TEMPERATURE CHANGES OVER EURASIA DURING THE LATE SUMMER OF 1979

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ABSTRACT

During the late summer of 1979, massive changes occurred in the distribution of temperature over Eurasia north of 15°N. At 300 hPa, zonal mean temperature averaged over Eurasia along 20°—25°N decreased sharply around 23 August. An abrupt decrease in 300 hPa zonal mean temperature also occurred over extensive mid-latitude zones (40°—55°N) around 18 August, i. e., about 5 days prior to the monsoon withdrawal over South Asia.

The intensity and location of N—S oriented, vertical overturning underwent significant changes over Eurasia during the transition from summer to fall. Near 20°—25°N, zonal mean updrafts weakened considerably during the transition period (18—27 August). Around 45°N, zonal mean downdrafts and the associated cooling (radiative) rate increased considerably during the transition period.

Near 15°N, 300 hPa zonal mean temperature fluctuated nearly periodically with an approximate 40-day period. These fluctuations appear to be associated with a small imbalance between 40-day filtered adiabatic cooling (heating) and diabatic heating (cooling).

I. INTRODUCTION

Limited research indicated a possible link between circulation changes over northern Eurasia during late spring and the onset of summer monsoon over South Asia. Winston and Krueger[1] found a large albedo difference between the early spring seasons of 1975 and 1976 over Soviet Central Asia. They postulated that this large interannual albedo difference may influence the onset and intensity of the following summer monsoon circulations.

The onset of the 1979 summer monsoon over central India was declared to be on 19 June[2]. Before and during the monsoon onset massive changes occurred in the wind and temperature fields over extensive areas of the Eurasian continent north of 15°N[3]. For example, the Afghanistan-western Tibetan Plateau region was characterized by an abrupt increase in 300 hPa temperature and intensification of anticyclone above 300 hPa around 4 June; i. e., about two weeks prior to the Indian monsoon onset. Concurrently, rapid development of upper-tropospheric anticyclone and increase in temperature also occurred in association with the establishment of intense monsoon rains (heating) over the East China Sea-Japan region. Thus, the increase in upper-tropospheric temperature and heating over these mid-latitude regions may be a prerequisite to the establishment of the summer monsoon near India.

Chinese meteorologists[4] made an extensive study concerning principal weather systems over Eurasia during the transition from summer to fall. During this transition period, the tropospheric temperature decreases rather sharply over and around the Tibetan Plateau. It is highly probable that this rapid temperature decrease is related in some way to the monsoon withdrawal from the Indian region. The present study examines this possibility. Other problems to be investigated include: 1) Is the rapid temperature decrease limited to the immediate vicinity of the Tibetan Plateau? 2) What influence does the N—S vertical overturning have in changing the latitudinal distribution of temperature over Eurasia during late summer? and 3) What role does the eastward propagating disturbances in the mid-latitude westerlies play in temperature change, via eddy sensible heat transport, over Eurasia?

II. DATA AND COMPUTATIONAL PROCEDURES

This study utilized twice-daily, u, v and T data at eight levels (100, 200, 300, 400, 500, 700, 850 and 1000 hPa) for the late summer (1 July—30 September) of 1979. These data were extracted from the operational objective analyses of the National Meteorological Center (NMC), Washington, D. C. at 2.5° latitude-longitude resolution over Eurasia from 0° to 150°E and 15° to 70°N. The computational procedures used here are identical to those described in the paper by Murakami and Ding[1]. The reader is directed to their paper for a complete description of the computational procedures.

III. WIND AND TEMPERATURE CHANGES DURING THE LATE SUMMER OF 1979

The 1979 summer monsoon over central India commenced on about 19 June. The Indian monsoon was quite active until the middle of July. During this active monsoon period, the 700 hPa westerlies over the central India-Indochina region (15°N, 70°—100°E) were substantially strong and occasionally exceeded 10 m s^{-1} (Fig. 1, left). During the following 2 weeks, the 700 hPa westerlies were very weak. This period of break monsoon was characterized by below-normal rainfall over central India[5]. On about 29 July the summer monsoon revived and lasted for approximately 3 weeks. The second break monsoon with weak 700 hPa westerlies began about 20 August. The summer monsoon never revived again over most areas in South and Southeast Asia. Thus, it was difficult to determine an exact withdrawal date of the 1979 summer monsoon[2]. The prolonged break was believed to be one important cause for the deficit of monsoon rains over many areas in India.

The equally obvious intraseasonal changes were also seen in the longitude-time section for 300 hPa u along 15°N (Fig. 1, right). It reveals that the active monsoons were characterized by weak upper easterlies or even upper westerlies, whereas the break monsoons were associated with stronger 300 hPa easterlies. Around 20 August when the second active phase ended, the 300 hPa easterlies became substantially strong. They remained strong until the end of August, then they gradually diminished.

Another feature of interest in Fig. 1 (right) during the active monsoon period (29 July to 20 August) is the westward propagation of westerly (or weak easterly) perturbations along 15°N from the western Pacific (150°E), across the Philippines, Indochina and India to as far west as western Africa (0°). Similarly during the break monsoon (15 July to 28 July) 300 hPa easterly perturbations tend to propagate westward from 150°E to 0°. Thus, the break monsoon near India appears to correspond to a phase change from a westerly (or weak easterly) to an easterly regime in the 300 hPa u along 15°N[1)]

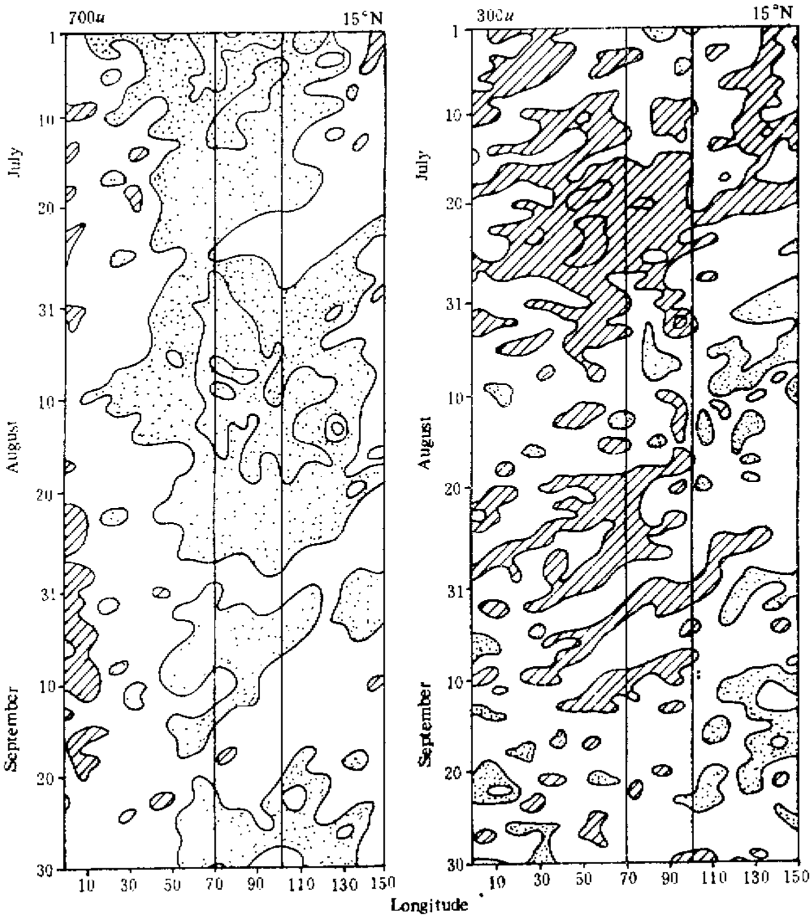

Fig. 1. Longitude-time sections of zonal winds with 10 units (m s^{-1} unit) per interval. Shading indicates regions of westerly zonal winds. Hatching denotes regions of greater than 10 m s^{-1} easterly winds. Left: For 700 hPa u at 15°N. Right: For 300 hPa u at 15°N.

In Fig. 2 (left), the 300 hPa T along 20°N indicates significant fluctuations over the entire Asia continent (0°—150°E), with fluctuations most prominent over the monsoon region (70°—100°E), namely, high T during active phases and low T during breaks.

Interestingly, 300 hPa T at 20°N appears to be positively correlated with 700 hPa u (Fig. 1, left) and 300 hPa u (Fig. 1, right) along 15°N. The Indian monsoon active (break) phases are associated with strong (weak) 700 hPa westerlies along 15°N south of the monsoon

1) Based on satellite-measured outgoing longwave radiation data during the three summers of 1975—1977, Murakami[4] confirmed that rainfall fluctuations between active and break monsoon periods over the Indian region were strongly associated with westward propagating, long-period (~20 days) perturbations. Their origin was near the dateline at around 15°—20°N.

trough, weak easterlies or westerlies (strong easterlies) at 300 hPa along 15° N, and relatively high (low) 300 hPa temperatures at 20° N. Apparently latent heat released from enhanced convection during an active monsoon is sufficient to raise 300 hPa temperatures nearly 2.5°C along 20° N over the Indian monsoon region (70°—100° E). Over this region, a sharp 300 hPa temperature decrease occurred around 23 August. Therefore, in this study, the date of summer monsoon withdrawal was defined as 23 August.

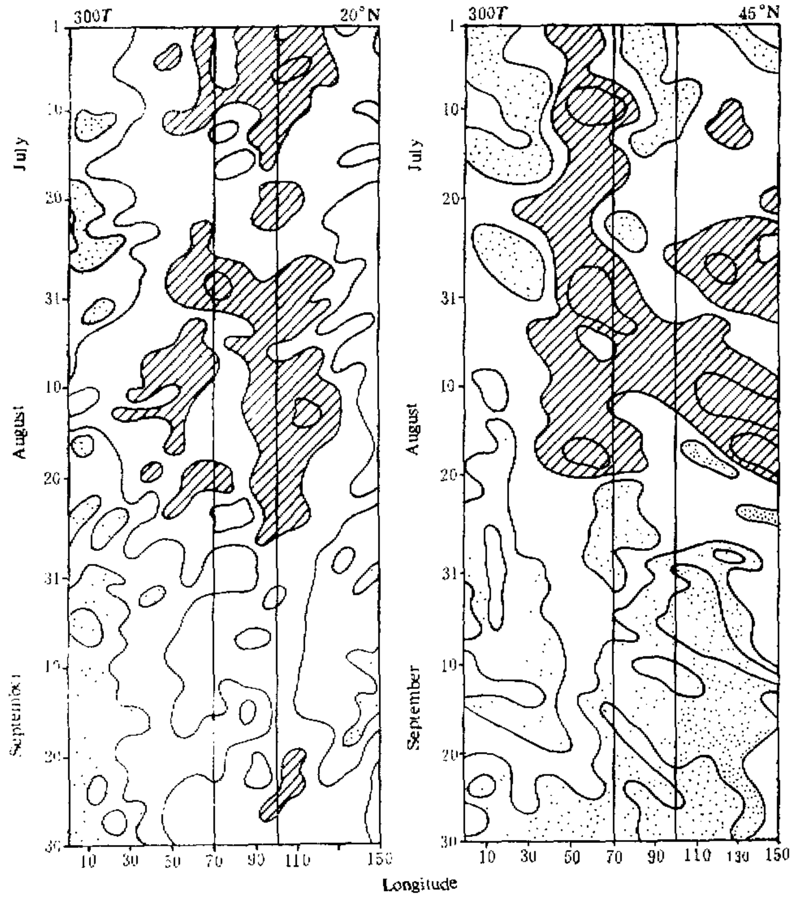

Fig. 2. Longitude-time sections of temperature (°C) at 300 hPa. Left: At 20°N, intervals are 2.5°C. Hatching (shading) denotes regions of higher (lower) than −27.5° (−32.5°) C. Right: At 45°N, intervals are 5°C. Hatching (shading) denotes regions of higher(lower) than −35° (−40°) C.

The 300 hPa temperature along 45° N (Fig. 2, right) exhibits an even more drastic drop than along 20° N on 20 August (approximately 3 days earlier than the monsoon withdrawal). Temperature changes are most pronounced near Kazakh (60°—70° E), where temperatures decreased from above −35° C prior to 20 August to below −40° C after that date.

What mechanisms are responsible for such distinct temperature decreases at these middle-latitudes? What is the relationship between sudden temperature decreases in middle latitudes and monsoon withdrawal over South and Southeast Asia? Before examining these problems, an attempt was made to more clearly identify the spatial character of temperature decreases by computing differences between T averaged over 20 days (29 July—17 August) of active monsoon and 20 days (11—30 September) after monsoon withdrawal. In Fig. 3, immediate-

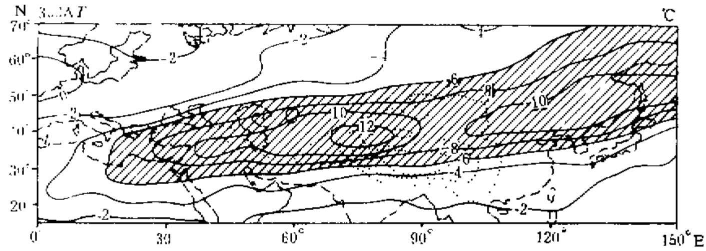

Fig. 3. Differences in the 20 day mean 300 hPa Temperature between 29 July—17 August and 11—30 September, 1979. 2°C intervals with hatching representing regions of less than −6°C temperature differences. Dotted lines represent smoothed topographic heights at 1 km intervals (Berkofsky and Bertoni[7]).

ly evident is an elongated WSW-ENE band of negative ΔT extending from northeast Africa through the northern Tibet to northeast China and beyond. Hence, these temperature decreases represent a very large-scale phenomenon. The largest temperature decrease (exceeding −12°C) occurs over the northwest Tibetan Plateau. This probably indicates local orographic enhancement of temperature changes. Large ΔT decreases are also observed over northern China and eastern Siberia.

The distribution of ΔT in Fig. 3 indicates a zonal character in T changes. Therefore, in the next section, we will examine the processes by which the zonal mean temperatures, over Eurasia, change during late summer.

IV. CHANGES IN ZONAL MEAN TEMPERATURES AT 300 hPa

The time-series of twice-daily, 300 hPa zonal mean T, averaged between 0° and 150°E at 5° latitudinal intervals from 15°N to 60°N, are shown in Figs. 4 and 5, respectively. In Fig. 4, note that there is a sharp decrease in 300 hPa T at 20° and 25°N around 23 August. The monsoon withdrawal from South and Southeast Asia (50°—100°E) as confirmed in Fig. 2 (left), contributes most to this sharp decrease. Further to the north, 300 hPa T at 30° and 35°N decreases gradually after 24 August (Fig. 4). The most interesting feature in Fig. 5 is the near-simultaneous occurrence of sharp decrease in 300 hPa T over extensive latitudinal zones around 18 August. This decrease is most clearly defined near 45°—50°N. Thus, 300 hPa T at higher latitudes decreases sharply about 5 days prior to the monsoon withdrawal. These features may be an indication of some form of lateral coupling between the high and low latitudes during the monsoon withdrawal phase.

To further investigate the nature of 300 hPa T fields, the zonal mean thermodynamic equation can be written as follows:

$$\frac{\partial \bar{T}}{\partial t}=\bar{Q}_m+\bar{Q}_e+\bar{Q}_d, \tag{1}$$

where

$$\bar{Q}_m=-\bar{v}\frac{\partial \bar{T}}{\partial y}-\bar{\omega}\left(\frac{\partial}{\partial p}-\frac{\kappa}{p}\right)\bar{T}, \tag{2}$$

$$\bar{Q}_e=-\left\{\frac{\partial \overline{T'u'}}{\cos\varphi\partial x}+\frac{\partial \overline{T'v'}\cos\varphi}{\cos\varphi\partial y}+\left(\frac{\partial}{\partial p}-\frac{\kappa}{p}\right)\overline{T'\omega'}\right\}, \tag{3}$$

$$\bar{Q}_d=\bar{Q}/C_p, \tag{4}$$

where $\bar{Q}_m$ denotes the sum of latitudinal and vertical T advections due to the mean meridional circulation. Therefore, $\bar{Q}_m$ may be regarded as an "apparent heating" associated with the mean N—S vertical overturnings over Eurasia. Similarly, $\bar{Q}_e$ represents "apparent heating" due to eddy perturbations. The first term on the right side of Eq. (3), which is related to the temperature fluxes across the western (0°) and eastern (150°E) boundaries of our region, is small in comparison to the other quantities in the equation. In this study $\bar{Q}_d$ was evaluated as a residual in Eq. (1). (A list of symbols is given in Appendix).

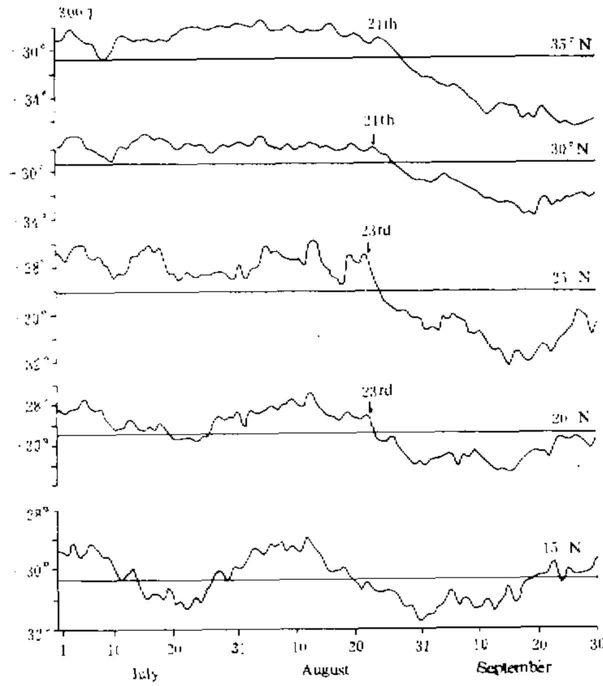

Fig. 4. Time-series of 300 hPa zonal mean temperature averaged between 0° and 150°E at 15°, 20°, 25°, 30°, and 35°N, respectively. Arrows indicate the initial date of sharp T decreases.

Eq. (1) was applied to twice-daily 300 hPa wind and temperature data during the active monsoon period (29 July to 17 August). Twice-daily values were then averaged over this

20 day period. This corresponds to an integration of the thermodynamic equation using a time increment of 12 hr over the active monsoon phase.

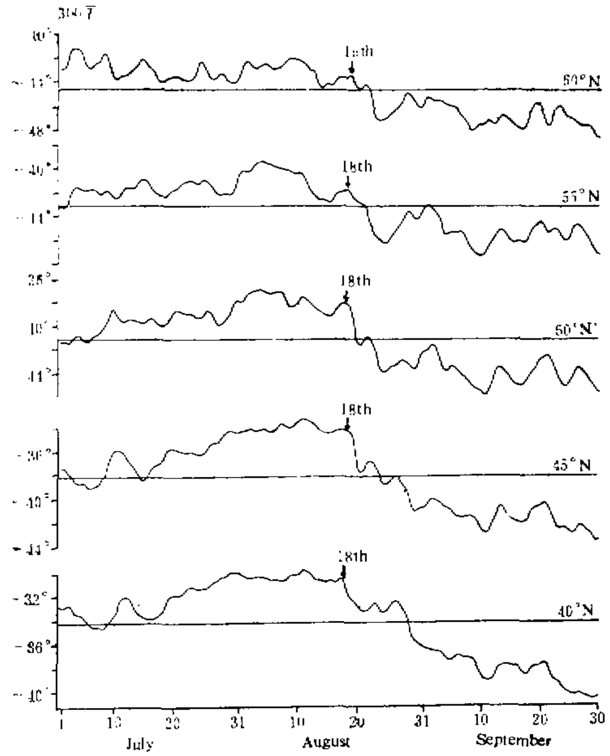

Fig. 5. As in Fig. 4, except at 40°, 45°, 50°, 55°, and 60°N.

Table 1 Computed Results for the Terms (°C day^{-1} unit) in Eq. (1) at 300 hPa, Averaged over the 20 Day Period from 29 July to 17 August, 1979.

Lat. (°N)	$\partial \overline{T}/\partial t$	$\overline{Q}_m$	$\overline{Q}_e$	$\overline{Q}_d$
60	0.03	0.7	0.8	−1.5
55	−0.01	0.1	−0.2	0.1
50	0.00	−0.2	0.2	−0.1
45	−0.02	1.2	0.0	−1.2
40	−0.08	1.0	−0.6	−0.5
35	−0.04	−0.0	−0.4	0.4
30	−0.02	0.1	0.2	−0.3
25	−0.01	−0.5	−0.3	0.8
20	0.00	−1.7	−0.0	1.7

In Table 1, the $\frac{\partial \overline{T}}{\partial t}$-tendency term is quite small. Thus, $\overline{Q}_m$, $\overline{Q}_e$ and $\overline{Q}_d$ effects approximately balance each other. Along 20°N where $\overline{Q}_e$ is near zero, large negative $\overline{Q}_m$-adiabatic cooling ($-1.7°C$ day^{-1}) is offset by large $\bar{Q}_d$-diabatic heating ($+1.7°C$ day^{-1}).D uring the active monsoon phase, an updraft (negative $\bar{\omega}$) is most pronounced near 20°N (Fig. 6, bottom). Associated with this strong ascending motion is an upper (lower) northerly (southerly) $\bar{v}$ equatorward of 25°N (Fig. 6, top). This implies a north-south oriented vertical overturning.

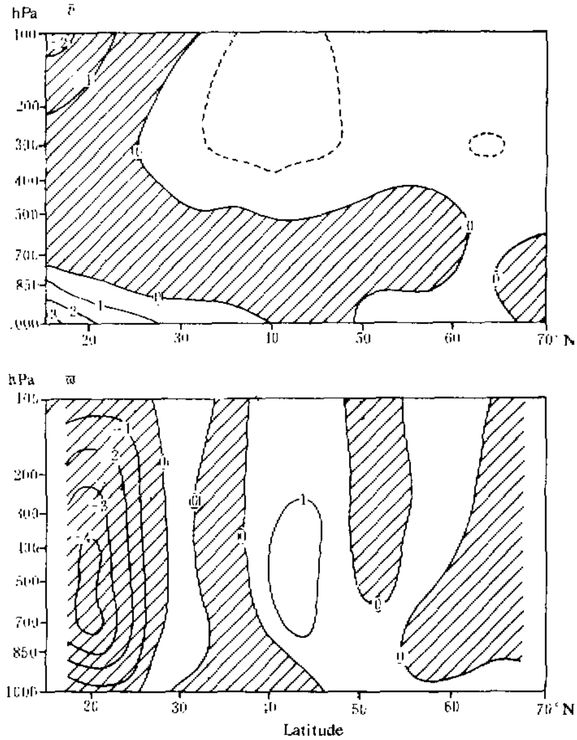

Fig. 6. Latitude-height sections of $\bar{v}$ (top; m s^{-1} unit) and $\bar{\omega}$ (bottom; 10^{-4} hPa s^{-1}unit) averaged over Eurasia (0—150°E) for the 20 day period from 29 July to 17 August, 1979. Intervals are 1 unit. Hatching indicates regions of northerly $\bar{v}$ (top) and regions of upward $\bar{\omega}$ (bottom). Dashed lines represent $+0.5$ m s^{-1} $\bar{v}$.

Fig. 6 (top) also reveals that significant upper outflow is directed northward as southerly $\bar{v}$, with a maximum (0.9 m s^{-1}) near 200 hPa along 40°N. Near 45°N strong downdrafts are associated with pronounced adiabatic heating of about 1.2°C day^{-1} (Table 1). Again this adiabatic heating is primarily offset by diabatic (presumably radiation) cooling of $-1.2°C$ day^{-1} (hereinafter denoted as d^{-1}).

Another feature of interest in Table 1 is the substantially large negative $\bar{Q}_e$ near 35°—40°N. This is associated with pronounced $\overline{v'T'}$ flux divergence at 300 hPa, and strong upward heat flux ($\overline{\omega'T'}$) between 400 and 200 hPa (Fig. 7). In comparison, at 45°N, $\overline{v'T'}$ is maximum (zero divergence) and $\overline{\omega'T'}$ is small near 300 hPa. Consequently, the $\bar{Q}_e$ effect is nearly zero at 45°N, whereas $\bar{Q}_m$ and $\bar{Q}_d$ effects are the dominant terms in achieving 300 hPa $\bar{T}$

balance. Similarly, equatorward of 25° N, both $\overline{v'T'}$ and $\overline{\omega'T'}$ are small near 300 hPa (Fig. 7), and thus the associated $\bar{Q}_e$ effect is not a dominant factor in the $\bar{T}$-balance along 20°—25° N (Table 1).

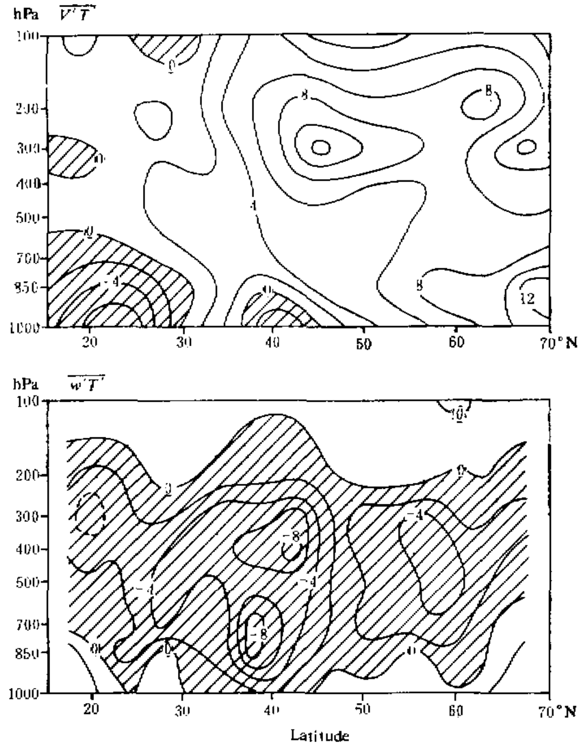

Fig. 7. Latitude-height sections of zonal mean $\overline{v'T'}$ (top; °C m s^{-1} unit) and $\overline{\omega'T'}$ (bottom; 10^{-4} °C hpa s^{-1} unit) averaged over the 20 d period from 29 July to 17 August, 1979. Intervals are 2 units. Hatching denotes regions of southward $\overline{v'T'}$ transports (top) and regions of upward $\overline{\omega'T'}$ transports (bottom).

We also applied Eq. (1) to 300 hPa data during the transition period between 18 and 27 August[2]. Computed 10 d mean values at 300 hPa are shown in Table 2. During this transition period, tendencies exhibit considerable negative values at all latitudes, with a minimum (−0.5°C d^{-1}) near 45°—50°N. Interestingly, $\bar{Q}_m$ along 55°N is a large negative value (−1.6°C d^{-1}). This is due to the development of a strong meridionally oriented $\bar{\omega}$ updraft (Fig. 8, bottom). Pronounced $\bar{Q}_d$—diabatic heating of 1.1°C d^{-1} probably reflects

2) As shown in Figs. 4—5, the initial date of sharp $\bar{T}$ decrease differs considerably along different latitudes. Thus, an attempt was made to integrate Eq. (1) with the initial date as assigned in Figs. 4—5 at each latitude. However, these 10 d mean values are nearly identical to those in Table 2. Computations were also made for the 3 and 5 d periods beginning with the initial date. Again, results are essentially the same as Table 2.

extensive rainfall and associated latent heat release, near the polar front extending across Eurasia. The sum of $\bar{Q}_m$ and $\bar{Q}_d$ effects amounts to only $-0.5°C\ d^{-1}$. Thus, the net effect of $\bar{Q}_m$, $\bar{Q}_d$ and $\bar{Q}_e$ results in the observed $\bar{T}$ decrease of $-0.2°C\ d^{-1}$ at 55°N.

Table 2 As in Table 1, except for the 10 d Period from 18 to 27 August, 1979.

Lat. (°N)	$\partial\bar{T}/\partial t$	$\bar{Q}_m$	$\bar{Q}_e$	$\bar{Q}_d$
60	−0.1	0.2	1.6	−1.9
55	−0.2	−1.6	0.3	1.1
50	−0.5	−0.3	−0.2	0.0
45	−0.5	2.4	−0.7	−2.2
40	−0.3	1.7	−0.5	−1.5
35	−0 2	0.5	−0.5	−0.2
30	−0.1	0.0	−0.0	−0.1
25	−0.1	−0.6	0.1	0.4
20	−0.1	−0.8	0.2	0.5

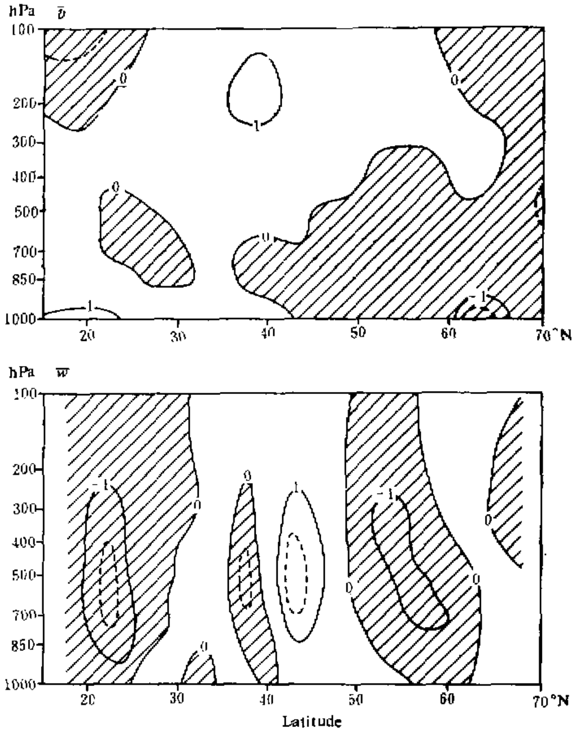

Fig. 8. As in Fig. 6, except for the 10 d period from 18 to 27 August, 1979.

A comparison between Figs. 6 and 8 reveals that downdrafts near 45°N are stronger during the transition period than during the active phase. Accordingly, the transition period is associated with large $\bar{Q}_m$ heating ($+2.4°C\ d^{-1}$). This $\bar{Q}_m$ heating is largely compensated by $\bar{Q}_d$ cooling of $-2.2°C\ d^{-1}$. At this latitude, $\bar{Q}_e$ ($-0.7\ °C\ d^{-1}$) appears to be the major contributor to $\bar{T}$-decreases during the transition period. The negative $\bar{Q}_e$ effect is associated with substantial $\overline{v'T'}$ flux divergence at 300 hPa near 45°N (Fig. 9, top). Undoubtedly this is indicative of the importance of eddy perturbation activity in decreasing the zonal mean temperature during the transition period.

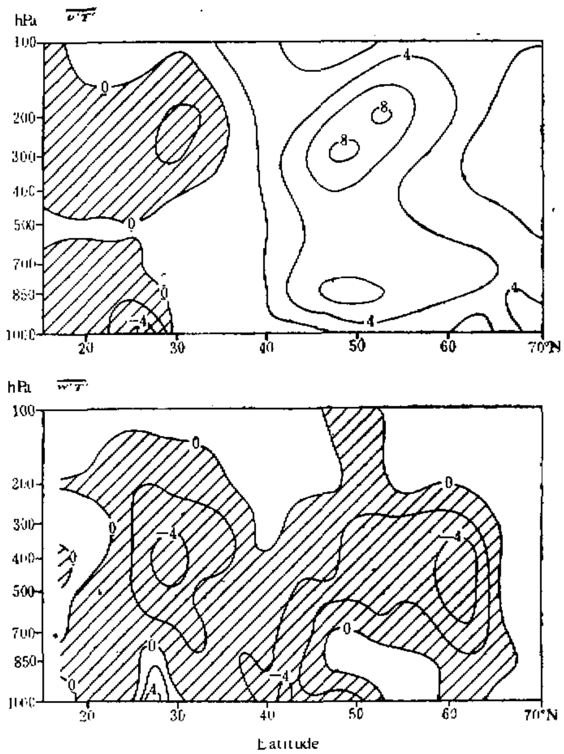

Fig. 9. As in Fig. 7, except for the 10 d period from 18 to 27 August, 1979.

Occurring at 20°N is a major updraft portion of the north–south vertical overturning (Fig. 8). Here, large negative $\bar{Q}_m$ (-0.8°C d^{-1}) is primarily offset by large positive $\bar{Q}_d$ heating ($+0.5$°C d^{-1}). The magnitude of $\overline{v'T'}$ flux convergence (Fig. 9, top) is about 0.2°C d^{-1}. Although $\bar{Q}_e$ is small, it is the same order of magnitude as $\frac{\partial \bar{T}}{\partial t}$ (-0.1°C d^{-1}) and thus, is an important component in long-range temperature prediction (>10 d) near 20°N.

A comparison between Tables 1 and 2 reveals that adiabatic $\bar{Q}_m$ cooling at 20°N decreases from -1.7°C d^{-1} during the active monsoon to -0.8 °C d^{-1} during the transition period. This is consistent with the weakening of updrafts from the active to transition phases (Figs. 6 and 8). Thus, reduced $\bar{Q}_m$ cooling cannot be a controlling factor in $\bar{T}$ decreases along 20°N during the transition period. More important is the reduction of diabatic heating from 1.7 to 0.5°C d^{-1} between the two periods; thus, the reduced $\bar{Q}_d$ heating rate appears to be the major contributor toward $\bar{T}$ decreases during the transition period.

Since $\frac{\partial \bar{T}}{\partial t}$ is near zero during the active monsoon, an approximate balance exists between $\bar{Q}_m$, $\bar{Q}_e$, and $\bar{Q}_d$. Assuming $\frac{\partial \bar{T}}{\partial t}=0$ during the active monsoon phase, we obtain the following relation:

$$\frac{\partial \bar{T}}{\partial t}=\Delta\bar{Q}_m+\Delta\bar{Q}_e+\Delta\bar{Q}_d, \tag{5}$$

where $\frac{\partial \bar{T}}{\partial t}$ denotes the rate of changes in $\bar{T}$ at the nth day during the transition period from 18 ($n=1$) to 27 ($n=10$) August 1979; $\Delta\bar{Q}_m$, $\Delta\bar{Q}_e$, and $\bar{Q}_d$ represent the departure of $\bar{Q}_m$, $\bar{Q}_e$, and $\bar{Q}_d$ at the nth day from the corresponding 20-d mean $\bar{Q}_m$, $\bar{Q}_e$, and $\bar{Q}_d$ values for the active monsoon period.

In Figs. 4 and 5, it is worth noting that $\bar{T}$ poleward of 20°N remains fairly constant prior to 18 August [initial date ($n=1$) for the time integration]. In fact, $\bar{T}$ at 18 August is approximately equal to the 20 d mean $\bar{T}$ for the active monsoon phase. As the time integration proceeds, however, $\bar{T}$ will respond to the changes in $\Delta\bar{Q}_m$, $\Delta\bar{Q}_e$, and $\Delta\bar{Q}_d$ from $n=1$ to 10. Table 3 presents 10 d mean values for $\frac{\partial \bar{T}}{\partial t}$, $\Delta\bar{Q}_m$, $\Delta\bar{Q}_e$, and $\Delta\bar{Q}_d$ during the transition period (18—27 August). Incidentally, the $\frac{\partial \bar{T}}{\partial t}$ value is identical to that in Table 2, and the $\Delta\bar{Q}_m$, $\Delta\bar{Q}_e$, and $\Delta\bar{Q}_d$ values were obtained by subtracting $\bar{Q}_m$, $\bar{Q}_e$, and $\bar{Q}_d$ values of the active phase (Table 1) from the corresponding values during the transition period (Table 2).

Table 3 Comsuted Results for the Terms (°C d^{-1} unit) in Eq. (5) at 300 hPa*

Lat. (°N)	$\partial\bar{T}/\partial t$	$\Delta\bar{Q}_m$	$\Delta\bar{Q}_e$	$\Delta\bar{Q}_d$
60	−0.1	−0.5	0.8	−0.4
55	−0.2	−1.7	0.5	1.0
50	−0.5	−0.2	−0.4	0.1
45	−0.5	1.2	−0.7	−1.0
40	−0.2	0.7	0.1	−1.0
35	−0.2	0.5	−0.1	−0.6
30	−0.1	−0.1	−0.2	0.2
25	−0.1	−0.1	0.4	−0.4
20	−0.1	0.9	0.2	−1.2

* $\Delta\bar{Q}_m$, $\Delta\bar{Q}_e$ and $\Delta\bar{Q}_d$ were obtained by subtracting $\bar{Q}_m$, $\bar{Q}_e$ and $\bar{Q}_d$ Values in Table 1 from the corresponding values in Table 2.

In Table 3 for 20°N $\Delta\bar{Q}_d$ is negative and reflects substantial decrease in $\bar{Q}_d$ diabatic heating from the active monsoon to transition period. Since $\Delta\bar{Q}_m$ and $\Delta\bar{Q}_e$ are both positive, $\Delta\bar{Q}_d$ is essentially important for the negative $\partial\bar{T}/\partial t$ tendency at 20°N. Large negative $\Delta\bar{Q}_d$ is also primarily responsible for the $\bar{T}$ decreases at 35° and 40°N. In comparison, $\Delta\bar{Q}_e$ and $\Delta\bar{Q}_d$ are both negative and equally important for the negative $\frac{\partial \bar{T}}{\partial t}$ tendency at 45°N. Poleward of 50°N, $\Delta\bar{Q}_m$ is negative and contributes to the $\bar{T}$ decrease during the transition period.

Let us again take a look at Fig. 4. At 15°N, 300 hPa $\bar{T}$ fluctuates nearly periodically with an approximate period of 40 d and an amplitude of about 1°C. Similar fluctuations, although much less distinct, also occur at 20° and 25°N. The existence of a phase difference between fluctuations at 15°, 20°, and 25°N is noted. Based on daily satellite cloudiness data for the eight summers of 1965 to 1972, Murakami[8] and Yasunari[9] showed that cloudiness perturbations propagated northward over the Indian Ocean and the Indian subcontinent with an average phase speed of about 1° latitude per day.

To facilitate this study, the prefiltering process described by Eq. (3) of Murakami[6] was applied to the original, twice-daily $\bar{T}$, $\bar{Q}_m$, $\bar{Q}_e$ and $\bar{Q}_d$ data. By assigning $\omega_0=\frac{2\pi}{40}$ d^{-1} and $\omega_1=\frac{2\pi}{30}$ d^{-1}, it is possible to obtain 40 d filtered time series data. Since the $\bar{Q}_m$ ($\bar{Q}_e$) term is nonlinear with respect to zonal mean (eddy perturbation) quantities, the resulting $\bar{Q}_m(\bar{Q}_e)$ values are the rates of temperature (sensible heat) transfer to 40 d perturbations via nonlinear coupling between zonal (eddy) transient disturbances of all period ranges. Fig. 10 depicts the filtered results for 300 hPa $\bar{T}$, $\bar{Q}_m$, $\bar{Q}_e$ and $\bar{Q}_d$ at 17.5°N. The unavailability of ω-values at 15°N (southern boundary) prevents the computations of $\bar{Q}_m$, $\bar{Q}_e$ and $\bar{Q}_d$ terms at that latitude.

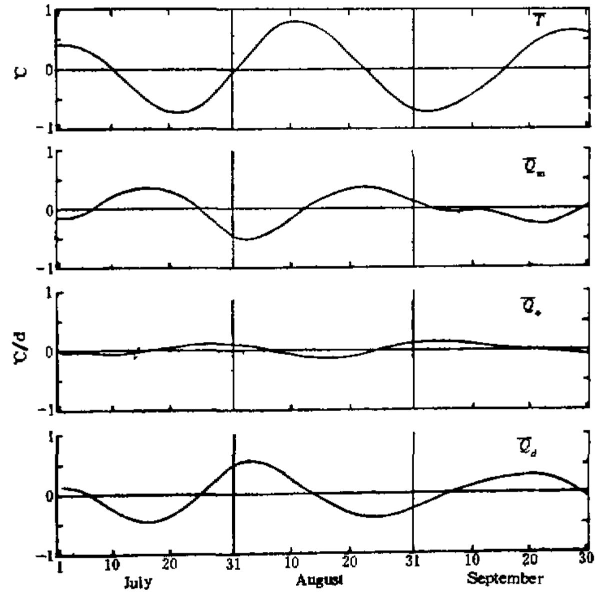

Fig. 10. Time-series of 40 d filtered $\bar{T}$ data (°C) and 40 d filtered $\bar{Q}_m$, $\bar{Q}_e$ and $\bar{Q}_d$ values (°C d^{-1}) at 300 hPa obtained from Eq. (1) for 17.5°N during the period from 00 GMT, 1 July to 12 GMT, 30 September, 1979. See text for further information.

In Fig. 10, $\bar{T}$ is a maximum (about 1°C above normal) on approximately 12 August. At this time, both $\bar{Q}_m$ and $\bar{Q}_d$ are near zero. $\bar{Q}_m$ is a minimum (above-normal updraft) around 2 August. Concurrently, $\bar{Q}_d$ is a maximum (above-normal heating). There exists about a 10 d (a quarter of a cycle) phase difference between the anomalous $\bar{Q}_d(\bar{Q}_m)$ heating (cooling) on 2 August and the $\bar{T}$ maximum on 12 August. Since anomalous $\bar{Q}_d$ heating is slightly more pronounced than $\bar{Q}_m$ cooling, the net sum $(\bar{Q}_m+\bar{Q}_d)$ results in a small positive value of about 0.1°C d^{-1} on 2 August, at which time the 40 d filtered $\bar{T}$ increases rapidly as shown in Fig. 10 (top). A similar 10 d (a quarter of a cycle) phase difference can also be found between the anomalous $\bar{Q}_d$ $(\bar{Q}_m)$ heating (cooling) on 22 August and the $\bar{T}$ minimum on 2 September. It appears that the changes in 40 d filtered $\bar{T}$ are primarily associated with a small imbalance between the large $\bar{Q}_m$ effect and the predominant $\bar{Q}_d$ forcing.

On 22 July when 300 hPa $\bar{T}$ is below normal, the atmosphere is probably more unstable than usual. This unstable atmosphere tends to increase upward motions(decreasing $\bar{Q}_m$) and associated rainfall (increasing $\bar{Q}_d$). The result is a small $(\bar{Q}_m+\bar{Q}_d)$ net heating which contributes to an increase in 300 hPa $\bar{T}$ around 2 August. The atmosphere becomes more stable than normal on 12 August due to increased 300 hPa $\bar{T}$. This is followed by the development of descending (anomaly) motions and anomalous $\bar{Q}_d$-cooling (below normal convection) between 12 August and 2 September. During this period, a small $(\bar{Q}_m+\bar{Q}_d)$ net cooling effectively decreases 300 hPa $\bar{T}$, reaching a minimum on approximately 2 September. This completes one cycle of 40 d oscillation for $\bar{T}$, $\bar{Q}_m$ and $\bar{Q}_d$ during late summer and early fall.

V. CONCLUDING REMARKS

Evidence has been presented that the NMC objectively analyzed u, v, and T data are adequate to describe some of the characteristic features of large-scale circulation over Eurasia (15°—70°N, 0°—150°E) during the late summer (July-September) of 1979. In summary, the main results of this study are:

(1) The summer monsoon region (15°—20°N,70°—100°E) is characterized by significant changes in wind and temperature in association with the phase shift between active and break monsoons. These changes are most clearly defined at the 300 hPa level, where temperature is above (below) normal and easterly winds are weaker (stronger) than usual during active (break) phases.

(2) Of particular interest is the sudden decrease in 300 hPa zonal mean $\bar{T}$ averaged over Eurasia (0°—150°E) along 20°—25°N, around 23 August 1979. The monsoon withdrawal from South Asia contributed the most to this sharp $\bar{T}$ decrease. A similar decrease in 300 hPa $\bar{T}$ also occurred over extensive mid-latitude zones (40°—55°N) around 18 August i. e. about 5 days prior to the monsoon withdrawal.

(3) These abrupt $\bar{T}$ decreases appear to be associated with changes in the intensity and locaiion of zonal mean vertical overturnings over Eurasia during the transition (18—27 August) from summe rto fall. Near 20°—25°N, zonal mean updrafts weakened considerably during thetransition period. This results in the reduction of diabatic heating rate and the decrease in $\bar{T}$ near 20°—25°N. Further to the north, the intensification of zonal mean downdrafts and the increased cooling (radiative) rate are primarily responsible for the sharp $\bar{T}$ decrease near 35°—45°N. The development of zoal mean updrafts occurs around 55°N during the transition period. Here, the adiabatic cooling in updrafts is the major factor affecting the $\bar{T}$ decrease.

The foregoing is admittedly somewhat speculative as it is based on data for the late summer of 1979 only. It should also be noted that the 1979 summer monsoon near India was not a typical monsoon season. The onset of the monsoon was delayed over 2 weeks. This, together with the early monsoon withdrawal, caused severe deficient rains over India and Indochina. Therefore, the results obtained from this study should be regarded as preliminary until tested and veiified for many years of data.

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APPENDIX: LIST OF SYMBOLS

λ, ϕ longitude and latitude
p pressure
t time
dx, dy $a\cos\phi d\lambda$ and $ad\phi$; a=radius of earth
u, v zonal and meridional wind components
ω vertical p-velocity
T temperature
Q diabatic heating rate per unit mass
Cp heat capacity of air at constant pressure
k R/c_p; R, gas constant of air and C_p, heat capacity of air at constant pressure
$(\overline{\ })$ zonal mean, averaged between 0° and 150° E
$(\ )'$ departure from zonal mean