Seasonal Heating of the Tibetan Plateau and Its Effects on the Evolution of the Asian Summer Monsoon

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Abstract

Using the objectively analyzed FGGE II-b upper-air data, the large-scale circulation, heat sources and moisture sinks over the Tibetan Plateau and surrounding areas are examined for a 9-month period from December 1978 to August 1979. In addition to the FGGE data, special soundings obtained during the Chinese Qinghai-Xizang (Tibet) Plateau Meteorological Experiment (QXPMEX) from May to August 1979 are also used in the objective analyses.

The evolution of the large-scale flow patterns, temperature, outgoing longwave radiation (OLR) and vertical circulation is described in order to identify the distinct seasonal changes from winter to summer that lead to the onset of the Asian summer monsoon. The Tibetan Plateau maintains a large-scale thermally driven vertical circulation which is originally separated from the planetary-scale monsoon system. The rising motion exists only on the western Plateau in winter and then spreads to the whole Plateau as the season progresses. The monsoon onset over Asia is an interaction process between the Plateau-induced circulation and the circulation associated with the principal rainbelt migrating northward.

During winter the Plateau is a heat sink, but it is surrounded by regions of more intense cooling. In spring the Plateau becomes a heat source, but the cooling in the surrounding areas continues. The sensible heat flux from the surface provides the major source of heating on the Plateau. However, additional contribution from condensation heating is observed in the western Plateau during all seasons and, more significantly, in the eastern Plateau during summer. The sensible heating of the elevated Plateau surface and the radiative cooling in the environment maintain the horizontal temperature contrast that drives the thermally direct vertical circulation.

The detailed examination of the warming process of the upper troposphere during two transition periods, i.e., the onset of the Southeast Asian monsoon in May and that of the Indian monsoon in June, reveals that the temperature increase over the eastern Plateau during the first onset was mainly the result of diabatic heating, whereas that over the Iran-Afghanistan-western Plateau region leading to the second onset was caused by intense subsidence.

There are large diurnal variations in the boundary layer and vertical circulation over the Plateau. As a result of diurnal heating of the surface, a deep mixed layer of nearly uniform potential temperature exists over the Plateau in the evening (1200 UTC), suggesting the role of thermal convection in the upward transport of heat. However, moisture is not well mixed vertically and there is a large horizontal temperature gradient in the boundary layer. From late spring to summer the boundary layer becomes more stable for dry convection. On the other hand, the vertical distributions of equivalent potential temperature in late spring and afterwards show a conditionally unstable stratification for moist convection with the increase of moisture of surface air.
1. Introduction

The importance of the Tibetan Plateau (the Qinghai-Xizang Plateau) as an elevated heat source for the establishment and maintenance of the Asian summer monsoon circulation has been discussed by many authors [see e.g., Yeh (1981, 1982), Gao et al., (1981), Murakami (1987a, b), Sumi and Murakami (1984) for reviews]. Recently, rapid progress has been made in observational studies using the data obtained during the First Global Atmospheric Research Program (GARP) Global Experiment (FGGE) and the Chinese Qinghai-Xizang Plateau Meteorology Experiment (QXPMEX) in 1979 (see, e.g., Johnson et al., 1987; Yeh, 1988; Zhang et al., 1988; Ji et al., 1989). In addition, a series of recent numerical experiments in which large-scale uplifts of the Tibetan Plateau were linked to the climatic changes on the geological time scale elucidated the effects of the Plateau on the planetary-scale circulations (Kutzbach et al., 1989; Manabe and Broccoli, 1990).

Yeh et al. (1957) [also in Staff Members of Academia Sinica (1958)], using the data for 1954–1956, discussed the influence of the Tibetan Plateau on the circulation over East Asia. They noted that, in summer, the wind blows cyclonically around the Plateau at the 3 km level, while at 6 km and above there is a huge anticyclone with its center over the southern periphery of the Plateau. This indicates that the temperature over the Plateau is higher than the surrounding air. They also estimated the mean long-wave radiative cooling, the short-wave absorption, the condensation heating and the sensible heat flux from the ground over the Tibetan Plateau in summer and in winter. From these they inferred that the Plateau is a heat source in summer and probably a heat sink in winter.

Flohn (1957, 1960), using observations from earlier expeditions, also concluded the existence of a thermal anticyclone above the Plateau. He suggested that the seasonal heating of the elevated surface of the Tibetan Plateau and the consequent reversal of the meridional temperature and pressure gradients south of 35°N trigger the large-scale change of the general circulation over East Asia and the monsoon burst over the Indian sub-continent. Similar suggestions were also made by Koteswaram (1958) and by Murakami (1958).

Since the 1950's a network of surface and upper-air stations on the Tibetan Plateau has been continuously expanded by the People's Republic of China. Using the Chinese observations, Flohn (1968) examined summer meteorological conditions on the Plateau and its vicinity. He suggested the importance of both the sensible heat flux in the semi-arid western portion of the Plateau and the latent heat release over the Himalayas, Assam, Bengal and adjacent mountains for the occurrence of the warm center of the Tibetan anticyclone in the upper troposphere.

Yeh and Gao (1979) summarized the Chinese research on the meteorology of the Tibetan Plateau prior to FGGE. They tabulated monthly mean values of the sensible heat flux and the precipitation amounts in the western and eastern Plateau and mean evaporation rates for the whole Plateau. These are based on long-term records obtained at surface stations with a mean elevation of about 4 km (600 hPa), and the border between the two parts of the Plateau was placed at 85°E. They also listed representative January and July values of solar absorption and long-wave radiation fluxes on the Plateau. Using these, Yeh and Gao estimated the monthly mean values of net atmospheric heat sources over the Plateau (Fig. 1). Their results showed that in the winter half-year the troposphere over the Plateau is a cold source, with maximum cooling in December. In the summer half-year it is a heat source, with maximum intensity in June. They noted also larger net heating rates in the western Plateau in all months. This is mainly due to the larger values of the sensible heat flux in the western Plateau, which attains its maximum value in June. On the other hand, the contribution from condensation heating is larger in the eastern Plateau, with the maximum value in July.

Meanwhile many numerical (Godbole, 1973; Manabe and Terpstra, 1974; Hahn and Manabe, 1975; Kuo and Qian, 1981, 1982; Chen and Dell’Osso, 1986; Zheng and Lion, 1986; Kitoh and Tokioka, 1987) as well as laboratory (Yeh and Chang, 1974; Chang et al., 1977; Yang et al., 1980; Chen and Li, 1982) experiments were performed to clarify the thermal and orographic effects of the Tibetan Plateau on the general circulation.

Since FGGE (December 1978–November 1979) renewed attempts have been made to examine the heat sources over the Tibetan Plateau and surrounding areas. In parallel to FGGE, Chinese scientists conducted the Qinghai-Xizang Plateau Meteorology Experiment (QXPMEX 79) during the 1979 summer (May–August). During QXPMEX 79, temporary surface and upper-air stations and radiation stations were added to the previously data-void region of the western and central Tibetan Plateau (see e.g., First Research Group of QXPMEX, 1984; Shen et al., 1984, 1987; Zhang et al., 1988). They conducted the second experiment from August 1982 to July 1983 (QXPMEX 82–83).

While the early studies of the Tibetan heat source were mainly based on the heating components measured at the surface, several attempts have been made to obtain the heat and moisture budgets of the troposphere using the enhanced upper-air observa-
tions during FGGE. Wei et al. (1983) and Schaack et al. (1990) obtained the seasonal global distributions of the mean tropospheric heating rate using the FGGE Level III-a and III-b analyses, respectively. They showed that in northern summer there is a maximum heating rate over the Plateau. Nitta (1983), using FGGE II-b station data, obtained the mean vertical profiles of the tropospheric heat source and moisture sink over several parts of the eastern Tibetan Plateau for a 100-day period from the end of May to early September 1979. He found that, over the eastern Plateau, heating occurs in a deep tropospheric layer and that the sensible heat flux from the surface and the release of latent heat of condensation contribute nearly equally to the total heating in this period.

Luo and Yanai (1983), using objectively analyzed FGGE II-b data, presented a detailed analysis of the time evolution of the large-scale precipitation, low-level (850 hPa) wind, moisture and vertical motion fields, and horizontal moisture transports over the Tibetan Plateau and surrounding areas during a 40-day period (26 May–4 July 1979) which included the onset of the Indian summer monsoon (Appendix C). They showed that the mean 850 hPa wind field exhibits a pronounced inflow towards the Plateau with diurnally varying intensity, confirming the earlier finding of Yeh et al. (1957).

Subsequently, Luo and Yanai (1984) examined the large-scale heat and moisture budgets of the same area during the 40-day period. They showed that there is a deep heating layer over the Plateau and that the mean heating rate of ~3K day^{-1} in the 200–500 hPa layer above the Plateau is as intense as that over the Assam-Bangladesh region. They also identified the principal components of the heat sources as the sensible heat flux from the surface and the additional condensation heat from the summer rains. They also showed that a deep, nearly mixed boundary layer with high potential temperature is observed over the Plateau in the evening (1200 UTC). The mixed layer was pronounced during the pre-onset period. They suggested that dry thermal convection originating near the heated surface in the afternoon hours may reach the upper troposphere and deposit the sensible heat there. Wang and Luo (1989) analyzed the heat and moisture budgets over the Plateau and its vicinity for a period from 15 May to 20 July using the ECMWF (European Centre for Medium Range Forecast) FGGE III-b analyses. They obtained the horizontal and vertical distributions of heat and moisture budget residuals and moisture transports similar to those discussed by Luo and Yanai (1983, 1984).

He et al. (1987) extended the heat and moisture budget analysis of Luo and Yanai to an 80-day period (16 April–4 July 1979) and over a larger domain (0°–50°N, 40°–130°E). They showed that, from spring to summer 1979, the general circulation over Asia underwent two distinct stages of abrupt transitions, resulting in the successive onsets of the Southeast Asian and Indian monsoons. The two onsets were related to similar two stages of upper tropospheric warming over the Asian land mass. The rapid warming occurred first over longitudes east of 85°E (the eastern Tibetan Plateau-South China), and then over longitudes west of 85°E (Iran-Afghanistan-the western Plateau). The successive warming over these regions is the primary cause of the successive reversals of the meridional temperature gradient on the south sides of these regions. They also showed the presence of a large-scale vertical circulation which is induced thermally by the Tibetan Plateau. This circulation was present separately from the migrating planetary-scale monsoon system. The existence of the Plateau-induced vertical circulation is evident in the distribution of subsidence in the surrounding desert areas.
We note, however, that the additional data obtained by QXPMEX-79 were not available to the works of Luo and Yanai (1983, 1984) and He et al. (1987). Only a very limited use of the QXPMEX data in conjunction with FGGE data has been attempted by Chen et al. (1985a, b), Feng et al. (1985) and Zhu and Fan (1988).

The main objectives of the present paper are 1) to review the observational studies of the heat sources and moisture sinks over the Tibetan Plateau, 2) to present the results of our recent work on the seasonal changes in the heat and moisture budgets over the Plateau and surrounding areas from winter to summer, and 3) to examine the roles played by the Plateau in the time evolution of the general circulation especially during the onset of the Asian summer monsoon. We have extended the heat and moisture budget analysis to a 9-month period (December 1978–August 1979). In addition to the FGGE II-b data, the data obtained at additional stations during QXPMEX (May–August 1979) have been fully included in the analysis.

In Section 2 we review the methods used in the literature to obtain the atmospheric heat sources and moisture sinks. Then we discuss the data and analysis procedures of the current work. Section 3 describes the seasonal changes in the large-scale flow patterns, temperature, outgoing longwave radiation (OLR) and the vertical motion fields from December 1978 to August 1979 to identify the major circulation changes that lead to the onset of the Asian summer monsoon. In Section 4 we discuss the spatial distributions and seasonal changes of heat sources and moisture sinks. Section 5 examines the roles played by the Tibetan Plateau in the onset of the Asian summer monsoon. In Section 6 we present a preliminary analysis of the planetary boundary layer on the Plateau in relation to the possible heating mechanisms. The summary and conclusions are given in Section 7.

2. Methods to determine the heat sources and moisture sinks

a. Review of the basic equations

In the literature three different methods have been used to deduce the heat sources and moisture sinks in the atmosphere.

1) METHOD A (TROPOSPHERIC BUDGETS)

The apparent heat source $Q_1$ and the apparent moisture sink $Q_2$ (Yanai, 1961; Nitta, 1972; Yanai et al., 1973) are defined by

\[ Q_1 = c_p \left( \frac{p}{p_0} \right)^{\kappa} \left( \frac{\partial \theta}{\partial t} + \mathbf{v} \cdot \nabla \theta + \omega \frac{\partial \theta}{\partial p} \right), \]  
\[ Q_2 = -L \left( \frac{\partial q}{\partial t} + \mathbf{v} \cdot \nabla q + \omega \frac{\partial q}{\partial p} \right). \]

In (1) and (2), $\theta$ is the potential temperature, $q$ the mixing ratio of water vapor, $v$ the horizontal velocity, $\omega$ the vertical $p$-velocity, $p$ the pressure. $\kappa=R/c_p$, $R$ and $c_p$ are the gas constant and the specific heat at constant pressure of dry air, $p_0=1000$ hPa, and $L$ the latent heat of condensation. $\nabla$ is the isobaric gradient operator. The overbar denotes the running horizontal average introduced by the objective analysis and subsequent smoothing.

$Q_1$ and $Q_2$ are the residuals of heat and moisture budgets of the resolvable motion, and may be interpreted as

\[ Q_1 = Q_R + L \left( \bar{\dot{e}} - \bar{e} \right) - \frac{\partial}{\partial p} s' \bar{\omega}', \]  
\[ Q_2 = L \left( \bar{\dot{e}} - \bar{e} \right) + L \frac{\partial}{\partial p} q' \bar{\omega}', \]

where $Q_R$ is the radiative heating rate, $\bar{e}$ the rate of condensation per unit mass of air, $e$ the rate of re-evaporation of cloud and rain water, $s=c_p T+g z$ is the dry static energy, $T$ the temperature and $g$ the acceleration of gravity. The prime denotes the deviation from the average due to unresolved eddies such as cumulus convection and turbulence (see Appendix A).

$Q_1$ and $Q_2$ are called the “apparent” heat source and moisture sink because of possible contributions resulting from unresolved eddies. The eddy vertical flux terms may have significant contributions to $Q_1$ and $Q_2$ in a highly convective situation. From (3) and (4) we find

\[ Q_1 - Q_2 - Q_R = \frac{\partial}{\partial p} h' \bar{\omega}', \]

where $h=s+L q$ is the moist static energy. (5) has been widely used to measure the activity of cumulus convection (e.g., Nitta, 1972, 1975, 1977; Yanai et al., 1973, 1976; Ogura and Cho, 1973; Cheng, 1989; Cheng and Yanai, 1989).

Integrating (3) and (4) from the tropopause pressure $p_T$ to the surface pressure $p_s$, we obtain

\[ <Q_1> = <Q_R> + LP + S, \]  
\[ <Q_2> = L(P - E), \]

where

\[ <\geq> = \frac{1}{g} \int_{p_T}^{p_s} \left( \int dp \right) dp, \]

$P$, $S$ and $E$ are respectively the precipitation rate, the sensible heat flux and the evaporation rate per unit area at the surface (see Yanai et al., 1973; Luo and Yanai, 1984). We note that

\[ <Q_1> - <Q_2> = <Q_R> + S + LE. \]
This method does not require separate estimates of \( <{Q_R}> \), \( S \), \( L_P \) and \( L_E \), but Eqs. (3)–(7) and (9) are useful for the interpretation of the budget results. Careful comparison between the horizontal distributions of \( <Q_1> \) and \( <Q_2> \) as well as comparison between the vertical distributions of \( Q_1 \) and \( Q_2 \) will yield valuable informations on the nature of heating processes. The method has been successfully used in the tropics when high-quality sounding data were available (e.g., Yanai et al., 1973; Nitta and Esbensen, 1974; Thompson et al., 1979).

2) METHOD B (SURFACE MEASUREMENTS)

This method gives only the vertically integrated quantities, \( <Q_1> \) and \( <Q_2> \). We note that (6) may be written as

\[
< Q_1 > = < Q_{SR} > + LR_1 - LR_2 + S + L_P \tag{10}
\]

where \( <Q_{SR}> \) is the absorption of solar radiation by the air column of a unit cross section area, \( LR_1 \) the effective longwave radiation from the earth’s surface, \( LR_2 \) the outgoing longwave radiation at the tropopause. In this method \( <Q_1> \) is obtained by the sum of individual terms on the right-hand side of (10). The advantage of this method is that \( <Q_1> \) can be obtained from \( S \) and \( L_P \) using routine surface observations with some standard estimates of radiation terms. This method has been used by many authors (e.g., Yeh and Gao; 1979; Yeh, 1981; Yao et al., 1984; Chen et al., 1985b; Ji et al., 1985, 1986).

A difficulty in this method is uncertainty in the values of the drag coefficient appropriate for the land surface that are needed to estimate \( S \) (e.g., Chen et al., 1985b). The representativeness of observing sites of \( S \) and \( L_P \) may pose another problem. Some authors obtain \( S \) from the following surface energy balance.

3) METHOD C (SURFACE ENERGY BALANCE)

We define the residual

\[
R = SR_1 - LR_1, \tag{11}
\]

where \( SR_1 \) is the flux of solar radiation at the surface. Then the energy balance at the surface requires

\[
S + LE = R - HG, \tag{12}
\]

where \( HG \) is the heat flux given to soil. Thus this method gives the total (sensible and latent) heat flux entering the atmosphere from the earth surface. If the Bowen ratio is known, we can separate \( S \) and \( LE \). If \( S \) is estimated independently, (12) may be used to obtain \( LE \) as the residual.

With (9) we find

\[
S + LE = < Q_1 > - < Q_2 > - < Q_{SR} >. \tag{13}
\]

Thus this method gives the flux of moist static energy supplied from the ground surface to the atmosphere, but it does not give the tropospheric heating rate \( <Q_1> \) separately.

Table 1 shows a selective list of observational studies on the heat sources and moisture sinks over the Tibetan Plateau, with the authors, year of publication, method used, domain and period of analysis, and data sources.

b. Data and analysis procedures of the present work

The principal data used in this study are the final FGGE Level II-b upper-air profiles in the domain 0°–50°N, 40°–130°E, from 1 December 1978 to 31 August 1979. The dataset includes rawinsonde and pilot wind data from land stations and ships, aircraft dropwindsonde data, and a limited amount of LIMS (Limb Infrared Monitor of the Stratosphere) temperature soundings from the Nimbus-7 satellite (e.g., Gille and Russel, 1984). In addition, special soundings obtained during the Chinese QXPMEX (May–August 1979) are also used. As shown in Fig. 2, the soundings from temporary QXPMEX stations greatly enhanced the data coverage over the western and central Tibetan Plateau.

A version of the successive correction method (Cressman, 1959) is employed to objectively analyze the horizontal wind components, potential temperature and water vapor mixing ratio at the earth surface and at standard pressure levels (850, 700, 500, 400, 300, 250, 200, 150, 100, 70, and 50 hPa) on a 2.5°×2.5° grid mesh. The ECMWF Level III-b analyses (Bengtsson et al., 1982) are used to provide the first-guess fields. The objective analyses are made twice a day (0000 and 1200 UTC). We also use the daily outgoing longwave radiation (OLR) data obtained from the Tiros N satellite (Gruber and Krueger, 1984) from January to August 1979.

In this work we follow Method A and obtain \( Q_1 \) and \( Q_2 \) using (1) and (2). The vertical p-velocity \( \tilde{\omega} \) is obtained from the horizontal divergence by vertically integrating the continuity equation

\[
\frac{1}{a \cos \phi} \left( \frac{\partial \tilde{u}}{\partial \lambda} + \frac{\partial \tilde{v}}{\partial \phi} (\tilde{v} \cos \phi) \right) + \frac{\partial \tilde{\omega}}{\partial p} = 0, \tag{14}
\]

with the surface boundary condition

\[
\tilde{\omega} = \tilde{\omega}_s = -gp_s \left( \frac{\tilde{u}_s}{a \cos \phi} \frac{\partial z_s}{\partial \lambda} + \frac{\tilde{v}}{a} \frac{\partial z_s}{\partial \phi} \right) \quad \text{at } p = p_s. \tag{15}
\]

In (14) and (15) \( \tilde{u} \) and \( \tilde{v} \) are the zonal and meridional components of the horizontal wind \( \vartheta \), \( a \) the mean earth radius, \( \lambda \) the longitude, \( \phi \) the latitude, \( \rho \) the density, and \( z_s \) the terrain height. The suffix \( s \) denotes the surface value. The values of smoothed terrain heights are taken from the NMC (National Meteorological Center) Level III-a data tape.
Table 1. Selected Studies of heat sources and moisture sinks over the Tibetan Plateau.

<table>
<thead>
<tr>
<th>Authors</th>
<th>Estimates (Method)$^1$</th>
<th>Domain</th>
<th>Period</th>
<th>Data Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1) Yeh and Gao (1979)</td>
<td>$&lt;Q_1&gt;$ and $&lt;Q_2&gt;$ (B)</td>
<td>E. and W. Plateau</td>
<td>Whole year</td>
<td>(Climatology)</td>
</tr>
<tr>
<td>(2) Gao and Lu (1979)</td>
<td>$S + LE$ (C)</td>
<td>Plateau and China Plain</td>
<td>Jan., Apr., July, October</td>
<td>(Climatology)</td>
</tr>
<tr>
<td>(3) Nitta (1983)</td>
<td>$Q_1$ and $Q_2$ (A)</td>
<td>E. Plateau</td>
<td>30 May–7 September 1979</td>
<td>FGGE II-b</td>
</tr>
<tr>
<td>(5) Yao et al. (1984)</td>
<td>$&lt;Q_1&gt;$ only (B)</td>
<td>Plateau and China Plain</td>
<td>May–August 1979</td>
<td>QXPMEX 79</td>
</tr>
<tr>
<td>(6) Luo and Yanai (1984)</td>
<td>$Q_1$ and $Q_2$ (A)</td>
<td>$60^\circ$–$130^\circ$E $10^\circ$–$50^\circ$N</td>
<td>26 May–4 July 1979</td>
<td>FGGE II-b</td>
</tr>
<tr>
<td>(7) Chen et al. (1985a)</td>
<td>$Q_1$ and $Q_2$ (A)</td>
<td>E. and W. Plateau</td>
<td>1 June–31 August 1979</td>
<td>QXPMEX 79</td>
</tr>
<tr>
<td>(8) Chen et al. (1985b)</td>
<td>$&lt;Q_1&gt;$ and $&lt;Q_2&gt;$ (B)</td>
<td>Whole Plateau</td>
<td>19 May–31 August 1979</td>
<td>FGGE II-b, QXPMEX 79</td>
</tr>
<tr>
<td>(9) Feng et al. (1985)</td>
<td>$Q_1$ and $&lt;Q_2&gt;$ (A) (B)</td>
<td>W. Plateau</td>
<td>19 May–31 August 1979</td>
<td>FGGE II-b, QXPMEX 79</td>
</tr>
<tr>
<td>(10) Ji et al. (1985)</td>
<td>$&lt;Q_1&gt;$ only (B)</td>
<td>Whole Plateau</td>
<td>August 1982–July 1983</td>
<td>QXPMEX 82-83</td>
</tr>
<tr>
<td>(11) Ji et al. (1986)</td>
<td>$&lt;Q_1&gt;$ only (B)</td>
<td>Whole Plateau</td>
<td>Nov. 1982–Jan. 1983</td>
<td>QXPMEX 82-83</td>
</tr>
<tr>
<td>(12) He et al. (1987)</td>
<td>$Q_1$ and $Q_2$ (A)</td>
<td>$40^\circ$–$130^\circ$E $0^\circ$–$50^\circ$N</td>
<td>16 April–4 July 1979</td>
<td>FGGE II-b, III-b</td>
</tr>
<tr>
<td>(13) Zhu and Fan (1988)</td>
<td>$Q_1$ and $Q_2$ (A)</td>
<td>Plateau and China Plain</td>
<td>May–August 1979</td>
<td>FGGE III-b, QXPMEX 79</td>
</tr>
<tr>
<td>(14) Zhang et al. (1988)</td>
<td>$S + LE$ (C)</td>
<td>Whole Plateau</td>
<td>May–August 1979</td>
<td>QXPMEX 79</td>
</tr>
<tr>
<td>(16) This work</td>
<td>$Q_1$ and $Q_2$ (A)</td>
<td>$40^\circ$–$130^\circ$E $0^\circ$–$50^\circ$N</td>
<td>December 1978–August 1979</td>
<td>FGGE II-b, III-b QXPMEX 79</td>
</tr>
</tbody>
</table>

$^1$ See the text for Methods A, B and C.

To reduce errors in estimates of heating rates in the upper troposphere, we impose an additional condition,

$$\bar{\omega} = \dot{\bar{\omega}} + \left( \frac{\partial \dot{\theta}}{\partial t} + \dot{\mathbf{v}} \cdot \nabla \dot{\theta} \right) \left/ \left( \frac{\partial \dot{\theta}}{\partial p} \right) \right. \text{ at } p = p_T.$$  
(16)

The values of the tropopause pressure $p_T (\lambda, \phi, t)$ are objectively analyzed using the tropopause level data reported at radiosonde stations (see Appendix B). The original estimates of the horizontal divergence $D_0$ are corrected by adding

$$D' = \frac{\bar{\omega}_T - \bar{\omega}_s - \int_{p_T}^{p_s} D_0 dp}{p_s - p_T}.$$  
(17)

Then $D = D_0 + D'$ is used to obtain $\omega$ from (14).

$Q_1$ and $Q_2$ are calculated for the domain $5^\circ$–$45^\circ$N, $45^\circ$–$125^\circ$E, and for the layer between the ground surface and the first standard pressure level above the surface, and for each successive layer between the standard pressure levels. Because of the horizontally varying terrain height and tropopause pressure, the number of layers may vary from a grid point to the next. The advection terms in (1) and (2) are evaluated at every observation time, and the local time change terms are evaluated using the centered difference of $\dot{\theta}$ and $\dot{q}$ over 24 hours.
Fig. 2. The 2.5°×2.5° grid, smoothed topography (dashed contours in m) and FGGE Level II-b rawinsonde (open circle) and pilot wind (open triangle) data distribution at 0000 UTC 1 June 1979. Additional QXPMEX radiosonde data are shown by dark circles. Cross sections along 15° and 32.5°N, and along 80° and 90°E appear in this paper. Boxes 1, 2, 3, 4, 5 and 6 represent six regions for detailed study (see Section 4).

Fig. 3. Monthly mean winds at 850 hPa from December 1978 to August 1979. Isotachs are drawn at 5 ms⁻¹ intervals. The ground surfaces above 1500 m are masked.

3. Seasonal changes in the large-scale circulations from winter to summer

a. The mean flow features

In Fig. 3 we show the sequence of monthly mean flow patterns at 850 hPa from December 1978 to August 1979. The mean streamlines and isotachs show remarkable orographic and thermal effects of the Tibetan Plateau on the low-level flow. As noted by Murakami (1981a, 1987a) and simulated by numerical models (e.g., Tokioka and Noda, 1986; Trenberth and Chen, 1988), the mean winds in winter tend to flow around the Tibetan Plateau except the inflow towards the southeastern slope of the Plateau.
However, the inflow towards the Plateau becomes a prominent feature of the low-level flow in spring and summer (Yeh et al., 1957; Luo and Yanai, 1983). The mean flow also exhibits 1) gradual weakening of the northeast winter monsoon in the Indian Ocean and the western Pacific (December–February), 2) a transition period characterized by an anticyclone over the Arabian Sea (March–April), and 3) the onset and intensification of the southwest summer monsoon (May–August). The southwesterlies commence over areas from the Bay of Bengal to the South China Sea in May, and then begin over the Arabian Sea and Indian subcontinent in June (He et al., 1987).

In the upper troposphere (Fig. 4), the most conspicuous seasonal changes in the mean flow are the weakening of the westerly jet stream in the downstream side of the Tibetan Plateau (April–May) and the development of the South Asian anticyclone accompanied by the easterly jet stream along its southern periphery (May–August). As shown by Zhu et al. (1980), Krishnamurti (1985) and He et al. (1987), the anticyclone migrates from the western Pacific to the Burma-Thailand area in May and moves towards the Tibetan Plateau in summer. Comparing to the 10-year mean flow patterns presented by Krishnamurti (1985), the intensity of the anticyclone in the 1979 summer appears to be subnormal, especially in July and August.

\[ b. \text{Temperature and outgoing longwave radiation (OLR)} \]

The orographic and thermal effects of the Tibetan Plateau on the seasonal change in the tropospheric temperature are clearly recognized in the sequence of monthly mean isotherms at 500 hPa (Fig. 5). During the winter months (December–March) there is a warm temperature ridge over the Plateau and a cold temperature trough along the southwestern periphery of the Plateau. In spring (April–May) both the warm ridge and the cold trough become pronounced. The cold temperature trough persists over North India until June. As discussed by He et al. (1987), a broad area of warm air ($\sim 4{\degree}C$) appears over the Thailand-Indochina area in May. At the same time a separate warm center forms on the Tibetan Plateau. This warm center is clearly seen in the evening (1200 UTC) with closed isotherms, but it is diffuse in the morning (0000 UTC) as will be discussed in Section 6.

The warming of the tropospheric air over the land area east of $85{\degree}E$ and south of $30{\degree}N$ in May results in the reversal of the meridional temperature gradient south of the eastern Plateau, and the onset of the low-level southwesterlies over the Bay of Bengal, Indochina and the South China Sea (see Fig. 3 and Section 5). In June the warm center on the southern periphery of the Plateau becomes more distinct, and another area of warm air forms over the northern Arabian Sea-Persian Gulf region corresponding to the formation of an anticyclone over this region.
Fig. 5. Monthly mean temperature (°C) at 500 hPa. The ground surfaces above 3000 m are shaded.

(see Fig. 4 and Section 5). In July and August the warm air (>−4°C) spreads over the entire longitudinal extent of the domain. With the increase of temperature over the Iran-Afghanistan-western Tibetan Plateau region, the reversal of the meridional temperature gradient and the onset of the southwesterlies over the Arabian Sea take place (see Fig. 3 and Section 5).

To reveal the effects of the Tibetan Plateau on the temperature field more clearly, we present the vertical cross sections showing the temperature anomaly (from the domain and 9-month mean) in Figs. 6a, 6b. The cross sections along 32.5°N (Fig. 6a) show that, when compared at the same level, the air immediately above the Plateau surface is warmer than the air of surrounding areas even in winter. An “explosive” growth of this warm boundary layer air occurs in May, and warm anomalies of 8° to 9°C occupy the whole tropospheric air column above the Plateau by August. The cross sections along 90°E (Fig. 6b) also show the development of a warm anomaly above the Plateau after May, and the consequent reversal of the meridional temperature gradient to the south of the Plateau.

The seasonal mean distributions of the OLR flux from winter to summer are shown in Fig. 7. From January to February (the data for December 1978 are missing) relatively large values (>260 Wm⁻²) of the OLR flux is observed along a broad latitudinal belt centered around 15°N. In spring this belt of high OLR splits over Indochina where monsoonal rains begin in May (see Section 5). We notice also that high OLR values (>280 Wm⁻²) persist over the Indian subcontinent, the Arabian Sea and Saudi Arabia. In summer very large OLR values (>300 Wm⁻²) are found over dry land areas including the Saudi Arabian Desert and the Iran-Turkestan region, reflecting high ground surface temperature there.

From winter to spring small OLR values (<200 Wm⁻²) are seen over the Tibetan Plateau, indicating the longwave emission from the elevated surface. The principal belt of the monsoon rainfall undergoes an annual migration from Indonesia to the foothills of Himalayas (e.g., Murakami, 1980; Krishnamurti, 1985; Murakami and Nakazawa, 1985a). In May the area of low OLR values (<240 Wm⁻²) associated with tall convective clouds moves over the Bay of Bengal-Indochina area (not shown). During the summer months (June–August) the areas of low OLR values (<240 Wm⁻²) correspond very well to the heavy rains over Southwest India, the Bay of Bengal, Assam, the eastern Tibetan Plateau, Indochina, the Philippines, and the South China coast (e.g., Krishnamurti and Ramanathan, 1982; Krishnamurti et al., 1983; Luo and Yanai, 1983; M. Murakami, 1984, 1987; Shinoda et al., 1986).

c. The mean vertical circulation

The influence of the Tibetan Plateau on the vertical circulation is clearly recognized in the horizontal distributions of the seasonal mean vertical p-velocity, ω, at the 400 hPa level (Fig. 8). Over the Tibetan Plateau the absolute value of ω usually
Fig. 6. Monthly mean temperature deviations (°C) for (a) along 32.5°N (contour intervals 1°C) and (b) along 90°E (contour intervals 2°C). The deviations are from the domain and 9-month average at each level.
Fig. 7. Seasonal mean OLR flux (day and night together, Wm\(^{-2}\)). The ground surfaces above 1500 m are shaded. Data for December 1978 are missing.

attains the maximum near this level.

During the winter and spring there are three major areas of large-scale subsidence: 1) the Iran-Persian Gulf region, 2) North India, and 3) the eastern Plateau-North and Northeast China. The subsidence over North India is a well-known feature and has been discussed by Das (1962). On the other hand, ascending motion takes place over the western Tibetan Plateau and its neighborhood in winter and it extends to the central Plateau in spring. After May the area of ascent expands over the whole Plateau. We also note that in spring another area of ascent associated with pre-monsoonal rains appear over the Indochina Peninsula, the South China coast and the Philippines. The two areas of ascent are clearly separated from each other in spring. After the onset of the Indian summer monsoon (June), the two ascending areas merge together, and the upward motion over the Plateau forms a northward ex-

Fig. 8. Seasonal mean vertical p-velocity (hPa h\(^{-1}\)) at 400 hPa. The ground surfaces above 1500 m are shaded.

tension of the large-scale ascent associated with the summer monsoon.

He et al. (1987) noted that the areas of an 80-day mean (16 April-4 July) descent surrounding the Plateau corresponded remarkably well to the desert areas of Turkestan, Iran and Saudi Arabia, the Great Indian (Thar), Taklamakan and Gobi Deserts. Recently, Manabe and Broccoli (1990) made a numerical experiment using a global climate model with and without orography, and showed that the orography produces the dry regions by induced subsidence.

The time evolution of the three-dimensional circulation over and around the Tibetan Plateau is illustrated in the vertical cross sections along 32.5\(^\circ\)N (Fig. 9a) and along 90\(^\circ\)E (Fig. 9b). In the cross sections streamlines are made parallel to the pseudovectors \((\bar{u}_D, -\bar{\omega})\) and \((\bar{v}_D, -\bar{\omega})\), respectively. Here \(\bar{u}_D\) and \(\bar{v}_D\) are the zonal and meridional components of the divergent part of horizontal wind, \(\bar{v}_D\). The values of \(\bar{\omega}\) are also shown in dashed contour lines.

Figure 9a shows that the vertical motion above the Plateau in December is downward except near
the surface of the western Plateau. However, from January to February the upward motion begins to organize and expands from the western to central Plateau. The seasonal change in the intensity of the upward motion over the Plateau is clearly recognized in the cross sections for spring and summer.
months. The center of rising motion shifts to the central Plateau as the season progresses.

In May, the ascending motion over the Plateau is accompanied by subsidence over Iran and Afghanistan (55°-65°E) and over the eastern periphery of the Plateau and the China Plain. In June, however, the ascending motion extends over most longitudes of the domain. The overall distribution of the vertical circulation in summer is similar to that of the daily mean (daytime and nighttime together) vertical motions along 35°N for July simulated by Kuo and Qian (1981). During summer this longitudinal circulation is part of the planetary-scale “east-west” circulation (Krishnamurti, 1971; Yeh and Gao, 1979; Chen, 1987; Magaña and Yanai, 1991).

The meridional cross sections along 90°E (Fig. 9b) shows that a weak ascent is present over the Plateau even in winter. During winter the ascent over the Plateau is accompanied by the descent in the south and north sides of the Plateau. In spring, the southerly flow sliding up along the southern slope of the Plateau is clearly observed. Until April the ascending motion over the Plateau is separated from the ascent with the southwesterly monsoon. These features are similar to those of the mean July circulations along 90°E presented by Yeh and Gao (1979) and simulated by Kuo and Qian (1981). In our analysis, however, the ascending motion over the Plateau in summer becomes a part of the intense vertical circulation associated with the monsoon.

The thermal origin of the Plateau-induced vertical circulation becomes evident when we analyze its difference between 1200 UTC and 0000 UTC (not shown). The mean maximum upward velocity over the Plateau for a 80-day period (16 April - 4 July) at 1200 UTC (1800 local time at 90°E) is ~1.5 times larger than that at 0000 UTC (0600 local time) (He et al., 1987).

4. Heat sources and moisture sinks

In this section we examine the spatial distributions and seasonal changes of heat sources and moisture sinks over and around the Tibetan Plateau. These will reveal regional differences of heating and drying processes and their possible effects on the time evolution of the large-scale circulation.

a. Horizontal distributions of heat sources and moisture sinks

The mean horizontal distributions of the vertically integrated heat source <Q1> and moisture sink <Q2> for three seasons are shown together in Fig. 10.

During winter values of <Q1> are generally negative (i.e., heat sink) except over areas covering the southern periphery of the Plateau, Assam, Burma and the Bay of Bengal. The Tibetan Plateau as a whole is a heat sink, but it is surrounded by regions of stronger cooling. The sign of <Q2> values is positive (precipitation exceeding evaporation) in the western Plateau, and negative (evaporation exceeding precipitation) in the eastern Plateau. These are consistent with the observation during this season that substantial rains fall over the western slope and southwestern edge of the Plateau, and dry weather persists downstream of the Plateau (Yeh and Gao, 1979; Murakami, 1987a).

Remarkable net cooling (radiative cooling exceeding sensible and condensation heating) and net moistening (evaporation exceeding precipitation) occur over 1) a large area including the Arabian Sea, Saudi Arabia, the northern Indian subcontinent, Iran and Turkestan, and 2) the region extending from the eastern Tibetan Plateau to the China Plain and to the Tarim Basin. Intense cooling (<−100 Wm⁻²) over Saudi Arabia, the Arabian Sea, and the northern India is comparable with the radiative cooling during January and February 1979 estimated by Ackerman and Cox (1987).

In spring the Tibetan Plateau becomes a heat source. The values of <Q1> are much greater than those of <Q2>. The <Q2> values are still positive in the western Plateau and negative in the eastern Plateau. The cooling over the regions surrounding the Plateau continues, although its magnitude is reduced especially over the China Plain. As pointed out by Charney (1975), desert regions act as heat sinks because of high albedo and because of the long-wave radiation emitted from hotter surfaces escaping to space through the less cloudy air. The radiative cooling over Saudi Arabia and the Arabian Sea prior to the onset of the Indian monsoon has been discussed by Ackerman and Cox (1982, 1987), Blake et al. (1983) and Ellingson and Serafino (1984). According to Ackerman and Cox (1982), the frequent presence of dust particles from deserts results in a significant radiative heating between 250 hPa and the surface for both over the Saudi Arabian Desert and the Arabian Sea. However, the 24-hour mean <QR> values were shown to be negative for both the desert and the ocean.

During summer the whole Plateau is a heat source. It is accompanied by generally positive values of <Q2> except over the western edge, suggesting larger contributions from rain in the eastern Plateau. Even though the vertically integrated values of <Q1> over the Plateau is smaller than those obtained for the regions affected by monsoonal rains, the heating rates over the Plateau are comparable to those over the Assam-Bengal region, because the air column above the Plateau is only 500 hPa thick (Section 5).

The summer distributions of positive <Q1> over the Indian subcontinent, the Bay of Bengal coast, South China and the East China Sea are very simi-
Fig. 10. Seasonal mean distributions of the vertically integrated heat source \(<Q_1>\) (left) and moisture sink \(<Q_2>\) (right) in Wm\(^{-2}\).

lar to those of \(<Q_2>\) showing the dominant role of condensation heating with the monsoon and Mei-yu rains in these regions. The minimum in \(<Q_2>\) in the western Arabian Sea is consistent with the large evaporation rates shown by Rao et al. (1981) and by Pearce and Mohanty (1984). The summer distribution of \(<Q_1>\) over land areas is qualitatively similar to those obtained by Yao et al. (1984) and by Wang and Luo (1989).

Figures 11a, 11b show the vertical distributions of the heating rate \(Q_1/c_p\) in the longitudinal plane along 32.5°N, and in the meridional plane along 80°E. These cross sections show the beginning of heating near the ground surface of the western Plateau in winter, its upward and eastward expansion in spring, and its coexistence with another heat source region spreading from the periphery of the Plateau to the Indian subcontinent in summer (see the summer distribution of \(<Q_1>\) in Fig. 10). Note that during the winter and spring, the heat source over the Plateau is surrounded by the intense cooling over the China Plain and Afghanistan (Fig. 11a) and the Indian Desert (Fig. 11b). As noted by He et al. (1987), the heating over the Plateau and the cooling in the surrounding areas maintain the temperature contrast (Fig. 6) that drives the large-scale vertical circulation (Fig. 9).

b. Time sequences of heat sources and moisture sinks

To identify the principal factors contributing to the mean heat budgets of different parts of the Tibetan Plateau and surrounding areas, we shall examine time sequences of the areal mean values of the vertically integrated heat source \(<Q_1>\) and moisture sink \(<Q_2>\) (denoted by \([<Q_1>]\) and \([<Q_2>]\), respectively) in six selected regions shown in Fig. 2. These are: Region I, the western Tibetan Plateau; Region II, the eastern Tibetan Plateau; Region III, the China Plain (the middle and downstream region of the Yangtze River); Region IV, Southeast Asia (the Thailand-Indochina region); Region V, the southwestern India; and Region VI, the Arabian Sea. Instead of showing daily values, we present the time sequences of 5-day mean values of \([<Q_1>]\) and \([<Q_2>]\).
1) REGION I (THE WESTERN PLATEAU) AND REGION II (THE EASTERN PLATEAU)

The time series of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ averaged over the Tibetan Plateau (Figs. 12a, 12b) clearly reveal the seasonal change of the heat source and its components. There is a noticeable difference in the progress of heating between the western and eastern parts of the Plateau. In Region I three positive peaks in $\langle Q_1 \rangle$ appear in February, March and April. These peaks are accompanied also by weak peaks of $\langle Q_2 \rangle$ and likely to be the results of rains over the western and southwestern periphery of the Plateau. The western Plateau becomes a heat source in February, whereas the eastern Plateau is a heat sink until late March. The $\langle Q_1 \rangle$ value on the western Plateau attains the maximum in May. In May the $\langle Q_1 \rangle$ value on the eastern Plateau is also a relative maximum. After the onset of the summer rains in May the heating is mostly due to the release of latent heat of condensation.

2) REGION III (THE CHINA PLAIN)

Figure 13a shows the time series of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ for Region III. Over the China Plain there is strong cooling with very small values of $\langle Q_2 \rangle$ during winter (December–February), and gradual increase of $\langle Q_1 \rangle$ from negative to positive values with fluctuations of a short period (~10 days) (March–May). During spring $\langle Q_2 \rangle$ values also exhibit similar fluctuations. These suggest that the radiative cooling dominates the heat budget in winter and that the condensation heating begins to offset the radiative cooling in spring. After the onset of the summer rains in May the heating is mostly due to the release of latent heat of condensation.

3) REGION IV (SOUTHEAST ASIA)

In region IV (Thailand-Indochina) the summer monsoon begins in May (see Section 5). Figure 13b shows that the values of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ are nearly identical to each other, indicating that the heating in this region is chiefly due to condensation heating. After May both $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ values oscillate with a period of 35 to 45 days. This is a manifestation of the modulation of rainfall intensity by the tropical “low-frequency” oscillation (Madden and Julian, 1971, 1972). The modulation of the monsoon activity by the Madden-Julian oscillation has been studied by many authors (e.g., Zangvil, 1975; Yasunari, 1979, 1980, 1981; Krishnamurti and...

4) REGION V (SOUTHWEST INDIA)

This region (Fig. 14a) was a heat sink with no evaporation until subperiod 33 (11–15 May) when a depression developed in the Bay of Bengal brought rains (e.g., Krishnamurti et al., 1980). After the onset of the Indian monsoon in subperiod 39 (see Appendix C), \(<Q_1>\) and \(<Q_2>\) values become large positive and oscillate with a 40 day periodicity, responding to the passage of the 30–50 day oscillation (e.g., Krishnamurti and Subrahmanyam, 1982; Murakami et al., 1984; Murakami and Nakazawa, 1985b).

5) REGION VI (THE ARABIAN SEA)

In this region (Fig. 14b) subsidence and radiative cooling prevail except subperiod 40 (15–19 June) when it experienced heavy rains accompanied by an “onset” vortex (Krishnamurti et al., 1981). The \(<Q_2>\) values are also negative except the short onset period, indicating strong evaporation from the sea surface.

c. Mean vertical profiles

In this section we present the seasonal mean vertical profiles of \([Q_1]\) and \([Q_2]\) for the six regions. They show distinct regional differences of heating processes.

Figure 15 shows seasonal mean vertical profiles of \([Q_1]\) and \([Q_2]\) for the western Plateau (Region I) and the eastern Plateau (Region II). During winter the \([Q_1]\) values for both parts of the Plateau are generally negative, suggesting the dominance of radiative cooling. In spring, \([Q_1]\) becomes positive in both regions. Note that during winter and spring \([Q_2]\) values are small positive (drying) in the western Plateau and small negative (moistening) in the eastern Plateau. These are consistent with the horizontal distribution of precipitation on the Plateau in winter (Yeh and Gao, 1979; Murakami, 1987a). In summer, \([Q_1]\) profiles for both regions show maximum heating (~2K d\(^{-1}\)) in the layer between 300 and 500 hPa. There are contributions from condensation heating especially in the eastern Plateau as shown previously by Nitta (1983) and by Luo and Yanai (1984). Approximately 40% of heating over the eastern Plateau is contributed from condensation heating.

A drastic change in the vertical heating profile takes place over the China Plain (Fig. 16a). In winter there is pronounced tropospheric cooling of ~2 to ~3K d\(^{-1}\) with negative \([Q_2]\) (net evaporation) near the surface. In spring, the net heating is nearly zero in the whole layer but the \([Q_2]\) values are clearly positive. This means that the radiative cooling is offset by condensation heating of nearly equal magnitude. The mean summer profiles of \([Q_1]\) and \([Q_2]\) for Region III are remarkably similar to each other, showing that the heating in this region is mainly the
result of stratiform rains (Mei-yu). However, the \([Q_1]\) values are slightly larger than the \([Q_2]\) values above the 500 hPa level and the reverse is true below this level, showing some contributions from eddy vertical transports due to cumulus convection [see Eq. (5)]. The upward shift of the \([Q_1]\) profile from the \([Q_2]\) profile is a phenomenon typical for a cumulus convective atmosphere (e.g., Yanai et al., 1973; Thompson et al., 1979).

Over Region IV (the Thailand-Indochina area)
seasonal changes in \([Q_1]\) and \([Q_2]\) profiles are the clearly recognized (Fig. 16b). In winter \([Q_1]\) is generally negative except near the surface and in the 100–250 hPa layer. The cause of this upper tropospheric peak is unclear at present. In this region pre-monsoonal rains begin near the end of April, and both \([Q_1]\) and \([Q_2]\) become positive in spring. In summer large \([Q_1]\) and \([Q_2]\) values appear with substantial rains.

In winter and spring, both \([Q_1]\) and \([Q_2]\) profiles over the southwestern India (Fig. 17a) and over the Arabian Sea (Fig. 17b) are similar. Large cooling in the middle and upper troposphere and large negative \([Q_2]\) (moistening) in the lower layer are consistent with the radiative cooling estimated by Ackerman and Cox (1987) and evaporation from the sea surface (e.g., Rao et al., 1981; Mohanty et al., 1983). After the onset of the Indian monsoon in June, the southwestern India receives rains but the Arabian Sea experiences subsidence (see Fig. 8), radiative cooling and intense evaporation. These are well reflected in the summer profiles of both regions.

d. Heat and moisture budgets of the Tibetan Plateau

The monthly mean values of \([<Q_1>]\) for Regions I and II are listed in Table 2. In this table the estimates given by Yeh and Gao (1979) are also shown in parentheses. Their values are based on long-term
Table 2. Regional mean tropospheric heating \(<Q_1>\) over the Tibetan Plateau (units: Wm\(^{-2}\)).

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Values in parentheses are those obtained by Yeh and Gao (1979).

Table 3. Regional mean tropospheric drying \(<Q_2>\) over the Tibetan Plateau (units: Wm\(^{-2}\)).

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The roles of the Tibetan Plateau for the onset of the summer monsoon

The results presented in the previous sections have shown that during spring and summer the Plateau is a heat source. The relationship between the Tibetan heat source and the evolution of the Asian summer monsoon is, however, not immediately obvious. Early studies (e.g., Flohn, 1957, 1960; Staff Members of Academia Sinica, 1958) suggested that the seasonal heating of the elevated surface of the Plateau is responsible for the warm center of the South Asian (Tibetan) anticyclone in summer. However, with the increased knowledge of the seasonal migration of the South Asian anticyclone, this simple view attributing the formation of the anticyclone directly to heating of the Plateau has been questioned. Murakami and Ding (1982) compared the large-scale circulation and temperature fields before and after the onset of the 1979 Indian monsoon and noted a large temperature increase over the western end of the Tibetan Plateau. They emphasized the importance of diabatic heating over the Eurasian continent as a whole in establishing the summer monsoon circulation. In this section we review (and improve) the detailed study of He et al. (1987) to reexamine the roles played by plateau in the onset of the summer monsoon of 1979.

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a. Onsets of the Southeast Asian and Indian monsoons

In a previous study covering an 80-day period (16 April–4 July 1979), the et al. (1987) showed that the general circulation over Asia underwent two distinct stages of abrupt transitions, resulting in the successive onsets of the Southeast Asian and Indian monsoons. The two onsets were clearly related to similar two stages of tropospheric warming over the Asian land mass. The rapid warming occurred first over longitudes east of 85°E (the eastern Tibetan Plateau-South China), and then over longitudes west of 85°E (Iran-Afghanistan-the western Plateau). The successive warming over these regions is the primary cause of the successive reversals of the meridional temperature gradient on the south sides of these regions (see Figs. 5 and 6b).

To emphasize this, He et al. presented the longitude-time section of the intensity \(\langle v \rangle\) of the southwesterly flow at 850 hPa along 15°N (Fig. 18a). For longitudes between 80°E and 125°E (the Bay of Bengal-the Philippines) the moderate (>5 ms\(^{-1}\)) and organized southwesterlies begin in subperiod 6\(^{1}\) (11–15 May), whereas for longitudes 50–80°E (the Arabian Sea and South India) the organized southwesterlies commence around subperiod 12 (10–14 June). We note that the southwesterlies over the longitudes 80°–115°E also intensify with the onset of the southwesterlies over the Arabian Sea. Then they examined the longitude-time section of the 5-
day mean rainfall rate at the same latitude (Fig. 18b). Corresponding to the onset of the southwesterlies over the Bay of Bengal and that over the Arabian Sea, intense (>10 mm day$^{-1}$) and organized summer rains also commence successively in subperiods 6 and 12 over the respective longitudes.

Corresponding to the abrupt changes in the 850 hPa flow and precipitation patterns, the upper-tropospheric circulation also undergoes drastic changes. The changes are the most clearly represented by the movement and development of the South Asian anticyclone at 200 hPa. Figure 19 shows the successive 5-day mean positions of the anticyclone center at 200 hPa for the sixteen subperiods (Appendix C) studied by He et al. From the movement of the 200 hPa anticyclone center we can also identify two transition periods. During the first transition (between subperiods 5 and 6) the center moves rapidly northward. The second transition (between subperiods 10 and 11) is characterized by the appearance of the western center over the Saudi Arabia-Iran region. These two periods nearly coincide with the two transition periods of the low-level flow and precipitation patterns.

The time evolutions of the southwest monsoon, rains and the South Asian anticyclone described above are closely linked to the changes in the tropospheric temperature. The presence of the Tibetan
Plateau has a profound effect on the horizontal distribution of the mean tropospheric temperature and its time evolution.

In Fig. 20a we show the longitude-time section of the deviation of the 200–500 hPa layer mean temperature (from the domain and 80-day average) along the 32.5°N latitude which cuts through the Tibetan Plateau. A sudden temperature increase over the longitudes 85°–115°E (from the eastern Plateau to the central China Plain) occurs at the first transition (subperiod 6). From subperiod 6 to subperiod 10 there is a remarkable temperature contrast between the eastern and western longitudes bordering at 85°E. The 200–500 hPa layer-mean temperature over Iraq, Afghanistan and the western Tibetan Plateau (40°–85°E) increases rapidly during the second transition and afterwards (subperiod 11–15) with the weakening of the cold temperature trough over the western Plateau (see Fig. 5). After the second transition the upper tropospheric temperature over the longitudes east of 85°E also increases.

The dramatic changes of the mean tropospheric temperature over land are contrasted to the much smaller temperature changes along 15°N (mostly over oceans) shown in Fig. 20b. The warm temperature over Southeast Asia (east of 95°E) does not show appreciable time change, and there is a weak east-west contrast of temperature between the warmer air over Southeast Asia and the South China Sea and the colder air over the Bay of Bengal and the Arabian Sea before subperiod 11 (5–9 June). After the onset of the Indian monsoon (subperiod 12), the mean temperature over these oceanic areas increases and the east-west temperature contrast disappears.

Figure 20c shows the longitude-time section of the mean meridional gradient of the 200–500 hPa layer-mean temperature between 5° and 25°N. The reversal of the meridional temperature gradient occurs first over the longitudes east of 85°E and then over the longitudes west of 85°E. The two stages of the reversal of the temperature gradient coincide with the two stages of the onsets of the low-level southwesterlies and organized rains over the Bay of Bengal and the Arabian Sea (Fig. 18a, 18b). The dominant role played by the temperature increases over land areas extending from the eastern Tibetan Plateau to South China in this reversal is evident.

b. Mechanism of the observed temperature increase

To identify the mechanisms which are responsible for the temperature increases during the two onset periods, He et al. (1987) analyzed the terms of the thermodynamic energy equation averaged over the 200–500 hPa layer,

\[
\frac{\partial T}{\partial t} = -\mathbf{v} \cdot \nabla T - \left( \frac{p}{p_0} \right) \frac{\partial \tilde{\omega}}{\partial p} + \frac{Q_1}{c_p},
\]

(18)

where

\[
\tilde{\omega} = \frac{1}{300 \text{ hPa}} \int_{200 \text{ hPa}}^{500 \text{ hPa}} \xi \text{ dp}.
\]

(19)

(18) interprets the observed local time change of \( \dot{T} \) in terms of the time changes due to the horizontal and vertical advection processes and diabatic heating. Here, we shall repeat this analysis for the same periods, since we now have more reliable analysis because of the QXPMEX data and also we have improved the computational scheme.

Figure 21 shows the horizontal distributions of the four terms of (18), averaged over the first onset period (a 15-day period from 3 May to 18 May) during
which the early summer rains commence over Southeast Asia. Large increases of the mean upper tropospheric temperature (0.2–0.3K day$^{-1}$) take place over a large area extending from the eastern Tibetan Plateau to the South China Plain. The warming of this region was instrumental for the sudden north-
ward movement of the South Asian anticyclone (see Fig. 19). We also observe that a large temperature increase (0.2–0.4K day$^{-1}$) took place over the Saudi Arabia, Iran and Turkestan.

Comparing the observed temperature increases with the temperature changes due to the horizontal advection, vertical advection and diabatic heating terms, we recognize that the warming over the eastern Plateau is indeed the result of diabatic heating, while the warming over the South China Plain is mainly due to the warm horizontal advection. We also note that the warming over the Saudi Arabia-Iran-Turkestan region and over the Gobi Desert is the result of large-scale subsidence. Most of the large diabatic heating over the central and western Tibetan Plateau is compensated with the cooling effects due to the horizontal and vertical advection terms with the resulting temperature changes over this region being very small.

Figure 22 shows similar analyses for the second transition period (from 28 May to 13 June) which corresponds to the onset of the Indian monsoon. Large temperature increases (0.3–0.5K day$^{-1}$) are observed over a huge area covering Saudi Arabia, Iran, Afghanistan, the northern India and the western Tibetan Plateau, while temperature decreases (−0.3K day$^{-1}$) are seen over a large area from the eastern Plateau to the China Plain. The warming over the area from Iraq to the western Plateau is related to the development of the 200 hPa anticyclone over this region (see Fig. 19). A similar pattern of temperature increase was noted by Pearce and Mohanty (1984) during the monsoon onset of 1982.

The examination of each term of (18) reveals very clearly the following: 1) The intense warming centered over the Iran-North India-western Plateau region is the result of adiabatic heating due to the large-scale subsidence. 2) The temperature decrease over the eastern Plateau and the China Plain is mainly the result of adiabatic cooling due to the large-scale ascent.

6. Mechanisms of heating and the boundary layer on the Plateau

We have demonstrated the presence of positive temperature anomalies over the Tibetan Plateau and the large-scale vertical circulation induced by the Plateau throughout the nine months from winter to summer. Before the onset of the summer monsoon, the heat source on the Plateau is surrounded by intense cooling in the adjacent regions.

Both the ascending motion and the warm center on the Plateau exhibit pronounced diurnal variations. In Fig. 23 we show the mean differences of temperature between 1200 UTC (evening) and 0000 UTC (morning) at 500 hPa for the three seasons. Large differences of 1–3K are seen over the Plateau. The diurnally varying heat low on the Plateau has been discussed by Yeh and Gao (1979), Gao et al. (1981) and Murakami (1981b), and its presence reflects in the diurnal variation of the low-level inflow towards the Plateau (Staff Members of

M. Murakami (1983), from the analysis of data obtained by the GMS-1 geostationary satellite, showed that the maximum convective activity in the southern half of the Tibetan Plateau occurs around 1800 local time (1200 UTC at 90°E) and the minimum activity occurs around 0900 local time (0300 UTC at 90°E). Nitta (1983) showed that the heating rate over the eastern Plateau is also larger at 1200 UTC than at 0000 UTC.

As a result of intense diurnal heating, a deep mixed layer of nearly uniform potential temperature is observed over the Plateau in the evening (Fig. 24a). The mixed layer is most well developed in spring. The vertical gradient of water vapor mixing ratio is also reduced at 1200 UTC, but the moisture is not well mixed vertically. Furthermore, the mixing ratio is smaller at 1200 UTC than at 0000 UTC (Fig. 24b). These features are commonly observed at all radiosonde stations located on the Plateau (including the temporary QXPMEX stations on the western Plateau). Similar decrease of moisture with height in the mixed layer over the high plains region of the United States has been noted by Mahrt (1976). He noted that the entrainment of dry air from the top and the moisture flux at the surface may create such a vertical gradient of moisture. He also suggests that, in the entrainment-drying boundary layer, dry air from the upper part of the layer may occasionally reach the surface (Mahrt, 1991).

There are several studies which aimed to clarify the mechanism of the heat low formation over the Saudi Arabian Desert (Ackerman and Cox, 1982; Blake et al., 1983; Smith, 1986a, b). These studies examined the balance between the solar heating including the effect of dust particles, infrared cooling and subsidence warming for the maintenance of the mixed layer. Because of the lack of large-scale subsidence, the mechanisms maintaining the mixed layer on the Tibetan Plateau must be very different from those operating in typical desert areas.

We must note that the boundary layer on the Plateau is highly baroclinic because of the strong horizontal temperature gradient. Its evening structure resembles that of a convective mixed layer that develops during the transformation of cold air mass over a large lake (Lenschow, 1973) or warm ocean surface (e.g., Ninomiya, 1972; Ninomiya and Akiyama, 1976; Agee and Lomax, 1978; Lenschow et al., 1980). A combined effect of the cold (and also dry) advection due to the westerlies blowing aloft and the intense diurnal heating of the Plateau surface appears to be responsible for the generation of this unique boundary layer.

Luo and Yanai (1984) speculated the activity of dry convection penetrating into the upper troposphere in the afternoon hours during the premonsoon period. From late spring to summer, however, the vertical distributions of potential temperature show a more stable stratification for dry convection. Instead, the vertical distributions of equivalent potential temperature at 1200 UTC during summer show a conditionally unstable stratification for moist convection with the increased moisture of surface air (Fig. 24c). Luo and Yanai (1984) also examined the
vertical heating profiles before and after the monsoon onset and showed the cumulus-convective nature of the summer rains on the eastern Plateau. Yeh and Gao (1979) mention frequent occurrences of thunderstorms and hailfall over the Plateau in summer.

7. Summary and conclusions

We have described the seasonal changes in the large-scale flow patterns, temperature and OLR fields, vertical circulation, and those of heat sources and moisture sinks over the Tibetan Plateau and surrounding areas during a 9-month period from December 1978 to August 1979, using the objectively analyzed FGGE Level II-b and QXPMEX-79 datasets. The major findings of this work may be summarized as follows:

1) The thermal effect of the Tibetan Plateau is evident from the positive temperature anomalies over the Plateau. During the three seasons, the air above the Plateau is always warmer than the air over the surrounding areas when compared at the same level and at the same latitude.

2) There is a large-scale vertical circulation thermally induced by the Plateau. This circulation is originally separated from the circulation associated with the principal monsoon system which migrates northward from winter to summer. The rising motion is seen near the surface of the western Plateau in December and spreads over the entire Plateau as the season progresses. The existence of the Plateau-induced vertical circulation is also evident in the distribution of subsidence in the surrounding desert areas, as pointed out by He et al. (1987).

3) An “explosive” expansion of the warm air over the Plateau occurs from late spring to early summer, leading to the reversal of meridional temperature gradient south of the Plateau. This warming causes the South Asian anticyclone to shift northward and the low-level southwesterlies to develop in the Indian Ocean. The area of ascent over the Plateau and that with the migrating rainbelt merge together, extending the monsoon rains to the north.

4) During winter the Plateau is a weak heat sink and it changes into a heat source in spring. However, the Plateau is surrounded by regions of intense cooling, i.e., the desert areas of Saudi Arabia and northern India, the Arabian Sea and the China Plain. The differential heating between the Plateau and its neighborhood maintains the temperature contrast that drives ascent over the Plateau and subsidence in the surrounding areas.

5) The sensible heat flux from the ground surface is the dominant factor in the heat budget of the Tibetan Plateau. The contribution of condensation heating is found in the western Plateau in all three seasons, and in the eastern Plateau after the onset of summer rains.

6) A detailed study of He et al. (1987) showed that the onset of the Asian summer monsoon exhibits two distinct periods of abrupt changes resulting in the successive commencements of summer rains over Southeast Asia and the west coast of India. The major tropospheric warming occurs first in May over the areas extending from the eastern Plateau to the South China Plain, which results in the reversal of the meridional temperature gradient south of the eastern Plateau and the commencement of the low-level southwesterlies over the Bay of Bengal. Then, the onset of the Indian monsoon occurs in June with the increase of temperature over a wide area covering Saudi Arabia, Iran, Afghanistan and the western Plateau.

7) The analyses of the warming processes of the upper troposphere during the two transition periods show that the diabatic heating is the primary cause of the temperature increase over the eastern Tibetan Plateau during the first onset. During the second onset, however, the adiabatic warming due

Fig. 23. The seasonal mean differences of temperature (°C) at 500 hPa between 1200 UTC (evening) and 0000 UTC (morning).
to subsidence is the key mechanism for the temperature increase to the west of the Tibetan Plateau.

8) There is a large diurnal change of temperature over the Plateau. A deep, nearly mixed boundary layer with high potential temperature is observed in the evening (1200 UTC), suggesting the role of thermal convection in the upward transport of heat. The mixed layer is most pronounced during spring. However, the moisture is not well mixed and a large horizontal gradient of temperature exists in this layer, suggesting the importance of the cold (and dry) horizontal advection for its maintenance. The boundary layer becomes conditionally unstable for moist convection in summer with the increase of moisture of the surface air.

In conclusion, the time evolution of the general circulation over Asia from mid-spring to early summer 1979 can be described in terms of an interaction of the vertical circulation induced by the heated Tibetan Plateau with the principal monsoon system which migrates northward. This interaction, manifested itself as the “explosive” warming over the Plateau and the “merger” of the Plateau-induced and monsoonal vertical circulations, appears to be the key to the understanding of the monsoon onset.

This suggestion is, however, based on qualitative interpretations of the synoptic and thermodynamic budget analyses. We need a more dynamical study of this interaction process. Dynamical processes associated with the transitions of the monsoon circulation such as the mechanism of the onset vortices (Krishnamurti et al., 1981) and the vorticity balance of the South Asian anticyclone (c.g., Holton and Colton, 1972; Fein, 1977; Sardeshmukh and Held, 1984) also require additional investigations. Recently, Boer (1991) analyzed the budget of the vertically integrated vorticity using the output of a general circulation model and found a strong sink over the Tibetan Plateau in summer.

It is also evident that we need a more extensive and quantitative study of the boundary layer and convection on the Plateau. Previously, the lack of the upper-air data over the western Plateau, the infrequent soundings which were made only at 0000 and 1200 UTC and the absence of fine vertical resolution of the FGGE II-b data severely limited a detailed examination of the boundary layer on the Plateau. Because the soundings from the QXPMEX include those of temporary stations on the western Plateau and also provide data at 600 hPa and many significant levels, some of these difficulties may be overcome.
Acknowledgments

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Appendix

A. Interpretation of $Q_1$

We write an approximate thermodynamic energy equation averaged over a horizontal area:

$$\frac{\partial \bar{s}}{\partial t} + \nabla \cdot \bar{\mathbf{v}} + \frac{\partial}{\partial p} \bar{s} \omega = Q_R + L (\bar{e} - \bar{\epsilon}).$$  \hfill (A1)

Eq. (A1) may be rearranged to give

$$\frac{\partial \bar{s}}{\partial t} + \bar{v} \cdot \nabla \bar{s} + \bar{\omega} \frac{\partial \bar{s}}{\partial p} = Q_R + L (\bar{e} - \bar{\epsilon})$$  

$$- \nabla \cdot (s' \bar{\mathbf{v}}') - \frac{\partial}{\partial p} s' \omega'$$  \hfill (A2)

In deriving Eq. (A2) we have used

$$\nabla \cdot \bar{v} + \frac{\partial \omega}{\partial p} = 0,$$  \hfill (A3)

and assumed that the Reynolds conditions and their consequences such as

$$\bar{A}B = \bar{AB}, \quad \bar{A}B' = 0,$$  \hfill (A4)

hold with sufficient accuracy (e.g., Kampe de Feriet, 1951; Monin and Yaglom, 1971, p. 207).

Even though the validity of (A4) for the given data cannot be shown, it is considered plausible when we imagine a horizontal area that is large enough to contain an ensemble of cumulus clouds, but small enough to be regarded as a fraction of the large-scale system (Yanai et al., 1973).

Noting

$$\frac{\partial \bar{s}}{\partial p} = c_p \left( \frac{p}{p_0} \right)^n \frac{\partial \bar{\theta}}{\partial p},$$  \hfill (A5)

and approximating

$$\left( \frac{\partial}{\partial t} + \bar{v} \cdot \nabla \right) \bar{s} = c_p \left( \frac{\partial}{\partial t} + \bar{v} \cdot \nabla \right) \bar{T},$$  \hfill (A6)

(this expression is exact only in height coordinates), we find that the left-hand side of (A2) is practically identical to $Q_1$ defined by (1).

Traditionally, the eddy horizontal transport term $- \nabla \cdot s' \bar{\mathbf{v}}'$ has been ignored. This is justified when the net lateral transports across the boundary of the fixed area by cumulus convection are negligible compared to the horizontal transports by the large-scale motion (Arakawa and Schubert, 1974), although we cannot rule out possible contributions from mesoscale eddies.

B. The Upper Boundary Condition for $\omega$

To obtain reliable estimates of $Q_1$ and $Q_2$ especially $Q_1$ in the upper troposphere where the static stability is large, we need a careful calculation of $\omega$ to be used in (1) and (2). Above the tropopause where the convective heat transport vanishes, we have

$$Q_1 = Q_R.$$  \hfill (B1)

Therefore,
\[ \tilde{\omega} = \tilde{\omega}_T = -\left[ \frac{\partial \tilde{\theta}}{\partial t} + \tilde{v} \cdot \nabla \tilde{\theta} - \frac{(p_0/p)^5}{c_p} Q_R \right] \left( \frac{\partial \tilde{\theta}}{\partial p} \right) \] 

at \( p = p_T \) (Nitta, 1977).

Inspecting the climatological profiles of \( Q_R \) (Katayama, 1967; Dopplick, 1972, 1979) as well as the profiles for Saudi Arabia and the Arabian Sea (Ackerman and Cox, 1982, 1987) and solar and infrared heating profiles calculated for the Tibetan Plateau (Cai et al., 1987; Chen and Kuo, 1987) we find that \( Q_R \) is small near the tropopause. Thus we use (16) as the additional condition for \( \tilde{\omega} \).

The tropopause pressure \( p_T (\lambda, \phi, t) \) is objectively analyzed at every map time using the tropopause data reported at radiosonde stations. Figure 25 shows three examples of the distribution of the tropopause pressure. The tropopause pressure varies greatly in space and time. There is a zone of strong gradient of \( p_T \) between the tropical tropopause and the subtropical tropopause. This zone moves northward from winter to summer.

C. 5-Day Subperiods

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