ANALYSES OF THE DYNAMIC EFFECTS ON WINTER CIRCULATION OF THE TWO MAIN MOUNTAINS IN THE NORTHERN HEMISPHERE

II. VERTICAL PROPAGATION OF PLANETARY WAVES

Zou Xiaolei (邹晓蕾), *Wu Guoxiong* (吴国雄) and *Ye Duzheng* (叶笃正)

Institute of Atmospheric Physics, Academia Sinica, Beijing 100080

Received May 11, 1992

ABSTRACT

A linear, hemispheric and stationary spectral model with multilayers in the vertical was employed to simulate the vertical propagation of waves triggered by mountains. Results show that, in cooperation with the East Asia zonal mean flow, Tibetan Plateau can excite a strong wavenumber 1 perturbation in the stratosphere with its ridge and trough located over the Pacific and Atlantic Oceans respectively. On the other hand, the stratospheric wavenumber 1 perturbation caused by the mechanical forcing of the Rocky Mountains in cooperation with the North America zonal mean flow is very weak. Calculations from observational data of the vertical profile of critical wavenumber for vertically propagating waves imply that the tropospheric wavenumber 1 perturbation can hardly penetrate the North America tropopause upwards, whereas it can freely propagate through the East Asia tropopause into the stratosphere. Two-dimensional E-P cross-sections obtained from both observational data and simulated results also demonstrate that waves excited by the Rocky Mountains are refracted towards low latitudes in the troposphere during their upward propagation; whereas, in addition to the above mentioned equatorward leaning branch, the wavenumber 1 and 2 planetary waves excited by the Tibetan Plateau possess another branch which is refracted to high latitudes during upward propagation and penetrates the tropopause into the stratosphere. It is therefore concluded that the difference in the horizontal and vertical wave propagations in the two hemispheres is a result of the different dynamical forcing induced by the two main mountains in the Northern Hemisphere.

Key words: planetary waves, vertical wave propagation, E-P flux cross-section, orographic forcing

I. INTRODUCTION

Charney and Drazin (1961) have studied the relation between vertical wave propagation and zonal mean basic flow in the troposphere. Their results indicate that waves can propagate upward when the zonal basic flow is westerly and less than a critical value, or will be trapped or refracted when the basic zonal flow is easterly or strong westerly. According to the typical winter data in the Northern Hemisphere, the mid-latitude waves with wavenumbers 1 and 2 (probably also 3) are able to propagate upward into the stratosphere. Waves with wavenumber greater than 3 will decay in vertical direction. Based on this theory, Matsuno (1971) concluded that the sudden warming of the stratosphere in spring results from the decrease of the zonal flow due to the upward propagation of planetary waves. Dunkerton (1981), Palmer (1981) and Shiotani et al. (1985) confirmed this conclusion by conducting a two-dimensional E-P flux analysis of the real data. They found that the vertical propagation of waves is related to the vertical structure of the zonal basic flow.

Another feature of the winter stratosphere in the Northern Hemisphere is that there exists a huge anticyclone over the North Pacific, i.e. the Aleutian high. Matsuno (1970) found that the Aleutian high is maintained by the vertical propagation of stationary Rossby wave with wavenumber 1 in the troposphere. The simulation of a general circulation experiment with and without topography by Kasahara et al. (1973) showed that the vertical propagation of the planetary waves with wavenumbers 1 and 2 is obvious only when topography is present. In the comparison simulations of Manabe and Terpstra (1974), they also found that the anticyclone over Aleutian in the stratosphere will develop only when topography was included. Held (1983) pointed out that the tropospheric response of the atmosphere to topographic forcing has an equivalent barotropic character. The amplitudes of the planetary waves would continue to increase in the stratosphere 10 to 30 degrees in longitude downstream of the topographic source only when this topography has a scale comparable to that of the Tibetan Plateau (Held ,1983; Fig.6). Their work indicates that in the stratosphere the existence of the Aleutian high at high latitudes is related to the topographic forcing at the atmospheric boundary.

Since the vertical propagation of waves is related to the structure of basic flow, and the latter is strongly affected by the topographic forcing, one may speculate that the behaviour of vertical wave propagation downstream of the Tibetan Plateau and the Rocky Mountains will be different due to the obvious difference of the topographically induced basic zonal flows. If this is true, the fact that the high pressure center in high latitudes in the stratosphere only occurs downstream of the Tibetan Plateau is another evidence of the different dynamic effects of the two main topographies in the Northern Hemisphere. Based on this consideration, we will in Section II use a steady, quasi–geostrophic and baroclinic model to investigate the propagation behaviour of the waves forced by the Eastern and Western Hemispheric topographies in various zonal flows. Section III will present the analysis of different wave behaviour in East Asia and North America by comparing the different vertical structures of critical wavenumber in the two areas. In Section IV we will further study the wave behaviour in the real and model atmosphere by means of two–dimensional E–P flux. Some conclusions will be presented in Section V.

II. NUMERICAL SIMULATIONS OF THE TOPOGRAPHIC FORCING

1. *Amplitude and Phase Distributions of Perturbations in the Model Atmosphere*

The quasi–geostrophic, linear, steady, and multilayer semi–spectral model developed by Huang and Gambo (1982) was employed for the following study. This model uses spherical coordinates and divides the atmosphere into N vertical levels ($N=34$) from the top $P=P_t$ ($P_t=1.140\times10^{-3}$ hPa in winter) to the ground $P=P_s$(P_s is the earth's surface pressure). The effect of topography is introduced by the vertical velocity term at the lower boundary, $\mathbf{V}_s\cdot\nabla P_G$, where $\mathbf{V}_s$ and P_G are the ground velocity and the surface pressure respectively.

In order to study the stationary response of the model atmosphere to the topographies in the Eastern and Western Hemispheres with different basic flow regimes, we designed the following two experiments.

Experiment A:

i) Eastern Hemispheric topographic forcing;

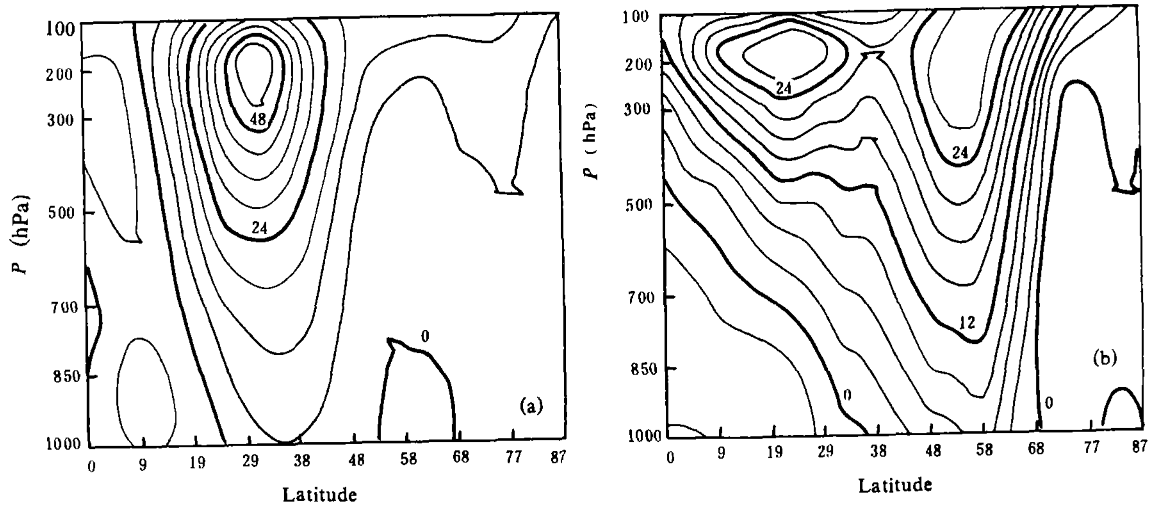


Fig. 1. Vertical distribution of zonal flow in January 1983 in (a) Area A (80°E—140°W), contour interval in 6 m / s, and (b) Area B (120°W—5°W), contour interval in 3 m / s.

ii) below 100 hPa the basic zonal flow is defined as the mean flow in January 1983 averaged over East Asia and West Pacific: 80°E—140°W (the same definition as given in part I of this study, see this issue, referred as ZYW hereafter), as shown in Fig.1a;

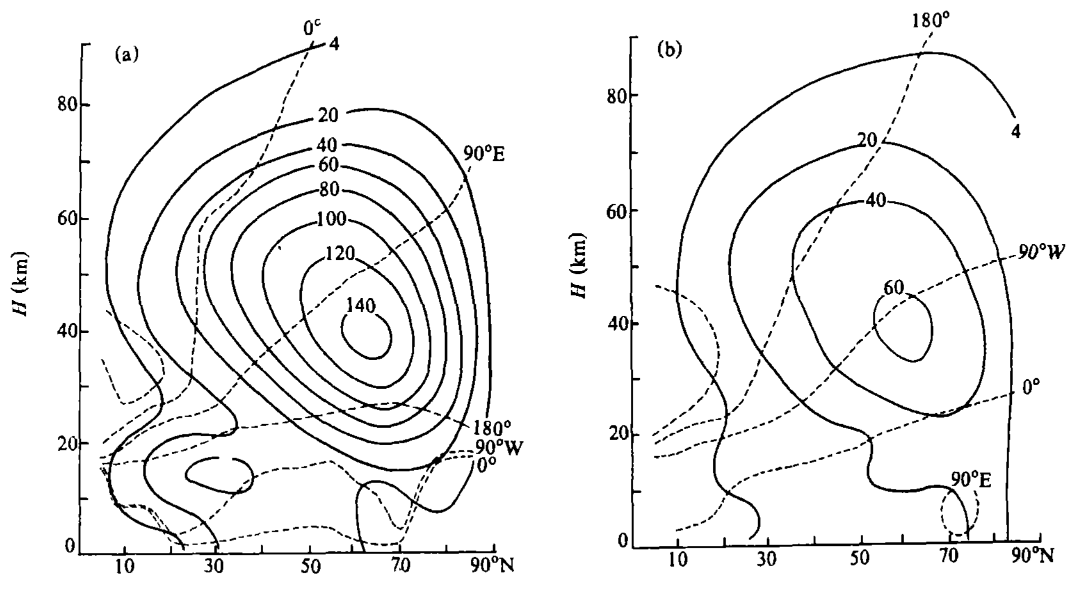
Experiment B:

i) Western Hemispheric topographic forcing;

ii) below 100 hPa the basic zonal flow is defined as the mean flow in January 1983 averaged over North America (120°W—5°W), shown in Fig.1b.

In these experiments the basic zonal flows above 100 hPa are obtained from the analyses of Holton (1976) (above 60 km) and Matsuno (1970) (from 100 hPa to 60 km).

Fig.2 shows the amplitude and ridge distributions of stationary waves (wavenumbers 1 to 3) in Experiments A and B. Wave 1 has the following main features (Figs. 2a,b):



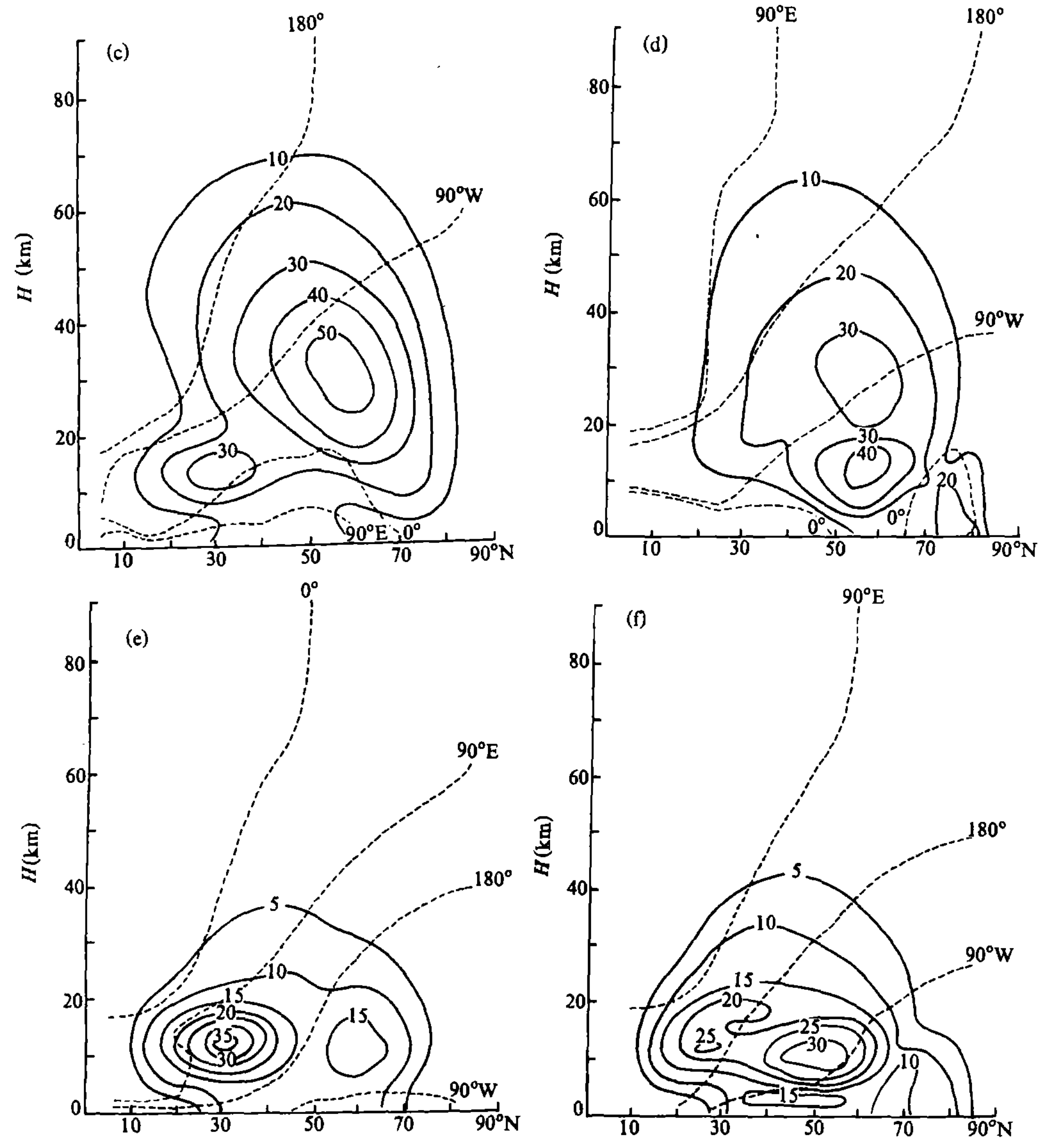


Fig. 2. Vertical distributions of amplitude (solid line, units in gpm) and the position of ridge (dashed line) of planetary waves forced by topography using a linear and stationary wave model. (a) wavenumber 1 forced by the Eastern Hemispheric topography with the zonal flow in Area A; (b) wavenumber 1 forced by the Western Hemispheric topography with the mean zonal flow in Aera B; (c) wavenumber 2 forced by the Eastern Hemispheric topography with the mean zonal flow in Aera A; (d) wavenumber 2 forced by the Western Hemispheric topography with the mean zonal flow in Area B; (e) wavenumber 3 forced by the Eastern Hemispheric topography with the mean zonal flow in Area A; (f) wavenumber 3 forced by the Western Hemispheric topography with the mean zonal flow in Area B.

(1) In the upper layers, wave 1 forced by the topography in the Eastern Hemisphere is out of phase with that forced by the topography in the Western Hemisphere. In the former case,

there is an intensive high pressure cell over the Pacific Ocean and an intensive low pressure cell over the Atlantic Ocean, whereas in the latter case the opposite holds. The phases are tilted westward with increasing height and eastward with increasing latitude. However in the troposphere and north of 50°N, the wave forced by the topography in the Eastern Hemisphere tilts westward with increasing latitude.

(2) The amplitude of wave 1 forced by the topography in the Eastern Hemisphere is much larger than that forced by the topography in the Western Hemisphere. The latter is only 43% of the former.

(3) The amplitudes of wave 1 forced either in the Eastern Hemisphere or in the Western Hemisphere all reach their maximum values near (37km, 65°N). In addition, the amplitude of the forced wave 1 in the Eastern Hemisphere has a secondary maximum value near (15km, 30°N). Webster (1972) pointed out that the time mean general circulation in low latitudes results mainly from the combined effects of diabatic heating and topographic forcing in low and mid–latitudes. Huang and Gambo (1982) found that the topographic forcing in high latitudes has an important impact on stationary waves in the upper troposphere at low latitudes. The results obtained here show that the topographic forcing of Tibetan Plateau is important for the development of wave 1 in the upper troposphere at low latitudes.

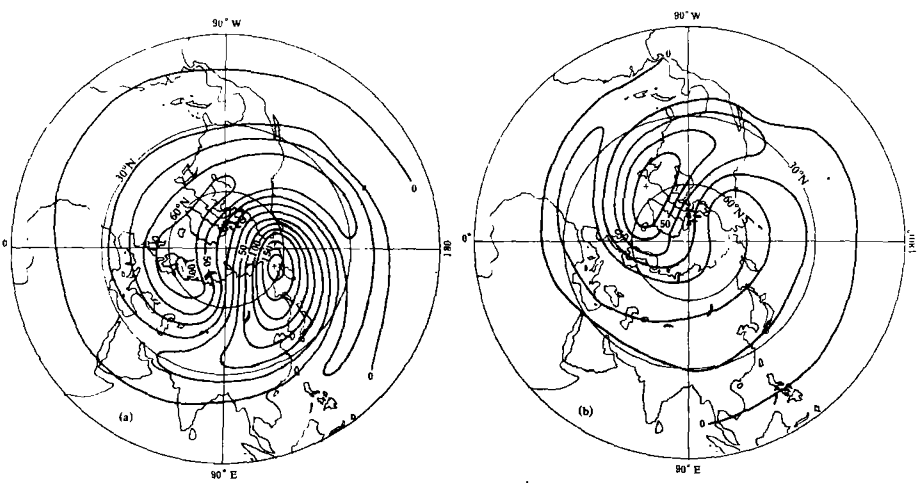
The amplitude and phase distributions of wave 2 in the two experiments are displayed in Figs.2c and 2d, respectively. The variation of phase with height and latitude is similar to wave 1. Compared to wave 1, their amplitudes in the stratosphere are much weaker and their heights of the maximum center are lower. The largest amplitude of wave 2 forced by the topography in the Eastern Hemisphere at (29km, 60°N) is as large as one third of wave 1 (Fig.2a). There is a secondary maximum near (13km, 30°N). The amplitude of wave 2 forced in the Western Hemisphere, on the other hand, is about half that of wave 1. Its secondary maximum is located near (29km, 55°N). Furthermore, over the Atlantic Ocean in the stratosphere, the intensity of the ridge of wave 2 (30 gpm) is much weaker than that of the trough of wave 1 (140 gpm). This result indicates that most of the energy of wave 2 forced by the Tibetan Plateau is able to propagate into the stratosphere, while most of the energy of wave 2 forced by the Rocky Mountains is trapped near the tropopause.

Figs.2e and 2f show the distributions of the amplitude and phase of wave 3 forced by topography. The maximum and the secondary maximum values are less than those of wave 2 and are concentrated near the tropopause. There is no maximum value in the stratosphere. Therefore, wave 3 forced by the topography either in the Eastern Hemisphere or in the Western Hemisphere can not propagate into the stratosphere, at least for the present cases.

In order to study the effect of basic zonal flow on the vertical wave propagation, we replaced the single year, area averaged zonal flow below 100 hPa in Experiments A and B by the global mean climate data given in Matsuno (1970), and re–ran the corresponding experiments denoted as Experiments A′ and B′ respectively. Results show that the difference in the distribution of zonal flow will mainly affect the intensity of the amplitude (figures not shown). For example, the amplitude of wave 1 in Experiment A′ is about 71% of that in Experiment A. On the other hand, the amplitude of wave 1 in Experiment B′ is about one and half times as large as that in Experiment B, while the amplitude of wave 2 in Experiment B′ is doubled. Besides, the intensities of the secondary maximum are also changed. Therefore, the variation of the basic zonal flow does affect the propagation of stationary waves, but it does not affect the qualitative conclusions about the propagation of stationary waves forced by topography.

2. *Distributions of Potential Height Perturbation on* 10 *hPa*

Fig.3a shows the distribution of geopotential height on 10 hPa in Experiment A. There is a strong positive potential height perturbation over the North Pacific, i.e., a strong Aleutian high appears in the stratosphere. In Experiment B (Fig.3b), there is also a positive potential height perturbation over the North Atlantic. However, it is very weak and its absolute value is less than that of the negative perturbation over the North Atlantic in Fig.3a, and Fig.3c is a combination of Figs.3a and 3b. In Fig.3c, there is only a very strong positive anomaly over the Aleutian area, but no positive anomaly over Iceland. To the east of the North American continent, a very weak positive anomaly zone is observed. In addition, over the Northern Europe there exists a negative anomaly. It is thus apparent that the dynamical effect of Tibetan Plateau is very important for the formation of the Aleutian high.



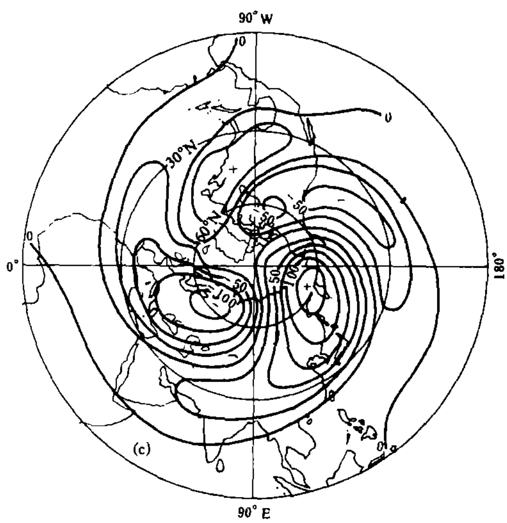


Fig.3. The perturbation of geopotential height on 10 hPa, forced by topography in a linear and stationary wave model, units in gpm. (a) By the Eastern Hemispheric topography with the mean zonal flow in Area A; (b) by the Western Hemispheric topography with the mean zonal flow in Area B; and (c) the sum of (a) and (b).

In Experiment A′, the center of the strong positive geopotential perturbation on 10 hPa observed in A over the North Pacific forced by the topography in the Eastern Hemisphere has moved westward. In Experiment B′, a strong positive center of geopotential perturbation appears over the western side of the North Atlantic (figures not shown), which is not consistent with observation . These results show that the model with the local zonal mean flow can simulate the behaviour of stationary waves better than the global zonal mean flow. In other words, the vertical structure of the local zonal flow is very important for the vertical wave propagation and for the formation of the Aleutian high at high latitudes in the stratosphere. In the next section, we will analyze the vertical wave propagation using real data.

III. CRITICAL WAVENUMBER OF VERTICAL WAVE PROPAGATION

Charney and Drazin (1961) and Held (1983), using the quasi–geostrophic vorticity equation and the thermodynamical equation, obtained an equation of amplitude of stream function $\tilde{\Psi}(z)$ for stationary perturbations, namely

$$\frac{f_0^2}{\rho_0}\frac{\partial}{\partial t}\left(\frac{\rho}{N^2}\frac{\partial\tilde{\Psi}}{\partial z}\right)=\tilde{\Psi}\left(K^2-\frac{\partial[q]/\partial y}{[u]}\right), \tag{1}$$

where $K^2=k^2+l^2$. Let

$$\tilde{\xi}=\tilde{\Psi}\exp(-z/2H). \tag{2}$$

The equation governing the behaviour of the vertical propagation of the stationary waves becomes

$$\frac{\partial^2\tilde{\xi}}{\partial z^2}=\tilde{\xi}\frac{N^2}{f_0}\left(K^2+\gamma^2-\frac{\partial[q]/\partial y}{[u]}\right), \tag{3}$$

where $\gamma=f_0/(2HN)$. N is the Brunt–Vaisala frequency. Obviously, the condition that Eq.(3) has a wave solution is that the horizontal wavenumber must be less than a critical value, namely

$$K^2\langle K_c^2=\frac{(\partial[q]/\partial y)}{[u]}-\gamma^2. \tag{4}$$

where K_c is the critical wavenumber for vertical wave propagation. Eq.(4) shows that waves with wavenumber larger than K_c will decay in the vertical direction and only the planetary waves with wavenumber less than K_c will be able to propagate upward. We will study the kinematic behaviour of wave propagation by calculating K_c. Hereafter, we will use the ECMWF data to calculate the critical wavenumber K_c downstream of the Tibetan Plateau and the Rocky Mountains. We take $H=10$ km, and

$$N^2 \doteq \begin{cases} 1.0\times10^{-4}\text{s}^{-2}, & \text{if} \quad z\leqslant 10\text{km};\\ 2.5\times10^{-4}\text{s}^{-2}, & \text{if} \quad z>10\text{km}. \end{cases} \tag{5}$$

Dividing the Asian Pacific area (80°E—140°W, Area A) and the American Atlantic area (120°W—5°W, Area B) into three latitudinal zones, i.e., high, middle and low latitude zones, the zonal mean westerly and the critical wavenumber K_c for each latitude zone are shown in Fig.4. At low latitudes and downstream of the Tibetan Plateau, waves with wavenumber $K<K_1=4.4$ can propagate vertically on both sides of the tropopause, waves with wavenumber

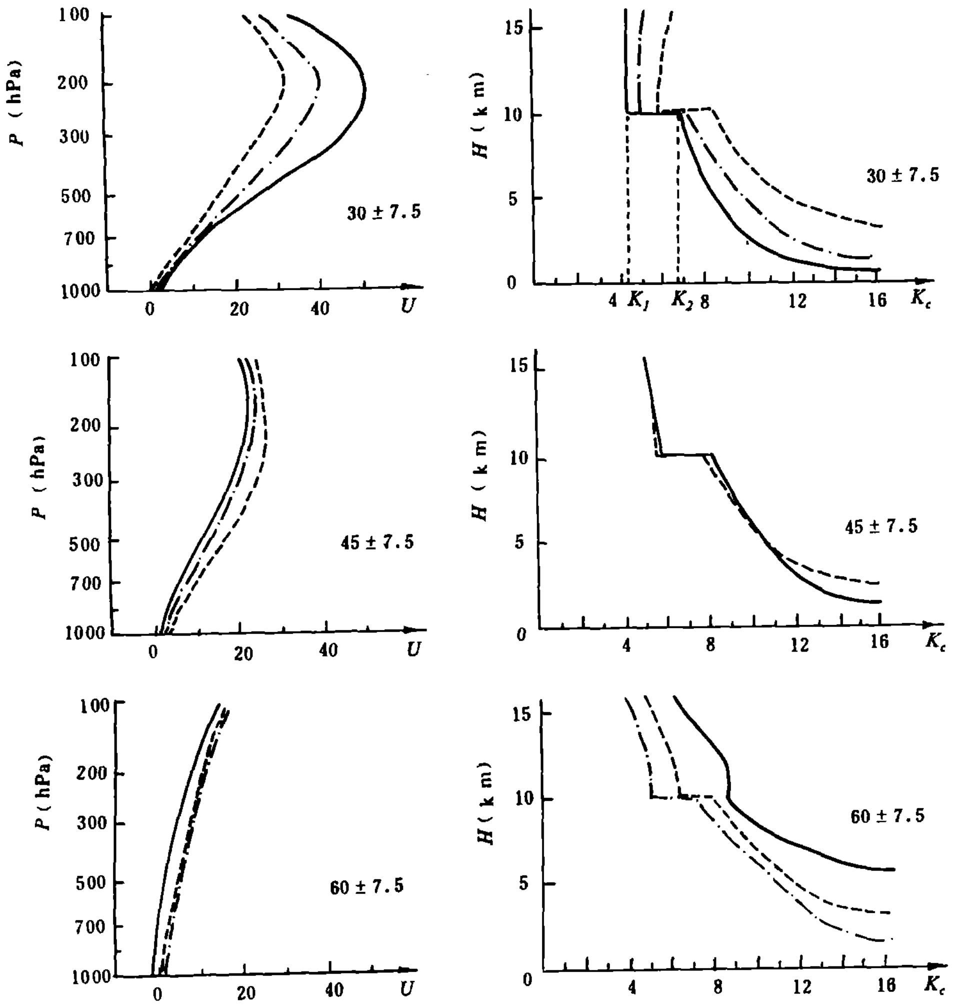


Fig. 4. The mean zonal flow profile (left) and the corresponding critical wavenumber K_c (right) at different latitudinal zones of (30 ± 7.5)°N (upper), (45 ± 7.5)°N (middle) and (60 ± 7.5)°N (lower) in different areas. Solid curve denotes Area A (80°E—140°W), dashed curve corresponds to Area B, and dot–dashed curve is zonally averaged one.

$K_1 < K < K_2 = 6.8$ have a turning point at the tropopause, while waves with wavenumber $K > K_2$ are trapped within the troposphere. Here K_1 and K_2 are the minimum and maximum values of the critical wavenumber of the discontinuous zone at the tropopause, respectively. The same arguments hold for the case of the Rocky Mountains, except that here $K_1 = 6.0$ and $K_2 = 8.4$. Therefore, at low latitudes, waves which are able to propagate vertically in Area A are less numerous than those in Area B on the same height. At middle latitudes, the distributions of critical wavenumber in the two areas are similar. Results at high latitudes are opposite to those obtained at low latitudes. At high latitudes near 60°N, $K_1(=(k^2+l^2)^{1/2})=8.3$ in Area A. By calculating the north–south wavenumber (l) from the width of westerly zone (about 30°), the zonal wavenumber can be estimated to be about 3. This means that downstream of the Tibetan Plateau waves with zonal wavenumbers 1 and 2 will be able to propagate upward from troposphere into stratosphere at high latitudes. In Area B, on the other hand, $K_l = 6.1$ and $k_l = 0.6$. That is to say, downstream of the Rocky Mountains waves can not propagate upward into the stratosphere in high latitudes. Moreover, in Area A, the critical wavenumber increases with latitudes, whereas in Area B it decreases with latitude. Fig.4 also shows that the intensity of zonal flow and horizontal shear will affect the behaviour of the vertical wave propagation (see also Eq.(4)). Near 45°N, there is a strong horizontal shear in both Areas A and B, which will cancel the β–effect and make the difference in vorticity gradient $(\partial[q]/\partial y)$ between the two areas small. Thus, the profiles of critical wavenumber in the two areas are similar. Near 60°N, the zonal flow is small in both areas. However, there is a strong horizontal shear in North America and a weak shear in East Asia, which results in an obvious difference in the distribution of critical wavenumbers between these areas. The reason that planetary wave can propagate from the troposphere into stratosphere in high latitudes of East Asia seems to be related to the fact that the Tibetan Plateau is located at low latitudes, which makes the zonal flow in high latitudes small and the horizontal shear weak.

In our study of the horizontal wave propagation (ZYW), we pointed out that waves in the North American continent propagate consistently towards low latitudes, whereas the wave chains in East Asia display a bifurcation: waves with high wavenumber propagate towards low latitudes and planetary waves with low wavenumber propagate towards high latitudes. We can thus conclude that the planetary waves which propagate northward in East Asia will propagate into stratosphere in high latitudes as well. Thus the high latitude region in East Asia is an important window for the planetary wave propagation from troposphere into stratosphere. This conclusion is consistent with the results of our numerical simulation in the previous section: the strong anticyclone on 10 hPa in the stratosphere appears only over East Asia at high latitudes.

IV. E–P FLUX

1. 2–*D E–P Flux and Its Physical Meaning*

Andrews and McIntyre (1976) defined the E–P flux as

$$\vec{E} = \left(-[u^* v^*],\ f_0 \Theta_p^{-1} [v^* \theta^*] \right). \tag{6}$$

Edmon, Hoskins and McIntyre (1980) also pointed out that the two–dimensional E–P flux is parallel to the group velocity. This property can be widely applied to the study of wave propagation and wave evolution.

Here we use the January mean real data in 1983 to compute the two–dimensional E–P flux averaged over Area A downstream of the Tibetan Plateau and Area B downstream of the Rocky Mountains (Fig.5). In Area A, the E–P flux displays a bifurcation in the upper layer with one branch propagating equatorward, which is weaker than the one propagating poleward. In Area B, however, all the E–P flux propagates equatorward. This result agrees with our previous conclusion in Part I about the horizontal wave propagation and the distribution of critical wavenumbers in North America and East Asia.

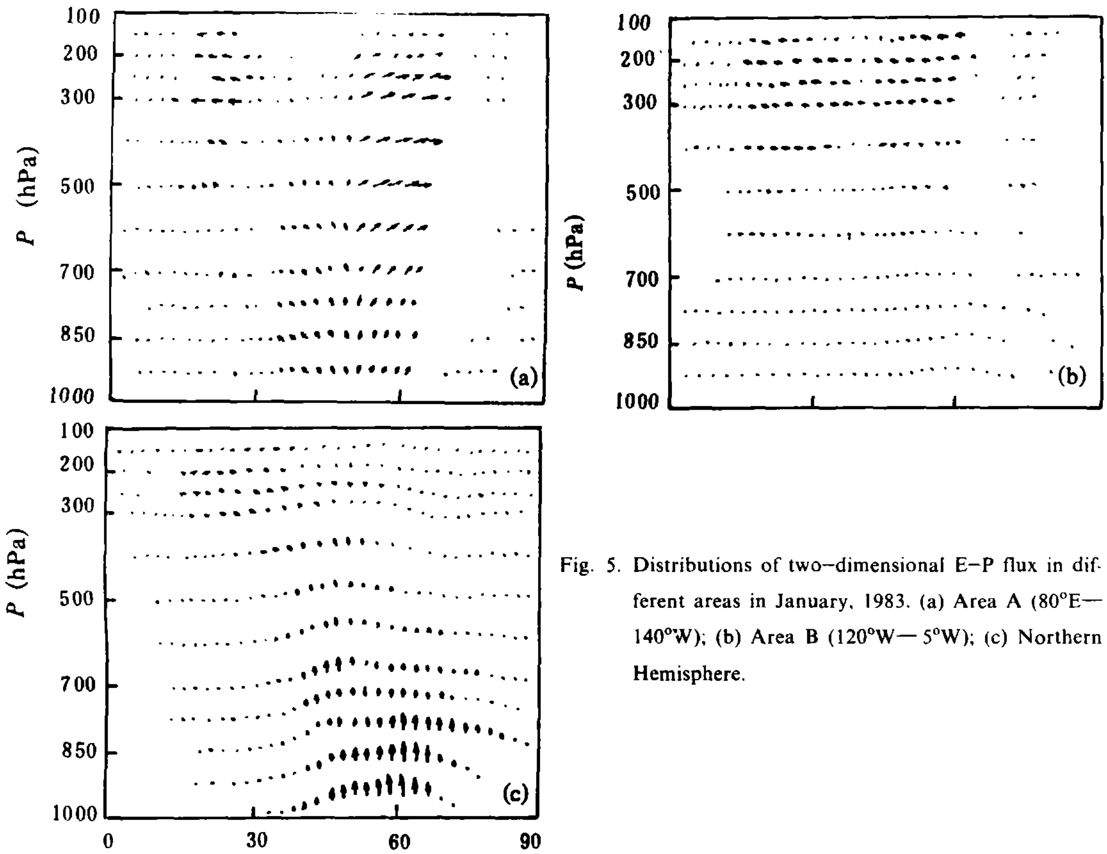


Fig. 5. Distributions of two-dimensional E-P flux in different areas in January, 1983. (a) Area A (80°E—140°W); (b) Area B (120°W—5°W); (c) Northern Hemisphere.

2. *2-D E-P Flux in Baroclinic Atmosphere Forced by Various Topographies*

In the analysis of the E-P flux in different wavenumber domain, Wu et al. (1988) pointed out that the synoptic E-P flux in the troposphere turns to the subtropics when propagating upward, showing an "anvil" configuration, while the planetary scale flux bifurcates into high and low latitudes when propagating upwards, displaying a "cauliflower" configuration. This means that only the planetary scale fluxes are able to propagate upward and poleward simultaneously. In order to further understand the relationship between the planetary waves and the dynamical effect of topography, we use here a two-dimensional E-P flux diagnostic to further analyse the results of topographic simulation in Section II. Figs.6a—d illustrate the E-P fluxes of wave 1 and wave 3 forced by the topography in the Eastern and Western Hemispheres, respectively in correspondence to Figs.2a, b and 2e, f. All the waves with wavenumber equal to and larger than 3 in the Eastern and Western Hemispheres will turn to low latitudes when propagating upward (Fig.6c—d), and will not affect the stratospheric circulation in high latitudes. Wave 1 (Figs.6a—b) and wave 2 (figures not shown) have different features under the different topographic forcings. Under the forcing in the Western Hemisphere wave 2 propagates also toward equator while wave 1 has a weak upward propagation. The E-P flux of wave 1 forced in the Eastern Hemisphere has a bifurcation in the upper layers of the troposphere near 50°N, with a stronger branch propagating upward and poleward into the stratosphere. This result agrees with the amplitude distribution of wave 1 shown in Fig.2a. Wave 2 also has an obvious

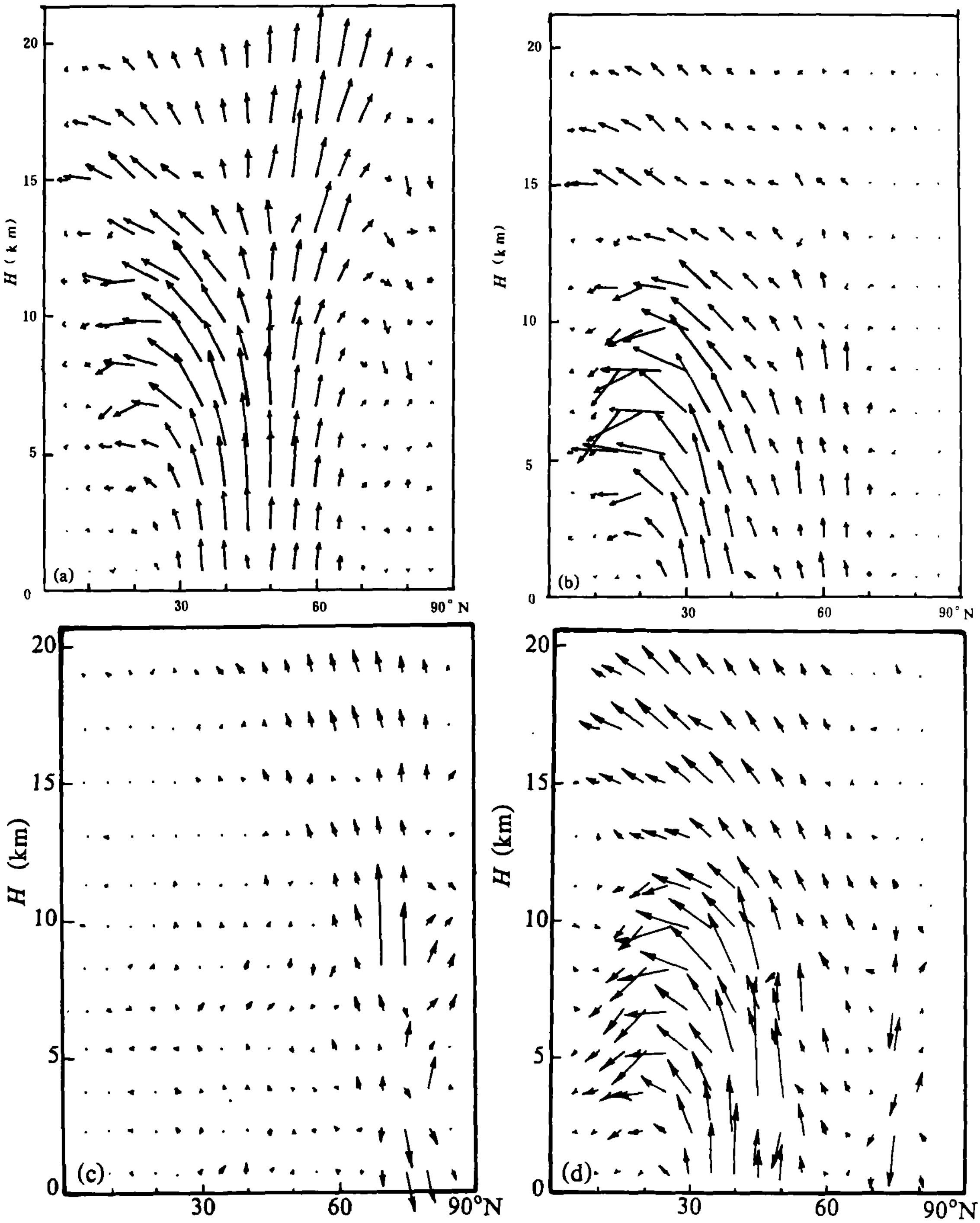


Fig. 6. The two-dimensional E–P flux excited by topographic forcing in a linear and stationary wave model.

(a) Wavenumber 1, forced by the Eastern Hemispheric topography with the mean zonal flow in Area A;

(b) wavenumber 1, forced by the Western Hemispheric topography with the mean zonal flow in Area B;

(c) wavenumber 3, forced by the Eastern Hemispheric topography with the mean zonal flow in Area A;

(d) wavenumber 3, forced by the Western Hemispheric topography with the mean zonal flow in Area B.

bifurcation phenomenon (figures not shown). But its northern branch is rather weak, the bifurcation being limited near the tropopause, and above the tropopause the E–P flux turns to the equator again. Since they are forced by the Eastern Hemispheric topography, it is obvious that the Tibetan Plateau plays a dominate role in the vertical propagation of planetary waves into the stratosphere.

V. CONCLUSIONS

The mechanical effect of the Rocky Mountains and the related zonal flow structure makes the waves turn towards low latitudes during their upward propagation and makes them hard to penetrate into the stratosphere at high latitudes. The forcing of the Tibetan Plateau and the related zonal flow structure allows the planetary waves (especially wave 1) able to propagate into the stratosphere at high latitudes. The effect of the Tibetan Plateau on the formation of Aleutian high in high latitudes of the stratosphere originates from the following two aspects: (i) the zonal flow structure in East Asia makes waves dispersive in their northward propagation and waves 1 and 2 can propagate toward high latitudes; (ii) the resulting axis of jet is located at subtropics, which in turn results in a rather weak zonal flow and a larger critical wavenumber in high latitudes. Therefore, the high latitudes in East Asia region become a window for the planetary wave 1 (sometimes also wave 2) to propagate into the stratosphere.

Under the Eastern Hemispheric topography and the East Asian zonal wind, one branch of the two–dimensional E–P flux of wave 1 turns to subtropics near 300 hPa in the upper layers of the troposphere. Another branch can propagate poleward into the stratosphere. The equatorward branch of wave 2 displays similar behaviour to that of wave 1, but its poleward branch is weaker. The E–P flux of wave 3 has only one branch propagating equatorwards and upwards to a rather low altitude. On the other hand, under the Western Hemispheric topographic forcing and the North American zonal flow, both the E–P fluxes of waves 1 and 2 are weak, while the E–P flux of wave 3 is large and propagates mainly towards lower latitudes.

Combining our previous analysis on horizontal wave propagation, we found that the long waves ($k \geqslant 3$) in North America and East Asia have similar propagation behaviour: they have a tendency of turning to low latitudes when propagating upward. The main difference in vertical wave propagation occurs in the planetary wave domain, especially wave 1 and wave 2. Due to the different local zonal flows forced by topography, the planetary waves in North America turn to the equator while they propagate upward, whereas in East Asia there is an additional branch which propagates into the stratosphere. Therefore, the formation of the Aleutian high in the winter stratosphere is closely related to the influence of the mechanical forcing of the Tibetan Plateau upon atmospheric motions.

We would like to thank Dr.Huang Ronghui and Mr.Yan Bangliang for their assistance in running the numerical experiments.

REFERENCES

Andrews, D. G. and McIntyre, M. E. (1976), Planetary waves in horizontal and vertical shear: the generalized Eliassen–Palm relation and the mean zonal acceleration, *J. Atmos. Sci.*, **33**:2031—2048.

Charney, J. G. and Drazin, P. G. (1961), Propagation of the planetary scale disturbance from the lower into the upper atmosphere. *J. Geophys. Res.*, **66:** 83—109.

Dunkerton, T., Hsu, C.-P. F. and McIntyre, M.E. (1981), Some Eulerian and Lagrangian diagnostics for a model stratospheric warming, *J. Atmos. Sci.*, **38:**819—843.

Edmon, H. J., Hoskins, B. J. and McIntyre, M. E. (1980), Eliassen-Palm cross section for the troposphere, *J. Atmos. Sci.*, **37:** 2600—2616.

Held, I. M. (1983), Stationary and quasi-stationary eddies in the extratropical troposphere: Theory, *Large-scale Dynamical Processes in the Atmosphere*, B.J. Hoskins and R. Pearce, Eds, Academic Press, pp.127—168.

Holton, J. R. (1979), *Introduction to Dynamical Meteorology*, Academic Press, 142 pp.

Huang Ronghui and Gambo, K. (1982), The response of a hemispheric multi-level model atmosphere to forcing by topography and stationary heat sources: (I) Forcing by topography, *J. Meteor. Soc. Japan*, **60:** 78—92.

Kasahara, A., Sasamori, T. and Washington, W.N. (1973), Simulation experiments with a 12-layer stratospheric global circulation-model: I. Dynamical effect of the earth's orography and thermal influence of continent, *J. Atmos. Sci.*, **30:**1229—1251.

Manabe, S. and Terpstra, T. B. (1974), The effects of mountains on the general circulation of atmosphere as identified by numerical experiments, *J. Atmos. Sci.*, **31:** 3—42.

Matsuno, T. (1970), Vertical propagation of stationary planetary waves in the winter Northern Hemisphere, *J. Atmos. Sci.*, **27:** 871—883.

Matsuno, T. (1971), A dynamical model of the stratospheric sudden warming, *J. Atmos. Sci.*, **28:** 1479—1494.

Palmer, T. N. (1981), Diagnostic study of a wavenumber 2 stratospheric sudden warming in a transformed Eulerian-mean formalism, *J. Atmos. Sci.*, **38:** 844—855.

Shiotani, M. and Hirota, I. (1985), Planetary wave-mean flow interaction in the stratosphere: a comparison between the Northern and Southern Hemispheres, *Quart. J. Roy. Meteor, Soc.*, **111:** 309—334.

Webster, P. J. (1972), Response of the tropical atmosphere to local, steady forcing, *Mon. Wea. Rev.*, **100:** 518—541.

Wu Guoxiong, Chen Biao and Wu Zhengxian (1988), E-P cross sections and residual circulations in different wavenumber range in dry and moist models, *Chinese Journal of Atmospheric Sciences*, **Special Issue**, pp.94—160.

Zou Xiaolei, Yeh, T. C. and Wu Guoxiong (1992), Analyses of the dynamic effects on winter circulation of the two main mountains in the Northern Hemisphere: (I) Horizontal propagation of planetary waves, *Acta Meteor. Sinica*, **6:** 395—407.