right)^k \omega\frac{\partial\theta}{\partial p}\right] \quad (7)$$

$$Q_2 = -L\left[\frac{\partial q}{\partial t} + \bar{V}\bullet\nabla q + \omega\frac{\partial q}{\partial p}\right] \quad (8)$$

where T is the temperature, p the pressure, q the mixing ratio of water vapor, θ potential temperature, ω the p-vertical velocity, L the latent heat of condensation, _=R/C_p, R and C_p the gas constant and the specific heat at constant pressure, $\bar{V}$ the horizontal wind vector and p_0=1000 hPa. The vertical integration of (7) and (8) from Pt to Ps gives

$$\langle Q_1\rangle = \langle Q_R\rangle + LP + S\,, \quad (9)$$

$$\langle Q_2\rangle = L(P-E)\,, \quad (10)$$

where $\langle\ \rangle = \frac{1}{g}\int_{p_s}^{p_t}(\ )dp,$ (11)

Q_R is the radiative heating rate, P, S and E are respectively the precipitation rate, the sensible heat flux and the evaporation rate per unit area at the surface. For the convenience of description, expression (7) can be rewritten as

$$Q_{11} = Q_1 - Q_{12} - Q_{13}, \quad (12)$$

where

$$Q_{11} = C_p\frac{\partial T}{\partial t}, \quad (13)$$

$$Q_{12} = C_p\bar{V}\bullet\nabla T, \quad (14)$$

$$Q_{13} = C_p\left(\frac{p}{p_0}\right)^K \omega\frac{\partial\theta}{\partial p}, \quad (15)$$

Thus the warming or cooling caused by horizontal temperature advection and adiabatic processes are represented by -Q_{12} and -Q_{13} respectively.

The computation was carried out on a daily basis, but the analysis was based on 5-day means within the period from March 1st to August 2nd, 1994, which was divided into 31 pentads. The onset time of the tropical summer monsoon in this year occurred in early May (25th pentad) over the EAM region and early June (31st pentad) over the INM region (He et al. 2000).

Figure 9 shows the height-time cross sections of 5-day mean -Q_{12} (°C day^{-1}) averaged over the representative areas: 22.5-27.5°N, 70-80°E in the INM region (a) and 22.5-27.5°N, 100-110°E in the EAM region (b) during the period from March 1st to August 2nd in 1994.

A: 100-110°E, 22.5-27.5°N					B: 70-80°E, 22.5-27.5°N				
Q_{11}	$-Q_{12}$	$-Q_{13}$	Q_1	Q_2	Q_{11}	$-Q_{12}$	$-Q_{13}$	Q_1	Q_2
0.10	0.71	-1.05	0.44	-0.36	-0.03	-0.87	0.52	0.31	0.19

The above results are qualitatively similar to those of other case studies of the warming processes in the upper troposphere for different individual years (He 1998; Jian et al., 1998, 2000). We believe that the spatial difference of the reversal of meridional temperature gradient in the upper troposphere is unambiguously related to the spatial contrasts of the thermal processes, which are determined by several factors including the thermal and dynamical effects of the Tibetan Plateau, the contrast between the sea and land, and the large scale circulations.

6. CONCLUSION AND DISCUSSION

The main results of this work are summarized as follows:

(1) The seasonal shift of the South Asian High in the upper troposphere and the establishment of the Asian summer monsoon are closely related to the evolution of the upper tropospheric temperature field. Seasonal warming during late spring or early summer in the Asian monsoon region results in the reversal of both meridional and zonal gradients of the upper tropospheric temperature south of 30°N. Based on the 15-year mean annual cycle, the reversal time of the meridional gradient of the upper tropospheric temperature is exactly the same as the mean monsoon onset time in the EAM region (28th pentad), and one pentad earlier than the mean monsoon onset time in the INM region (32st pentad). In most years (87%) of the 15-year period, the reversal time of the meridional temperature gradient coincides or precedes the onset time of the summer monsoon in both the EAM and the INM regions. Thus the reversal time of the meridional gradient of the upper tropospheric mean temperature is a precursor of the summer monsoon onset. The results suggest that thermodynamics is the major cause of the earlier onset of EAM than INM. Note that, in the INM region, the reversal time of the meridional gradient (dashed in Fig.6) and that of the zonal gradient (solid in Fig. 7) of the upper-tropospheric temperature are the same. However, the two are quit different in the EAM region (comparing the solid line in figure 6 with the solid line in Fig.7).

(2) Over the Asian monsoon region, the mean zonal temperature gradient in the upper troposphere during the summer season is oriented from west to east, which is opposite to that in the lower troposphere. This suggests that such seasonal warming in upper troposphere is mainly resulted from internal thermal and dynamical processes other than surface heating.

(3) In the EAM region, budget analysis showed that horizontal warm advection and diabetic heating are the main warming processes in the upper troposphere against the intense adiabatic cooling due to the upward motion. In the INM region, adiabatic heating due to the prevailing downward motion and the diabetic heating are the main warming processes against strong horizontal cold advection.

The prevailing northwesterlies over the southwestern periphery of the Tibetan Plateatu and the Indian sub-continent and the southwesterlies over the East Asia are related to the sub-tropical westly jet in the upper troposphere to the south of the Plateau. They contribute to the cold advection over the INM region and warm advection over the EAM region. The strong downward motion over the INM region is probably related to the downward branch of the Plateau induced vertical circulation (He et al, 1987). Thus the plateau-induced circulation and

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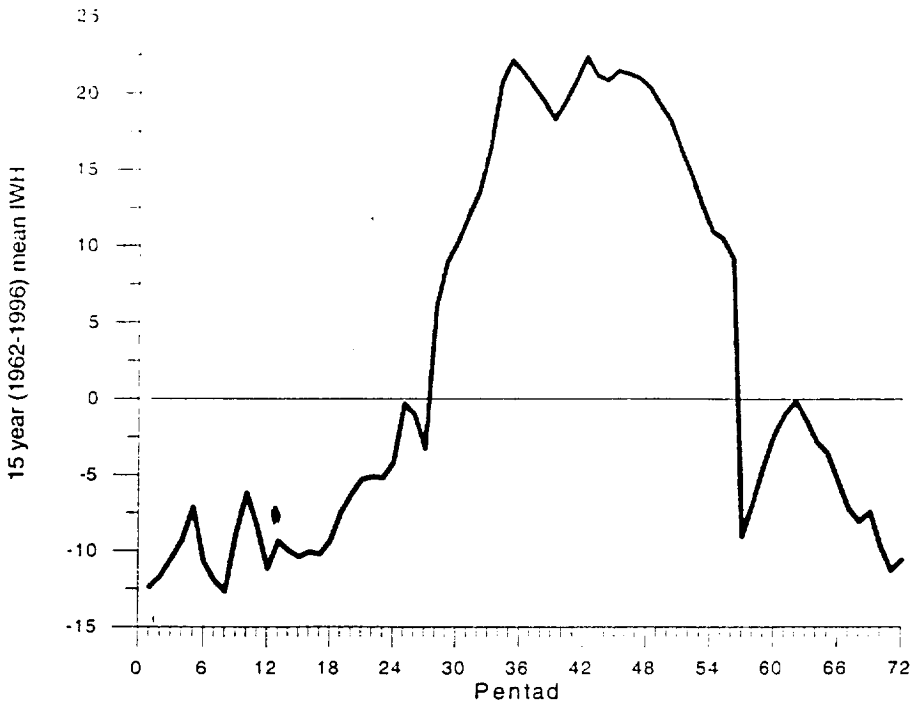

Fig. 1 15-year (1982-1996) mean of the onset index (*Iwh*) in the EAM region (5 N-20 N, 90 E-120 E)

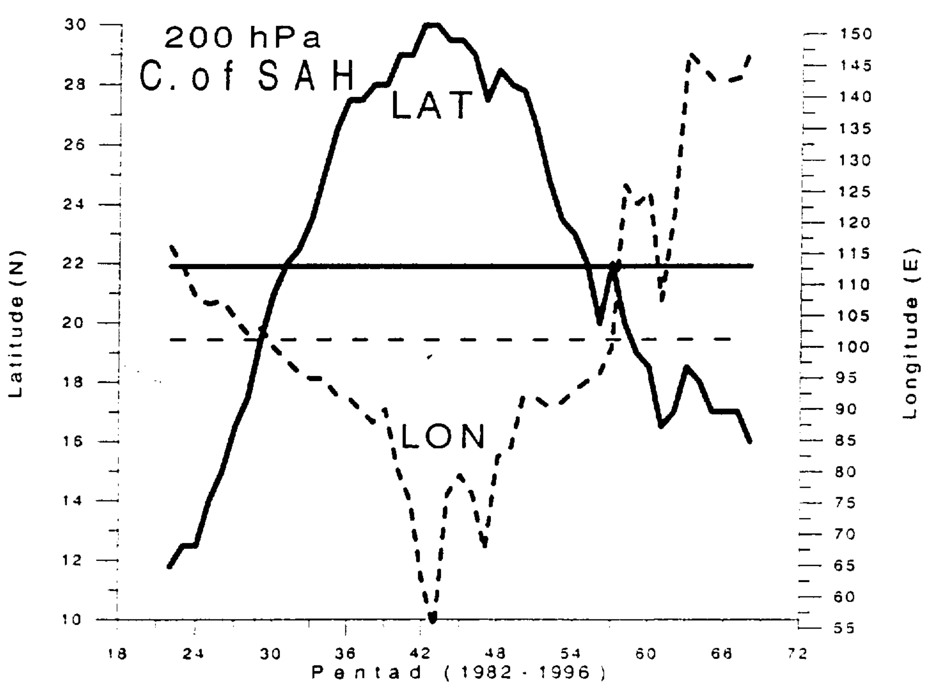

Fig. 2 15-year (1982-1996) mean movement of the center of South Asian anticyclone at 200 hPa.

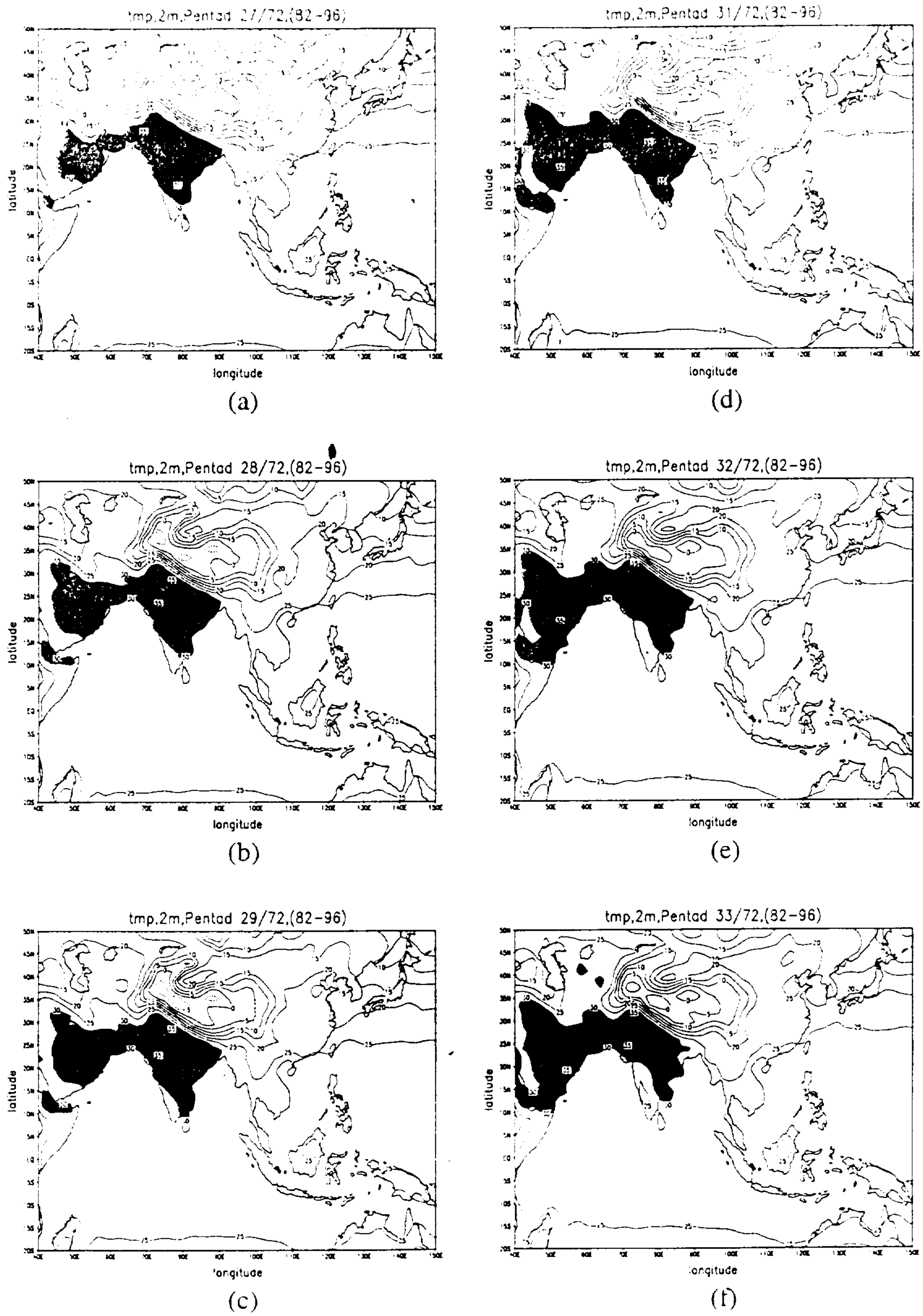

Fig. 4 Same as figure 3 except for temperature at the height of 2 meters.

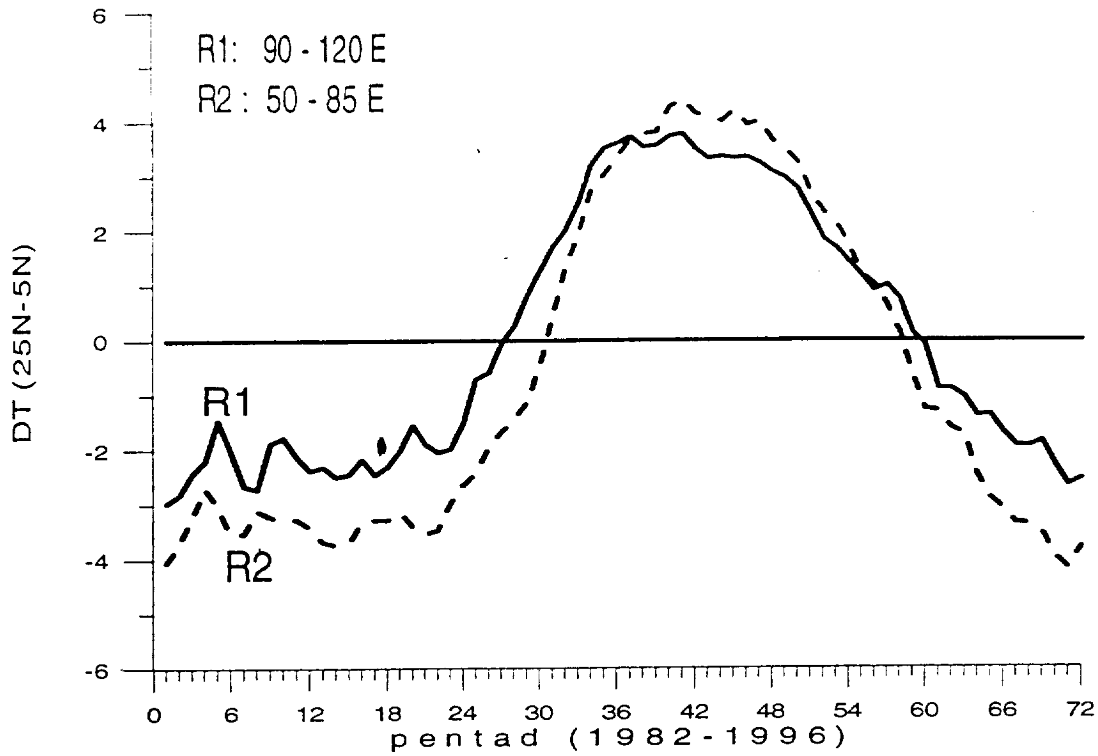
R1: 90 - 120 E
R2 : 50 - 85 E
R1
R2
DT (25N-5N)
6
4
2
0
-2
-4
-6
0
6
12
18
24
30
36
42
48
54
60
66
72
pentad (1982-1996)

Figure 6

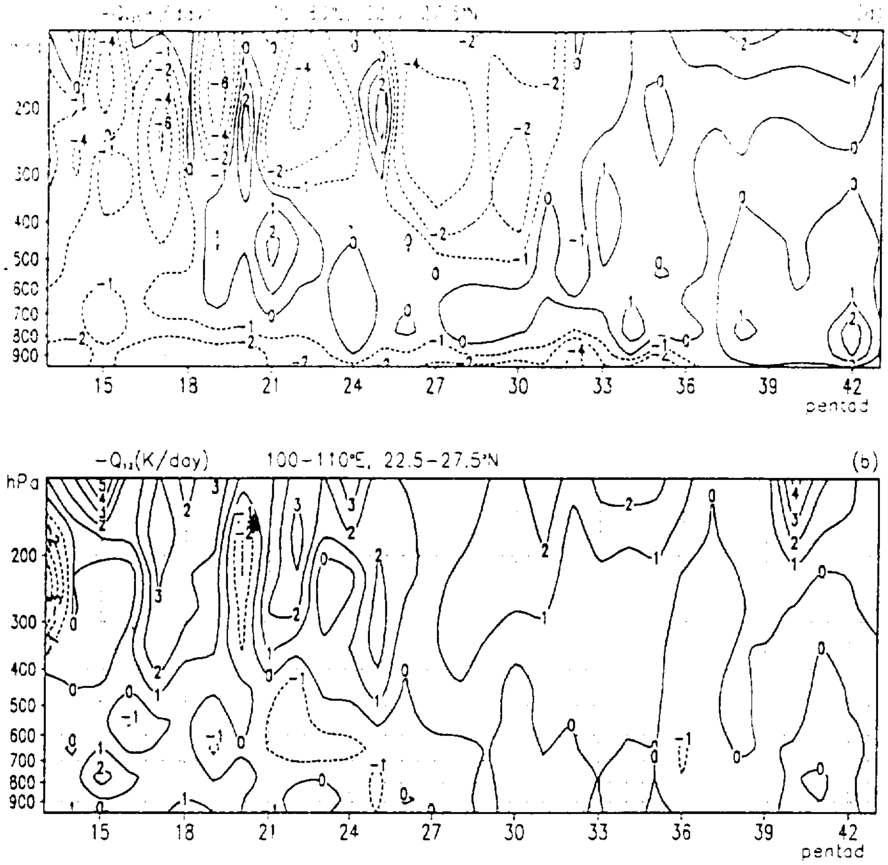

Figure 9

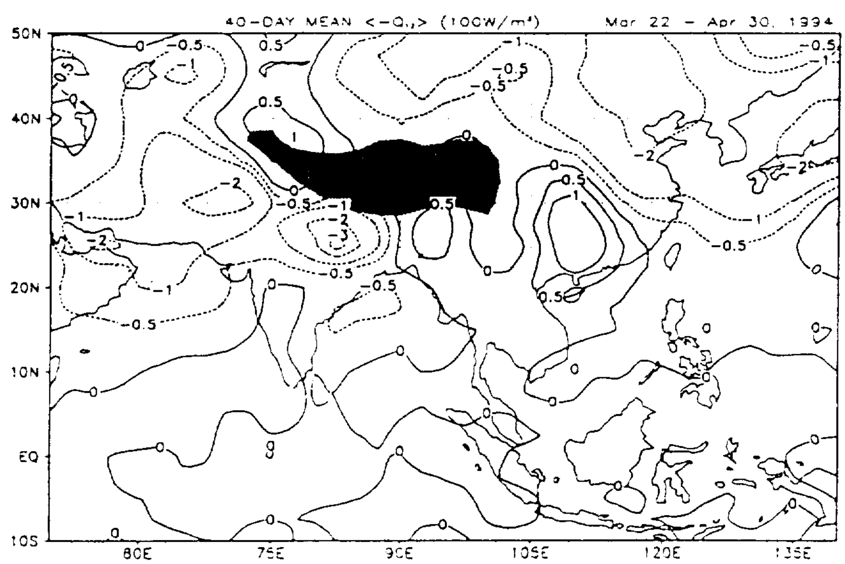

Figure 10