Suppression of Stationary Planetary Waves by Internal Gravity Waves in the Mesosphere

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ABSTRACT

The suppression of stationary planetary waves by internal gravity waves in the mesosphere is treated using a quasi-geostrophic model on a midlatitude beta-plane. The drag forces due to internal gravity waves are parameterized based on the wave breaking assumption recently proposed by Lindzen. In the present model the vertical propagation of internal gravity waves is affected not only by mean zonal wind distribution but also by eastward and northward velocity perturbations associated with stationary planetary waves, viz., the total local velocity.

Numerical results show that the drag force due to breaking internal gravity waves acts like a Rayleigh friction, and the amplitudes of stationary planetary waves in the mesosphere are much reduced by this effect. Equivalent Rayleigh friction coefficients are also presented.

1. Introduction

Stationary planetary waves can propagate vertically in a somewhat restricted mean zonal wind, in the form of internal waves (Charney and Drazin, 1961). If vertical propagation is possible, wave amplitudes may in theory increase with height approximately as $e^{z/2H}$ in the absence of dissipation. The situation in the real atmosphere, however, is complicated. Not only vertical variation but also lateral variation of the mean zonal wind seriously affects the propagation of planetary waves (e.g. Dickinson, 1968; Matsuno, 1970; Schoeberl *et al.*, 1979). In realistic mean zonal winds in the winter hemisphere, part of the stationary planetary waves are ducted towards the equator where they are absorbed at critical lines (e.g. Dickinson, 1968; Matsuno, 1970; Dunkerton *et al.*, 1981; Huang and Gambo, 1982; Kanzawa, 1982; Karoly and Hoskins, 1982; Lin, 1982). This ducting is an important cause of inhibiting stationary planetary wave activity in the mesosphere. Infrared radiative cooling also affects the propagation of planetary waves by dissipating wave energy (Dickinson, 1969).

Schoeberl *et al.* (1979) studied the structure of stationary planetary waves in the middle atmosphere using realistic atmospheric models. Their results indicate that wave amplitudes grow almost continuously with height up to the lower thermosphere. In case of zonal wavenumber 1, the amplitude of geopotential height exceeds 1000 m above the upper mesosphere.

Recently, planetary wave amplitudes in the mesosphere have become available based on satellite data (Barnett, 1980). Because of limited vertical resolution of the satellite data in the mesosphere, the derived amplitude in the mesosphere may not be definitive. However, the deduced planetary wave structure is completely different from the above-mentioned numerical result. The maximum amplitude of geopotential height occurs around stratopause and its value is about 600 m, and the amplitude is reduced in the mesosphere.

One possible mechanism for more reduction of the amplitude in the mesosphere may be infrared radiative cooling provided that it is stronger than that believed earlier (Wehrbein and Leovy, 1982). However, as will be shown in the present study, infrared radiative cooling alone seems to be inadequate to reduce the amplitude in the mesosphere.

Another mechanism that has been used to reduce the wave amplitude is Rayleigh friction (e.g. Holton and Wehrbein, 1980b). In their model, a large Rayleigh friction with a damping rate of the order of 10^{-6} s^{-1} is used. As a result of this strong damping effect the maximum wave amplitude occurs in the lower mesosphere, and the amplitude in the upper mesosphere is greatly reduced.

It has been speculated that the origin of the required large Rayleigh friction in the mesosphere may be the breakdown of internal gravity waves due to exponential growth of their amplitudes with height (Houghton, 1978; Holton and Wehrbein, 1980a). Recently Lindzen (1981) and Matsuno (1982) revealed that the physical origin of the Rayleigh friction can be attributed to the momentum transport by upward propa-

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gating internal gravity waves. By introducing the effect of internal gravity waves, zonally-symmetric models of the middle atmosphere were able to simulate the weak mean zonal winds around the mesopause (Holton, 1982, 1983; Matsuno, 1982; Miyahara, 1984).

It has also been suggested that the momentum transport by upward propagating internal gravity waves interacting with stationary planetary waves will act to attenuate them (Lindzen, 1984; Schoeberl and Strobel, 1984). Schoeberl and Strobel (1984) have pointed out that the gravity wave drag acts like a Rayleigh friction.

In the present paper we shall show that the momentum transport by upward propagating internal gravity waves interacting with stationary planetary waves acts like a Rayleigh friction on the stationary waves and reduces the planetary wave amplitudes in the mesosphere. The parameterization of drag forces due to internal gravity waves used in the present paper is based fundamentally on the wave breaking assumption proposed by Lindzen (1981). In what follows we shall treat effects of drag forces of breaking internal gravity waves on stationary planetary waves by using a very simplified quasi-geostrophic model on a midlatitude beta-plane.

2. Breaking and drag forces of internal gravity waves

We consider an internal gravity wave that is propagating in a mean zonal wind and a stationary planetary wave. We assume that the horizontal and vertical wavelengths of an internal gravity wave under consideration are much smaller than representative length scales of the mean zonal wind and the planetary wave. We also assume that a ray path of an internal gravity wave is almost vertical. This assumption is valid for small-scale internal gravity waves under consideration because the vertical group velocity is relatively large except for a critical level, so that the wave cannot propagate much horizontally while it is propagating from the ground to the mesosphere. Under these assumptions, we treat the propagation of an internal gravity wave based on WKB method. We express the vertical velocity perturbation of an internal gravity wave of zonal and meridional wavenumbers k and l respectively and frequency ω in the approximate form

$$w'' = W''(z)\exp(z/2H)\exp[i(kx + ly + \omega t)], \quad (1)$$

where H is the scale height, $z = H\ln(P_0/p)$, and $W''(z)$ satisfies the following equation

$$\frac{d^2W''}{dz^2} + \frac{N^2}{\hat{\omega}^2}(k^2 + l^2)W'' = 0. \quad (2)$$

Here we assume that the local vertical wavenumber $m = N/\hat{\omega}(k^2 + l^2)^{1/2}$ is much larger than k, l and 1 H^{-1}. In the above equation, the Doppler shifted wave frequency of an internal gravity wave is given by

$$\hat{\omega} = \omega + k(\bar{u} + u') + lv', \quad (3)$$

where $\bar{u}$ is a basic mean zonal wind and u' and v' are the eastward and northward velocity perturbations respectively associated with the stationary planetary wave. Hereafter, perturbations associated with planetary waves are denoted by single prime and those associated with internal gravity waves are denoted by double prime.

The WKB solution of (2) is given by

$$W''(z) = W_0\left(\frac{\hat{\omega}}{\hat{\omega}_0}\right)^{1/2}\exp\left(i\int_{z_0}^{z} mdz'\right), \quad (4)$$

where W_0 is the amplitude of W'' at $z = z_0$ and $\hat{\omega}_0$ is the Doppler shifted wave frequency at $z = z_0$ (Holton, 1982).

Following Lindzen (1981) we assume that an internal gravity wave begins to break at the height that the total lapse rate exceeds the dry adiabatic one; then the breaking height is given by

$$z_b(x, y) = 3H\ln\left(\frac{|\hat{\omega}|}{\tilde{\omega}}\right), \quad (5)$$

where

$$\tilde{\omega}^{3/2} = \frac{W_0N(k^2 + l^2)^{1/2}}{|\hat{\omega}_0|^{1/2}}, \quad (6)$$

and the drag forces due to the momentum flux divergence by a breaking internal gravity wave becomes

$$\begin{aligned} F_x(x, y, z) &= -\frac{1}{p}\frac{dp\langle u''w''\rangle}{dz} \\ &= -\frac{\epsilon k\hat{\omega}^3}{2N(k^2 + l^2)^{3/2}}\left(\frac{1}{H} - \frac{3}{\hat{\omega}}\frac{d\hat{\omega}}{dz}\right), \\ F_y(x, y, z) &= -\frac{1}{p}\frac{dp\langle v''\omega''\rangle}{dz} \\ &= -\frac{\epsilon l\hat{\omega}^3}{2N(k^2 + l^2)^{3/2}}\left(\frac{1}{H} - \frac{3}{\hat{\omega}}\frac{d\hat{\omega}}{dz}\right), \end{aligned} \quad (7)$$

where angle brackets denote the average over one wavelength, and ϵ is an efficiency factor which represents an intermittence of gravity wave activity (Lindzen, 1981; Holton, 1982, 1983). Equation (7) is appropriate above $z = z_b(x, y)$. Below $z_b(x, y)$, stable gravity waves are still capable of breaking and inducing nonlinear cascade to smaller scales (Lindzen and Forbes, 1983). Thus, following Holton (1982), we assume an exponentially decaying drag force

$$F_{x,y}(x, y, z < z_b) = F_{x,y}(x, y, z_b)\exp[(z - z_b)/H]. \quad (8)$$

As shown in (3), the intrinsic frequency of an internal gravity wave in the present model is Doppler shifted not only by a mean zonal wind but also by eastward and northward velocity perturbations associated with stationary planetary waves, so that the drag forces due to an internal gravity wave are modulated by stationary planetary waves.

3. Numerical model and results

We consider here stationary planetary waves and the influence of breaking gravity waves on them on a midlatitude beta-plane. The distribution of mean zonal winds in the mesosphere itself is greatly controlled by drag forces due to internal gravity waves (Holton, 1982, 1983; Matsuno, 1982; Miyahara, 1984). However, the purpose of the present paper is to discuss effects of internal gravity waves on stationary planetary waves, so that we prescribe a distribution of mean zonal wind and do not consider a change of mean zonal winds due to internal gravity waves. For simplicity of the mathematical treatment, we assume that the basic mean zonal wind varies only with height.

In log-pressure coordinate system, the linearized quasi-geostrophic potential vorticity equation on a midlatitude beta-plane can be written as

$$\bar{u}\frac{\partial}{\partial x}\left[\nabla^2\psi' + \frac{1}{p}\frac{\partial}{\partial z}\left(\frac{pf_0^2}{N^2}\frac{\partial\psi'}{\partial z}\right)\right] + \left[\beta - \frac{1}{p}\frac{d}{dz}\left(\frac{pf_0^2}{N^2}\frac{d\bar{u}}{dz}\right)\right]\frac{\partial\psi'}{\partial x} = -\frac{f_0^2}{p}\frac{\partial}{\partial z}\left(\frac{p\alpha}{N^2}\frac{\partial\psi'}{\partial z}\right) + \frac{\partial\tilde{F}_y}{\partial x} - \frac{\partial\tilde{F}_x}{\partial y}, \quad (9)$$

where α is the coefficient of Newtonian cooling and $\tilde{F}_x$ and $\tilde{F}_y$ are deviations of drag forces from the zonal mean drag forces due to internal gravity waves.

Breaking gravity waves also generate vertical eddy diffusion (Lindzen, 1981). In his model of the circulation of the middle atmosphere, Holton (1983) treated the effects of planetary waves dissipating by a vertical eddy diffusion generated by breaking gravity waves. However, as will be shown later, the wave drag is essential in the present mechanism, so that the effects of the vertical eddy diffusion are neglected in the present model for simplicity.

As suggested by Lindzen (1981), in the winter hemisphere topographically-generated stationary internal gravity waves may dominate. In the present model, we consider only two stationary upward-propagating internal gravity waves with the same horizontal wavelength 57 km, but whose wavenumber vectors are different. One component has a wavenumber vector $\mathbf{k} = (k, l) = (7.85 \times 10^{-5}, 7.85 \times 10^{-5})$ m^{-1}, and the other has $\mathbf{k} = (7.85 \times 10^{-5}, -7.85 \times 10^{-5})$ m^{-1}, thus the former is directed towards northeast, and the latter is southeast. We must also specify the amplitudes of gravity waves and the efficiency factor ϵ. According to recent observations (Balsley and Ecklund, 1983), breaking of gravity waves occurs near 70 ± 9 km in winter hemisphere. In the present paper, the wave amplitude W_0 is chosen to be 2.4×10^{-2} m s^{-1} so that the active region of the breaking gravity waves coincides with the observation. The efficiency factor is chosen as 0.025. This choice gives a smaller but still the same order of magnitude of zonally-averaged drag force used by Holton (1982, 1983).

The parameters of gravity waves used in the present model are not necessarily based on observational results, because only little is yet known about the climatology of gravity waves in the middle atmosphere. For instance, it is possible to choose any other wavelength as long as the WKB approximation remains valid. However, when the wavelength is changed, we also have to change the values of the gravity wave amplitude W_0 and the efficiency factor ϵ to provide appropriate breaking level and drag forces in the mesosphere. By the tuning of the values of W_0 and ϵ, the present results will become insensitive to the choice of the wavelength. We must await observational results elucidating the nature of internal gravity waves that are relevant to this mechanism in the real atmosphere.

The drag forces due to these two breaking internal gravity waves are given by summing up (7) for these two wave components. Thus

$$F_x = -\frac{\epsilon k}{2N(k^2+l^2)^{3/2}}\left[\frac{1}{H}(\hat{\omega}_{+l}^3 + \hat{\omega}_{-l}^3) - 3\hat{\omega}_{+l}^2\frac{d\hat{\omega}_{+l}}{dz} - 3\hat{\omega}_{-l}^2\frac{d\hat{\omega}_{-l}}{dz}\right],$$

$$F_y = -\frac{\epsilon l}{2N(k^2+l^2)^{3/2}}\left[\frac{1}{H}(\hat{\omega}_{+l}^3 - \hat{\omega}_{-l}^3) - 3\hat{\omega}_{+l}^2\frac{d\hat{\omega}_{+l}}{dz} + 3\hat{\omega}_{-l}^2\frac{d\hat{\omega}_{-l}}{dz}\right], \quad (10)$$

where $\hat{\omega}_{+l}$ and $\hat{\omega}_{-l}$ are Doppler shifted wave frequencies of (k, l) and $(k, -l)$ wave components, respectively. The $\tilde{F}_x$ and $\tilde{F}_y$ in (9) are deviations from zonal mean of (10).

Even though an excitation of a stationary planetary wave is monochromatic, the resulting stationary wave may contain many wavenumbers because of a nonlinear dependence of the drag force terms due to internal gravity waves on planetary waves, as shown in (3)–(7) and (10). However, as shown later, departures of F_x and F_y from a sinusoidal form are not large, so that we may assume as a first approximation that a stationary planetary wave has a form

$$\psi' = \Psi'(z)e^{z/2H}e^{ik_px}\sin l_py, \quad (11)$$

where k_p and l_p are wavenumbers of the forced stationary planetary wave. This assumption is not appropriate for large-amplitude planetary waves whose velocity perturbations are comparable to those of the mean zonal wind. In such cases, Doppler effects due to stationary planetary waves are so large that a critical level may exist. Even if Doppler effects are not large enough to form a critical level, infrared cooling will easily damp gravity waves, and their propagation into the mesosphere may be possible only in a confined region, in which case departures of F_x and F_y from a sinusoidal form become large (Schoeberl and Strobel, 1984).

When planetary wave amplitudes are not so large, (9) is written as

$$\frac{f_0^2}{N^2}\left(\bar{u}-\frac{i\alpha}{k_p}\right)\frac{d^2\Psi'}{dz^2}-\frac{i}{k_p}\frac{f_0^2}{N^2}\frac{d\alpha}{dz}\frac{d\Psi'}{dz}$$
$$+\left\{\beta-\frac{f_0^2}{N^2}\frac{1}{p}\frac{d}{dz}\left(p\frac{d\bar{u}}{dz}\right)-\bar{u}(k_p{}^2+l_p{}^2)\right.$$
$$\left.-\frac{1}{4H^2}\frac{f_0^2}{N^2}\left(\bar{u}-\frac{i\alpha}{k_p}\right)-\frac{i}{2Hk_p}\frac{f_0^2}{N^2}\frac{d\alpha}{dz}\right\}\Psi'$$
$$=\left(F'_y-\frac{il_p}{k_p}F'_x\right)e^{-z/2H}, \quad (12)$$

where F'_x and F'_y are $e^{ik_px}\cos l_py$ and $e^{ik_px}\sin l_py$ components of the drag forces F_x and F_y, respectively.

Equation (12) is still highly nonlinear because the drag force terms depend on planetary wave solution ψ' as shown in (3)–(7) and (10). We solve this nonlinear equation using an iterative method. First we solve (12) without drag forces. Once we obtain a planetary wave solution, we calculate the drag forces using (3)–(8) and (10). Then, the calculated drag forces F_x and F_y are Fourier transformed to find F'_x and F'_y, which are the projections on the planetary wave mode under consideration. Equation (12) is solved again using F'_x and F'_y to obtain an approximate solution. Applying this process iteratively, we converge to a solution of (12).

Assumed profiles of mean zonal winds and a Newtonian cooling coefficient used in the present model are shown in Fig. 1. In the present paper, we consider only effects of internal gravity waves on stationary planetary waves, so that we prescribe the vertical profiles of the basic zonal wind which decreases rapidly with height in the mesosphere. The radiative relaxation time varies from about 20 days at 20 km height to about 4.6 days in the mesosphere. The large damping rate through the mesosphere assumed in the present paper is based on the recent calculation of infrared radiation by Wehrbein and Leovy (1982). The large coefficient is extrapolated toward the lower thermosphere in order to reduce the reflection of planetary waves at the upper boundary.

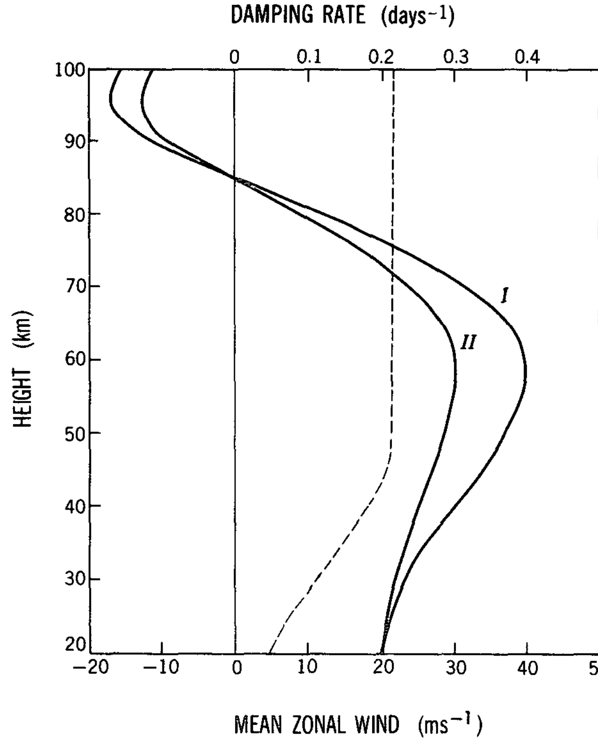

FIG. 1. Vertical profiles of the zonal mean winds (solid lines) and the Newtonian cooling coefficient (dashed line).

Although the assumed basic mean zonal wind $\bar{u}$ varies with height, we assume that the basic state is isothermal $T_0 = 240$ K. This assumption may be permissible when discussing the stationary planetary wave structures because these structures are mainly controlled by zonal mean winds (Schoeberl and Geller, 1977).

As a stationary planetary wave source, we specify the amplitude of the geopotential height at the bottom,

$$\Psi' = 50 \text{ m} \quad \text{at} \quad z = 20 \text{ km}. \quad (13)$$

At the upper boundary we assume that

$$\frac{d\Psi'}{dz} = 0 \quad \text{at} \quad z = 100 \text{ km}. \quad (14)$$

Although this condition is artificial, the effects of this assumption will be small because in the present model the wave amplitude will become small around the top boundary due to the assumed large Newtonian cooling and the easterly winds in the lower thermosphere.

In the present paper, planetary waves with zonal wavenumber 1 are considered, so that k_p is chosen to be 2.22×10^{-7} m^{-1}, and the meridional wavenumber l_p is assumed to be 3.14×10^{-7} m^{-1}.

Figures 2a, b show the vertical variations of amplitude and phase structures of calculated stationary planetary wave geopotential height in the case of wind profile I. When effects of drag forces due to

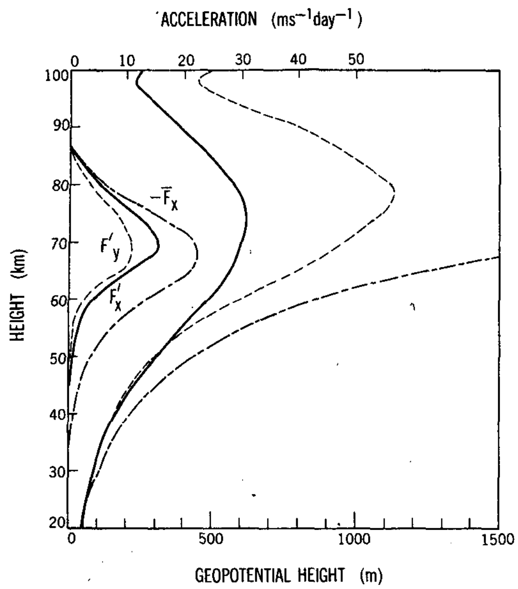

FIG. 2a. Height variations of amplitudes of stationary planetary wave geopotential height with (solid line) and without (dashed line) effects of internal gravity waves, and height variations of mean zonal drag force and $e^{ik_px}\cos l_py$ and $e^{ik_px}\sin l_py$ components of the zonal and the meridional drag forces, respectively. Dash-dotted line show $5 \times 10e^z/2Hm$ for reference.

internal gravity waves are neglected, although the wave energy is dissipated by Newtonian cooling, the amplitude grows gradually with height up to 78 km due to the effect of density stratification; above that height the amplitude decreases with height because of the weak mean zonal wind ($\bar{u} < 15$ m s^{-1}) and the large Newtonian cooling effect. The maximum amplitude is 1130 m at 78 km height. This result indicates that the Newtonian cooling effect alone is not enough to suppress the stationary planetary wave amplitude in the mesosphere, at least not in a laterally uniform flow.

If the effects of the drag forces due to internal gravity waves are taken into account, the amplitude of the stationary planetary wave in the stratosphere is almost unchanged. However, the amplitude in the mesosphere is greatly reduced and the maximum amplitude achieved is 610 m at 74 km height. However, the phase difference between these two cases is less than 20° for all heights, so the effects on phase distribution are much smaller than those on the amplitude.

Figure 2a also shows the vertical profiles of the amplitudes F'_x and F'_y, and the profile of $-\bar{F}_x$ (zonally-averaged zonal drag force). These profiles are qualitatively similar to the VHF echo profile in the winter mesosphere which is considered to represent the breaking gravity wave activity (Balsley and Ecklund, 1983). The magnitude of $\bar{F}_x$ is somewhat smaller than that in Holton's model (1982, 1983) which is required to obtain a realistic mean zonal wind distribution in the mesosphere. Thus, the breaking gravity wave activity in the present model is not overestimated. Nevertheless, its effects on the stationary planetary wave amplitude are still significant.

In order to elucidate the mechanism of suppression, the longitudinal distributions of zonal wind $\bar{u} + u'$, meridional wind v', zonal drag force F_x and meridional drag force F_y at 70 km height at $y = \frac{3}{4}\pi$ are shown in Fig. 3. The forces F_x and F_y vary almost sinusoidally and have obvious negative correlations with the perturbations of zonal wind and the meridional wind associated with the planetary wave. This means that F'_x and F'_y are almost out of phase with u' and v'

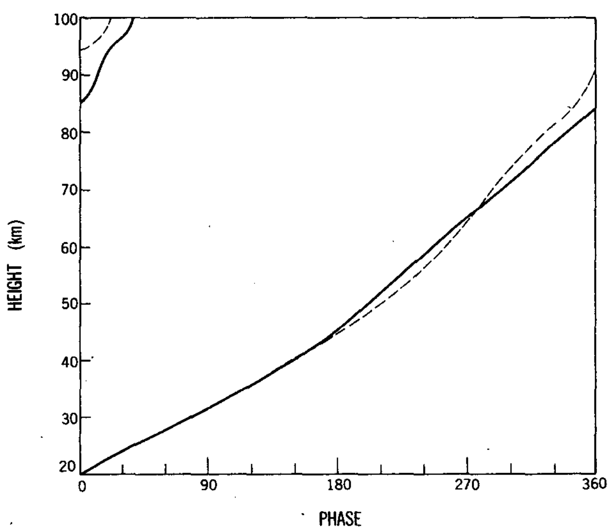

FIG. 2b. Height variations of phase of stationary planetary wave geopotential height with (solid line) and without (dashed line) effects of internal gravity waves.

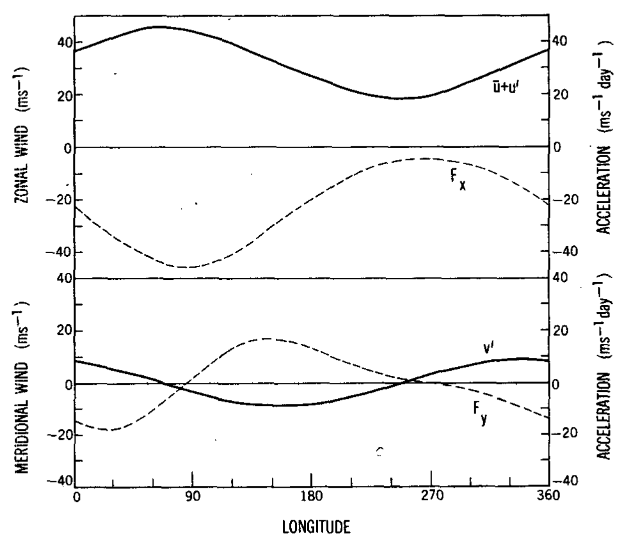

FIG. 3. Longitudinal distributions of the zonal wind, meridional wind, zonal drag force and meridional drag force at 70 km height and $y = \frac{3}{4}\pi$.

respectively. Thus the effects of the drag forces of the internal gravity waves act like Rayleigh friction as pointed out by Schoeberl and Strobel (1984). We can define coefficients of equivalent Rayleigh friction by

$$\alpha_x = -\frac{F'_x}{u'}, \qquad (15a)$$

$$\alpha_y = -\frac{F'_y}{v'}. \qquad (15b)$$

The real part of the coefficient gives the Rayleigh friction coefficient and the imaginary part gives the change of the planetary wave frequency. Table 1 lists equivalent coefficients of Rayleigh friction at several heights. The real parts of α_x and α_y are all positive through the mesosphere and act like Rayleigh friction. These values (the maxima are about 9×10^{-6} s^{-1} at 70 km height) are much larger than the Rayleigh friction coefficient (5×10^{-7} s^{-1}) used by Schoeberl *et al.* (1979), and they are somewhat larger than that required in a zonally-symmetric model to get a reasonable zonal wind profile (see Holton, 1983). The imaginary parts of α_x and α_y are all negative through the mesosphere and act to reduce the planetary wave frequency. The magnitude of the reduction is smaller than the Doppler effect ($k_p\bar{u} = 8.2 \times 10^{-6}$ s^{-1} at 65 km height), but it is not negligible.

The equivalent Rayleigh friction coefficients calculated in the present model depend not only on the activities of internal gravity waves, but also on the magnitude of mean zonal winds and the amplitude of planetary waves, so that even though the activities of internal gravity waves are specified, the coefficients can not be determined uniquely.

Figure 4 shows the numerical result in the case of wind profile II. Other parameters are the same as for the previous case. In this case the Doppler shifted planetary wave frequency and the vertical group velocity is smaller than in the former case, so that the amplitude is much reduced by Newtonian cooling alone. However, the suppression effect of internal gravity waves is still clear. Table 2 lists coefficients of equivalent Rayleigh friction in this case. Maxima of the real parts of the coefficients are about 5×10^{-6} s^{-1} at 70 km height, about half those of the former case. They are, however, still one order larger than those used by Schoeberl *et al.* (1979).

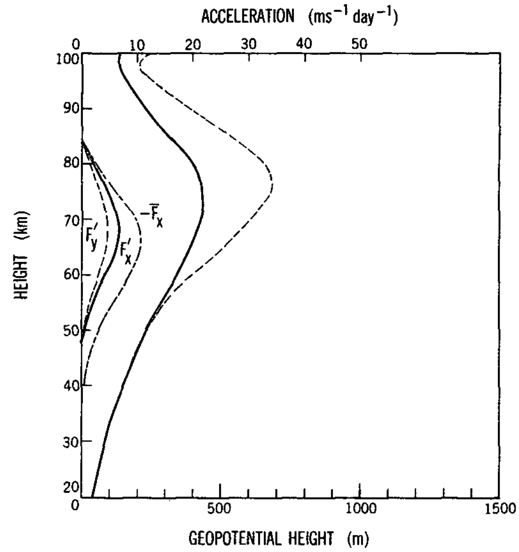

FIG. 4. As in Fig. 2a but for the case of wind profile II.

The vertical eddy diffusion coefficient induced by breaking internal gravity waves is given by

$$D_{\mathrm{eddy}} = -\bar{u}F_x/N^2, \qquad (16)$$

(Lindzen, 1981; Holton, 1982, 1983). In the present case, the maximum value of D_{eddy} is about 5×10^1 m^2 s^{-1}, and the maximum value of the eddy momentum diffusion of planetary waves, $D_{\mathrm{eddy}}|\partial^2 V'/\partial z^2|$, is estimated to be about 8×10^{-6} m s^{-2} ($\sim$0.69 m s^{-1} day^{-1}). This is much smaller than the magnitude of the drag forces. Thus, the primary effect of breaking internal gravity waves on stationary planetary waves is due to the wave drag forces, and the effect of induced eddy diffusion on them is secondary, at least under the present situation.

The EP flux divergence of the suppressed planetary wave in the present model is about -10 m s^{-1} day^{-1} at 70 km height, and it is smaller than the deceleration

TABLE 1. Equivalent Rayleigh friction coefficients for the wind profile I.

Height (km)	α_x ($\times 10^{-6}$ s^{-1}) Re α_x	Im α_x	α_y ($\times 10^{-6}$ s^{-1}) Re α_y	Im α_y
55	1.2	−0.6	1.3	−0.4
60	2.2	−2.3	2.2	−2.2
65	5.8	−4.0	6.2	−4.1
70	9.2	−3.5	9.4	−3.5
75	7.1	−1.9	7.1	−1.9
80	3.7	−0.7	3.6	−0.6

TABLE 2. As in Table 1 but for the wind profile II.

Height (km)	α_x ($\times 10^{-6}$ s^{-1}) Re α_x	Im α_x	α_y ($\times 10^{-6}$ s^{-1}) Re α_y	Im α_y
55	1.6	−1.9	2.0	−1.9
60	3.5	−3.3	3.8	−3.8
65	5.0	−3.3	5.2	−3.4
70	5.1	−2.4	5.2	−2.4
75	4.0	−1.4	4.1	−1.3
80	2.1	−0.5	2.1	−0.4

due to the breaking internal gravity waves. However, it is not negligibly small. As shown by Holton (1983), the dissipating stationary planetary waves may be able to modify the mean zonal wind in the winter mesosphere.

4. Summary and remarks

The suppression of stationary planetary wave amplitudes by internal gravity waves in the mesosphere by use of a quasi-geostrophic model on a midlatitude beta-plane has been discussed. Drag forces due to internal gravity waves are parameterized by using a wave breaking assumption proposed by Lindzen (1981). In the present model, the propagation and breaking of internal gravity waves are affected not only by mean zonal wind but also by velocity perturbations associated with stationary planetary waves.

The calculated results show that the amplitudes of stationary planetary waves in the mesosphere are much reduced by the effects of drag forces due to internal gravity waves. The drag forces are anticorrelated with velocity perturbations of stationary planetary waves. Thus, they act like a Rayleigh friction as pointed out by Schoeberl and Strobel (1984). The equivalent Rayleigh friction coefficients at 70 km height are about 1/day to 1/2 days. These values depend on the mean zonal wind profile and the planetary wave amplitudes (which, needless to say, greatly depend on the gravity wave activities), so that the coefficient can not be independently prescribed, even though the gravity wave activities are given.

In the present paper, we have confined ourselves to the treatment of only moderately large-amplitude planetary waves. If the amplitude of planetary waves is extremely large, the propagation of internal gravity waves into the mesosphere is possible only in a confined region due to the critical level absorption and/or the effects of radiative damping (Schoeberl and Strobel, 1984). In this case, the highly truncated numerical model discussed herein is not applicable, and we cannot easily estimate the equivalent Rayleigh friction coefficient.

The amplitude distribution of the stationary planetary waves is very sensitive to the distribution of mean zonal wind, and it also greatly depends on the model equations and boundary conditions (e.g. Dickinson, 1968; Matsuno, 1970; Schoeberl *et al.*, 1979; Lindzen *et al.*, 1982; Lin, 1982; Kawahira, 1983). However, the present results suggest that the amplitudes of stationary planetary waves in the mesosphere are greatly suppressed by the drag forces that are due to breaking internal gravity waves.

In order to reveal structures of stationary planetary waves and effects of internal gravity waves on them, the present mechanism will be discussed in a more realistic situation using a primitive equation system on the spherical earth, in a separate paper.

To verify the present mechanism, it would be desirable to conduct observational studies that investigate correlations between the internal gravity wave activity and the stationary planetary waves in the mesosphere.

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