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Downward migration of extratropical zonal wind anomalies

R. Alan Plumb and Kirill Semeniuk[1]

Dept. of Earth, Atmospheric and Planetary Sciences, M. I. T., Cambridge, MA

Abstract. We show (in confirmation of previous work) using one- and three-dimensional models that extratropical zonal wind anomalies, produced by fluctuating Rossby wave forcing in the troposphere, appear first in the stratosphere, and migrate downward into the troposphere. By systematically eliminating wave reflection and "downward control" through an induced meridional circulation, it is shown that the downward migration is dependent on neither process. Rather, the mechanism appears to rely on local wave, mean-flow interaction just as in the similar downward migration evident in the tropical quasi-biennial oscillation. In particular, these results imply that the similar downward migration observed in the Arctic Oscillation should not be taken to indicate any controlling influence of the stratosphere on the troposphere.

1. Introduction

The Arctic Oscillation (AO) has been shown to have a clear signal through the stratosphere in winter, as an intensification or weakening of the polar vortex [Thompson and Wallace , 1998; Thompson and Wallace, 2000; Baldwin and Dunkerton, 1999, 2001]. The anomalous zonal mean winds appear first in the upper stratosphere, and then migrate all the way down to the troposphere, an observation that has been interpreted as indicating possible stratospheric influence over the tropospheric AO [Baldwin and Dunkerton, 1999, 2001; Kuroda and Kodera, 1999; Shindell et al., 1999].

In fact, it has long been recognized that the rapid appearance of easterly wind anomalies in the stratosphere, of the kind seen to be associated with the AO and which in their most intense manifestations take the form of "major warmings", follow, and are produced by, anomalously large fluxes of wave activity from the troposphere. Indeed, since the classic study by Matsuno [1971] of the dynamics of major warmings, downward migrating stratospheric easterlies have been seen as symptomatic of the stratospheric response to tropospheric forcing, rather than vice-versa. Therefore it is not clear to what extent, if at all, the observed downward migration of AO zonal wind anomalies can be interpreted in terms of downward influence.

Another, well known, stratospheric phenomenon exhibiting such downward migration is the equatorial quasi-biennial oscillation (QBO), which is understood to be driven by upward-propagating internal gravity or equatorial waves (for a recent wide-ranging review of the QBO, see Baldwin et al. [2001]). In a simple nonrotating, one-dimensional model of the QBO, in which the vertical wavelength of the forcing waves is assumed to be sufficiently small for a WKB assumption to be valid, it can be argued [Plumb, 1977] that no downward influence is possible. The mean flow response in a nonrotating system is purely local, in the sense that the induced acceleration of the mean flow is spatially coincident with the convergence of the wave's momentum flux. In the WKB limit there is no wave reflection, and so propagation of information via the waves is exclusively upward and thus the wave-induced acceleration of the mean flow is dependent only on conditions at and below the level of interest. In this simplest system, wind regimes must migrate downward, towards the wave source, but the arrow of influence is uniquely upward.

While there are some similarities between the equa-

[1] Dept. of Earth and Atmospheric Science, York Unversity, Toronto, ON, Canada.

torial QBO problem and that of the interaction between upward-propagting Rossby waves and the mean flow in the extratropical stratosphere, there are two key differences. First, extratropical Rossby waves have large vertical wavelengths (comparable with the depth of the stratosphere), which makes them susceptible to internal reflection. Indeed, such reflections have been documented in both hemispheres [Harnik and Lindzen, 2001; Perlwitz and Graf, 2001]. Second, because of the effects of planetary rotation, the response of the mean flow itself is nonlocal: a localized rearrangement of potential vorticity (PV) by the waves will be felt, through the agency of an induced meridional circulation, throughout the region within a vertical distance D of the rearrangement, where $D \sim fL/N$, with f, L, and N being respectively the Coriolis parameter, the latitudinal scale of the PV anomaly, and the buoyancy frequency. The induced meridional circulation will be biased downward (exclusively so, in steady state), a fact that has been referred to as "downward control" [Haynes et al., 1991]. Inversion of observed PV anomalies [Hartley et al., 1998; Black, 2002] has revealed a potentially significant induced signal in the troposphere.

It is of some interest, therefore, to ask how much "downward influence" is at work in the low-frequency variability of the Rossby wave, mean flow, interaction of the extratropical stratosphere. We will not try to address all aspects of this issue here, but ask: can extratropical zonal wind regimes which, like those seen as a component of the AO, migrate from the upper stratosphere downward to the troposphere, be produced in simple models in which the stratosphere is unequivocally responding to the troposphere, rather than vice-versa? If so (and it will be shown that they can, both in one- and three-dimensional models) then the observations of such downward migration in the observations is consistent with the AO being a tropospherically produced phenomenon and does not constitute evidence of any stratospheric influence on tropospheric behavior. Moreover, we show that neither internal wave reflection nor "downward control" via an induced meridional circulation plays a significant role in the downward migration: rather, the migration seems to arise from the local wave, mean-flow interaction mechanism discussed by Matsuno [1971], and much as in the QBO.

2. Forced oscillations in the Holton-Mass model

The simplest model of the interaction between upward-propagating Rossby waves and the stratospheric zonal flow is the one-dimensional model of Holton and Mass [1976]. In fact, the actual calculation used here is a slight variant of the Holton-Mass model, discussed in Plumb [1989], to which the reader is referred for details. In outline, the model is a quasigeostrophic representation of a mean zonal flow

$$\bar{u}(y,z,t) = U(z,t)\sin\frac{\pi y}{L} ,$$

and a single Rossby wave whose geopotential height and PV perturbations are

$$\begin{pmatrix} \phi'(x,y,z,t) \\ q'(x,y,z,t) \end{pmatrix} = \mathrm{Re}\left\{ \begin{pmatrix} \Phi(z,t) \\ q(z,t) \end{pmatrix} e^{ikx} \right\} \sin\frac{\pi y}{L} ,$$

in a β-channel bounded on $y = 0, L$, where

$$q = -\frac{g}{f}\left(k^2 + \frac{\pi^2}{L^2} - \frac{1}{\rho}\frac{\partial}{\partial z}\rho\frac{f^2}{N^2}\frac{\partial}{\partial z}\right)\Phi , \qquad (1)$$

$\rho = \rho_0 \exp(-z/H)$ being the basic state density. Wave and mean flow departures from a specified and steady "radiative equilibrium" state are damped by Newtonian cooling with rate coefficient $\alpha(z)$, which increases linearly from $(25\mathrm{d})^{-1}$ at and below 10 km altitude to $(5\ \mathrm{d})^{-1}$ at and above 50 km. This radiative equilibrium state has the corresponding balanced wind profile

$$U_e(z) = \begin{cases} 5 + 0.2z , & z < 10\ \mathrm{km} \\ 25 , & 10 < z < 20\ \mathrm{km} \\ 25 + 75\sin\left(\pi\frac{(z-20)}{80}\right) , & 20 < z < 60\ \mathrm{km} \\ 100 , & z > 60\ \mathrm{km} \end{cases} ,$$

where here U_e is expressed in ms^{-1} and z is in km. The geopotential amplitude $\Phi(0,t)$ is specified at the lower boundary.

The calculation proceeds as follows. The predicted variables are q and Q_y, the mean PV gradient, which is related to U through

$$Q_y = \left[\frac{\pi^2}{L^2}U - \frac{1}{\rho}\frac{\partial}{\partial z}\left(\rho\frac{f^2}{N^2}\frac{\partial U}{\partial z}\right)\right] . \qquad (2)$$

All spectral components are truncated to the single mode $\sin(\pi y/L)$ in the y-direction. The mean flow responds to the Rossby wave through the latter's Eliassen-Palm (EP) flux divergence, and the wave responds to the changing mean flow. The vertical EP flux (there is no latitudinal component), truncated in y, is $F \sin \pi y/L$ where

$$F(z) = \gamma\rho\frac{g^2 k}{2N^2}\,\mathrm{Re}\left(i\Phi\frac{\partial\Phi^*}{\partial z}\right) , \qquad (3)$$

where g is gravity, and $\gamma = 8/(3\pi)$ expresses the mapping of $\sin^2 \pi y/L$ onto $\sin \pi y/L$. The evolution equations for wave and mean state are

$$\frac{\partial q}{\partial t} = -ik\gamma U q - i\frac{gk}{f}\left(\beta + \gamma Q_y\right)\Phi - \frac{g}{\rho}\frac{\partial}{\partial z}\left(\rho \frac{\alpha f}{N^2}\frac{\partial \Phi}{\partial z}\right), \tag{4}$$

and

$$\frac{\partial Q_y}{\partial t} = -\frac{\pi^2}{\rho L^2}\frac{\partial F}{\partial z} + \frac{1}{\rho}\frac{\partial}{\partial z}\left[\rho \frac{\alpha f^2}{N^2}\frac{\partial}{\partial z}\left(U - U_e\right)\right]. \tag{5}$$

Completing each time step involves inverting the mean PV gradient, using (2), to determine $U(z,t)$.

It is now well known, following Holton and Mass [1976], that such models may exhibit spontaneous vacillations, even when $\Phi(0,t) = \Phi_0$ is constant, for sufficient large amplitude. With our choice of parameters, however, vacillations do not occur at any reasonable value of Φ_0; the model always equilibrates to steady state. This fact gives us confidence in ascribing the time-dependence of the response to that of the external forcing itself, and not to any internal fluctuations.

The model is forced at the lower boundary with the amplitude-modulated cycle

$$\Phi(0,t) = \frac{1}{2}\Phi_0\left[1 - \cos\frac{2\pi t}{T}\right]$$

where the period T of the modulation is 200 days. Not surprisingly, even in the equilibrated state, the model response is periodic. As shown in Fig. 1, this periodic response comprises occasional strong amplification of the wave, followed by reversal of the zonal wind and subsequent wave collapse, at altitudes above the stratopause. The behavior is, in fact, very much like that of Holton and Mass [1976] and others, despite the fact that, in this case, the vacillation is clearly externally forced, rather than being internal.

Fig. 1 gives the impression that the anomalies descend with time through the model domain from the upper mesosphere. This is seen more clearly in Fig. 2, in which the wave geopotential height amplitude Φ and the zonal wind anomaly (departure from the time mean) are composited over the final three 200-day cycles of the forcing. Both show systematic downward migration of the signal from the mesosphere to near the tropopause. At high altitudes, maximum wave amplitude and rapid development of easterlies occur at about day 65, 35 days prior to maximum forcing amplitude at the lower boundary. Subsequently, the wave shuts down at high altitudes as anomalous easterlies develop below.

Interpretation of the origins of the zonal wind oscillation throughout the depth of the model, as a direct

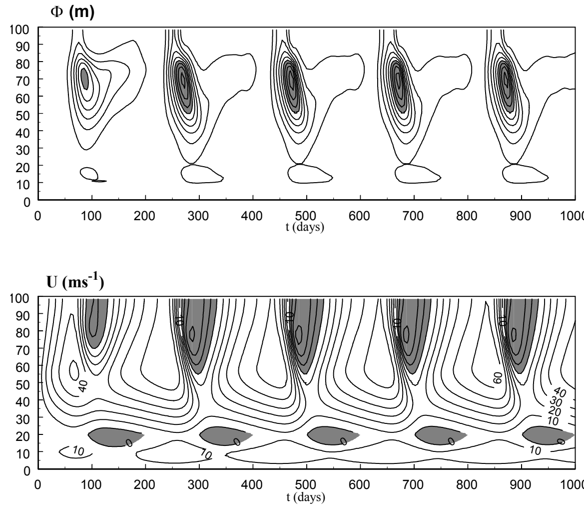

Figure 1. Response of the β-channel model to modulated forcing with $\Phi_0 = 60$m. Top: geopotential wave amplitude (contour interval: 100m; $\Phi > 600$m shaded). Bottom: zonal wind (contour interval 10ms^{-1}; easterlies shaded).

response to the modulation of the lower boundary condition, seems clear in this case. The origins of the downward migration of wind anomalies may not be so clear. In particular, is this migration indicative of a downward transfer of information that might make the behavior at lower altitudes sensitive to conditions above? As discussed in the Introduction, two processes that might be responsible for downward influence in this situation are wave reflection and downward penetration of the meridional circulation. In terms of the mathematics of the problem at hand, these two possibilities manifest themselves in separate ways: the first, through solution of the second-order wave problem—(4) together with (1)—and the second through the mean flow problem, which is nonlocal both through the nonconservative second term on the right hand side of (5), and especially through the inversion of (2) by which U is obtained from the updated distribution of Q_y. In order to investigate the role of these two processes in the downward migration, each was suppressed, in turn.

We have investigated the first possibility in the context of these experiments by replacing the calculation of the wave's vertical structure with an instantaneously

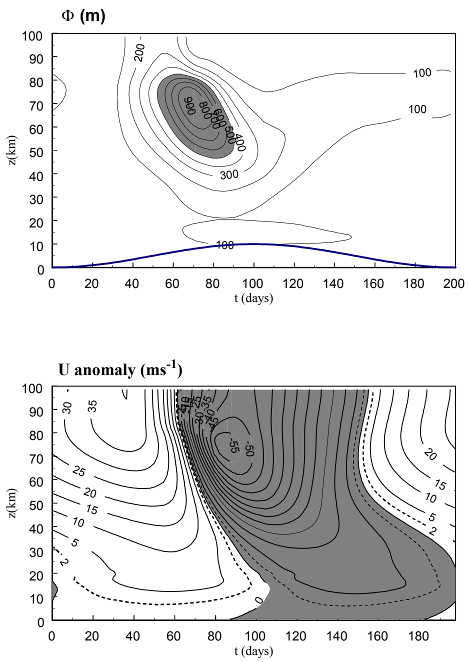

Figure 2. The case of Fig. 1, composited over the final three forcing periods. Top: wave geopotential amplitude (m). Bottom: zonal wind anomaly (departure from the time mean) (ms^{-1}). The heavy curve at the lower boundary shows the boundary forcing $\Phi(0,t)$ (amplitude is arbitrary).

steady state[1], WKB solution in which, amongst other things, wave reflections are neglected. The WKB approach rests on the assumption that the mean state varies over height scales long compared with the vertical wavelength and on time scales longer than the time (a few days, in our case) for the wave to propagate through the system. In our case, as can be judged from the results to be presented below, this is only a fair approximation. Nevertheless, the essential point of this exercise is to demonstrate that downward migration of anomalous mean winds still occurs in a system in which no downward reflection is permitted: the accuracy of the WKB approximation is not crucial to this demonstration. Rather than being determined from (4) and (1), the wave solution is now specified to be

$$\begin{pmatrix} \Phi \\ q \end{pmatrix} = \mathrm{Re}\begin{pmatrix} \tilde{\Phi}(z) \\ \tilde{q}(z) \end{pmatrix} \exp\left(\frac{z}{2H} + \int_0^z i\, m(z)\, dz\right) , \tag{6}$$

where $\tilde{\Phi}$ and $\tilde{q}$ are slowly varying functions of z, related by

$$\tilde{q}\,(z) = -\frac{g}{f}\left[k^2 + \frac{\pi^2}{L^2} + \frac{f^2}{N^2}\left(m^2 + \frac{1}{4H^2}\right)\right]\tilde{\Phi} , \tag{7}$$

and where, for an upward propagating stationary wave, the local vertical wavenumber is

$$m(z) = \frac{N}{f}\sqrt{\frac{\beta + \gamma Q_y}{\gamma U} - k^2 - \frac{\pi^2}{L^2} - \frac{f^2}{4N^2H^2}} .$$

Instead of solving directly for the wave amplitude $\Phi(z)$, since it is the EP flux divergence that appears in (5), we use the WKB solution for F, which we obtain by multiplying (4) by Φ^*, taking the real part, evaluating the remaining terms to leading order in the WKB expansion, and finally using (3) to obtain

$$\frac{\partial F}{\partial z} = -\Lambda F ,$$

where

$$\Lambda(z) = \frac{\alpha}{\gamma k U m}\left(m^2 + \frac{1}{4H^2}\right) \tag{8}$$

so

$$F(z) = F(0)\exp\left[-\int_0^z \Lambda(z')\, dz'\right] . \tag{9}$$

The expression (9) is then used as the forcing term in the mean PV tendency equation (5). It is evident from

[1]Since the wave propagation time across the domain (a few days) is much shorter than the externally imposed periodicity, the wave is always close to steady state balance. We have confirmed this by repeating a case like that of Fig. 1 with the wave solution from a steady state calculation.

(9) that the wave forcing of the mean state at any altitude is independent of conditions above that altitude, a simple consequence of the suppression of downward reflection inherent in the WKB approach.

Results of substituting the WKB solution (9) in place of the full wave calculation, for the experiment of Fig. 1, are shown in Fig. 3. Although some differences are

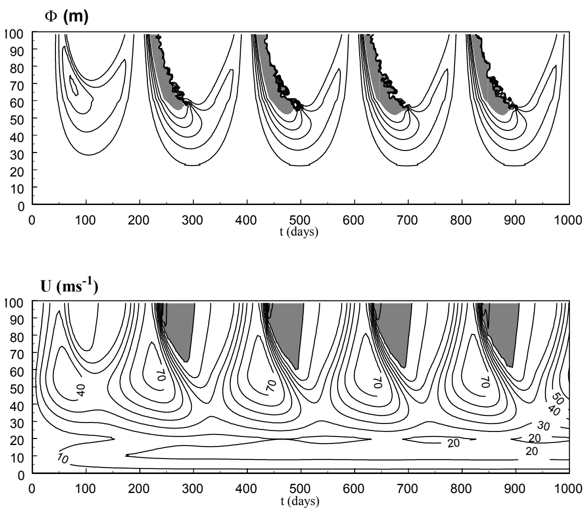

Figure 3. As Fig, 1, but with the calculation of wave vertical structure based on a WKB approximation. (See text for discussion.)

apparent (the noisiness at high altitude occurs where U is near zero and (8) is almost singular), the downward migration of the zonal wind anomalies is still present, and in fact is accentuated by the WKB calculation: the wave amplification and the consequent easterly acceleration at high altitudes occur earlier in the modulation cycle than in Fig. 1, while the timing at lower altitudes is essentially unchanged. Therefore, wave reflection is not responsible for the downward migration.

The second possibility for downward influence arises through the vertical non-localness of the mean flow problem. This effect was suppressed, artificially, by replacing (2) with

$$Q_y = \frac{\pi^2}{L^2} U \ , \qquad (10)$$

to make the inversion vertically local (by making the dynamics of the mean flow, but not of the wave, barotropic). Making this change eliminates the coupling between mean wind and temperature fields and a PV evolution equation that is consistent with (10) has no term corresponding to the last term in (5). Since that term provides another route for downward information transfer, its removal serves the interests of this exercise. However, its absence removes the thermal relaxation toward equilibrium that is an essential part of the system's dynamics. To remedy this without introducing any new avenue for vertical information transfer, a Rayleigh friction was added to relax the mean flow back toward the equilibrium wind distribution $U_e(z)$, with a rate coefficient α_R that varied from $1/(200)$ day^{-1} at the ground to $1/2$ day^{-1} above 60 km. The tendency equation (5) for the mean state is replaced by

$$\frac{\partial Q_y}{\partial t} = -\frac{\pi^2}{\rho L^2}\frac{\partial F}{\partial z} - \alpha_R \frac{\pi^2}{L^2}\left(U - U_e\right) \ . \qquad (11)$$

Thus, while the wave may now carry information vertically (and we revert to the full wave solution here, rather than using the WKB approach), the mean flow solution defined by (10) and (11) is now local in z, except for a small viscosity that was added to prevent noisiness in the vertical. (Results were found to be insensitive to its value.) Unlike the base case of Fig. 1, this case does exhibit internal variability (in response to a constant forcing) at the forcing amplitude used above, and so we used $\Phi_0 = 30$m here. Results from this "local" calculation are shown in Fig. 4. The characteris-

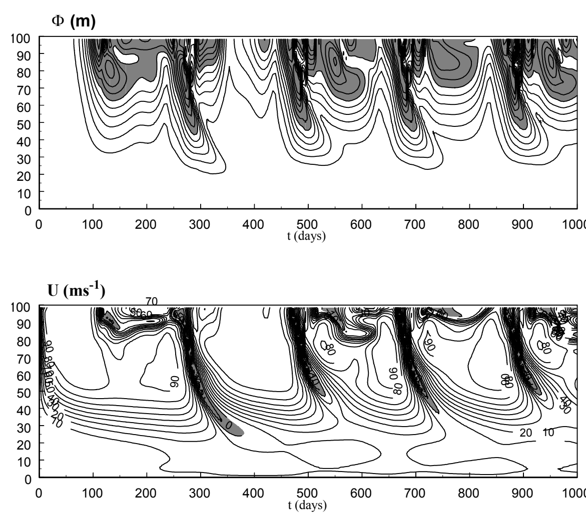

Figure 4. As Fig. 1, but using the local PV inversion, and a wave forcing amplitude $\Phi_0 = 30$m.

tics of the response differ in many ways from those of the original calculation; nevertheless, the easterly wind regimes once again migrate downward, much as in the first case.

3. 3D model results

In order to show that these results are not a consequence of the simplified nature of the model used, as well as to investigate latitudinal as well as vertical migration of zonal jets, we performed similar calculations in a three-dimensional stratospheric model. The model is based on the spherical harmonic decomposition of the primitive equations in pressure coordinates. The domain extends from the surface to 65 km with 40 levels spaced uniformly in log-pressure. The spectral truncation is trapezoidal with 42 meridional modes and five zonal modes. The diabatic heating is approximated by Newtonian relaxation with a radiative damping rate having the same height-dependence as that in the β-channel model. A zonal jet is produced via a prescribed time-independent radiative equilibrium temperature. The vertical profile of the radiative equilibrium zonal jet is similar to that in the 1-D case, and it peaks near 60°N. The geopotential height is specified at the surface to be

$$\Phi(\lambda,\phi,t) = \frac{1}{2}\Phi_0\left(1-\cos\frac{2\pi t}{T}\right)\exp[-(\frac{\phi-45}{15})^2]\cos(\frac{\pi\lambda}{180})$$

where λ and ϕ represent latitude and longitude (in degrees). Because of the latitudinal spreading of the wave away from the source latitudes and consequent dilution of wave activity, response to a given forcing is weaker than in the 1-D case. Therefore the forcing amplitude was increased in the 3-D model, to $\Phi_0 = 120$m. The model was run for 1000 days.

The time-height behavior of this model is qualitatively similar to the 1-D case, as illustrated in Fig. 5. The timing of the mean flow response is similar to that shown in Fig. 2 (note the different vertical scale on Fig. 5), with anomalous easterlies appearing at 50 km about 25 days prior to maximum forcing forcing amplitude at the lower boundary, and about 20 later at lower altitudes.

Meridional cross-sections of anomalous zonal wind and EP flux are shown in Fig. 6, at four phases of the 200 day cycle. Successive easterly and westerly wind regimes first appear in the low latitude upper stratosphere, before migrating downward and poleward, in a manner qualitatively similar to that described by Kuroda and Kodera [1999] from composites

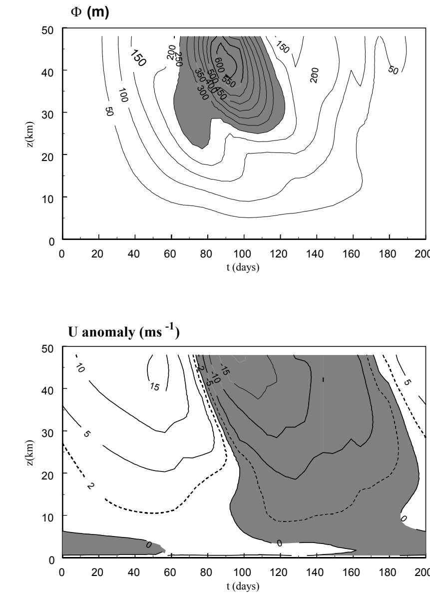

Figure 5. As Fig. 2, (but note the different vertical scale) for the 3D model with $\Phi_0 = 120$m.

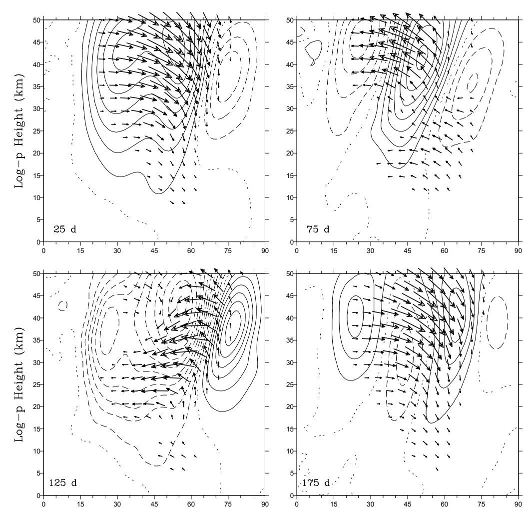

Figure 6. Zonal wind anomaly (contour interval 2 m/s) and Eliassen-Palm flux anomaly divided by density ($\mathbf{F}/\rho_0$) at days 25, 75, 125 and 175 of the forcing cycle in the 3D model.

of stratospheric data. The anomalous EP fluxes[2] also show similar behavior to the observations, in particular anomalously large upward and equatorward fluxes during and just prior to maximum easterlies. Thus, at this more detailed level, results suggest that the observed behavior is consistent with the stratosphere responding passively to fluctuations of wave activity in the troposphere.

4. Conclusions

In the one-dimensional model, we have found that the downward migration through the stratosphere of extratropical wind anomalies is not dependent on "downward control" via induced meridional circulations, nor on downward reflection of upward-propagating waves. Rather, the mechanism of downward propagation is determined by the purely local interaction between the upward-propagating wave and the zonal mean flow. Under increasing wave forcing, anomalies easterlies first appear at high altitudes. This causes a subsequent collapse of the wave amplitude at those altitudes, but deceleration of the mean flow continues lower down, and thus the locus of maximum easterly acceleration migrates downward. As such, the mechanism for downward migration of the easterlies much as described by Matsuno [1971], and entirely analogous to that found in simple models of the QBO. As Fig. 2 shows, anomalous westerlies also descend, just because the developing, migrating, easterly anomalies block further wave propagation, leaving the flow above to relax back toward equilibrium.

Under this scenario, the downward migration cannot be interpreted as indicative of any downward influence. This has been confirmed explicitly in calculations with the 1D model, in which changes (such as severely damping the zonal flow fluctuations) made above some level were found to have no significant impact on the evolution of the flow at lower altitudes. As an example, the calculation of Fig. 1 was repeated with the mean PV gradient Q_y held fixed, after day 50, with its value at day 50 at all altitudes above z_0, thus removing the low frequency variability in Q_y above z_0. Results for $z_0 = 25$km are shown in Fig. 7. There is some modest impact on the evolving mean flow within a scale height or so below z_0, but nevertheless the oscillation at lower altitudes is almost unchanged from the case of Fig. 1.

We have not discussed here the internal vacillations that stratospheric models, especially 1D models, have

[2]The EP fluxes shown here are calculated from the full, non-geostrophic definition.

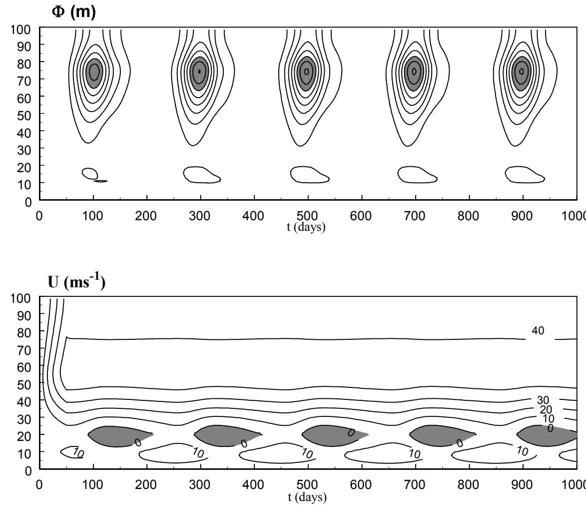

Figure 7. As Fig. 1, but with the mean PV gradient Q_y held fixed after day 50 (at its day 50 value) above $z_0 = 25$km.

long been known to exhibit in the presence of constant forcing, provided the wave amplitude exceeds some threshold value. These vacillations also display downward migration, and for the same reason as the externally forced oscillations discussed here. In fact, we have found in controlled experiments similar to that shown in Fig. 7 that, just as for the externally forced oscillations, evolution of the vacillation at low altitudes is very insensitive to behavior more than a scale height or so above. The only exception we have found to this statement is when the imposed wave amplitude at the lower boundary of the model is very close to the threshold value for internal vacillations, in which case changes at high altitude can have a significant impact throughout the model by eliminating the vacillations altogether.

All in all, results from the various 1D model experiments and from the 3D model explicitly confirm that downward migration of zonal wind anomalies through the stratosphere and into the troposphere occurs in situations for which the oscillation itself is imposed at the lower boundary. The observed similar behavior in successive phases of the AO is therefore quite consistent with what would be expected if the key dynamics of the AO were purely tropospheric, with the stratosphere responding in an entirely passive way to fluctuations in tropospheric Rossby wave activity. More generally, it is wrong to argue that the observed fact that stratospheric wind regimes appear before those in the troposphere indicates, of itself, stratospheric influence on tropospheric climate.

This is not the same thing, of course, as saying that no such influence exists. While we have found that wave reflection and remote action via the mean meridional circulation do not play an essential role in the downward migration, they may still produce a quantitatively significant coupling of the two regions. That the meridional circulation may do so in principle is indicated by the results of piecewise PV inversion of stratospheric PV anomalies [Hartley et al., 1998]. In the context of the AO, Black [2002] has shown that the AO signal in stratospheric PV induces zonal wind anomalies in the lower troposphere comparable with those observed. Moreover, numerical experiments by Polvani and Kushner [2002] provide explicit evidence of the potential sensitivity of the surface AO signal to stratospheric conditions. By what mechanism the stratosphere might play a role in the dynamics of the AO, however, remains to be clarified.

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References

Baldwin, M. P., and T. J. Dunkerton, Propagation of the Arctic oscillation from the stratosphere to the troposphere. *J. Geophys. Res., 104*, 30,937-40,946, 1999.

Baldwin, M. P., and T. J. Dunkerton, Stratospheric harbingers of anomalous weather regimes, *Science*, *294*, 581-584, 2001.

Baldwin, M., L. Gray, T. Dunkerton, K. Hamilton, P. Haynes, W. Randel, J. Holton, M. Alexander, I. Hirota, T. Horinouchi, D. Jones, J. Kinnersley, C. Marquardt, K. Sato, and M. Takahashi: The Quasi-biennial Oscillation. *Rev. Geophys.*, *39*, 179-229, 2001.

Black, R. X., Stratospheric forcing of surface climate in the Arctic Oscillation, *J. Climate*, *15*, 268-277, 2002.

Harnik, N., and R. S. Lindzen, The effect of reflecting surfaces on the vertical structure and variability of stratospheric planetary waves. *J. Atmos. Sci.*, *58*, 2872–2894, 2001.

Hartley, D. E., J. Villarin, R. X. Black, and C. A. Davis, A new perspective on the dynamical link between the stratospheric and troposphere, *Nature*, *391*, 471-474, 1998.

Haynes, P.H., C.J. Marks, M.E. McIntyre, T.G. Shepherd, and K.P. Shine, On the "downward control" of extratropical diabatic circulations by eddy-induced mean zonal forces, *J. Atmos. Sci.*, *48*, 651-678, 1991.

Holton, J. R., and C. Mass, Stratospheric vacillation cycles, *J. Atmos. Sci.*, *33*, 2218-2225, 1976.

Kuroda, Y., and K. Kodera, Role of planetary waves in the stratosphere-troposphere coupled variability in the northern hemisphere winter. *Geophys. Res. Lett.*, *26*, 2375-2378, 1999.

Matsuno, T., A dynamical model of the sudden stratospheric warming phenomenon. *J. Atmos. Sci.*, *28*, 1479-1494, 1971.

Perlwitz, J., and H.-F. Graf, Troposphere-stratosphere dynamic coupling under strong and weak polar vortex conditions. *Geophys. Res. Lett.*, *28*, 271-274, 2001.

Polvani, L.M., and P.J. Kushner, Tropospheric response to stratospheric perturbations in a relatively simple general circulation model. *Geophys. Res. Lett.*, *29(7)*, 10.1029/2001GL014284, 2002.

Plumb, R.A., The interaction of two internal gravity waves with the mean flow; implications for the theory of the quasi-biennial oscillation. *J. Atmos. Sci.*, *34*, 1847-1858, 1977.

Plumb, R.A., On the seasonal cycle of stratospheric planetary waves, *PAGEOPH*, *130*, 233-242, 1989.

Shindell, D. T., R. L. Miller, G. A. Schmidt, and L. Pandolfo, Simulation of recent northern winter climate trends by greenhouse gas forcing. *Nature*, *399*, 452-455, 1999.

Thompson, D., and J. M. Wallace, The Arctic oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.*, *25*, 1297-1300, 1998.

Thompson, D., and J. M. Wallace, Annular modes in the extratropical circulation. Part I: Month-to-month variability. *J. Climate*, *13*, 1000-1026, 2000.

R. A. Plumb, Room 54-1712, Dept. EAPS, M. I. T., Cambridge MA 02139. Email: rap@rossby.mit.edu.

K. Semeniuk, Dept. of Earth and Atmospheric Science, York University, Petri Science Building, 4700 Keele Street, Toronto, Ontario, MJ3 1P3, Canada. Email: kirill@nimbus.yorku.ca.

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