Effects of the Qinghai–Xizang (Tibetan) Plateau on the Circulation Features over the Plateau and Its Surrounding Areas

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Received November 22, 1990; revised July 30, 1991

ABSTRACT

Areview of the effects of theTibetan Platean on circulation features over the plateau and its surrounding areas has been made, with a special emphasis upon the monsoon circulations in South Asia and East Asia. This includes estimates of heat sources, dynamic and thermal effects of the plateau, adn effects of the plateau on summer and winter monsoons. Major progresses made in this aspect by Chinese meteorologists have been specifically described and are compared with the achievements made by the meteorologists of other countries.

I. INTRODUCTION

The Tibetan Plateau is one of the most complex geographical features in the world, with an average elevation of about 4 kilometers. Thus, it not only exerts a significant mechanical barrier effect,But also serves as an elevated heat source that generates a temperature contrast with the surrounding free atmosphere. These two important effects of the Tibetan Plateau to a great extent determine the monsoon circulation over South and East Asia, with the mechanical barrier effect mainly affecting the Asian winter monsoon circulation and the thermal effect mainly affecting the summer monsoon over Asia.

In fifties of this century, a network of surface and upper–air stations was established with most of stations situated above an altitude of 3000 m. One may use the data obtained with this network to study the large–scale aspects of the circulation systems over the Tibetan Plateau. These achievements are mainly reflected in the pioneering book on the Meteorology on the Tibetan Plateau (Yang et al., 1960). During the time period from 1961 to 1970, data coverage and quality continued to improve with the estabishment of other radiosonde and surface stations. The main results during this period were well summarized in the book on the Meteorology over the Tibetan Plateau (Yeh, Gao et al., 1979). In FGGE year, the observational network was further upgraded with new and some temporary stations set up over central and eastern Tibetan Plateau. In particular, Chinese scientist carried out the Pinghai–Xizang Plateau Meteorology Experiment (QXPMEX) during the 1979 summer (May to August), the same period as the summer MONEX. The main objectives of QXPMEX were to study the thermal and dynamic effect at the surface of the Tibetan Plateau on the general circulation in East Asia, the radiation and heat budget at the surface of the plateau, and the weather systems under various synoptic conditions.

Utilizing the EXPMEX data with a number of additional observations included, investigators at the first time are able to analyse and study the heat budget and circulation features over the Tibetan Plateau with the greater reliability adn details than ever, this significantly improving our understanding of the orographic influence of the Tibetan lateau upon monsoon circulations.

Following the QXPMEX, the Chinese meteorologists once again carried out another field experiment over the Tibetan Plateau (Qinghai–Xizang Land Surface Heat Source Experiment) in the time period from August 1982 through July 1983, with a special emphasis on the geographic distributions of the heat sources and sinks over the Tibetan Plateau and their seasonal variations. For this purpose, four Heat Source Observing Stations (HSOS) were specially set up over the plateau, at each of which the components of the surface radiation balance were directly measured. Based on these data one may readily calculate the surface radiation balance over the plateau that is part of the heat sources. In this review paper, we shall concentrate on the thermal and dynamic effects of the Tibetan Plateau on the winter and summer monsoon circulation on the large–scale basis.

II. HEAT SOURCES OF THE TIBETAN PLATEAU AND NEIGHBORING AREA

The efforts to investigate diabatic heat sources over the Tibetan Plateau have been made by several researchers by employing FGGE / MONEX and QXPMEX data on both global scales and regional scales. Two fundamental problems have been dealt with: (1) is the Tibetan Plateau heat sources or heat sinks and (2) is it the sensible heat or latent heat to make the dominating contribution to the heat sources? The evidence that the Tibetan Plateau is a heat source in summer was previously indicated, but it was not conclusive for winter situation. According to the recent studies (Yeh, Gao et al., 1979) the Tibetan Plateau on the average deemed to be the heat source relative to the atmosphere above it for both winter and summer. Namely, on the whole year, the heat is transported from the surface of the plateau upward to the atmosphere in the different form. Here, the heat source is defined as upward heat transport from underlying surface to the atmosphere. On the whole year, the turbulent sensible heat transfer from the earth's surface is the maximum component of the plateau heat source; the effective radiation is next to the sensible heat; the evaporation is at the minimum. In July and August, the surface evaporation reaches its peak, but still much less than the turbulent sensible heat. In the rest of the months, the latent heat carried upward by the evaporation may be negligible. During the winter, the surface effective radiation becomes the maximum component and the turbulent sensible heat transfer is of secondary importance.

For the annual mean, the upward heat transfer from the surface of the Tibetan Plateau to the atmosphere is in the order of nearly 340 cal / cm^2 • d (Yeh, Gao et al., 1979), which accounts for 65% of the global maximum surface heat source near the eastern coast of North America ($\sim$ 575 cal / cm^2 • d). This fact demonstrates the importance of the Tibetan Plateau for the general circulation from the viewpoint of energetics. The total energy supplied by the Tibetan Plateau has its maximum in late spring and early summer, with the peak of about 500 unit occurring in May. The minimum heat transfer of nearly 200 unit is observed in December. The heat to be actually used to warm the local atmosphere of the Tibetan Plateau has five components: The effective radiation coming from the surface (LR_1), the surface turbulent sensible heat transfer (SH), the solar radiation (SR), the condensation heat released through local precipitation (LP) and the outgoing long wave radiation at the top of the atmosphere (LR_2). The sum of these five components of diabatic heating (cooling) may be written as follows:

$$E = SH + LR_1 + LP + SR + LR_2 \ , \tag{1}$$

where E is defined as the atmospheric heat source. If $E > 0$. it is the heat source; if $E < 0$, it is the cold source. The computational results are shown in Table 1. For the annual mean, the tropospheric air column per squared cm can receive the heat of 40–50 cal / d. Therefore, the atmosphere over the Tibetan Plateau is the heat source on the annual basis. The significant seasonal variations can be found. From March through September, the atmosphere over the Tibetan Plateau has a net heat gain, showing the nature of heat source. A part of heat thus acquired is used to heat the atmosphere over the Tibetan Plateau and another part of it is transported outside the plateau. In late fall and winter, the atmosphere of the Tibetan Plateau is the heat sink, with the maximum observed in December and January. One may note that the months when the atmosphere of the Tibetan Plateàu has the maximum net heat gain are not July or August, but June.

Finally, the monthly heat transfer from the earth–atmosphere system of the Tibetan Plateau to the surrounding atmosphere (F) may be easily estimated below:

$$F = SH + L_e + LR_1 + SR + LP - LR_2 - C_p M \Delta T \ . \qquad (2)$$

Here L_e is the surface evaporation latent heat, M is the total air mass above the Tibetan Plateau in the layer of 600–100 hPa, and ΔT is the month–to–month variations of temperature. The calculated results are also presented in Table 1 (Yeh, Gao et al., 1979). On the average for the whole year, the earth–atmosphere system of the Tibetan Plateau transports 80 cal / cm^2 • d outside to the surrounding atmosphere. Therefore,this earth–atmosphere system is a heat source on the annual basis. In winter (December through February) it is a heat sink, with the maximum observed in December and January when the heat of 140–150 cal / cm^2 • d is transported into the earth–atmosphere system from the outside. In spring and summer, a huge heat source exists over the Tibet, with the maximum observed in June when the heat of 210 cal / cm^2 • d is transported outside. Thus, one may conclude that the thermal effect of the Tibetan Plateau should be very significant.

Tabel 1. The Atmospheric Heat Source (E) and Energy Budget of the Earth–Atmosphere System (F) of the Tibetan Plateau. Unit: cal / cm^2 • d

	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.	Annual average
E	−48	−87	51	124	193	224	208	152	91	−20	−112	−159	43
F	−147	−88	44	119	181	207	203	154	101	−9	−98	−152	43

There have been for long time the different opinions about the nature of heating of the Tibetan Plateau or the question as to which component of diabatic heating is the dominating one. Previously, Flohn (1968) suggested that the Tibetan Plateau may be considered as an elevated sensible heat source, and estimated the input of sensible heat into the atmosphere from a semi–arid surface in the western Tibet based on the surface heat budget, and obtained the heat flux of about 120 w / m^2. Flohn, however, then changed his viewpoint with a special stress on the release of condensational heat by convection, especially over the southeastern region of the Tibetan Plateau. This is consistent with Riehl's (1967) speculation that latent heating, rather than sensible heating is the primary factor in atmospheric heating sources over the eastern Tibetan Plateau. Thus, these works emphasized the importance of latent heating. Based on stations data for ten years (1961–1970), Yeh, Gao et al. (1979) computed various components of long–term mean heat balance over the Tibetan Plateau and their seasonal variations, and pointed out that there is a marked difference in the nature of the diabatic heating between the eastern part and western part of the Tibetan Plateau. They .has obtained an extremely large sensible heat flux over the arid western Tibetan Plateau with a maximum of 450 cal / cm^2 • d (~219 w / m^2). This is about twice the magnitude of Flohn's estimate of 250 cal / cm^2 • d. As compared with SH over the western Tibet, SH over the eastern Tibet is much less pronounced; yet it exceeds LP until June, which indicates that the sensible heating predominates over the entire Tibet until the onset of summer monsoon or rainy season occurs.

Then during the period of summer (July and August), LP becomes slightly larger than SH. Bacause of the predominance of SH, the net heating E is quite large over the western Tibetan Plateau, with the results that this region contributes most to the net heat balance over the entire Tibetan Plateau. The contribution from the eastern Tibetan Plateau is much less significant. During FGGE years and QXPMEX, data coverage over the plateau was much improved. These data provide an excellent opportunity to reestimate the heat and moisture budgets over the Tibetan Plateau. Weng (1986) calculated the surface heat balance of the Tibetan Plateau from May to August of 1979.

In May, which is the dry season for the plateau, the whole plateau acts as a strong heat source with the maximum exceeding 105 w/m^2 found at the surface of the central and southern Tibet, where there exists abundance of solar radiation due to lack of rainfall and cloudiness. In June, the general pattern of the heat source is similar to that of May, but with maximum increasing up to 120 w/m^2. In July when the rainy season sets in over the Tibet, the surface heat source further intensifies all over the Tibet except for the eastern plateau. The distinct increase in the Yarlung Zangbo River in the southern Tibet and the Sichuan Basin is evident. It is worthwhile to point out that this increase is due to the increase in the latent heat transfer that may be caused by nocturnal rainfalls and moist ground surface there. The maximum latent heat component was observed over the Yarlung Zangbo River that is coincident with the maximum of the surface heat source. In contrast to the seasonal variation of the latent heat, the sensible heat greatly decreased from the dry season (May) to the wet season (July), especially over the eastern Tibet, with the maximum drop occurring in the Yarlung Zangbo River. This trend of seasonal variation is in broad agreement with those of the previous result for the long–term mean obtained by Yeh, Gao et al. (see Table 1), but with different magnitudes. For example, the calculated values of latent heat are considerably larger than those of Yeh, Gao et al. In addition, the surface sensible heat transfer is found to be the primary process for heating the atmosphere during the dry period, especially over the western Tibetan Plateau.

Combining QXPMEX data together with the satellite data from Nimbus–7 giving the radiation balance at the top of the atmosphere, Chen, Reiter and Feng (1985) calculated the monthly mean atmospheric heat source over the Tibet from May to August of 1979. Their calculated values of the atmospheric heat source turned out to be considerably smaller than those provided by Yeh, Gao et al. (1979), about half of the previously estimated values (64 w/m^2 for the average over the whole Tibet; 82 and 46 w/m^2 for eastern Tibet and western Tibet, respectively. The main reason for the large discrepancy may be the different assumption of the drag coefficient, which was estimated to be 4×10^{-3} during QXPMEX, nearly half of the value used by Yeh, Gao et al. Another possible reason is that the net radiation observed by Nimbus–7 is smaller than the previous value, but the net radiation at the ground surface is greater than the previously used value. But, they also obtained a similar result that the atmospheric heat source over the Tibet is mainly modulated by the release of latent heat.

Table 2 and Table 3 demonstrate the comparison of total heat and moisture budget over the eastern and western Tibetan Plateau obtained by the various investigators. Over the western Tibet, Luo and Yanai (1984) using a different approach obtained a smaller SH value (109 w/m^2) than the June mean value of Yeh, Gao et al. (219 w/m^2). The latent heating there is very insignificant, so that the net heating $[Q_1]$ value (142 w/m^2) derived by Yeh, Gao et al. is one third larger than that obtained by Luo and Yanai (101 w/m^2). For the moisture balance, the evaporation is the maximum component, with the value over the western Tibet is very close to the average value for the whole Tibet for June. Over the eastern Tibetan Plateau, the SH derived by the three groups are nearly identical to each other. The latent heating obviously becomes large, but still smaller than SH. The net heating $[Q_1]$ is at the range of 94–120 w/m^2 with the contributions of LP and SH both being rather significant. For the $[Q_2]$, the contributon from LP exceeds that made from LE, thus creating the positive $[Q_2]$. From the above descriptions, one may clearly conclude that the sensible heat transfer from the surface is the dominating factor in the heat balance over the whole plateau, in particular over the western part of the Tibet in June of 1979; the latent heating is very important for the heat budget over the eastern Tibetan Plateau, especially during and post to the summer rainy season.

Tabel 2. The Comparison of Heat and Moisture Balances over the Western Tibetan Plateau Obtained by Various Investigators. Unit: w/m^2

	$<[Q_1]>$	$<[Q_R]>$	$[LP]$	$[SH]$	$<[Q_2]>$	$[LP]$	$[LE]$
Luo and Yanai (Dry season prior to rainy season)	101	−77	9	(169)*	−22	9	(31)
Yeh and Gao (June)	142	−94**	17	219	−22	17	39**

* estimated as the residual of heat balance equation

** average value for the whole plateau

Tabel 3. Same as Table 2, but for the Western Tibetan Plateau Unit: w/m^2

	$<[Q_1]>$	$<[Q_R]>$	$[LP]$	$[SH]$	$<[Q_2]>$	$[LP]$	$[LE]$
Lou and Yanai (39–day mean)	113	−62	71	(104)	44	71	(27)
Yeh and Gao (June)	(94)	−94**	86	(102)	(47)	86	39**
Nitta (100–day mean)	120	−75	90	(105)	(25)	90	(65)

The horizontal distributions of the vertically integrated heat source $<Q_2>$ and moisture sink $<Q_2>$ averaged for the 40–day period from May 26 to July 4, 1979 show that the most pronounced heat source of $\approx 300\ w/m^2$ is not located over the Tibetan Plateau as previously thought, but over the Assam–Bengal region. Consistent with the moisture sink of the same order of magnitude, this maximum heat source is undoubtedly related to the heavy rains in this region. This result is in basic agreement with that obtained by Chen and Li (1985) who estimates the atmospheric heat source by use of the direct approach. Their maximum heating rate of 8.0 °C/d is located over the northeastern part of the Bay of Bengal whereas the long–term mean heating rate (1961–1974) over the Tibetan Plateau is only of about 1 °C/d that can not be regarded as a major heat source in summer monsoon region. By use of similar method, Yao and Luo (1982) also estimated the monthly mean heating fields of the atmosphere over the Tibetan Plateau and its surrounding areas and their annual variation utilizing the climatological data of 1961 to 1970 for more than 180 meteorological stations. Major heat source is located over the region of 60°E–100°E longitudinal range to the south of 45°N, with the maximum over northern part of the Bay of Bengal, Assam and North India. Average heating rate in this region is about 8 °C / day, as same as nearly the value obtained by Chen and Li. The heating rate over the Tibetan Plateau is only of 1.0–1.5 °C / d. In winter, the heat sink over the Tibetan Plateau seems to be weaker than that to the south and that to the north of the plateau.

III. THE THERMAL EFFECT OF THE TIBETAN PLATEAU ON THE CIRCULATION FEATURES OVER THE PLATEAU AND ITS SURROUNDING AREAS

Large plateau and mountain ranges generally constitute elevated heat or cold sources which generate baroclinicity that results in the great variety of circulation systems. The Tibetan Plateau with the significant seasonal variations of atmospheric heat source or sink can exercise a marked thermal effect on the flow processes and circulation features over the plateau and its surrounding areas. Over the plateau region, the winter mean pressure pattern at low–level shows that high pressure systems dominate. One may observe a closed high at 600 hPa, which is basically separated from the Siberian cold high on the long–term mean chart. Therefore, they are two independent circulation systems. This fact reflects the prominent thermal impact of the Tibetan Plateau on the regional circulation system. In summer, there is a closed heat low over the Tibet at low level. This heat low over the plateau is also independent from the heat low or monsoon through over India on the long–term mean 600 hPa chart. There are semi–permanent meso–scale features in the low pressure area whose impact is felt in the development of convective precipitating systems. The apparent seasonal variation of the height patterns at low–level described above (the dominance of low–pressure conditions in summer and of high–pressure conditions in winter, respectively) can cause the monsoonal flow around the plateau that may be considered as a self–maintaining circulation system. The layer characterized by a monsoonal wind reversal (> 120 degrees between winter and summer) is rather distinct over the Tibetan Plateau, with the thickness of monsoon layer greater than 4 km over main body of the plateau . The high elevation of the Tibetan Plateau obviously generates a great degree of baroclinicity with its surroundings, causing a very vigorous circulation system to establish itself. The active layer of monsoonal flow may extend downstream far away from the plateau region (up to the central China).

Thus, the monsoon system over the Tibet may also exert an important impact on the winter and summer monsoon in Asian region. Occasionally, the two independent cold highs on the mean chart for winter and the two independent heat lows on the mean chart for summer may merge together, thus leading to blowing of northeast monsoon in winter and southwest monsoon in summer up to the Tibetan Plateau. Under these conditions, one may observe outbreaks of cold air over the Tibet in winter and vigorous transportation of abundant moisture over the Tibet in summer. In winter, there is a belt of low perssure surrounding the surface cold high over the plateau whereas there is a belt of high pressure surrounding the heat low over the Tibet in summer.

There are four types of weather systems at low and middle levels over the Tibetan Plateau in summer, i.e., the low–level vortex, wind shear–line, moving anticyclone and cold air outbreak. The vortex at 500 hPa is one of the sub–synoptic scale rain–bearing weather systems on the plateau in summer. The vortex occurred mainly in the central and western plateau during QXPMEX with the total number of 54 cases. The vortex moved eastward generally along the shear–line and dissipated in the eastern plateau and at its eastern periphery. Among them, only a few can move eastward out from the plateau. Most of vortices originated in the central plateau. Based on the statistics of frequency of occurrence and eastward movement of vortices over the plateau for May through August of 1975–1982 (Zhang et al., 1988), it is seen that there were 54 vortices during QXPMEX, among them only 7 moved out from the plateau, with 5 vortices dissipating at 102–110°E and 2 vortices moving to the east of 110°E. This is broadly similar to the mean condition for 1975–1982, under which there were 37 vortices per year, of them only 9 vortices could move out from the plateau. Seven vortices further dissipated at 102–110°E and two vortices moved to the east of 110°E.

The upper level weather system in summer over the plateau is predominantly the Tibetan high that is part of the huge South Asia high. The Tibetan high is of thermal nature, with its peak intensity, horizontal extent and high stability at 100 hPa. It has a very great effect on the circulation condition in the Northern Hemisphere. In general, when the Tibetan high at 200 or 100 hPa superposes over the high at lower–and middle level (500 hPa), the plateau would have a dry season or spell, whereas the plateau would often have the rainy season or spell when the upper–level Tibetan high lies over a low–level low. The Tibetan high is the major center of action in the Northern summer and its formation and maintenance are to a great extent related to the effect of heat sources over the Tibet and its surrounding areas.

The high demonstrates the quasi–periodic oscillation about these positions, with the biweekly oscillation often observed. It is not clear yet as to the reason of the zonal oscillation of the Tibetan high. On the one hand, this oscillation reflects the thermal effect of th Tibetan high, mainly related to the distributions of latent heat and sensible heat over the Tibet and its eastern periphery. Previously, some investigators stressed on the role of release of latent heat for the Tibetan high. On the other hand, the zonal oscillation of the high is believed to be associated with the adjustment of long–wave pattern in the middle and high latitudes, in particular the interaction between the troughs in westerlies and the Tibetan high (Yeh, Gao et al., 1979).

According to recent studies (Murakami and Ding, 1982; Yanai, 1987), the seasonal advance and retreat of the Tibetan high has a more complicated behavior. From both the climatological mean and case study (Ding et al, 1988), it has been found that while the Tibetan high advances northward there is a center of high to move over Iran–Afghanistan and the western part of the Tibet from the Arabian Peninsula and the latter has a greater intensity and a farther northern position than the eastern one. This result, on the one hand, indicates the fact that the Tibetan high should be viewed as one cell out of the upper tropospheric belt of anticyclonic circulation. This cell is regionally intensified due to the effect of the plateau. On the other hand, the thermal effect of the Tibetan Plateau on the formation of the Tibetan high should not be overemphasized. Now it is not clear on the cause of formation of the high center over the western part of the Iranian plateau. This is an important problem to be further studied.

The meridional circulation over the Tibetan Plateau and its surrounding areas is closely associated with the characteristics of distribution of heat source and sink over the plateau and continental region. The cross–sections of mean meridional circulation along 90°E show the condition across the Tibetan Plateau (Yeh, Gao, et al., 1981). The Hadley cell in winter is very vigorous, extending northward all the way to about 30°N. This indicates that the atmosphere over the Tibetan Plateau acts as the cold source in winter. Further, this heat sink is found in the middle troposphere, thus greatly enhancing the intensity of the Hadley circulation in the monsoon region. As a result, one may observe the strong Hadley meridional circulation at the longitudinal range of the Tibetan Plateau. Eastward from this region, the Hadley circulation gradually weakens and is not detectable in the central Pacific. Therefore, the mean Hadley circulation in the Northern winter is likely to be mainly a phenomenon over the continent.

The summer vertical circulation induced by the Tibetan Plateau is significantly different from that for winter mean condition. On the whole, there exists a huge monsoon meridional cell, with upward flow of its northern leg extending up to 40–50°N. The descending motion is predominantly observed in the Southern Hemisphere. Also, there are two smaller meridional cells on the northern slope and southern slope of the plateau, respectively. The northern one only reaches 300 hPa in the vertical extent while the southern one extends upward up to over 200 hPa. These two meridional cells only show up at the longitudinal range of the Tibetan Plateau, indicating that they result from the heating effect of the plateau. For the individual year, the above condition is also true. For instance, the similar meridional circulation was observed for the period prior and post to the onset of summer monsoon of 1979 (from May to the early part of July of 1979). The east–west and north–south vertical cross sections of the 80–day mean meridional circulation show that 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 distribution of the vertical circulation in this plane is similar to that of the daily mean vertical motions along 35°N for July simulated by Kuo and Qian (1981).

As indicated above, the maximum sensible heating occurs during the seasonal transition period from spring to summer (April to June). Affected by the heat source, the atmosphere over the plateau would warm up rapidly, so that the temperature gradient on both sides of the plateau can be changed. Correspondingly, the flow pattern around the plateau would be also changed. The atmosphere over the plateau in the mid–summer has abnormally high temperature, with the warmest region in July and August found in the region of the longitudinal range of 50–110°E, including the Tibetan Plateau and the Iranian Plateau. During the transition season from spring to summer, the warming in this region occurs also earlier than the other zones of the same latitude. In March, the increase in thickness is very evident and attains its maximum in May and June. This drastic change in temperature during the transition season is believed to cause a corresponding change in wind field.

Yin (1949) earlier pointed out that the westerly wind at upper level to the south of the Tibetan Plateau assumes the feature of sudden change. In the early part of June, the upper–level westerly wind drastically decreases. This decrease in wind speed is related to the onset of Indian southwesterly monsoon. Later, Yeh et al. (1958) found that the above–described sudden decrease and northward withdrawal of upper–level westerly jet–stream in East Asia may be observed in the whole Northern Hemisphere, but this change in wind field occurs earliest over the Tibetan Plateau. This fact reflects the particularity of the plateau region. Yeh et al. further indicated that this change in wind field is closely associated with a number of weather events and processes, including (1) the sudden northward jump of the subtropical jet in westerlies from the region to the south of the plateau to region to the north of the plateau, and at the same time moving of the upper–tropospheric anticyclone (Tibetan high) over the Tibetan Plateau and the establishment of the upper–level easterly jet to the south of the plateau; (2) meanwhile, onset of Indian southwest monsoon with much monsoon rainfall occurring along the west coast of the Indian peninsula; (3) the northward shift of polar front over the East China and the Yangtze River valley, causing the simultaneous onset of plum rains in the Yangtze River valley and rainy season in Japan.

Based on the long–term mean data, the fact that the jet stream in westerlies suddenly retreats, has been also documented. On the average, June is a time period when the subtropical westerly wind jet undergoes the dramatic change. About in the third pentad, the jet significantly diminishes to the south of the plateau, and soon disappears (northward retreat). At the same time, the west wind rapidly shifts to the steady upper–level easterlies at Calcutta (22.5°N, 87°E). The interannual variability in the northward retreat of jet is very great, with earlier retreat observed at the end of May or the beginning of June. Therefore, May and June is the time period of sudden decrease in westerly wind speed and alternation of west wind and east wind. This process is fully consistent with the seasonal variation of heating field over the plateau and the resulting seasonal evolution of temperature and height fields. As previously indicated, May is the month of the maximum sensible heating and also the month of the maximum increase in the air temperature over the plateau. Under the influence of this very strong heating, the atmospheric temperature over the plateau sharply increases, this leading to weskening of the nomal meridional temperature gradient or the reverse direction, and further the above change in the upper–level westerly wind jet.

According to the study of summer monsoon of 1979 (Murakami and Ding, 1982), this seasonal northward jump of westerly wind jet clearly showed up over the extensive Eurasian region of the Tibetan Plateau (Fig.1). For instance, the 300 hPa U wind component along 75°E exhibits a distinct northward shift of strong (20 m / s) westerlies from about 30°N

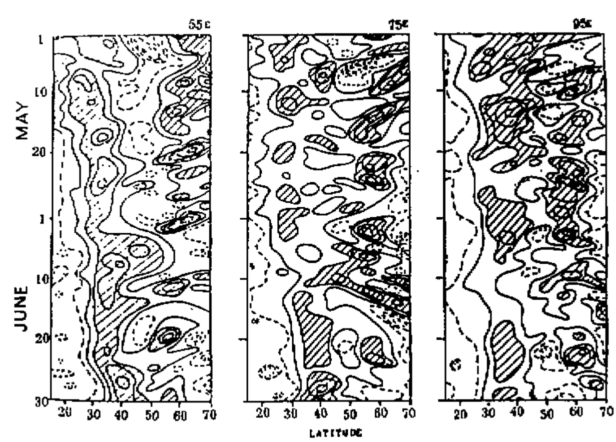

Fig.1. Time–latitude section of 300 hPa U, averaged over a region extending 10 degree of longitude east and west of 55°E (left), 75°E (middle) and 95°E (right),respectively, during the early summer (1 May to 30 June) of 1979. Intervals are 10 m / s with full and negative wind, respectively. Hatching denotes regions of greater than 20 m / s westerlies. Dashed hatching denotes regions of greater than 10 m / s easterlies. (After Murakami and Ding, 1982).

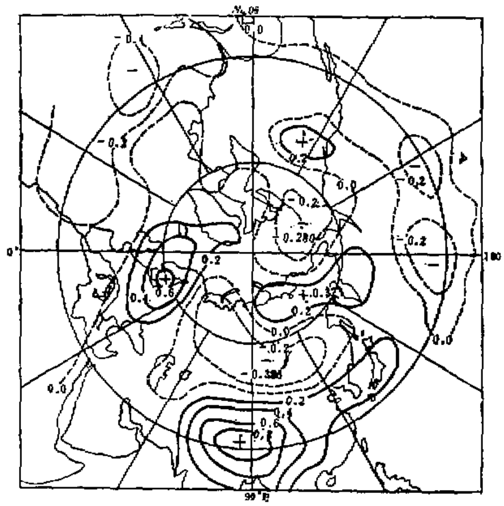

Fig.2. The correlative pattern of 500 hPa disturbed height field for the Northern summer, with the point of 90°E, 30°N as the base point.

during May to 35–40°N after mid–June. A similar northward shift of the 300 hPa jet stream also takes place very far upstream (55°E) of the Tibetan Plateau on about 3 June. Yin (1948) postulated that the onset of the Indian monsoon occurs almost simultaneously with a sudden shift in the location of the upper–tropospheric subtropical jet stream from the southern to northern periphery of the Tibetan Plateau. However, this is not evident in their study as the time–latitude section along 95°E (eastern Tibetan Plateau) does not exhibit a similar northward shift of 300 hPa U jet stream. This supplies the possible existence of the significant interannual variability of the seasonal change which is a problem to be further studied. In order to study the role played by the heating field over the plateau and its surrounding areas in the build –up of mean circulation in winter and summer, a number of numerical simulations on this problem have been done by using 5–level p–σ incorporated coordinates (Qian et al., 1988). They have obtained that the contribution of summer heating field to the horizontal flow pattern seems to be more significant in the upper troposphere than in the middle and lower troposphere. If the heating effect is not included in summer, the huge anticyclone in the upper troposphere can not be simulated out. This may result from the fact that the vertical flux of latent and sensible heat may reach up to very high level. However, the case for winter is opposite with the effect of heating field (heat sink) being more significant in the middle and lower troposphere, i.e., the heat sink over the plateau causes the deepening of major trough off coast of East Asia and weakening of the topographic ridge over the plateau.

The thermal condition of the Tibetan Plateau may not only cause the variations of circulation systems over the plateau and its surrounding areas, but also may cause the plenatary–scale circulation anomalies. Huang (1985) has simulated the effect of anomalous heat sources over the Tibetan Plateau on the general atmospheric circulation in the Northern Hemisphere, indicating that when the heat sources over the Tibetan Plateau enhance, does so the Tibetan high; then the trough in North China would intensify; the trongh over Alaska would deepen; the high ridge over North America would enhance. This sequence of downward development is further documented by the studies of teleconnection correlation with observational data (Fig. 2). This kind of teleconnection may reflect a dynamic response to thermal forcing originating in the Tibetan Platerau region in the domain of low–frequency mode. Recently, Sun and Chen (1988) have indicated the possibility that the 30–40 day low–frequency oscillation has its origin region over the Tibetan Plateau.

IV. THE DYNAMICAL EFFECT OF THE TIBETAN PLATEAU ON THE CIRCULATION FEATURES OVER THE PLATEAU AND ITS SURROUNDING AREAS

It is well known that the width of westerlies in the lower troposphere is only 30–40 latitude degrees or even less. However, the longitudinal range of the Tibetan Plateau has 15 latitude degrees, accounting for a half of the width of the westerlies. The zonal extent of the Tibetan Plateau is about 3000 km, accounting for one tenth of the entire latitude circle; vertically, the average height of the Tibetan Plateau is about 4 km, nearly one half –one third of the tropospheric layer. It is very natural to believe that such a huge obstacle is able to exert a significant dynamic effect on the westerlies.

The large–scale dynamic effect of the Tibetan Plateau is another important problem associated with the topographic effect of the plateau. It is related to the formation of the jet stream in westerlies in East Asia, the establishment of the major trough in East Asia and so on. The dynamic effect of the large–scale topography includes the two following aspects: (1) The mechanic barrier effect that can force the airflows to cross over or go around the Tibetan Plateau. This forced flow, either going over or around, is not only associated with the shape of large–scale topography, but also associated with the strength of the airflows that cross over the mountains. Therefore, although the mountain ranges are fixed and unchanged as geographical features, the dynamic effect of the mountains on airflows assumes seasonal variation. (2) The inhomogeneous distribution of friction is caused by the large–scale topography. In this section, only (1) will be discussed.

The dynamic effects of the plateau and mountain ranges have been studied by a number of investigators. Among them, Charney and Eliassen (1949) the earliest estimated the perturbations in the westerlies along 45°N produced by the mechanic effect of large–scale topography. But, their study lent little support to illustration of the dynamic effect of the Tibetan Plateau, as the Tibetan Plateau is mainly located to the south of 45°N. Bolin (1950) studied the theoretical model of flow pattern caused by the circle–shaped mountain in the uniform westerly winds and found that the ridge over the mountain and the first trough downstream of the mountain are nearly consistent with the observations. Using linear models, Zhu (1957) discussed theoretically the contributions of dynamic effects caused by huge mountains to the mean troughs and ridges in the Northern Hemisphere. In recent years, numerous researchers have further studied the dynamic effect of the mountains, especially utilizing the methods of numerical experiments. For example, Okamura (1976) and Nakamura (1978) have studied the dynamic effects of various ideal terrains in different seasons.

When the airflows impinge on mountains, they may go over or go around them. These two dynamic effects are different. Based on the studies of numerical experiment, the purely dynamic effect of the Tibetan Plateau in summer shows up mainly in splitting of the airflows while going around the plateau. Many circulation features are associated with the effect of going around the plateau of airflows, for instance, mean flow pattern of two troughs and one ridge in between to the north of 35°N, the intensity and location of major trough in East Asia, low–pressure systems and the belt of high pressure to the north of it. The effect of sloping and crossing over the mountain of airflows is of the secondary importance, but it may make greater contribution to breaking up of the subtropical high. In winter,in the latitude belt to the south of 40°N where exist highs, the flow patterns are considerably distorted due to the influence of topography. Over the main body of the plateau, the flow pattern assumes significant anticyclonic curvature (ridge), thus creating trough zones. The trough at the eastern periphery of the plateau is of travelling character, and this movement of trough is related to the new development of low trough in the northwesterlies to the northeast of the plateau. This kind of process is somewhat similar to the developmental process of the major trough in East Asia. The trough in the southern branch of westerlies is well simulated numerically and its genesis and amplification mainly come from the contribution of the effect of flow around the plateau, while the anticyclonic circulation feature in the western part of the plateau seems to result from the non–linear interaction between the air flows going over and around the plateau. Thus, either sloping effect or flow around the plateau can not appropriately reflect the influence of topography on the circulation features.

The dynamic forcing can also influence the vertical structure and wavelengths of the disturbance in the various basic zonal currents (Chen, Ji and Shen, 1985). They found that in a baroclinic atmosphere, the dynamic disturbances in different basic westerly currents caused by the Tibetan Plateau bear a westward tilting structure, associated with a well–developed vertical circulation. The strength weakens rapidly with height. The wavelength at mid–latitudes is about 70–90 longitudes, longer than that of corresponding quasi–stationary free wave.

The disturbed energy propagates mainly downstream, but its amplitude decays rapidly. So, its influence is largely restricted to the west of 140°E.

Finally, a comparison of the difference in dynamic effect of large–scale topography between winter and summer may be made as follows:

(1) The dynamic effect of the topography of the plateau for winter and summer is different. In summer, the dynamic effect plays an important role in the maintenance of the belt of high pressure to north of the plateau. The belt of low pressure over the plateau seems to be associated with the thermal effect. In winter, however, the high pressure ridge and thermal ridge over Lahsa are associated with the dynamic effect. In comparison with the climatological mean condition, the simulations for winter are more realistic than those for summer, possibly indicating that the dynamic effect of the plateau in winter may be more important, whereas the influence of the plateau in summer may possibly show up in the thermal effect; (2) Owing to the fact that the difference in numerical simulations between winter and summer results from different zonal mean fields at initial state, one may infer that there exists some sort of interaction between the pure dynamic effect of the plateau and the circulation conditions; (3) In summer, the dynamic effect of the plateau is largely reflected in flowing of airstreams around the plateau, while in winter the flows both over and around the plateau are rather important. When weak basic current impinges upon the plateau, the flow over the plateau would dominate, whereas the flow around the plateau is very important when the aircurrents are strong; (4) The dynamic effect of the plateau for both winter and summer has the most significant influence in the lower and middle troposphere. But this dynamic influence thus produced may propagate upward up to very high level.

V. THE EFFECT OF THE TIBETAN PLATEAU ON THE SUMMER AND WINTER MONSOONS OVER EAST ASIA

As indicated in Section 2, the Tibetan Plateau receives a large amount of the solar radiation in summer, thus causing a significant contrast of heating rate between the plateau region and the surrounding free atmosphere. This kind of horizontal heating difference in the mid–troposphere is even greater than heating contrast between continent and ocean on the plain region. Therefore, one is naturally led to the conclusion that it is far from complete to illustrate the formation of the Asian monsoon according to Halley's classical theory. The Halley's theory, only taking into consideration the continent–ocean heat contrast, interprets the Asian monsoon circulation as a huge sea breeze system produced by the continent–ocean heat contrast. In order to elucidate the formation and development, it is necessary to consider the thermal effect of the plateau by itself. But in recent years, the apparent effect of the Tibetan Plateau upon the summer monsoon circulation was questioned with analysis of 500 hPa temperature field over Eurasia and it was found that an east–west elongated band of high temperatures extended along latitudes 25–30°N from Egypt, through northern India, into southern China. This evidence implies that the Tibetan Plateau is not directly responsible for high temperature over Eurasia through its thermal effect. Thus, one has to reemphasize the importance of the classical theory in explaining the maintenance of the upper tropospheric monsoon anticyclonic system, namely, the Tibetan high is a part of the upper–tropospheric anticyclonic system which extends along 25°N–30°N over the entire Eurasian continent. Nevertheless, these high mountains undoubtedly contribute to the enhancement of local circulation around the entire highland complex. This point will be examined in the following.

Based on the 80–day (for 1979) mean streamlines and isotachs at the 200, 500 and 850 hPa levels(He et al., 1987), the mean 850 hPa field shows southwesterlies over a large area extending from the Arabian Sea to the Bay of Bengal and to Indochina peninsula and South China Sea. There is a cyclonic inflow toward the Tibetan Plateau, which 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, Gao et al, 1979). The inflow exhibits a remarkable diurnal variation. The mean wind fields at this level and 700 hPa suggest the presence of the mean inflow center on the eastern Tibetan Plateau. The inflow is related to the heat low on the plateau. On the east side of the plateau, the confluent zone between the northwesterlies and southwesterlies is the dominant feature of the mean 850 hPa flow.

At 500 hPa the mean winds are generally westerly except for the regions influenced by the subtropical ridges over Saudi Arabia and western Pacific. The wind speed of the 500 hPa flow above the Tibetan Plateau is noticeably small, and the westerly jet stream splits into northern and southern branches between approximately 75 and 100°E. The reduction of wind speed above the plateau well extends up to the tropopause.

A spectacular feature of the Asian summer monsoon circulation is the development of the South Asian anticyclone in the upper troposphere over the low–level southwesterlies. The 80–day mean 200 hPa wind field shows that the mean center of South Asian anticyclone is located over the border of Burma and Thailand.

Next, it is interesting to investigate how these winds are established during the onset of the summer monsoon. In Assam and the southeastern corner of Tibet summer rains start much earlier in central India. The frequency and amount of rainfall during April and May in Assam are remarkable, generally exceeding 200–500 mm of rain for these two months alone. The combined effect of the direct heating of the elevated surface and the release of latent heat in the ascending air causes the upper–tropospheric warming. This upper–tropospheric warming and its associated thermal wind balance result in a weakening of the upper westerlies which circulate around the southern periphery of the Tibetan Plateau during winter and spring. By early June, the continual upper–tropospheric warming causes the westerly jet to

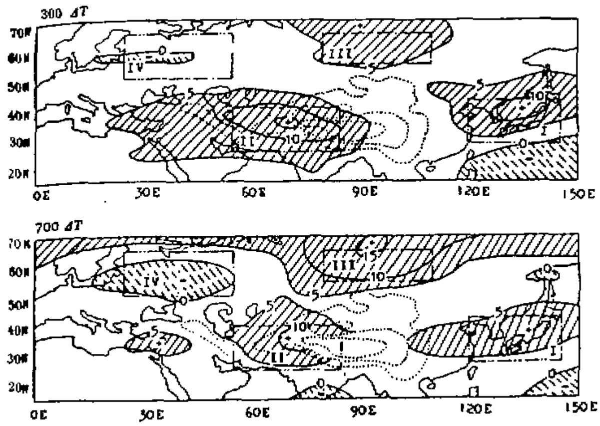

Fig.3. Mean temperature difference between the pre–onset (May 15–30) and post–onset (June 20–30) phases of the 1979 summer monsoon as a function of latitude and longitude at 300 hPa with 5°C interval. Topographic contours (1.5 km interval) are shown by the dotted lines. Areas of T greater than 5℃ are hatched and T less than 5℃ are indicated by dashed–hatching. (After Murakami and Ding, 1982).

disappear and to be replaced by the easterly jet. In general, the establishment of this upper–level easterly jet occurs nearly simultaneously with the onset of monsoon rains over central India. As indicated previously, the shift from the westerly to easterly jet takes place in an abrupt manner. Yin (1949) suggests that the thermal effects of the Tibetan Plateau are responsible for these sudden changes in the zonal wind fields. But, other authors have then postulated different views, finding that the northward retreat of the westerly jet is not only observed around the Tibetan Plateau, but also nearly simultaneously occurs over the Middle East and at locations some distance away from the Tibetan Plateau regions. Therefore, Yeh et al. (1958) postulated that as the elevation of sun increases northward from winter to summer in the Northern Hemisphere, the temperature contrast from pole to equator decreases until it reaches a threshold value, at which time a certain type of instability in the atmosphere appears, causing an abrupt change in the upper–air circulation which is not directly related to mountain effects.

From the differences in the 300 hPa mean temperature fields (DT) between the pre–and post–onset phases summer monsoon of 1979, it is immediately evident that there are well–organized east–west oriented bands of positive temperature DT at approximately 30–40°N across Eurasia with two distinct centers:near Japan and the Middle East and the western part of the Tibet (Fig.3). Hence, these temperature changes between the pre–and post–monsoon onset periods represent a very large–scale phenomenon. The largest temperature increase occurs over the western Tibetan Plateau where the northward retreat of the westerly jet is most prominent. In contrast, over the eastern plateau where atmospheric warming occurs much earlier than the monsoon onset over central India, the temperature increase is small and the associated northward jump of the westerly jet is not clearly defined. Murakami and Ding's study (1982) indicates the importance of the thermal effect of the Eurasian continent as a whole. However, the western Tibetan Plateau is instrumental in causing a local enhancement of the temperature increase during the monsoon onset phase.

From the above discussions, it has turned out that our observational knowledge is insufficient to adequately describe the exact role of the Tibetan Plateau in maintaining the summer monsoon circulation. There have been no any unified results and viewpoints. This may be partly due to the different data sources used in these studies; on the one hand, this fact also indicates that the study of the topographic effect of the Tibetan Plateau is a difficult problem to solve. To overcome this difficulty, the observational studies should be further enhanced. On the other hand, one may also utilize other methods. Among them, an effective method of identifying the role of mountain in the monsoon circulation is to perform numerical simulation experiments. In addition, the rotating annulus experiment is also an important method to study this problem.

Murakami et al.(1970) used an eight–layer, two–dimensional model to study the summer monsoon, with the primary purpose of studying the influence of the Himalayas on the zonal wind components (the low–level westerlies and upper–level easterly jet) of monsoon circulation. Without mountains, the low–level westerly wind maximum was less than 5 m / s and the upper–level easterly wind maximum was less than 10 m / s. When the mountains were included, the low–level westerly jet exceeded 10 m / s and the upper–level easterly jet became stronger than 35 m / s. These model–computed westerly and easterly jets are slightly stronger than the corresponding observed speeds in July along 90°E, but generally the results confirm the strong impact of the Tibet on the monsoon circulation. Hahn and Manabe (1975) performed numerical experiments using a three–dimensional, global general circulation model to study the effect of the Himalayas on the establishment of the monsoon circulation. In the M model with mountains, the rapid northward progress of the subtropical jet across the Tibetan Plateau during the monsoon onset was well simulated. Without mountains (NM model), the subtropical jet does not abruptly jump northward to its summertime position, but rather it slowly moves northward throughout May and June, stabilizing its position in July at a latitude approximately 10° farther south than in the M model. Maximum temperature found over the Tibetan Plateau in the M model agrees with observations, while in the NM model, temperatures at 500 hPa over Tibet are approximately 10–12°C lower than in the M model. This clearly indicates that the mountains, at least in the model, act as a mid–tropospheric heat source and contribute to the formation of warm core low pressure area near the surface of the Tibetan Plateau as well as the warm core high pressure system aloft. The M model and NM model have significant differences in the heating fields. In the M model, the latent heating tends to be most important over the Tibet while in the NM model, sensible heating tends to dominate. In short, the above two numerical experiments both emphasize the crucial role played by the Tibetan Plateau in the development of the monsoon circulation. If there were no these mountain barriers, the monsoon flows in South Asia and East Asia would not be as it is now. Qian et al. (1988) used a five–level limited area model to simulate the role played by the Tibetan Plateau in the monsoon development in South Asia and East Asia. They have drawn the following conclusion: the sea–level pressure and near–surface wind fields are mainly determined by the diabatic heating field associated with the continent–ocean contrast, and the topography only exercises a secondary role in development of these fields. However, the comparison of the mean precipitation rate made for these two experiments demonstrates a remarkable difference, especially in East Asia. If the Tibetan Plateau were nonexistent, the precipitation over the Bay of Bengal would disappear and have a wetter region along the East

Asian coast. This suggests the importance of the topography in precipitation distribution of monsoon region.

All of the above simulations have emphasized the remarkable importance of the Tibetan Plateau in the development of monsoon circulation. Using a different approach, Webster and Chou (1982)performed an experiment which included the Asiatic continent and adjacent oceans, but not the Tibetan Plateau, thereby focusing on the heat contrast between the Asiatic continent and adjacent cooler oceans rather than on the topography. Their results have shown that the interactive nature of oceans adjacent to the Asiatic continent appears to be a crucial element in determining the magnitude and spatial variation of the upper–level monsoon circulation. But, the above numerical experiment can not determine the relative importance of the topographic influence and heat contrast between the Asiatic continent and adjacent oceans.

The Tibetan Plateau can also exert a significant impact on the winter monsoon circulation in Asia. In January, one may observe the strongest winter monsoon in Asia which is the most vigorous part of the general circulation in the Northern Hemisphere. In winter, the Tibetan Plateau acts as a heat sink for the monsoon circulation, and at the same time it has an important mechanical influence that mainly includes two aspects: the splitting of the westerlies and the frictional effect.

It was pointed out in fifties by Chinese studies that the westerlies at low and middle level splits into two major streams as it encounters the western end of the Tibetan Plateau; the upper–level westerly jet stream has the maximum wind speed at the southern flank of the plateau, rather than at the northern flank. Based on the studies for recent ten years in which the more reliable and greater volume of data over the plateau are available, especially the data for 1978–1979 winter (Global Weather Experiment), the previous major results remain to be correct and reasonable, although a quite few new evidences have been provided (Murakami, 1981).

From the 1978–1979 winter mean wind and vorticity fields at 700 hPa, it may be seen that the 700 hPa mean wind splits into the southern and northern branches of major streams starting from the western end of the western plateau. The northern branch is directed northeastward approximately along 2 km height contours, then turns eastward into the saddle area between the Tibetan Plateau and the Altay Mountains. The southern branch, which is slightly weaker than its northern counterpart, flows eastward nearly parallel to the southern periphery of the Tibetan Plateau. To the east of the plateau, these branches tend to converge into a single major stream near 120°E. Of particular interest is the zone of extremely weak winds which extends approximately 1000 km downstream from the eastern slope. Associated with this wind minimum is a vorticity pair, with anticyclonic vorticity to the north and cyclonic vorticity to the south. A similar feature, although not as well defined, is observed near the western end of the Tibetan Plateau. These features are called Karman vortex street. They mainly develop in the stagnation region downstream of a sharp–edged obstacle. Staff Members (1956) had already indicated these phenomena based on the early study.

At 500 hPa, the splitting of the main westerly flow is still evident at both 25°N and 45°N between 70°E and 120°E. An interesting feature is that the winter mean winds at 500 hPa are weakest (less than 10 m / s) directly over the plateau. These extremely weak winds may be an indication of the frictional effects of the Tibetan massif. Murakami (1981) estimated the average height of the planetary boundary layer to be about 1.5 km above the plateau. Thus, over the plateau, the top of the planetary boundary layer may reach up to about 6 km above sea level. It is natural to conceive that the some kind of weather systems, generated in the planetary boundary layer mainly through the frictional effect, once brought out of the plateau region, would have a considerable influence on the weather over the region downstream, especially when they obtain the growth due to some dynamical or thermal mechanisms.

This high planetary boundary layer acts as extending the horizontal and vertical extents of the Asian high mountain ranges. Winter mean winds at 300 hPa (not shown) are also weak over the plateau, indicating that the mechanical effect of the Tibetan Plateau may extend that high. Nakamura and Murakami(1983)simulated flow fields over and around a prescribed obstacle using a six–level primitive equation model, with flows exhibiting features that were similar to the Taylor column phenomenon; namely, lower–tropospheric winds tended to flow around the obstacle, and upper–tropospheric flows were characterized by weak winds above the obstacle. In fact, even at 200 hPa, winds are strongest not above, but along the southern periphery of the Himalayas. This characteristic feature of the winter mean flow cannot be explained by the conservation theorem of absolute potential vorticity. According to this theory, if air encountering an obstacle rises, acquiring anticyclonic curvature as the depth of the air column decreases on its windward side, the wind speed should increase markedly above and to the north of the barrier. This is not the case in the observed winter mean upper–tropospheric flows for 200 hPa.

Another important effect of the Tibetan Plateau on the winter monsoon circulation shows up in the distinct eastward acceleration of the upper–level jet stream south of the Himalayas at the 200 hPa level. Murakami (1981) pointed out that this substantial eastward acceleration is maintained primarily by southerly ageostrophic winds. This southerly flow may correspond to an upper branch of the thermal, local Hadley circulation which is associated with the heat contrast between the cooler continent and high mountains to the north and the warmer ocean and equatorial rainfall to the south. Chinese scientists (Yeh, Gao et al., 1979) demonstrated the importance of a heat sink over the Tibetan Plateau in inducing a strong jet stream to its south. It is reasonable to expect that the available potential energy represented by these temperature contrasts is converted more or less in situ to fuel the acceleration of jet stream.

The existence of a direct solenoidal circulation to accomplish this conversion has been known for many years. Staff members of the Academia Sinica (1957–1958) deduced mean vertical motions indicating a large–scale direct circulation. Murakami and Unninayar (1977), using operational analyses from the National Meteorological Center (NMC), outlined a large–scale direct circulation over East Asia for the 1970–1971 winter. They showed the updraft part of the circulation over the Malaysia–Indonesia area and the sinking motion centered over North China. They pointed to the wet and cloudy weather over the Malaysia–Indonesia area as a major heat source to maintain the supply of available potential energy. Blackman et al. (1977) concluded from a study of momentum flux that in the time–averaged flow, there is a direct meridional circulation in the region of the jet stream acceleration. Cressman (1981) also showed a strong single–celled direct circulation centered on the East Asian–West Pacific upper front and jet stream. Later, his estimates of kinetic energy budgets further provides a quantitative description of the sources and sinks of kinetic energy for the jet stream of that region. In the location of jet stream acceleration, corresponding to a large–scale direct solenoidal circulation, the mean generation of kinetic energy was 95×10^{10} kw or 34 wm^{-2}. Kinetic energy conversion at these rates greatly exceeds that in a typical vigorous extratropical cyclone, which could be in the range of $10–20\times 10^{10}$ kw. Orographic forcing over the east edge of the Himalayan plateau is suggested as a process partly responsible for the geographic reliability of generation region in China.

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