THE PHYSICAL EFFECTS OF TOPOGRAPHY AND HEAT SOURCES ON THE FORMATION AND MAINTENANCE OF THE SUMMER MONSOON OVER ASIA

Huang Ronghui (黄荣辉) and *Yan Bangliang* (严邦良)

Institute of Atmospheric Physics, Academia Sinica, Beijing

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ABSTRACT

The physical effects of topography and heat sources on the formation and maintenance of the summer monsoon over Asia are discussed in this paper by using the transformed Eularian-mean equations and a quasi-geostrophic 34-level spherical coordinate model.

The computed results of the divergence of the E-P flux, the induced meridional circulation and the perturbation geostrophic wind speed induced by the forcing of topography and heat sources show that the diabatic heating effect over the Tibetan Plateau may play an important role for the formation and maintenance of the summer monsoon over Asia, which is much greater than the dynamical effect of topography.

The computed results also show that, of the physical effects of topography and heat sources on the formation and maintenance of the summer monsoon over Asia, the effect of forced meridional circulation is larger than that of the divergence of E-P flux of the induced waves.

I. INTRODUCTION

Monsoon is a climatic system in which wind systems change with seasons. Over South Asia and East Asia, in particular, it has obviously different wind systems in wintertime and summertime. Generally speaking, the monsoon circulation prevaling over South Asia and East Asia has three specific features: (1) there is an anticyclone in the upper troposphere over the Tibetan Plateau, i.e. the so-called South Asian high, and there is a very strong easterly jet over South Asia and East Aisa; (2) a southwest flow prevails in the lower troposphere over South Asia, while over East Asia the flow from South Asia changes into southerly due to the influence of southeast flow from the west of the West Pacific subtropical high; and (3) this monsoon circulation brings the monsoon rainfall in India and Bengladesh and causes the plum rains in China and Japan.

Recently, many investigations show that the said monsoon circulation is in connection with the strong diabatic heating over the Tibetan Plateau. Yeh and Gao (1979) show that the heat source over the southeast of the Tibetan Plateau is the largest heat source during the Northern Hemispheric summer. It is not only a heat source over Eurasian continent that is in contradiction to the cold source over the ocean, but also like a "heat island" directly heating the atmosphere, as the Plateau is very high. Using a numerical model, Ji and Tibaldi (1984) calculated the effect of diabatic heating over the Tibetan Plateau for the onset of the monsoon over Asia. Kuo and Qian (1982) made some numerical experiments on the maintenance of the monsoon circulation over Asia by GCM. The computed results show that the diabatic heating over the Tibetan Plateau plays an important role in the maintenance of the summer monsoon circulation over Asia. However, the physical mechanism of the heat source over the Tibetan

Plateau for the formation and maintenance of the summer monsoon over Asia has not been deeply discussed so far. In this paper, the transformed Eularian-mean equations are used to discuss the influence of planetary-scale eddies induced by the Plateau topography and heat source on the flow, and thus explain the effect of the heat source over the Tibetan Plateau on the formation of the mean monsoon circulation in summer. In addition, a geostrophic 34-level model is used for studying the thermal effect of the diabatic heating over the Tibetan Plateau on the mean monsoon circulations over South Asia and East Asia.

II. THE TRANSFORMED EULARIAN-MEAN EQUATIONS AND THE INDUCED MERIDIONAL CIRCULATION

According to Edmon's (1980) derivation, taking the Eularian-mean of the equations of motion, continuity equation, and thermodynamic equation in the spherical coordinates, and neglecting the small terms, we can obtain the following equations:

$$\frac{\partial \bar{U}}{\partial t} - f\bar{v}^* - \bar{D} = \frac{1}{a\cos\varphi}\nabla\cdot F, \tag{1}$$

$$f\bar{U}_p - \frac{R}{a}\bar{\theta}_\varphi = 0, \tag{2}$$

$$\frac{1}{a\cos\varphi}(\bar{v}^*\cos\varphi)_\varphi + \bar{\omega}^*_p = 0, \tag{3}$$

$$\frac{\partial\bar{\theta}}{\partial t} + \bar{\theta}_p\bar{\omega}^* - \bar{S} = 0. \tag{4}$$

Eqs. (1)—(4) are the so-called transformed Eularian-mean equations. R is the gas constant, $\bar{D}$, $\bar{S}$ is the Eularian-means of friction diabatic heating, respectively. $\bar{\omega}$ is the Eularian-mean of vertical velocity. $(\bar{v}^*, \bar{\omega}^*)$ is the residual meridional circulation and can be written as

$$\bar{v}^* = \bar{v} - \frac{\partial(\overline{\theta' v'}/\bar{\theta}_p)}{\partial p}, \tag{5}$$

$$\bar{\omega}^* = \bar{\omega} + \frac{\partial(\overline{\theta' v'}/\bar{\theta}_p \times \cos\varphi)}{a\cos\varphi\partial\varphi}. \tag{6}$$

From Eqs. (5) and (6) we can see that the so-called residual meridional circulation consists of the meridional circulation of the basic state and that induced by topography and heat source. Because $\bar{v}=0$ and $\bar{\omega}=0$ for the meridional circulation of the basic state in this investigation, $(\bar{v}^*, \bar{\omega}^*)$ is the induced meridional circulation. Thus (5) and (6) can be written as

$$\bar{v}^* = -\frac{\partial}{\partial p}\left(\frac{\overline{\theta' v'}}{\bar{\theta}_p}\right), \tag{7}$$

$$\bar{\omega}^* = \frac{1}{a\cos\varphi}\frac{\partial}{\partial\varphi}\left(\frac{\overline{\theta' v'}}{\bar{\theta}_p}\cos\varphi\right). \tag{8}$$

Moreover, F in Eq. (1) is the divergence of the Eliassen-Palm flux and may be expressed as

$$\nabla\cdot F = \frac{1}{a\cos\varphi}\frac{\partial}{\partial\varphi}[F(\varphi)\cos\varphi] + \frac{\partial}{\partial p}[F(p)], \tag{9}$$

$$F=(F(\varphi),F(p))=\left(-a\cos\overline{u'v'},\ fa\cos\varphi\frac{\overline{\theta'v'}}{\bar{\theta}_p}\right). \tag{10}$$

In the above equations, θ' is the perturbation potential temperature. *From Eq. (1) we can* obtain the following results:

(1) If $\bar{D}=0$, $\bar{S}=0$, and there is no heat transfer, the basic flow will not vary when the E–P flux due to perturbation vanishes.

(2) *If there is heat transfer due to perturbation, the westerly flow will be accelerated,* when the E–P flux due to perturbation is divergent and the induced meridional circulation is northward.

(3) When the E–P flux due to perturbation is convergent and the induced meridional *circulation is southward, the westerly flow will decrease, while the easterly flow accelerates.*

III. MODEL AND PARAMETERS

1. *Model*

In order to discuss the effects of divergence of E–P flux and the induced meridional circulation due to perturbation by the forcing of topography and heat source on the flow, we have to use a numerical model to compute the planetary waves and quasi-stationary perturbations at isobaric surfaces induced by topography and heat source. In this paper, a quasi-geostrophic 34-level spherical coordinate model is used, in which Rayleigh friction, the effect of Newtonian cooling and the horizontal thermal diffusivity are included (Huang and Gambo, 1982). For the planetary-scale motion, the divergent component of motion may play an important role in the meridional transfer of potential vorticity. Therefore, the effect of non-geostrophic wind on the meridional transfer of potential vorticity is considered and we can obtain the following model equations,

$$\hat{\Omega}_{n-\frac{1}{2}}\frac{\partial}{\partial\lambda}\left\{\frac{1}{2\Omega_0\sin\varphi}\frac{1}{a^2}\left[\frac{\sin^2\varphi}{\cos\varphi}\frac{\partial}{\partial\varphi}\left(\frac{\cos\varphi}{\sin^2\varphi}\frac{\partial\phi'}{\partial\varphi}\right)+\frac{1}{\cos^2\varphi}\frac{\partial^2\phi'}{\partial\lambda^2}\right]\right\}_{n-\frac{1}{2}}$$

$$+\frac{1}{a}q_{n-\frac{1}{2}}\frac{1}{2\Omega_0\sin\varphi}\frac{1}{a\cos\varphi}\frac{\partial\phi'_{n-\frac{1}{2}}}{\partial\lambda}=f\left(\frac{\partial\omega}{\partial p}\right)_{n-\frac{1}{2}}-(R_f)_{n-\frac{1}{2}}\frac{1}{2\Omega_0\sin\varphi}\frac{1}{a^2}$$

$$\times\left[\frac{\sin\varphi}{\cos\varphi}\frac{\partial}{\partial\varphi}\left(\frac{\cos\varphi}{\sin\varphi}\frac{\partial\phi'}{\partial\varphi}\right)+\frac{1}{\cos^2\varphi}\frac{\partial^2\phi'}{\partial\lambda^2}\right]_{n-\frac{1}{2}}, \tag{11}$$

$$\hat{\Omega}_n\frac{\partial}{\partial\lambda}\left(\frac{\partial\phi'}{\partial p}\right)_n-\left(\frac{\partial\hat{\Omega}}{\partial p}\right)_n\frac{\partial\phi'_n}{\partial\lambda}+\sigma_n\omega_n=-\left(\frac{RH}{C_p p}\right)_n-(\alpha_R)\left(\frac{\partial\phi'}{\partial p}\right)_n+(K_T)_n$$

$$\times\frac{1}{a^2}\left[\frac{\partial^2}{\partial\varphi^2}-\tan\varphi\frac{\partial}{\partial\varphi}+\frac{1}{\cos^2\varphi}\frac{\partial^2}{\partial\lambda^2}\right]\left(\frac{\partial\phi'}{\partial\varphi}\right)_n, \tag{12}$$

$$\hat{\Omega}_{n+\frac{1}{2}}\frac{\partial}{\partial\lambda}\left\{\frac{1}{2\Omega_0\sin\varphi}\frac{1}{a^2}\left[\frac{\sin^2\varphi}{\cos\varphi}\frac{\partial}{\partial\varphi}\left(\frac{\cos\varphi}{\sin^2\varphi}\frac{\partial\phi'}{\partial\varphi}\right)+\frac{1}{\cos^2\varphi}\frac{\partial^2\phi'}{\partial\lambda^2}\right]\right\}_{n+\frac{1}{2}}+\frac{1}{a}q_{n+\frac{1}{2}}$$

$$\times\frac{1}{2\Omega_0\sin\varphi}\frac{1}{a\cos\varphi}\frac{\partial\phi_{n-\frac{1}{2}}}{\partial\lambda}=f\left(\frac{\partial\omega}{\partial p}\right)_{n+\frac{1}{2}}-(R_f)_{n+\frac{1}{2}}\frac{1}{2\Omega_0\sin\varphi}\frac{1}{a^2}\left[\frac{\sin\varphi}{\cos\varphi}\right.$$

$$\left.\times\frac{\partial}{\partial\varphi}\left(\frac{\cos\varphi}{\sin\varphi}\frac{\partial\phi'}{\partial\varphi}\right)+\frac{1}{\cos^2\varphi}\frac{\partial^2\phi'}{\partial\lambda^2}\right]_{n+\frac{1}{2}}, \tag{13}$$

where H is the diabatic heating per unit time per unit mass,

R the gas constant, C_p the specific heat at constant pressure, α_R the Newton cooling coefficient, K_T the horizontal eddy thermal diffusivity, R_f the Rayleigh friction coefficient of perturbation, n indicates levels in the model, and q is expressed as

$$q=\left[2(\Omega_0+\hat{\Omega})-\frac{\partial^2\hat{\Omega}}{\partial\varphi^2}+3\tan\varphi\frac{\partial\hat{\Omega}}{\partial\varphi}\right]\cos\varphi$$

where Ω_0 is the angle velocity of the earth rotation and $\hat{\Omega}$ is the angle velocity of the basic flow, i.e.,

$$\hat{\Omega}=\frac{\bar{U}}{a\cos\varphi}$$

2. *Parameters*

(1) Static stability parameters σ_n: The static stability used in this model is calculated from the mean temperature and density at 45°N in July of the U.S. Standard Atmosphere and is assumed not to change with latitudes.

(2) The vertical profile of the basic zonal mean flow: The profile computed by Murgatrogid is used. In order to eliminate small-scale features, a smooth operator is used. The result is shown in Fig. 1

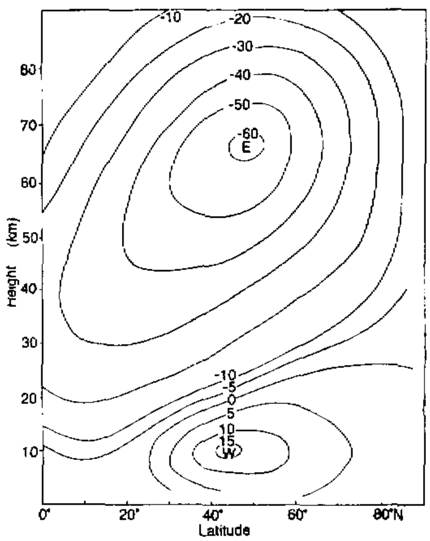


Fig. 1. The vertical distribution of zonal mean flow (m/s) in summer.

The coefficients of Rayleigh friction R_f and Newtonian cooling α_R are the same as these in Huang and Gambo (1982).

(4) The horizontal eddy thermal diffusivity coefficient K_T: Because the zonal mean winds in the tropics and subtropics in summer are very weak, the thermal diffusivity coefficient in the thermal balance equation should be smaller than that in winter and will be taken as a constant of 0.1×10^6 m^2 s^{-1}.

Thus, if the forcing source is given, the vertical distribution of stationary planetary waves and the stationary perturbation patterns at isobaric surface induced by the forcing source can be computed by using the model equations.

IV. DIVERGENCE OF E-P FLUX AND FORCED MERIDIONAL CIRCULATION INDUCED BY TOPOGRAPHY AND HEAT SOURCE

In order to investigate the physical roles of the diabatic heating over the southeast of the Tibetan Plateau in the formation and maintenance of the summer monsoon, we have made two numerical calculations: one is the divergence of E–P flux and the induced meridional circulation of the quasi-stationary planetary waves due to the forcing by topography during the northern summer; the other is the divergence of E–P flux and the induced meridional circulation of the quasi-stationary planetary waves due to the forcing by topography and heat source during the northern summer.

1. *A Case Considering Only Topographical Forcing*

The Northern Hemispheric topography arranged by Berkofsky and Bertni (1955) is used as the actual topography. It is expanded into Fourier series and is substituted into Eqs. (11)—(13). In this way we can obtain the distributions of amplitude and phase of quasi-stationary planetary waves due to the forcing by actual topography. Then, we can calculate the divergence of E–P flux and the forced meridional circulation of the guasi-stationary planetary waves induced by the forcing of topography.

Fig. 2 (a), (b), (c) shows the divergence of E–P flux of the quasi-stationary planetary waves due to the forcing by topography for wavenumbers 1, 2 and for composing wave components of $K=1-3$, resepectively. From Fig. 2 we can see that the E–P flux of the quasi-stationary planetary wave due to the forcing by topography is convergent in the subtropic troposphere, but it is divergent in the upper troposphere in middle latitudes. From Eq. (1) we can find that the forcing effect of topography is favorable for the easterly strengthening in the subtropics. However, in the case of middle latitudes, it is favorable for the westerly strengthening in the upper troposphere, but for the easterly strengthening or the westerly weakening in the lower troposphere.

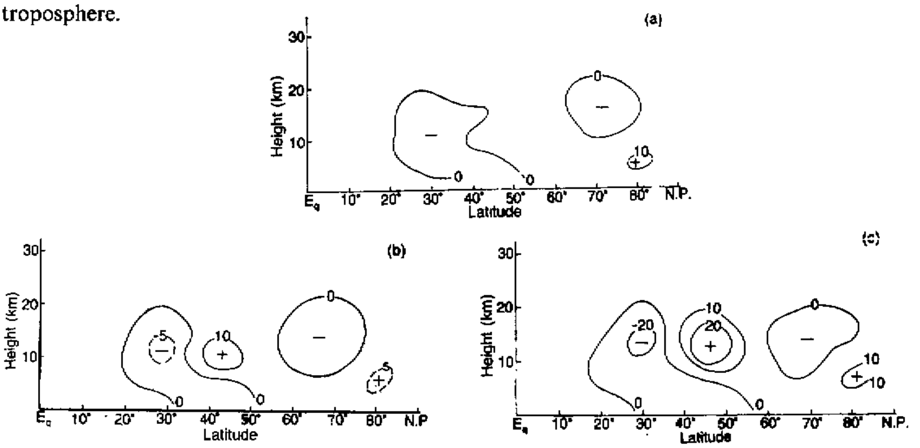


Fig. 2. The divergence (units, m/s) of E–P flux of the quasi-stationary planetary waves due to the forcing by topography (a) for wavenumber 1, (b) for wavenumber 2, (c) for composing wave components of $k=1-3$.

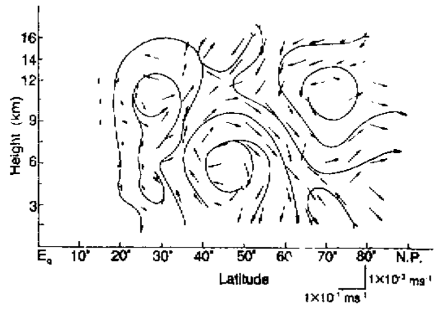


Fig. 3. The Induced meridional circulation due to the topographical forcing, for composing wave components of $k=1-3$.

In order to see clearly the influence of the forcing effect of topography on the flow, we have computed the meridional circulation induced by topography. Fig. 3 that shows the meridional circulation induced by topography, we can see that there is a direct circulation to the north of 30°N. Moreover, there is a very weak meridional circulation in the middle and upper troposphere in the region of 20°—30°N, and in the middle and lower troposphere in the region of 20°—25°N, while a downward flow occurs in the region near 20°N. In order to investigate the effect of this induced meridional circulation on the flow, we have also computed the value of $f\bar{v}^*$ in this circulation. Fig. 4 shows the Coriolis force $f\bar{v}^*$ in the mean meridional circulation induced by topography in summer. From Fig. 3 and Fig. 4 we can see that the topographical forcing effect makes the induced meridional circulation flow southward and the value of $f\bar{v}^*$ become negative in the upper subtropical troposphere. The flow accelerates westward due to the Coriolis force, and the easterly enhances, while the westerly weakens. However, in the lower subtropical troposphere, the induced meridional circulation flows northward, $f\bar{v}^*$ has a positive value, the flow accelerates eastward, and the westerly enhances, while the easterly weakens. In middle-latitude areas it is in contrary to that in the subtropics.

So far we have discussed theoretically the role of plateau topography in the formation of the Asian monsoon. Now we shall discuss the physical roles of topography and heat source for the formation of summer monsoon in Asia, using model Eqs. (11)—(13).

2. *A Case Considering Both Topographical and Thermal Forcing*

As mentioned in the introduction, the southeast part of the Tibetan Plateau is a region in which the diabatic heating rate is the largest in the atmosphere over the Northern Hemisphere. In order to compute the roles of heat source over the Tibetan Plateau in the formation and maintenance of the mean monsoon circulation over Asia, we first assume that the vertical distribution of the heat source over the Northern Hemisphere is given by

$$H_0(\lambda,\varphi,p)=\hat{H}_0(\lambda,\varphi)\exp\left(-\left(\frac{p-\tilde{p}}{d}\right)^2\right), \tag{14}$$

where $d=300$ hPa. Observations show that the height of maximum heating rate decreases with increasing latitude. At middle latitudes, the maximum heating rate is located at 500 hPa, while over the Tibetan Plateau it is at 400 hPa. Thus, we take $\tilde{p}=500$ hPa for $\varphi\geqslant 40°$N, and

$\bar{p}=400$ hPa for $\varphi<40°$N, i.e., at middle and high latitudes, the maximum heating rate is located at 500 hPa, while in the subtropics it is at 400 hPa.

We expand the actual heat sources computed by Ashe (1979) into Fourier series and substitute them into the model equations to compute the quasi-stationary planetary waves and guasi-stationary disturbance pattern at constant height level, induced by both topography and heat source.

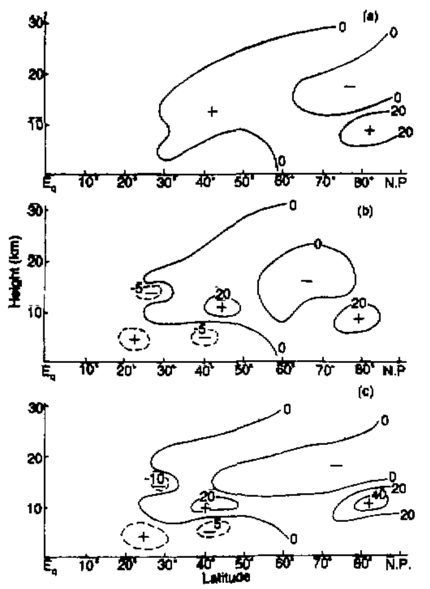


Fig. 4. fv^* in the meridional circulation induced by topography (units: 10^{-6} m s^{-2}) (a) for wavenumber 1; (b) for wavenumber 2; (c) for composing wave components of $k=1-3$.

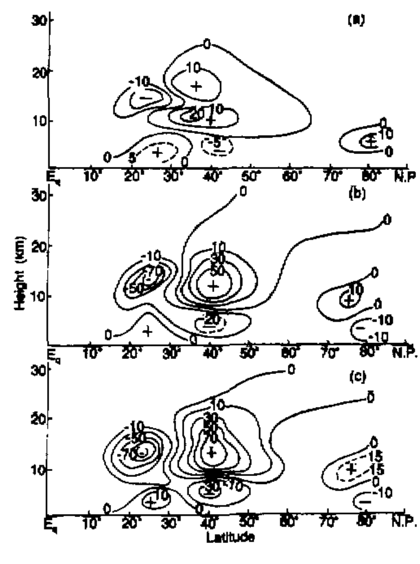


Fig. 5. As in Fig. 2, except that these are for the forcing by both topography and heat source.

Fig. 5 a, b, c is the divergence of E-P flux of the quasi-stationary planetary waves due to the forcing by topography and heat source for wavenumber 1, wavenumber 2, and composing wave component of $K=1-3$. Comparing Fig. 5 with Fig. 2, we can see that in the subtropics, the divergence of E-P flux taking into account both topography and heat source is obviously different from that taking into account only topography. When the forcing effect of heat source is considered, the E-P flux is convergent in the upper subtropical troposphere and much larger than that due to topography. However, it is divergent in the lower subtropical troposphere. Moreover, the convergence of E-P flux in the upper troposphere and the divergence in the lower troposphere in middle-latitude areas are much stronger than those due to the topographical effect. That is to say, in summer, the forcing effect of heat source over the Tibetan Plateau is largely favorable for the easterly acceleration in the upper subtopical troposphere, while it is so for the westerly strengthening or the easterly weakening in the lower subtropical troposphere, In order to see clearly the forcing effect of heat source on the zonal mean flow, we have computed the meridional circulation induced by topography and heat source. Fig. 6 shows the meridional circulation induced by the forcing of both topography and heat source for composing wave component of $K=1$—3. Comparing Fig. 6 with Fig. 3, we can see that, when considering both topography and heat source, the monsoon circulation in the subtropics extends southward significantly, the upward flow is over 30°N, and the downward flow is over 10°N. Moreover, the meridional circulation is stronger than that induced only by topography. This induced meridional circulation is obviously divided into two circulations: one is located in the upper and middle troposphere; the other in the middle and lower troposphere. The northerly flow in the upper troposphere and the southerly flow in the lower troposphere induced by topography plus heat source are stronger than those induced only by topography. In order to further study the effect of the induced meridional circulation due to topography and heat source on the flow, we have also computed the Coriolis force $f\bar{v}^*$ in the meridional circulation. Fig. 7 a,b,c show the values of $f\bar{v}^*$ due to topography and heat source for wavenumbers 1, 2, and the composing wave component of $K=1$—3. Comparing Fig. 7 with Fig. 4, we find a great difference between them. $f\bar{v}^*$ is negative in the upper subtropical troposphere and it is positive in the lower subtropical troposphere, and its values are larger than those in Fig. 4. Moreover, $f\bar{v}^*$ is positive in the upper troposphere and it is negative in the lower troposphere at middle latitudes. Its values are also larger than those in Fig. 4. Thus, because of the Coriolis torque, the induced meridional circulation due to the thermal effect of plateau will create a strong easterly acceleration in the upper subtropical troposphere and the westerly acceleration in the lower subtropical troposphere. This is favorable for the formation and maintenance of the southwest monsoon.

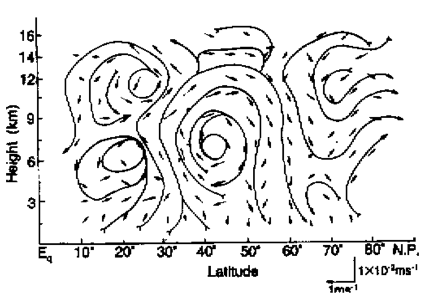
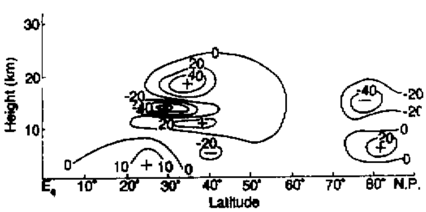
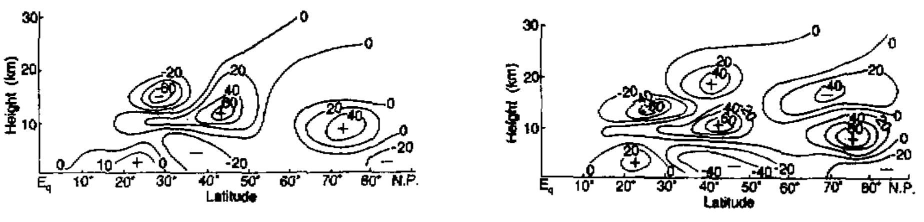


Fig. 6. As in Fig. 3, except that it is for the forcing by both topography and heat source.



(a)



(b) (c)

Fig. 7. As in Fig. 4, except that it is for the forcing by both topography and heat source.

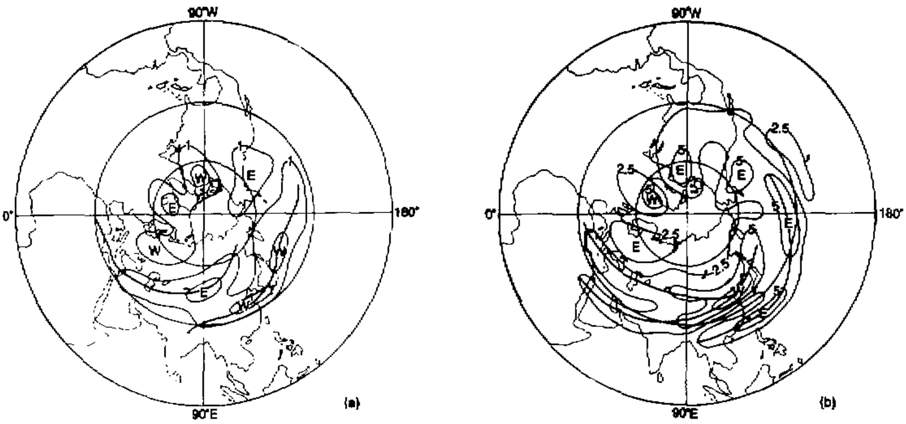


Fig. 8. The perturbation geostropic wind speed (units: m s^{-1}) induced by topography. (a) at the 3 km height level; (b) at the 12 km height level.

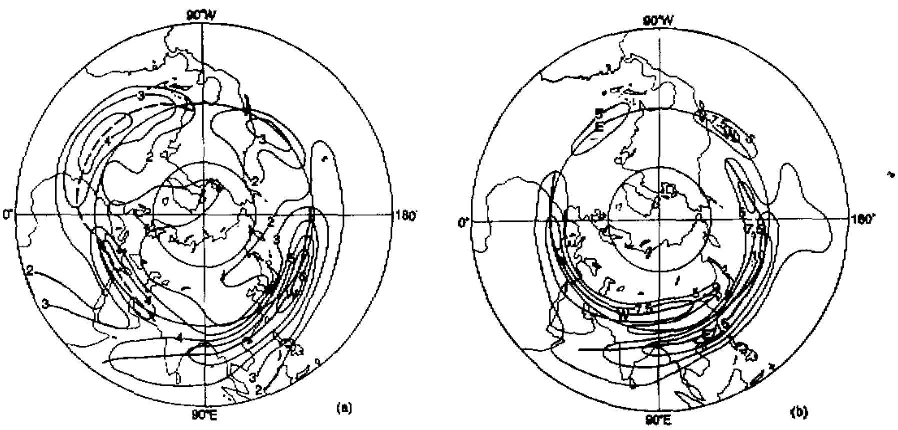


Fig. 9. As in Fig. 8, except that this is for the forcing by both topography and heat sources.

V. THE PERTURBATION FLOW INDUCED BY TOPOGRAPY AND HEAD SOURCES

In the above we have discussed theoretically the forcing effect of topography and heat sources on the zonal mean flow. Now we shall compute the stationary disturbance patterns and perturbation flow at constant height level, using the numerical model. Fig. 8 (a) shows the perturbation geostrophic wind speed induced only by topography at the 3 km height level; Fig. 8(b) shows the pertubation geostrophic wind speed induced only by topography at the 12 km height level. In Fig. 8 we can see that the southwest perturbation flow, due to only topographical forcing, is weak in the lower troposphere over South Asia. Moreover, the east perturbation flow is also weak in the upper troposphere over South Asia. Thus, if only the topographical effect is considered, the monsoon circulation should be very weak over South Asia.

In order to see clearly the effect of heat source over the plateau on the summer monsoon over Asia, we have also computed the quasi-stationary disturbance patterns and perturbation flow at constant-height levels using the numerical model. Fig. 9 (a) shows the perturbation geostropic wind induced by both topography and heat sources at the 3 km height level; Fig. 9(b) shows the perturbation geostrophic wind induced by both topography and heat sources at the 12 km height level. Comparing Fig. 9 with Fig. 8 we find a large difference between them. When both topography and heat sources are considered, a strong southwest wind belt passes through the India Peninsula, the Bay of Bengal, and the south of China to the south of Japan. This strong southwest wind belt plays an important role in the maintenance of the summer monsoon over Asia. Moreover, in Fig. 9 we can also see that a strong easterly belt passes through the West Pacific Ocean, the south of China, the Indian Peninsula to West Africa in the upper troposphere. The maximum strength of this flow may reach 15 m s^{-1} and even larger. This easterly also plays an important role for the maintenance of the monsoon circulation in the upper troposphere over Asia. Besides, in Fig. 9 we can find a strong westerly belt from North Africa along the Mediterranean Sea to the Caspian Sea.

The computed results show that the heat sources over the Plateau play an important role for the formation and maintenance of the summer monsoon over Asia.

VI. CONCLUSIONS AND DISCUSSIONS

The physical effects of topography and heat sources on the formation and maintenance of the summer monsoon over Asia are discussed by computing the divergence of the E-P flux of the planetary waves, induced meridional circulation, and perturbation geostrophic wind speed due to the forcing by topography and heat source. The computed results show that in summer the divergence of E-P flux of the quasi-stationary waves and the induced meridional circulation due to the forcing by the plateau topography are very weak. Moreover, in the distribution of perturbation geostrophic wind speed induced by only topography, the strongest southwest flow is located in the lower troposphere over East Asia at middle latitudes, but it is very weak; in the upper troposphere the strongest easterly flow is located over the south of China and the West Pacific Ocean and it is also very weak. Thus one can see that the forcing effect of the Plateau's topography alone can not form the monsoon circulation. The computed results also show that, when the forcing effect of diabatic heating over the plateau are considered, both the divergence of E-P flux of the quasi-stationary planetary waves and the induced meridional circulation due to the forcing by topography and heat source are very strong in the upper and lower subtropical troposphere. These make the easterly accelerate largely in the upper subtropical troposphere and the westerly accelerate in the lower subtropical troposphere.

The results computed by the model also show that, in the distribution of perturbation geostrophic wind speed induced by both topography and heat source, a strong southwest wind belt passes through the Indian Peninsula, the Bay of Bengal and the south of China to the West Pacific Ocean, while a strong easterly passes the West Pacific Ocean, the south of China, and

India to West Africa. Thus, the heat source over the Tibetan Plateau may play an important role in the formation and maintenance of the monsoon circulation over South Asia and East Asia.

From the computed results it can be seen that, for the forcing effects by topography or by topography plus heat source, the order of magnitude of Coriolis torque $f\bar{v}^*$ in the induced meridional circulation in the same as that of the divergence of E-P flux, i.e., $\nabla \cdot F/a\cos\varphi$, in the upper troposphere. However, the Coriolis torque $f\bar{v}^*$ in the meridional circulation is much larger than the divergence of E–P flux in the lower troposphere. Thus, in the discussion of wave-flow interaction in the lower troposphere the effect of the induced meridional circulation should be considered in addition to the divergence of E–P flux.

Because the model used in this paper is a hemispheric one, we can only discuss the effects of topography and heat sources over the Northern Hemisphere on the formation and maintenance of the Asia monsoon. The effect of the cross-equator flow from the Southern Hemisphere on the formation of the Asian monsoon can not be addressed here.

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