Onset of the Asian Summer Monsoon in 1979 and the Effect of the Tibetan Plateau

HAIYAN HE*, JOHN W. MCGINNIS, ZHENGSHAN SONG** AND MICHIYO YANAI

Department of Atmospheric Sciences, University of California, Los Angeles, CA 90024

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ABSTRACT

The time evolution of the general circulation over Asia during an 80-day period from mid-April to early July 1979 is studied using objectively analyzed FGGE Level II-b data. Through the analyses of the wind, temperature, precipitation, outgoing longwave radiation (OLR) flux, and heat and moisture budgets in the area 0°–50°N, 40°–130°E, the major changes of the circulation leading to the onset of the summer monsoon and the role of the Tibetan Plateau in these changes are examined.

During the analyzed period the general circulation underwent two distinct stages of abrupt transitions resulting in the successive onsets of early summer rains over Southeast Asia and the Indian summer monsoon. The first transition occurred in May in which the low-level southwesterlies began over the longitudes east of 80°E (from the Bay of Bengal to the South China Sea), resulting in the spreading of early summer rains over Assam, the Bay of Bengal coasts of Burma and the Malay Peninsula, Thailand, Indochina and the South China Sea. The center of the South Asian anticyclone at 200 mb abruptly moved northward from 10° to 20°N with an increase of the upper tropospheric temperature over the eastern Tibetan Plateau and the South China Plain and the reversal of the meridional gradient of temperature on the south side of the Plateau. During the second transition, which took place in June, the southwesterlies developed over the Arabian Sea and the monsoon rains commenced along the west coast of India. A new center of the 200 mb anticyclone formed over the Saudi Arabia–Iran region with a large increase of the upper tropospheric temperature over the Iran–Afghanistan–western Plateau region.

The analyses of the time sequences and 80-day mean distributions of the vertically integrated heat sources and moisture sinks, precipitation and the OLR flux reveal that the release of latent heat of condensation is the primary heat source driving the Asian summer circulation. Over the Tibetan Plateau, however, there is continuous sensible heating from the ground surface throughout the analyzed period.

A prominent feature seen in the general circulation during this period is the presence of the vertical circulation induced by the Tibetan Plateau which coexists with the principal monsoon system migrating northward. The sensible heating from the elevated surface and the radiative cooling in the environment maintain the temperature contrast that drives intense ascent over the Plateau and compensating descent in the surrounding areas. The examination of the warming process of the 200–500 mb layer during the two transition periods shows that the adiabatic warming due to large-scale subsidence is the dominant mechanism for the temperature increase. The subsidence on the northeast and west sides of the Plateau appears to provide a key mechanism for the observed warming associated with the distinct changes in the general circulation.

1. Introduction

The seasonal march of the general circulation leading to the establishment of the Indian summer monsoon is a classical problem in meteorology and has received attention over centuries (e.g., see Riehl, 1954; Ramage, 1971). Recent studies since the First Global Atmospheric Research Program (GARP) Global Experiment (FGGE) and the Monsoon Experiment (MONEX) of 1979 have substantially increased our knowledge of the monsoons over the globe. The seasonal migration of the planetary-scale monsoon has been well documented (e.g., Krishnamurti, 1985). Studies based on the satellite-measured outgoing longwave radiation (OLR) and precipitation patterns have shown that the principal belt of the monsoon rainfall undergoes an annual migration from Indonesia to the foothills of the Himalayas, as the season progresses from the northern winter to the northern summer (Murakami, 1980; Krishnamurti and Ramanathan, 1982; Krishnamurti, 1985; Murakami and Nakazawa, 1985).

The onset of the Indian summer monsoon is preceded by the early summer rains over Assam, the Bay of Bengal coasts of Burma and the Malay Peninsula, Indochina and the South China coast, which normally start in May with a conspicuous change of the low-level flow (e.g., Krishnamurti, 1985). It is also well recognized that the onset of the Indian summer monsoon is accompanied by distinct changes of the large-scale circulation and rainfall distribution over large parts of South and East Asia. These include 1) the northward
displacement of the subtropical westerly jet stream from the southern to the northern periphery of the Tibetan Plateau (the Qinghai-Xizang Plateau) (Yin, 1949; Staff Members of Academia Sinica, 1957), 2) the development of the upper tropospheric South Asian (Tibetan) anticyclone and the establishment of the easterly jet stream along its southern periphery (Diao and Chen, 1957; Koteswaram, 1958), and 3) the northward advance of the polar front over the China Plain and the East China Sea and the commencement of the rainy seasons of East Asia, “Mei-yu” in the Yangzi River Valley and “Baiu” in Japan (Suda and Asakura, 1955; Diao and Chen, 1957; Staff Members of Academia Sinica, 1957; Tao and Ding, 1981). In this paper we use the term “Asian summer monsoon” in a broad sense to include the early summer rains of Southeast Asia, the classical Indian monsoon, and some aspects of Mei-yu in China.

Undoubtedly the Asian summer monsoon is a response of the circulations to the differential heating between the Asian continental land mass and the oceans to the south. Recent studies of the global distributions of the mean tropospheric heating rate during FGGE elucidated the seasonal migration of heat sources and sinks (Wei et al., 1983; Johnson et al., 1985; Johnson and Wei, 1985). Johnson and Wei (1985) showed that in July the heating maximum is centered on the Tibetan Plateau and that the cooling maximum is over the northwest coast of Australia.

The possible importance of the Tibetan Plateau as an elevated heat source for the establishment and maintenance of the Asian summer monsoon has been discussed by many authors (e.g., Flohn, 1957, 1960; Staff Members of Academia Sinica, 1958; Murakami, 1958). Flohn (1957, 1960) 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 South and East Asia and the monsoon burst over the Indian subcontinent. Flohn (1968) then suggested the importance of both the sensible heat flux in the semi-arid western portion of the Tibetan Plateau and the latent heat release over the Himalayas, Assam, Bengal and adjacent mountains for the occurrence of the warm center of the South Asian anticyclone. Many numerical as well as laboratory experiments have been performed to clarify the orographic and thermal effects of the Tibetan plateau on the large-scale summer circulation over Asia [see Hahn and Manabe (1975), Yeh (1981), Luo and Yanai (1984) for more detailed reviews].

Yeh and Gao (1979) and Yeh (1981, 1982) summarized Chinese research on the meteorology of the Tibetan Plateau. Yeh and Gao (1979, Chap. 1) tabulated monthly mean values of the sensible heat flux, precipitation amounts and evaporation rates over the Plateau using long-term records of surface observations. Their results showed that there are large upward fluxes of sensible heat in the western Plateau from May to July and significant contributions from the latent heat release in the eastern Plateau from June to August. Moreover, they showed that the thermal effect of the Plateau induces a large seasonal change or even reversal of the wind direction in the boundary layer over the Plateau, which they called the “Plateau monsoon phenomenon” (Yeh and Gao, 1979, Chap. 6).

Since FGGE renewed attempts have been made to describe the heat sources over the Tibetan Plateau and surrounding areas. Nitta (1983) presented 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 late 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 contribute nearly equally to the total heating.

Luo and Yanai (1983) presented a detailed analysis of the time evolution of the large-scale precipitation, low-level (850 mb) wind, moisture and vertical motion fields over the Tibetan Plateau and surrounding areas during a 40-day period from late May to early July 1979. They showed that the mean 850 mb wind field exhibits a pronounced inflow towards the Plateau with diurnally varying intensity. Subsequently, Luo and Yanai (1984) examined the large-scale heat and moisture budgets of the same areas during the 40-day period. They showed that there is a deep heating layer over the Plateau and that the mean heating rate of \(\sim 3 \text{ K day}^{-1}\) in the 200–500 mb 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 addition of condensation heating after the onset of summer rains. Qualitatively similar results were obtained by Yao et al. (1984) and Chen et al. (1985) using the estimates of heating components at the surface.

However, the relationship between the Tibetan heat sources and the large-scale circulation changes during the onset of the Asian summer monsoon is not immediately obvious. As shown by Zhu et al. (1980) and by Krishnamurti (1985), the South Asian anticyclone migrates from the western Pacific to Southeast Asia in spring and moves towards the Tibetan Plateau in summer. With the withdrawal of the summer monsoon the anticyclone center retreats to the western Pacific in fall. Thus the immediate association of the formation of the anticyclone with the seasonal heating of the Plateau does not seem to be adequate.

Murakami and Ding (1982) compared the large-scale circulation and temperature fields before and after the onset of the 1979 Indian summer monsoon. They showed that the maximum warming took place over the Afghanistan–western Tibetan Plateau region and over the East China Sea–Japan region. They then emphasized the importance of diabatic heating over the
Eurasian continent as a whole in establishing the summer monsoon circulation. There is a need to examine more thoroughly the warming process in terms of the heat sources as well as the horizontal and vertical advection effects.

In the present paper, we extend the work of Luo and Yanai (1983, 1984) to examine the relationship between the heating processes and the time evolution of the Asian summer monsoon. To obtain more reliable mass, heat and moisture budgets, the objective analysis scheme has been improved and the analyses of the FGGE II-b data have been extended to cover a larger domain of Asia for an 80-day period from middle spring to early summer 1979.

The objectives of the present work are 1) to describe the time evolution of the large-scale circulation, heat sources and moisture sinks to give a coherent view of the Asian summer monsoon (especially the Southeast Asian and Indian monsoons), 2) to identify the transition periods during which the large-scale circulation exhibits major changes leading to the onset of the aforementioned components of the summer monsoon, 3) to relate the tropospheric warming during the transition periods to the heat sources and the heating due to the horizontal and vertical advection effects, and 4) to clarify the role played by the Tibetan Plateau in the evolution of the summer monsoon.

2. Data and analysis procedures

The principal data used in this study are the restructured FGGE Level II-b upper-air profiles in the domain 0°–50°N, 40°–130°E, from 16 April to 4 July 1979. The dataset includes vertical soundings from land stations and ships, aircraft dropwindsondes, and a limited amount of the LIMS (Limb Infrared Monitor of the Stratosphere) temperature soundings from the Nimbus-7 satellite (e.g., see Gille and Russel, 1984).

A modified version of the successive correction method (Cressman, 1959) is employed to 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 mb) on a 2.5° × 2.5° grid mesh. The ECMWF (European Centre for Medium Range Weather Forecasts) Level III-b analyses (Bengtsson et al., 1982) are used to provide the first-guess fields. The objective analyses are made twice daily (0000 and 1200 UTC).

Figure 1 shows the grid system and an example of the distribution of the II-b data (0000 UTC 26 May 1979). The data coverage is generally adequate for defining large-scale features of the motion, temperature and moisture fields except in oceanic regions and over the western Tibetan Plateau, Iran, Saudi Arabia, Somalia and Ethiopia. For the data-sparse regions, the ECMWF analyses which used additional data such as the satellite-derived cloud winds give useful first-guess fields.

We also analyze daily precipitation amounts for the 80-day period. The daily precipitation amounts south of 40°N and from 1 May to 4 July are taken from the analyses made by Krishnamurti et al. (1983), which are based on the FGGE Level II-c raingage data and satellite-measured brightness. The precipitation amounts between 40° and 50°N and those during the period 16–30 April are additionally analyzed using the level II-c raingage data only. (The distribution of the raingage stations is shown in Fig. 3a of Krishnamurti et al., 1984.) We also use the daily outgoing longwave radiation (OLR) data obtained from the Tiros N satellite (Gruber and Krueger, 1984).

The vertical p-velocity ω is obtained from the horizontal divergence by integrating the continuity equation

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

with the surface boundary condition

$$\omega = \omega_s = -g p_s \left( \frac{u_x}{a \cos \phi} \frac{\partial h}{\partial \lambda} + \frac{v_y}{a} \frac{\partial h}{\partial \phi} \right) \quad \text{at} \quad p = p_s. \quad (2)$$

In (1) and (2) u and v are the zonal and meridional components of the horizontal wind, a the mean earth radius, φ the latitude, λ the longitude, p the pressure, g the acceleration of gravity, ρ the density, and h the terrain height. The suffix s denotes the surface value. The values of smoothed terrain heights are taken from the National Meteorological Center (NMC) Level III-a data tape.

Assuming that the motion is approximately adiabatic in the layer between 100 and 150 mb, we impose the additional condition near the tropopause:

$$\omega = \omega_T = -\left( \frac{\partial \theta}{\partial t} + v \cdot \nabla \theta \right) \left( \frac{\partial \theta}{\partial p} \right) \quad \text{at} \quad p = p_T = 125 \text{ mb}, \quad (3)$$

where θ is the potential temperature. The original estimates of the horizontal divergence D_0 are corrected by adding

$$D' = \frac{\omega_T - \omega_s - \int_{p_T}^{p_r} D_0 dp}{p_s - p_T}. \quad (4)$$

Then D = D_0 + D’ is used to obtain ω from (1). The reasons which led to the use of (3) and (4) are given in Appendix.

The apparent heat source Q_1 and the apparent moisture sink Q_2 (e.g., Yanai et al., 1973) are calculated from

$$Q_1 = c_p \left( \frac{p}{p_r} \right) \left[ \frac{\partial \theta}{\partial t} + v \cdot \nabla \theta + \frac{\partial \theta}{\partial p} \right], \quad (5)$$

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1 The data from temporary radiosonde stations on the western Tibetan Plateau during the Chinese Qinghai-Xizang Plateau Meteorological Experiment (QXPMEX) were not available for this study.
\[ Q_2 = -L \left( \frac{\partial q}{\partial t} + v \cdot \nabla q + \omega \frac{\partial q}{\partial p} \right), \]  

where \( \kappa = R/c_p \), \( R \) and \( c_p \) are the gas constant and the specific heat at constant pressure of dry air, \( p_0 = 1000 \) mb, \( L \) the latent heat of condensation, and \( q \) the mixing ratio of water vapor.

As discussed by Yanai et al. (1973), the vertical integration of (5) and (6) from \( p_T \) to \( p_s \) gives approximately

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

where

\[ \langle \cdot \rangle = \frac{1}{\int_{p_T}^{p_s} \left( \cdot \right) dp}, \]

\( Q_R \) is the radiative heating rate, \( P, S \) and \( E \) are respectively the precipitation rate, the sensible heat flux and the evaporation rate per unit area at the surface. We note that

\[ \langle Q_1 \rangle - \langle Q_2 \rangle = \langle Q_R \rangle + S + LE. \]

3. Onset of the Asian summer monsoon

In this section we describe the time evolution of the wind and temperature fields, organized precipitation and OLR flux during the 80-day period and identify two distinct transitional stages characterizing the establishment of the 1979 Asian summer monsoon. Because of the large amounts of analyzed materials, the results are averaged (0000 and 1200 UTC together) for each of the following 5-day subperiods for brevity of description (Table 1). Subperiods 9–16 correspond to subperiods 1–8 in the papers of Luo and Yanai (1983, 1984).

a. Evolution of the low-level southwesterlies and organized precipitation

The southwesterly flow at low levels over the Arabian Sea, Bay of Bengal and South China Sea, and resulting heavy rains along the west coast of India, over Assam, the Bay of Bengal coasts of Burma and the Malay Peninsula, Indochina and the South China coast are the most well-known features of the Asian summer monsoon (e.g., Fu et al., 1983). The analysis reveals that during the 80-day period there are two clearly recognizable transitional stages during which abrupt changes occur in the low-level flow pattern and the distribution of organized precipitation.

Figure 2a shows the longitude–time section of the intensity (\( |v| \)) of the southwesterly flow (\( u > 0, v > 0 \)) at 850 mb along the latitudinal strip of 5° width centered on 15°N. For longitudes between 80° and 125°E
TABLE 1. Five-day subperiods for description.

<table>
<thead>
<tr>
<th>Subperiod</th>
<th>Dates</th>
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<tr>
<td>1</td>
<td>16 April–20 April</td>
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<td>2</td>
<td>21 April–25 April</td>
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<td>3</td>
<td>26 April–30 April</td>
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<td>15</td>
<td>25 June–29 June</td>
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<tr>
<td>16</td>
<td>30 June–4 July</td>
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* The onset of the Southeast Asian monsoon identified in this study.

The official onset of the Indian monsoon was declared on 12 June over the southwest coast of India and on 19 June at Bombay (Fein and Kuettner, 1980).

A break monsoon period over most parts of India.

(mainly because of the elevated surface. The gradual increase of the OLR values over the Plateau with the progress of the season reflects the increase of the surface temperature. However, daily OLR values over the Plateau show considerable time and space variations, suggesting the frequent presence of deep convection as discussed by M. Murakami (1983, 1984).

During the pre-onset period (Fig. 3a) weak southwesterly winds are seen over Bangladesh, the Burma–Thailand–Indochina region and a belt from South China to the East China Sea in association with the organized rains (>5 mm day⁻¹) in these regions. These pre-monsoon southwesterlies are located at the western periphery of the subtropical high in the western Pacific and the downstream side of the India–Burma trough. The rain spreading over the Indian Ocean between 80⁰ and 90⁰E is associated with a cyclonic circulation centered around 8⁰N, 88⁰E. However, the principal belt of the monsoon rainfall is still over the equatorial Indian Ocean and Indonesia during this period (e.g., Krishnamurti et al., 1980).

During the period between the two transitions (Fig. 3b), the southwesterlies (>5 m s⁻¹) are well established over areas from the Bay of Bengal to the South China Sea, bringing the heavy rains (>10 mm day⁻¹) over the Andaman Sea and the west coast of the Malay Peninsula, the Gulf of Thailand coast and the South China Sea. But a dry northwesterly flow with subsidence still prevails over the Indian subcontinent. There is no significant precipitation over the subcontinent in this period.

After the onset of the Indian monsoon (Fig. 3c), the intense southwesterlies (5–15 m s⁻¹) cover a broad belt extending from the Somalia coast to the Arabian Sea, India and the Bay of Bengal. There is a well-defined monsoon trough over Northeast India. The southwesterlies over the South and East China Seas have also been intensified. Heavy rains (>20 mm day⁻¹) fall along the west coast of India, the Bay of Bengal coasts and Indochina. Over the China Plain and adjacent waters the rain belt has reached the Yangzi River Valley and the East China Sea, indicating the commencement of Mei-yu. More detailed descriptions of the 850 mb flow and precipitation during subperiods 9–16 were given by Luo and Yanai (1983).

Krishnamurti et al. (1981) noted the appearance of the “onset vortex” (subperiod 13, not shown) during the shift of the northwesterlies to southwesterlies over the Arabian Sea. It is interesting that we observe a very similar appearance of an “onset vortex” in the Bay of Bengal (subperiods 4–6, see Fig. 3a) during the onset of the southwesterlies over Southeast Asia.

A remarkable feature of the 850 mb flow throughout the whole period is the presence of a cyclonic inflow toward the Tibetan Plateau. The inflow is pronounced especially along the southeastern periphery of the Plateau and the northern slope facing the Tarim Basin. This inflow is a manifestation of the Plateau monsoon.
(Yeh and Gao, 1979, Chap. 6). The inflow exhibits a remarkable diurnal variation (Staff Members of Academia Sinica, 1958; Luo and Yanai, 1983). The inflow is part of the vertical circulation produced by the thermal effect of the Plateau which will be shown in section 5.

b. The movement and development of the South Asian anticyclone

Corresponding to the abrupt changes in the 850 mb flow and precipitation patterns described above, the upper-tropospheric circulation also undergoes drastic changes. The changes are most clearly represented by the movement and development of the South Asian anticyclone at the 200 mb level. Figure 4 shows the successive 5-day mean positions of the anticyclone center at 200 mb for the 16 subperiods. The center is initially located near 10°N, 100°E (subperiods 1–5). Between subperiods 5 and 6 the center moves abruptly to 22.5°N and stays around this latitude until subperiod 13. In subperiod 11 another anticyclone center is formed at 23°N, 56°E, and two anticyclonic centers
Fig. 3. The mean streamlines and isotachs (m s⁻¹) at 850 mb (left) and the mean daily rainfall rate (mm day⁻¹) and daytime OLR flux (W m⁻²) (right) for (a) subperiods 1–5, (b) subperiods 6–10 and (c) subperiods 11–15.
coexist in subperiods 11–16 except subperiods 14–15 during which there is only one center or a ridge oriented in the east–west direction (Krishnamurti et al., 1979).

From the movement of the 200 mb 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 discussed in section 3a. The composite 200 mb streamlines and isotachs for subperiods 1–5, 6–10 and 11–15 are illustrated in Figs. 5a–c. With the development of the South Asian anticyclone, the upper-level easterly jet intensifies along the southern periphery of the anticyclone.

Figure 6 shows the longitude–time section of the 5-day mean zonal wind velocity at 200 mb along the \(5^\circ\)N–\(15^\circ\)N. At the first transition (between subperiods 5 and 6), the zonal wind changes from westerly to easterly for the longitudes east of \(65^\circ\)E. After this transition the core of easterlies is found near \(90^\circ\)–\(95^\circ\)E. At the second transition (between subperiods 10 and 11), the easterly flow establishes itself over the whole longitudinal extent in the domain.

c. Changes in the tropospheric temperature

The evolution of the monsoon circulation described above is clearly 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. 7a we show the longitude–time section of the deviation of the 200–500 mb layer mean temperature from the domain and time average along the \(32.5^\circ\)N latitude which cuts through the Tibetan Plateau. A sudden temperature increase over the eastern Plateau and the central China Plain (\(85^\circ\)–\(115^\circ\)E) occurs at the first transition (subperiods 5–6). From subperiods 6 to 10 there is a remarkable temperature contrast between the eastern and western longitudes bordering \(85^\circ\)E. The 200–500 mb layer mean temperature over Iraq, Iran, Afghanistan and the western Tibetan Plateau (\(40^\circ\)–\(85^\circ\)E) increases rapidly during the second transition and afterwards (subperiods 11–15) with the weakening of the cold temperature trough over the western Plateau (see Figs. 8b–c). After the second transition the upper tropospheric temperature over the eastern Plateau and the China Plain also increases.

The drastic changes of the mean tropospheric temperature along \(32.5^\circ\)N are contrasted to the much weaker temperature changes along \(15^\circ\)N shown in Fig. 7b. The warm temperature over Southeast Asia (east of \(95^\circ\)E) does not show appreciable time change, and there is an east–west contrast of temperature between the warm air over Southeast Asia and the cold air over the Bay of Bengal and the Arabian Sea before subperiod 11. After subperiod 11 the mean temperature over these
Fig. 5. The mean streamlines and isotachs (m s$^{-1}$) at 200 mb for (a) subperiods 1–5, (b) subperiods 6–10 and (c) subperiods 11–15.
oceanic areas increases and the east–west temperature contrast disappears.

Figure 7c shows the longitude–time section of the mean meridional gradient of the 200–500 mb 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 onset of the low-level southwesterlies and organized rains over the Bay of Bengal and the Arabian Sea (section 3a). The dominant role played by the temperature increases over land areas including the Plateau in this reversal is evident from Figs. 7a and 7b.

The composite temperature fields at 500 mb for subperiods 1–5, 6–10, 11–15 are illustrated in Figs. 8a–c. During the pre-onset period (Fig. 8a) the warmest air (−4°C) is located over the Thailand–Indochina area. There is a separate warm center (−7°C) on the Tibetan Plateau. To the west of this warm center, there is a pronounced cold temperature trough extending from the southwestern periphery of the Plateau to North India. Because the prevailing wind over the Plateau and its upstream side at 500 mb is westerly (not shown) and the mean vertical motion above the Plateau is upward (section 5), there must be heat sources maintaining the warm center against the effects of cold advection and adiabatic cooling. Between the two onsets (Fig. 8b), the warm temperature ridge has shifted northward and the maximum temperature over the Plateau has risen to −3°C. During the post-onset period (Fig. 8c), an additional warm area (−2 to −4°C) appears over the Saudi Arabia–Iran region, corresponding to the development of the anticyclone center over this region (Figs. 4 and 5c).

The warm center on the Tibetan Plateau is a unique feature observed at this level during most of the analyzed period and related to the low-level inflow discussed earlier. The warm center exhibits a pronounced diurnal variation. The center is always distinct at 1200 UTC but diffuse at 0000 UTC (not shown). The presence of the diurnally varying heat low on the Plateau has been shown by Yeh and Gao (1979) and Gao et al. (1981). The vertical structure of the warm air above the Plateau will be examined in section 5.

The Tibetan Plateau with its maintained heat source acts to offset the effects of the cold advection and adiabatic cooling (section 6). This favors the early warming over the areas extending from the eastern Plateau to the South China Plain and the consequent reversal of the meridional temperature gradient south of these areas.

**4. Heat sources and moisture sinks**

In this section we examine the time sequences and mean horizontal distributions of the vertically integrated heat source $\langle Q_1 \rangle$, moisture sink $\langle Q_2 \rangle$ and the OLR flux. These analyses reveal regional differences of the heating process in relation to the time evolution of the monsoon circulation.

**a. Time sequences of heat sources and moisture sinks**

Figures 9a–c respectively show the longitude–time sections of $\langle Q_1 \rangle$, $\langle Q_2 \rangle$ and the OLR flux along the latitudinal strip of 5° width centered on 15°N. At this latitude there is a remarkable contrast of heat sources between the longitudes west of ~75°E (the Arabian Sea) and the rest of the domain. The Arabian Sea is a heat sink [the radiative cooling exceeding the sensible
Fig. 7. Longitude–time sections for (a) the 5-day mean deviation of the 200–500 mb layer mean temperature (K) from the time and domain average at 32.5°N, (b) as in (a) but at 15°N, and (c) the 5-day mean meridional gradient of the 200–500 mb layer mean temperature (10⁻⁶ K km⁻¹) between 5° and 25°N. The vertical dashed-dotted lines in (a) indicate the boundaries of the Tibetan Plateau (>3000 m).
Fig. 8. The mean temperature (°C) at 500 mb for (a) subperiods 1–5, (b) subperiods 6–10 and (c) subperiods 11–15.
FIG. 9. Longitude–time sections at 15°N showing 5-day mean values of (a) the vertically integrated apparent heat source \( \langle Q_1 \rangle \), (b) the vertically integrated apparent moisture sink \( \langle Q_2 \rangle \) and (c) the daytime OLR flux (units: W m\(^{-2}\)).
Fig. 10. As in Fig. 9 but at 32.5°N. The vertical dashed-dotted lines indicate the boundaries of the Tibetan Plateau (>3000 m).
Fig. 11. Latitude-time sections at 75°E showing 5-day mean values of (a) the vertically integrated apparent heat source (\( \langle Q \rangle \)), (b) the horizontally integrated Loomis flux (units: W m\(^{-2} \)), and (c) the daytime OLR flux (units: W m\(^{-2} \)). The horizontal dashed-dotted lines indicate the boundaries of the Tibetan Plateau (>1000 m).

Fig. 12. As in Fig. 11 but at 100°E.
Fig. 13. The mean values of the vertically integrated apparent heat source ($Q_h$) (left) and apparent moisture sink ($Q_v$) (right) in units of W m$^{-2}$ for (a) subperiods 1–5, (b) subperiods 6–10 and (c) subperiods 11–15.
and latent heating, see Eq. (7)] until the onset of the southwesterly monsoon (subperiod 11). To the east of 75°E there are three heat source regions (>150 W m⁻²); 80°-90°E (the Bay of Bengal), 95°-105°E (the Andaman Sea–Thailand) and 110°-125°E (the South China Sea). Comparisons with the time sections of 〈Q₂〉 and the OLR flux show that these are mainly the results of precipitation. In particular, we note a sudden increase of heating with low OLR values over the three regions at the first transition period (subperiods 5–6). After the second transition (subperiods 10–11), the cooling over the Arabian Sea disappears and intense heating (∼300–450 W m⁻²) takes place with the onset of the monsoon rains. The heating to the east of 85°E also increases abruptly with the intensified precipitation over a large longitudinal extent from the Bay of Bengal to Thailand, Indochina and the South China Sea after the second transition.

At 32.5°N (Figs. 10a–c) the situation is quite different. Note that 〈Q₁〉 is continuously positive throughout the entire period between 75° and 100°E (the Tibetan Plateau). The positive 〈Q₁〉 values are accompanied by much smaller positive (east of 85°E) or negative (west of 85°E) values of 〈Q₂〉, showing the dominance of sensible heating over the Plateau. There is a suggestion, however, that the large 〈Q₁〉 value on the western Plateau in subperiod 6 is influenced by precipitation as indicated by a weak maximum of 〈Q₂〉 and low values of the OLR flux. Another remarkable feature in Fig. 10 is the continuous cooling accompanied by negative 〈Q₂〉 [evaporation exceeding precipitation, see Eq. (8)] over the longitudes west of 75°E (Iraq–Iran–Afghanistan–Pakistan) until about subperiod 11. These semiarid or arid regions are acting as heat sinks until the onset of the Indian monsoon. We note a remarkable increase of OLR values over these dry regions after subperiod 11. To the east of 100°E (along the Yangzi River) cooling prevails through subperiod 9, until intense heating (>150 W m⁻²) with an equally intense moisture sink and low OLR values appears during subperiod 10 and after subperiod 13.

We also examine the latitude–time sections of 〈Q₁〉, 〈Q₂〉 and OLR along 75°E which cuts through the western Tibetan Plateau (Figs. 11a, b, c). The features in the 〈Q₁〉, 〈Q₂〉 and OLR values are remarkably consistent with each other. These include the heating with positive 〈Q₂〉 and low OLR values over the western Plateau–Tien Shan region during subperiods 3 and 6, the cooling with negative 〈Q₂〉 and high OLR values over western India (15°–30°N), and the northward advance of heating due to monsoon rains along the west coast of India (5°–25°N) after subperiod 11. It is also evident that there is continuous heating over the western Plateau during the whole period. The heating over the western Plateau decreases from April to July.

Similar analyses of time sequences of 〈Q₁〉, 〈Q₂〉 and OLR are made along 100°E which cuts through the eastern Tibetan Plateau (Figs. 12a–c). In these sections we recognize a sharp contrast of heat source distribution between the northern and southern sides of the Plateau. The heating pattern clearly reflects the onset of monsoon rains over Thailand and the Malay Peninsula (5°–20°N) in subperiod 6 and the northward advance of rains to the eastern Plateau.

b. Horizontal distributions of heat sources and moisture sinks

The mean horizontal distributions of the vertically integrated heat source 〈Q₁〉 and moisture sink 〈Q₂〉 for subperiods 1–5, 6–10 and 11–15 are shown together in Figs. 13a–c. The distributions of 〈Q₁〉 over land areas are qualitatively similar to those obtained by Yao et al. (1984) for May and June using the estimates of heating components at the surface based on climatological data from 1961 to 1970. The distribution and time change of 〈Q₁〉 over the Arabian Sea are consistent with those obtained by Pearce and Mohanty (1984) for May and June 1979. With the observed mean precipitation rates and OLR fluxes (Figs. 3a–c), we find the following.

The pronounced heat sources of 200–300 W m⁻² centered over Burma and Thailand (subperiods 1–5, 6–10) and over the Bay of Bengal coast and the west coast of India (subperiods 11–15) are associated with moisture sinks of similar magnitudes and related to the heavy rains in these regions. Heat sources over the South China Sea and the South China coast also correspond well to moisture sinks and precipitation in these areas.

We find heat sources of ∼50–200 W m⁻² over most parts of the Tibetan Plateau during all of the subperiods. [The large positive 〈Q₁〉 values (∼300 W m⁻²)
FIG. 15. The streamlines of the mean divergent wind $v_p$ and the mean horizontal divergence ($10^{-4} \text{s}^{-1}$) at 200 mb for (a) subperiods 1–5, (b) subperiods 6–10 and (c) subperiods 11–15.
and large negative \( \langle Q_2 \rangle \) values (\( \sim -200 \text{ W m}^{-2} \)) along the southwestern periphery of the plateau for subperiods 6–10 are questionable because of poor data coverage and possible computational errors due to the steepness of the terrain.] The location of the heating maximum over the plateau shifts from the western plateau to the eastern plateau as the season progresses. The 80-day mean value of \( \langle Q_s \rangle \) over the whole plateau area (above 3000 m) is \( \sim 85 \text{ W m}^{-2} \) and close to the net heating estimated for May and June by Yeh and Gao (1979), but much larger than those obtained by Yao et al. (1984) and Chen et al. (1985). The heat sources are accompanied by moisture sinks of \( \sim 15 \text{ W m}^{-2} \). The large difference between \( \langle Q_s \rangle \) and \( \langle Q_2 \rangle \) over the plateau suggests the importance of the sensible heat flux from the ground surface [see Eq. (10)], even though the contributions from precipitation heating are seen over the western plateau during subperiods 1–5 and 6–10 and over the eastern plateau during subperiods 11–15 (see also Figs. 10–12).

Figures 13a–c also show that there are remarkable net cooling (radiative cooling exceeding sensible and latent heating) and net moistening (evaporation exceeding precipitation) over 1) large areas to the west of 70°E (including Turkestan, Saudi Arabia and the western Arabian Sea) for subperiods 1–10, 2) the Indian Desert, and 3) the region extending from the northeastern corner of the Tibetan Plateau to the Gobi Desert and to the Northeast China Plain. Note that the cooling generally occurs in desert areas under strong subsidence as pointed out by Charney (1975). Intense radiative cooling over Saudi Arabia and the Arabian Sea prior to the onset of the Indian monsoon has been shown by Ackerman and Cox (1982), Blake et al. (1983) and Ellingson and Serafino (1984). Blake et al. showed that for a day in May 1979 the radiative cooling generally exceeded the sensible heating between the surface and 600 mb level. The radiative cooling over the Indian Desert was discussed by Das (1962). The minimum in \( \langle Q_2 \rangle \) in the western Arabian Sea is consistent with the large evaporation for May and June shown by Rao et al. (1981) and by Pearce and Mohanty (1984).

The meridional profiles of \( \langle Q_s \rangle \) and \( \langle Q_2 \rangle \) averaged over the 80-day period at 75°E and 100°E are shown in Fig. 14. These profiles show distinct difference in the differential heating along the two meridians. At 75°E the maximum heating with smaller \( \langle Q_2 \rangle \) values occurs on the western Tibetan plateau region (32°–45°N) and net cooling is seen over North India (23°–32°N). The condensation heating is evident only along the west coast of India (south of 23°N). On the other hand, the heating distribution along 100°E is dominated by the strong condensation heating on the south side of the eastern plateau and the cooling on the north side. The maximum heating (270 W m\(^{-2}\)) is located over Thailand (17.5°N).

5. The circulation induced by the Tibetan Plateau

The analyses of the 850 mb wind field (section 3a) and temperature at 500 mb (section 3c) suggest the presence of a large-scale vertical circulation induced by the thermal effect of the Tibetan Plateau. In this section we examine the three-dimensional structure of

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**Fig. 16.** The 80-day mean vertical \( p \)-velocity (mb h\(^{-1}\)) at 40Q mb.
the Plateau-induced circulation and the related quantities.

a. The vertical circulation

We decompose the observed wind field into the rotational part \( v_R \) and the divergent part \( v_D \), i.e.,

\[
v = v_R + v_D, \tag{11}
\]

using the direct separation method (Endlich, 1971). In Figs. 15a–c, we illustrate the mean horizontal divergence and the divergent wind field at 200 mb for sub-periods 1–5, 6–10 and 11–15. These figures clearly show the presence of outflow over the Plateau and its relation to the outflow from the monsoon rain belt.

During the pre-onset period (Fig. 15a) there are conspicuous centers of outflow over the central and western Plateau which are clearly separated from the outflow from the premonsoon rain belt extending from Thailand to South China. We note also the presence of inflow centers over the Arabian Sea, North India and the Gobi Desert, suggesting subsidence over these areas. After the onset of the monsoon rains over Southeast Asia (Fig. 15b) the outflow centers over the Plateau and the Burma–Thailand area begin to join together. Intense upper-level inflow is seen over the Arabian Sea and Iran. After the onset of the Indian monsoon (Fig. 15c), a single source region of outflow extends from the Bay of Bengal coast to the Plateau region.

The horizontal distribution of the 80-day mean vertical \( p \)-velocity at 400 mb is shown in Fig. 16. Upward motions associated with the low-level southwesterlies cover a large domain to the east of 65°E and south of 35°N except North India, including the eastern half of the Arabian Sea, the Bay of Bengal, Southeast Asia, the South China Plain and the South and East China Seas. The maximum upward velocities (\(-3 \text{ mb h}^{-1}\)) are located over the Burma–Thailand region and along a belt extending from the South China Plain to the East China Sea.

The ascending motion with its maximum (\(-2 \text{ mb h}^{-1}\)) over the central Tibetan Plateau, and the descending motions over its surroundings including areas to the west of the Plateau, North India and the North and Northeast China Plain are a manifestation of the vertical circulation induced by the thermal effects of the Plateau. We note that the areas of descent correspond remarkably well to the desert areas of Turkestan, Iran and Saudi Arabia, the Great Indian (Thar), Takla Makan and Gobi Deserts.

The vertical circulation induced by the Tibetan Plateau is clearly seen in the longitudinal cross section of the 80-day mean vector field \( (u_D, -\omega) \) along 32.5°N (Fig. 17a). The intense ascending motion over the Plateau is accompanied by subsidence over its vicinity, i.e., Iran and Afghanistan (55°–65°E) and the eastern periphery of the Plateau and the China Plain. The overall distribution of the vertical circulation in this plane 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). This longitudinal circulation is more localized than the "east–west circulation" envisioned by Krishnamurti (1971) and Yeh and Gao (1979).

The meridional cross section along 92.5°E (Fig. 17b) shows that the intense ascending motion over the Plateau is accompanied by subsidence over the Tarim Basin (40°–45°N). There is also a narrow zone of weak ascent between the stronger ascending motion over the Plateau and the vigorous ascent over Bangladesh (\(-20°N\)) which is associated with the southwesterly monsoon. These are similar to the distribution of the mean July vertical circulation along 90°E presented by Yeh and Gao (1979) and simulated by Kuo and Qian (1981).

The thermal nature of the Plateau-induced vertical circulation becomes evident when we analyze the vectorial difference (\( \Delta u, -\Delta \omega \)) along 32.5°N (Fig. 18a) and (\( \Delta v, -\Delta \omega \)) along 92.5°E (Fig. 18b). Here

\[
\Delta = (\text{1200 UTC}) - (\text{0000 UTC}). \tag{12}
\]

The vectorial differences are approximate measures of the diurnal range of the vertical circulation. The mean maximum upward velocity over the Plateau at 1200 UTC (1800 local time at 90°E) is \(\sim 1.5\) times larger than at 0000 UTC (0600 local time at 90°E), showing the effect of the diurnally heated Plateau surface.

b. The vertical structure of the warm air over the Plateau

Figure 19a shows the vertical distribution of the 80-day mean temperature anomaly (the deviation of temperature from the horizontal and time mean in the whole analysis domain) in the longitudinal plane along 32.5°N. The maximum positive anomaly (\(-4°C\)) is located over the Plateau and positive anomalies (\(>2°C\)) extend upward and eastward. The southward tilt of the axis of positive anomalies is seen in the meridional vertical cross section along 92.5°E (Fig. 19b). Thus the axis of the warm air tilts southeastward with height. The center of the maximum anomaly (\(-5°C\)) near 350 mb is located at 20°N and it is closely associated with the South Asian anticyclone. This upper-level warm center is separated from the warm center immediately above the plateau and moves northwestward with the migration of the anticyclone center (not shown). The temperature contrast between the air over the Plateau region and that over the surrounding areas must be the primary cause of the vertical circulation described in section 5a.

c. The vertical distribution of heating

The 80-day mean vertical distribution of the heating rate \( Q_s/c_p \) along 32.5°N is shown in Fig. 20a. Over the Tibetan Plateau a deep layer of heating (2–3 K day\(^{-1}\))
occupies the whole troposphere. The heating over the Plateau is accompanied by a weak moisture sink ($\sim 1$ K day$^{-1}$) in the lower layer (not shown). To the west of the Plateau there is cooling in the upper troposphere over Iran, Afghanistan and Pakistan ($45^\circ$–$75^\circ$E). The cooling is accompanied by a weak moisture source in the lower layer. The weak heating over the eastern periphery of the Plateau near $110^\circ$E and over the China Plain is associated with a moisture sink of similar magnitude (not shown).

The mean distribution of $Q_l/c_p$ in the meridional plane along $92.5^\circ$E is shown in Fig. 20b. The heating rate above the southern slope of the eastern Plateau has a maximum value of $\sim 3$ K day$^{-1}$. This heating is accompanied by a moisture sink of $\sim 1$ K day$^{-1}$ (not shown), suggesting partial contributions from condensation heating (Nitta, 1983; Luo and Yanai, 1984). The heating above the Plateau is clearly separated from the equally intense latent heating associated with the heavy rains over Bangladesh. The sensible heating from the elevated surface and the radiative cooling in the environment maintain the temperature contrast that drives the Plateau-induced circulation.

6. Mechanisms of the observed temperature changes

To identify the mechanisms which are responsible for the temperature changes during the two transition
periods, we analyze the terms of the thermodynamic energy equation averaged over the 200–500 mb layer,

$$\frac{dT}{dt} = -\mathbf{v} \cdot \nabla T \left( \frac{P}{P_0} \right) \omega \frac{dT}{dp} + \frac{Q_i}{c_p},$$  \hspace{1cm} (13) \hspace{1cm}$$

where

$$\left( \right) = \frac{1}{300 \text{ mb}} \int_{200 \text{ mb}}^{500 \text{ mb}} \left( \right) dp.$$

Equation (13) interprets the observed local time change of $T$ in terms of the time changes due to the horizontal and vertical advection processes and diabatic heating.

Figures 21a–d respectively show the horizontal distributions of the four terms of (13), averaged over the first transition period (a 15-day period from the middle of subperiod 4 to the middle of subperiod 7) during which the early summer rains commence over Southeast Asia. Large increases of the mean upper tropospheric temperature (0.2–0.5 K day$^{-1}$) take place over the Iran–Saudi Arabia region and over the area including the Gobi Desert, the eastern Tibetan Plateau and the South China Plain (Fig. 21a). The warming over the eastern Plateau is related to the northward shift of the center of the South Asian anticyclone (see Figs. 4, 5, 7a, 8).

Comparing the observed temperature increases with the temperature changes due to the horizontal advection term (Fig. 21b), the vertical advection term (Fig. 21c) and the diabatic heating (Fig. 21d), we recognize...
Fig. 21. The analysis of the time change of the 200 mb layer mean temperature during the first transition (from the middle of subperiod 4 to the middle of subperiod 7).

(a) the local time change (10^(-6) K per day), (b) the horizontal advection (K day^{-1}), and (c) the diabatic heating (K day^{-1}).
Fig. 21. As in Fig. 20, but during the second transition (from the middle of subperiod 9 to the middle of subperiod 10).

Fig. 22. As in Fig. 21, but during the second transition (from the middle of subperiod 9 to the middle of subperiod 10).
that the warming over the Iran–Saudi Arabia region and over the Gobi Desert is the result of subsidence. Both the subsidence warming and direct diabatic heating contribute to the warming over the eastern Tibetan Plateau, while the warming over the South China Plain is mainly due to the warm horizontal advection. We note that most of the large diabatic heating over the central and western Tibetan Plateau is compensated by cooling effects due to the horizontal and vertical advection terms with the resulting temperature increase over this region being relatively small.

Figures 22a–d show similar analyses for the second transition period (from the middle of subperiod 9 to the middle of subperiod 12) which corresponds to the onset of the Indian monsoon. As shown in Fig. 22a, large temperature increases (0.2–0.5 K day$^{-1}$) are observed over large areas from Iraq, Iran, Afghanistan to the western Tibetan Plateau, while temperature decreases (−0.2 to −0.4 K day$^{-1}$) are seen over large areas from the Gobi Desert to the eastern Plateau, and to the China Plain. The warming over the area from Iraq to the western Plateau is related to the development of the 200 mb anticyclone over this region (see Figs. 4, 5, 7a, 8).

The examination of the horizontal advection term (Fig. 22b), the vertical advection term (Fig. 22c) and the diabatic heating term (Fig. 22d) reveals the following: 1) the warming centered over the Iran–western Plateau region is mainly the result of subsidence; 2) the cooling over the Gobi Desert is due to diabatic (radiative) cooling; 3) the cooling over the eastern Plateau and the China Plain is mainly the result of adiabatic cooling due to large-scale ascent.

Similar analyses are carried out for the post-onset period (subperiods 11–15, not shown). This period is characterized by large temperature increases over the Iraq–Iran–Afghanistan–western Tibetan Plateau region and over the North and Central China Plain and the East China Sea. These are similar to the temperature changes at 300 mb from a pre-onset period (15–30 May) to a post-onset period (20–30 June) of the Indian monsoon discussed by Murakami and Ding (1982). A similar horizontal distribution of the change of tropospheric mean temperature from 1–15 May to 1–15 June 1982 was shown by Pearce and Mohanty (1984). The examination of each term of (13) reveals the following: 1) The warming over the large area from Iraq to the western Tibetan Plateau is primarily the result of subsidence and partly due to diabatic heating; 2) The warming over North China is caused by subsidence; 3) The warming over the Yangzi River Valley and the East China Sea is the result of warm horizontal advection and condensation heating.

7. Summary and discussion

In this study we have described the time evolution of the large-scale circulation, temperature, precipitation and OLR fields, and heat and moisture budgets over Asia during an 80-day period from mid-April to early July 1979. During this period the large-scale fields underwent distinct seasonal changes that characterize the establishment of the Asian summer monsoon. The major findings of this work may be summarized as the following:

1) During the analyzed period, the distributions of the large-scale wind, temperature, and precipitation as well as heat sources and moisture sinks exhibit distinct changes in two transition stages. The first transition around subperiod 6 (11–15 May) is characterized by the establishment of the low-level southwesterlies over the longitudes east of 80°E covering the Bay of Bengal, Indochina and the South China Sea. Early summer rains spread over Assam, the west coast of the Malay Peninsula, Indochina and the South China Sea. The resulting circulation and rainfall distribution may be termed the "Southeast Asian summer monsoon". The second transition is the onset of the Indian monsoon to the west of 80°E. This occurs about a month later around subperiod 11 (5–9 June) during which the southwesterlies over the Arabian Sea develop and heavy rains commence at the southwestern tip of India. At about the same time the Mei-yü starts in the Yangzi River Valley.

2) The circulation in the upper troposphere also shows drastic changes during the two transition periods described above. The changes are most clearly recognized by abrupt movements of the center of the South Asian anticyclone at 200 mb. During the first transition period the anticyclone center over Southeast Asia moves suddenly from 10°E to 20°N with an increase of the 200–500 mb layer mean temperature over the eastern Tibetan Plateau and the China Plain. With the increase of the temperature over these regions, the mean meridional temperature gradient between 5° and 25°N reverses over the longitudes east of 85°E. During the second transition period another anticyclone center develops over the Saudi Arabia–Iran region. This period is characterized by the spreading of warm tropospheric air 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 mean meridional temperature gradient between 5° and 25°N takes place over the longitudes west of 85°E.

3) The time sequences of the vertically integrated heat source $\langle Q_v \rangle$ and moisture sink $\langle Q_v \rangle$, and the OLR flux in longitudinal and latitudinal sections show that the abrupt changes in the circulation and temperature during the two transition periods are generally accompanied by sudden increases of heating over the respective regions. Precipitation is the major heat source except over the Tibetan Plateau where continuous sensible heating is observed throughout the analyzed period.
4) The mean diabatic heating is mainly confined to the south of 30°N and east of 70°E with maximum heating over the Burma–Thailand region, the Bay of Bengal coast, the western coast of India, the South China Sea and the South China coast. The regions of heating generally coincide with those of large moisture sink and precipitation, demonstrating that the release of latent heat of condensation is the primary heat source. Over the Tibetan Plateau, however, intense heating is accompanied by a much smaller moisture sink and very little precipitation, showing the importance of the sensible heat flux from the ground surface. On the other hand, intense cooling with negative moisture sinks prevails over the desert areas of Saudi Arabia and North India, the Gobi Desert, and the western Arabian Sea, suggesting strong radiative cooling and evaporation exceeding precipitation in these regions.

5) A prominent feature of the large-scale circulation over Asia during this period is the presence of the circulation induced by the Tibetan Plateau, which coexists with the principal monsoon circulation characterized by the southwesterlies at low levels and the South Asian anticyclone at upper levels. The thermal effects of the Tibetan Plateau manifest themselves as low-level inflow towards the Plateau, a warm temperature anomaly and intense upward motion over the Plateau. The existence of the Plateau-induced vertical circulation is evident in the distribution of descending motions in the surrounding desert areas. The sensible heating from the elevated surface and the radiative cooling in its neighborhood maintain the temperature contrast that drives intense upward motion over the Plateau and subsidence in the surrounding areas.

6) The analyses of the warming of the 200–500 mb layer mean temperature during the two transition periods show that the dominant mechanism for the temperature increase is adiabatic warming due to subsidence. The direct effects of the sensible heating over the Tibetan Plateau and the condensation heating associated with monsoonal rains on the observed temperature changes are relatively small because of the compensating advection effects. However, these heat sources are exerting large influences upon the temperature of surrounding areas by inducing large-scale subsidence there. In particular, the adiabatic warming due to descent on the northeast and west sides of the Tibetan Plateau appears to be the primary mechanism responsible for the circulation changes leading to the establishment of the Asian summer monsoon.

The mean features and time evolution of the general circulation over Asia from midspring to early summer 1979 can be described in terms of the interaction of the vertical circulation induced by the heated Tibetan Plateau with the principal monsoon system which migrates northward.

During the analyzed period the general circulation exhibits two distinct periods of abrupt changes resulting in the successive commencements of summer rains over Southeast Asia and the west coast of India. Because the Tibetan Plateau with its maintained heat source acts to offset the effects of the cold advection and adiabatic cooling, the major tropospheric warming occurs first in May over the areas extending from the eastern Plateau to the South China Plain. The primary causes of this warming are the diabatic heating over the eastern Plateau, the subsidence over the northeastern periphery of the Plateau and the warm advection over the Yangzi River Valley. The meridional gradient of the mean tropospheric temperature between 5° and 25°N reverses over the longitudes east of 80°E. This results in the commencement of the low-level southwesterlies and early summer rains over Assam, the Bay of Bengal coast, Indochina and the South China Sea.

The release of latent heat of condensation associated with the rains over these regions and the continuous heating over the Plateau results in further intensification of the vertical circulation of the principal monsoon system whose ascending branch joins with that over the Plateau. The subsidence over the Iran–Afghanistan–western Plateau region continues. The warming over these regions then leads to the second transition of the circulation in June which brings southwesterlies over the Arabian Sea and the heavy rains over the west coast of India.

These suggestions are based on the results and interpretation of synoptic and thermodynamic budget analyses only. Dynamical processes responsible for the transitions of the monsoon circulation, such as the mechanism of the onset vortices and the vorticity balance of the South Asian anticyclone, require additional investigations.

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APPENDIX

The Upper Boundary Condition for ω and Corrections to the Horizontal Divergence

To obtain reliable estimates of Ω₁ and Ω₂, especially Ω₁ in the upper troposphere, we tested more than twenty combinations of various lower and upper boundary conditions for ω and the correction method for the horizontal divergence. A logical choice of the upper boundary condition for ω may be
\[ \omega = \omega_T = - \frac{\partial}{\partial t} + v \cdot \nabla \theta - \left( \frac{p_0}{p} \right)^{\gamma} \frac{\partial p}{\partial p} \left( \frac{p_0}{p} \right)^{\gamma} \left( \frac{\partial}{\partial p} \right), \]  
\begin{equation} \tag{A1} \end{equation}

at \( p = p_T \) (Nitta, 1977) where \( p_T \) should be chosen at a stratospheric level where the convective vertical heat transport can be ignored. However, our tests revealed that the use of the wind and potential temperature data at 70 and 50 mb levels gives rather erroneous \( \omega_T \) because of poor quality of the analyses at these levels. Therefore, we choose \( p_T \) in the 100–150 mb layer where the net radiative heating is known to be small (Katayama, 1967; Doplick, 1972; Cox and Griffith, 1979).

The final chosen combination of (2), (3) and (4) gives the most satisfactory results, i.e., 1) relatively small \( D' \), 2) reasonable \( Q \) values in the upper troposphere, and 3) mutually consistent \( \langle Q \rangle \) and \( \langle Q_2 \rangle \) values.

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