On the seasonal asymmetry of the low and middle latitude QBO circulation anomaly.

Jonathan S. Kinnersley
Dept. of Applied Mathematics,
University of Washington,
Seattle, WA 98195-352420

Date submitted to JAS: Jan 1998
Date revised: June 1998

Abstract

It is usually assumed that the circulation anomaly induced directly by the quasi-biennial oscillation (QBO) in the equatorial zonal wind is, approximately, a seasonally-independent, two-cell structure, symmetric about the equator and confined to the tropics.

It is shown here using a simple two-dimensional model, that although the two-cell structure exists at equinox, at solstice the summer cell disappears and the winter cell is greatly strengthened (about 3 times stronger than at equinox) and expanded. Strong cross-equatorial flow is induced near the shear zone where the QBO winds are easterly. This result may explain why the observed anomalies in trace gases are small in the summer hemisphere and also reduces the need for a modulation of the planetary wave fluxes at low and middle latitudes in order to explain the modulation of the circulation there. At low latitudes, the shape of the isopleths of a modeled tracer closely resemble those observed in both easterly and westerly phases of the QBO.

The difference between the hemispheres stems from the non-linear advection of the equatorial zonal wind anomaly into the winter hemisphere, where it leads to a large temperature anomaly due to the explicit latitudinal dependence of the thermal wind equation. An asymmetric circulation anomaly is induced both at steady state and in the transition to steady state: during transition, an asymmetric circulation works to produce the asymmetric temperature anomaly required by thermal wind balance, while at steady state the same circulation balances the Newtonian cooling induced by the (now established) temperature anomaly. Thus, in the real stratosphere, an asymmetric circulation may exist even in the absence of a large asymmetric temperature anomaly, though the circulation anomaly will eventually produce such a temperature anomaly. The poleward extent of the circulation anomaly is increased by moderate Rayleigh friction on the zonal wind polewards of about 15 degrees in the winter hemisphere. However, low-latitude friction reduces the low-latitude zonal wind anomaly and hence the circulation anomaly.

1 Introduction

After the discovery of the quasi-biennial oscillation (QBO) in the lower-stratospheric zonal wind over the equator (Reed 1960) QBO signals were discovered in the ozone column (eg. Angell and Korshover 1964, Hasebe 1984). Later, satellite observations were used to study the vertical distribution of the QBO in ozone (eg. Randel and Wu 1996) and other trace gases in the stratosphere. Features that these trace gases all show are small anomalies in the summer hemisphere and winter anomalies that extend from low to high latitudes. Initial modelling studies (Holton and Lindzen 1972, Plumb and Bell 1982, Dunkerton 1985) were able to produce fairly realistic vertical structures of the equatorial QBO in zonal wind by assuming it was driven within about 7 degrees of the equator by the dissipation of tropospherically-forced equatorial waves. A forcing of the zonal wind near the equator will induce a meridional circulation to alter the zonal-mean temperature and maintain thermal wind balance. The studies of Plumb and Bell (1982) and Dunkerton (1985) showed that the mean circulation induced by such a forcing was a two-celled structure symmetric about the equator, with ascent over the equator and descent at about 15 degrees when the equatorial shear in the zonal wind was easterly. These studies, however, did not consider the effect of the seasonal cycle on the circulation anomaly, and in certain other studies (eg Holton 1989, Tung and Yang 1994) it has been assumed that the circulation anomaly induced directly by the QBO is independent of season. This assumption is questioned in this paper and in the recent paper of Jones et al. (1998). The seasonal dependence of the trace-gas anomalies in the extra tropics is generally attributed either to the mean circulation (eg. Gray and Pyle 1989, Gray and Dunkerton 1990, Holton 1989, Gray and Ruth 1993) which, it is suggested, disrupts the summer anomaly and transports the winter anomaly to higher latitudes, or to the presence of a circulation anomaly at middle latitudes in winter, induced by a modulation of the planetary wave fluxes by the equatorial QBO (eg Tung and Yang 1994). The latter theory is supported by observations of strong correlations between temperature and ozone in the extratropics (Randel and Cobb 1994).

A third possibility (explored in this paper) is that the circulation anomaly induced by the equatorial zonal wind anomaly is *not* independent of season and can be influenced by the seasonally-dependent Brewer-Dobson circulation. It will be shown here that the summer-to-winter mean circulation may advect the equatorial zonal wind anomaly into the winter hemisphere where, because of the growth of the Coriolis parameter with latitude, it requires a larger temperature anomaly to maintain thermal wind balance. Such a seasonal asymmetry of the near-equatorial zonal wind has been observed (eg. Dunkerton and Delisi 1985, 1997) and has also been noted in model simulations of the QBO (Dunkerton 1997). The seasonally-asymmetric temperature anomaly leads to a large winter circulation anomaly which, it will be shown, can easily dominate the summer circulation anomaly. The resulting one-celled QBO circulation anomaly exists without any Rayleigh friction being applied to the zonal wind, but is confined to within about 30 degrees of the equator. If a moderate friction (with a relaxation time of 20 days) is applied to the zonal wind in the winter hemisphere the anomaly spreads to middle latitudes in the winter hemisphere.

The effect on ozone is similar to that of the theory of Tung and Yang 1994 (and hence is supported by the correlations observed between temperature and ozone anomalies), in that the ozone anomaly is produced by a circulation anomaly, but it differs in that it does not require an interaction with the planetary waves. While it appears that the planetary waves are influenced by the QBO (eg Holton and Tan (1982), Dunkerton and Baldwin (1991), Baldwin and Tung 1994), their fluxes are not as strongly correlated with the equatorial QBO (even after seasonally averaging the fluxes) as the ozone column is (Randel and Cobb 1995, Kinnersley and Tung 1998). Although this may be due to the difficulty in calculating the fluxes it nevertheless remains to be proved that their variability could produce the strong low and middle latitude signal observed in the ozone column. In fact, in the model study of Kinnersley and Tung (1998, 1999) planetary wave variability is not a major cause of the low and middle latitude ozone column QBO.

2 The Numerical Model

The model used is a simplified version of the zonal-mean part of the isentropic model described in Kinnersley (1996), with higher-resolution. It now has 39 evenly spaced boxes from pole to pole (about 5 degrees wide) and 31 layers (evenly spaced in its isentropic vertical coordinate) from the ground to about 49km. It uses the vertical coordinate $z = ln(\theta/\theta_s)$, where θ is potential temperature and θ_s is potential temperature at the ground (set here to equal 300K at all latitudes). The layers are about 1.2km thick in the model's stratosphere, and 5km thick in the troposphere. For the figures in this paper, the results have been interpolated onto a latitude–height grid and only the values between 24 and 36km displayed.

The model solves the same set of zonal-mean equations as in Kinnersley and Harwood (1993), using the same method with the improvement described in Kinnersley (1996). The model solves the zonal-mean prognostic equations for zonal wind

$$u_t + vG + wu_z = -Fu \ , \tag{1}$$

and for isentropic density ρ

$$\rho_t + \frac{(v\rho cos\phi)_y}{cos\phi} + (w\rho)_z = 0 \ , \tag{2}$$

where $\rho = -p_z/g$, p is pressure on an isentropic surface, $G = \tau_y/(acos\phi)$, τ is angular momentum, ϕ is latitude, a is the earth's radius and F is the friction coefficient which will be described later. A subscript denotes differentiation by that variable. The model's zonal wind and temperature

$$T = \theta \left(\frac{p}{1000}\right)^{0.288} \tag{3}$$

are constrained by the thermal wind equation (in isentropic coordinates)

$$c_pT_y = -\left[u\left(f + \frac{utan\phi}{a}\right)\right]_z \ , \tag{4}$$

where c_p is the specific heat of air at constant pressure and f is the planetary vorticity. This constraint is used to determine v, since w is determined by the heating rate

$$w = \frac{Q}{T} \ . \tag{5}$$

In order to solve for v, a stream-function ψ is introduced such that

$$\psi_z = -v\rho cos\phi \ , \tag{6}$$

and

$$\psi_y = \left(w\rho - \frac{p_t}{g}\right) cos\phi \ , \tag{7}$$

so that Eq.(2) is automatically satisfied. This allows us to write u_t and p_t (and hence T_t from Eq.(3)) in terms of w and ψ. Note that, since $w\rho cos\phi$ is the mass flux of air parcels across an isentropic surface and $-cos\phi p_t/g$ is the mass-weighted vertical velocity of the isentropic surface in pressure coordinates, then ψ_y must be the mass-flux of air parcels across an isobaric surface, and hence ψ describes the non-divergent circulation in pressure coordinates. We can thus write the vertical velocity of air parcels relative to isobaric surfaces, w^p, as

$$w^p = \frac{\psi_y\left(\frac{dz^*}{dz}\right)}{\rho cos\phi} \ , \tag{8}$$

where z^* is geometric height, so that the dz^*/dz term converts from an isentropic to a geometric velocity. Using Eq.(4) to write an expression for u and T at the next time step (which involves u_t and T_t) we are able to find a second-order partial differential equation for ψ. This equation is elliptic provided the zonal wind remains inertially stable (i.e. $y\tau_y < 0$) and so can be solved by the successive over-relaxation method. The heating rate in this version of the model is a simple relaxation

$$Q = \frac{T_{eqm} - T}{t_{rel}} \tag{9}$$

towards an equilibrium temperature T_{eqm} with a relaxation time $t_{rel} = 10$ days, which is a typical value for the middle stratosphere. The equilibrium temperature at the equator decreases from 300K at the ground to 200K at about 14km and from there increases to 280K at about 49km, thus approximating the stratospheric temperature gradient. The model results do not differ significantly if the equatorial equilibrium temperature is constant with height, however.

Away from the equator the equilibrium temperature is set to simulate either solstice or equinox. For solstice, the equilibrium temperature changes linearly with latitude, dropping 50K from the summer to the winter pole.

$$T(\phi)_{eqm} = T(0)_{eqm} - \frac{25\phi}{90}, \tag{10}$$

where ϕ is latitude in degrees. For equinox, T_{eqm} changes as the cosine of latitude, being 25K colder at the poles than at the equator.

$$T(\phi)_{eqm} = T(0)_{eqm} + 25(cos\phi - 0.5) \tag{11}$$

In the lowest three model layers the equilibrium temperature is constant with latitude to avoid numerical problems near the ground. Also in these three layers and in the uppermost four layers a friction of 10 days is applied to the zonal wind. In all the other model layers friction (with a relaxation time of 20 days) acts only on the westerly zonal winds (similar to the "one-sided" friction used by Dunkerton 1989, although he applied it only to strong westerlies). This is supposed to represent in a crude way the effect of Rossby waves on the zonal wind. However the results are not overly sensitive to this friction, as will be shown in section 6. Near the equator, in order to keep the equation for the meridional circulation elliptic, horizontal fluxes of angular momentum are invoked in order to ensure that angular momentum decreases monotonically away from the equator. The effect of these fluxes on the results are discussed in section 5.1. An explicit Adams-Bashforth time step of one day is used. Each of the model runs to be described are 60 days long and start from a state of rest with the temperature equal to the equatorial equilibrium temperature.

In order to investigate the effect of an equatorial QBO shear zone on the mean circulation, for most of the runs described here the equatorial zonal wind within 5 degrees of the equator and between about 17km and 45km is relaxed towards a specified profile, u_{qbo} with a relaxation time of 2 days. Below model level 15 (about 30km) u_{qbo} has one constant value u_{lower} while above 30km it has a different constant value u_{upper}. Thus a vertical shear is produced in the model's zonal wind over the space of approximately one model layer (about 1.2km).

To demonstrate the effect that the QBO circulation might have on a stratospheric trace gas with a long lifetime, a tracer was advected using the modeled meridional circulation and an upwind advection scheme.

3 Results of model runs

Two sets of experiments will be described in this section. Each set consists of three runs. In the first run there is no forcing of an equatorial QBO wind shear, while in the second a shear is forced (see section 2) with $u_{upper} = u_{westerly}$ and $u_{lower} = u_{easterly}$, and in the third the shear is reversed, with $u_{upper} = u_{easterly}$ and $u_{lower} = u_{westerly}$. For the runs of this section $u_{easterly} = -30$ m/s and $u_{westerly} = 20$ m/s. For the sake of convenience the second run shall be referred to as the westerly run since the vertical shear of the zonal wind at the equator is westerly, while the third run will be called the easterly run.

In the first set of experiments the equinoctal diabatic forcing described in section 2 (Eqn.(11)) is used, while the second set uses the forcing for a solstice (Eqn.(10)).

Note that when vertical velocity is referred to it will mean the vertical velocity relative to isobaric surfaces, w^p of Eq.(8), unless otherwise stated.

3.1 Equinox

The zonal wind on day 60 of the easterly run is shown in figure 1. Since the heating (given by Eqn.(9) and Eqn.(11)) and friction are symmetric about the equator, the zonal wind is also symmetric, and reach typical equinoctal values for the mid-stratosphere.

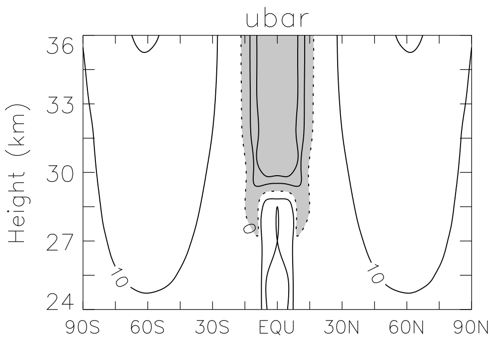

Figure 1: *Zonal wind from equinoctal run with easterly QBO forcing (contours at 10m/s intervals, negative values shaded).*

Figure 1 also illustrates the QBO forcing near the equator. The anomaly (the westerly run minus the easterly run) in the vertical velocity w^p (figure 2b) shows the classic two-celled structure, with descent over the equator and ascent between about 15 and 30 degrees from the equator in both hemispheres.

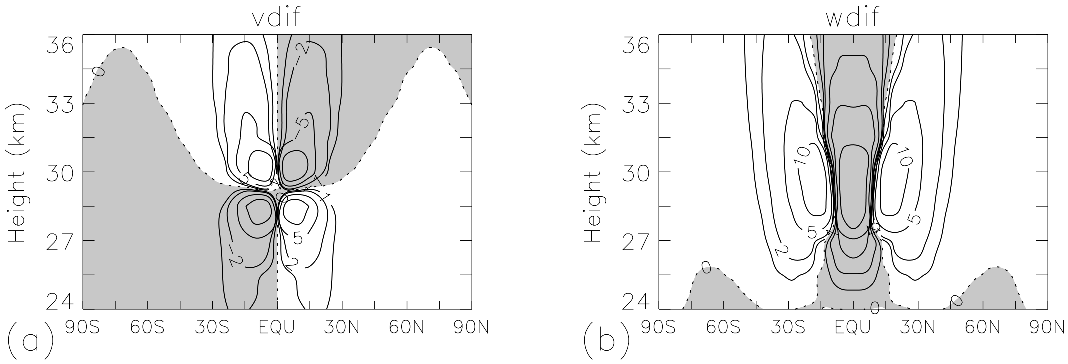

Figure 2: *(a) horizontal (in km/day) and (b) vertical (in m/day) velocity anomalies (westerly run minus easterly run) with equinoctal heating. Contours at 0,1,2,5,10,20. Negative values shaded.*

The anomaly in the horizontal velocity (figure 2a) simply closes off the two cells, and advects angular momentum in such a sense as to oppose the vertical gradient of zonal wind (poleward advection below the shear zone tending to decrease the easterly wind there while equatorward advection above tends to decrease the westerlies).

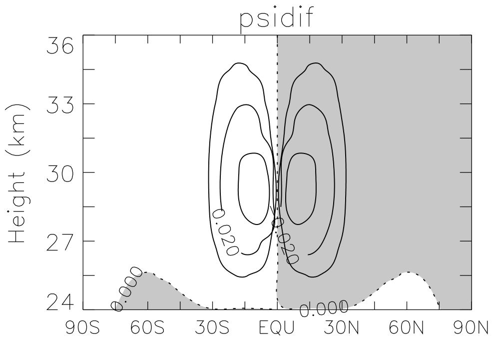

Figure 3: *Stream-function anomaly (westerly run minus easterly run) with equinoctal heating. Negative values shaded.*

The anomaly in the stream function ψ (figure 3) shows clearly the two-cell circulation anomaly modeled by Plumb and Bell (1982) and the one which has been used in models by Holton (1989) and Tung and Yang (1994) and conceptually by others to describe a seasonally-independent circulation induced directly (as opposed to a circulation anomaly produced by a modulation of planetary wave fluxes) by the equatorial zonal wind QBO. In the next section it will be shown that the anomally induced by the equatorial QBO even in this simple model is far from independent of season.

3.2 Solstice

The heating for the solstice runs is given by Eqn.(9) and Eqn.(10). The zonal wind produced by the first run, with no QBO forcing, (figure 4) is broadly typical of a northern winter.

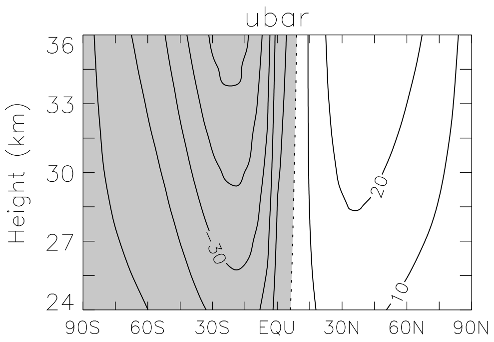

Figure 4: *Zonal wind from solsticial run with no QBO forcing (contours at 10m/s intervals, negative values shaded).*

The meridional flow (figure 5a) extends across the equator and well into the winter hemisphere because of the friction which is applied only to the westerly winds (see section 2). As explained by Dunkerton (1989) friction prevents the winds from reaching the thermal–equilibrium value (determined by T_{eqm} and thermal wind balance) and so prevents the temperature from reaching T_{eqm}. Hence there is heating and a meridional flow (which advects angular momentum and so balances the friction on the zonal wind).

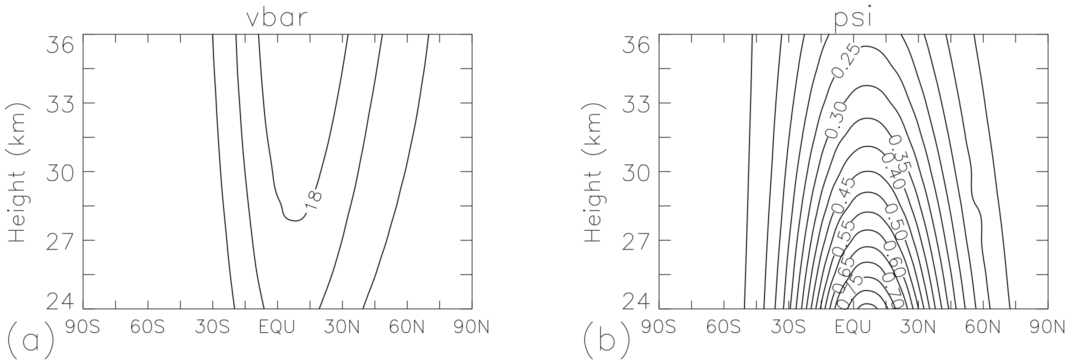

Figure 5: *(a) meridional velocity (in km/day) and (b) streamfunction from solsticial run with no QBO forcing.*

The cross-equatorial advection does not, as might be expected, produce a zero gradient of angular momentum at low latitudes on the summer side of the equator (this is clear from figure 4 since the strongest easterlies do not occur on the equator). This appears to be mainly because of the shape of the streamlines (figure 5b) which have a positive slope everywhere in the summer hemisphere (the slope reverses sign about 5 degrees north of the equator, which is also where the heating rate changes sign). If the horizontal gradient of angular momentum near 24km is positive in the southern hemisphere and angular momentum is advected along the streamlines, then at steady state the streamlines will also be isolines of angular momentum and (assuming the streamlines retain the same basic shape) the horizontal gradient of angular momentum will be positive everywhere in the summer hemisphere. After 60 days the model has reached an almost, but not exactly, steady state. After 130 days (even closer to a steady state) the zonal wind and streamfunction have qualitatively the same shapes, so that it seems reasonable to extrapolate this argument back from a steady state to an almost steady state.

The difference between the meridional and vertical velocities of the westerly and easterly runs (figure 6) shows clearly the one-celled QBO anomaly in the circulation.

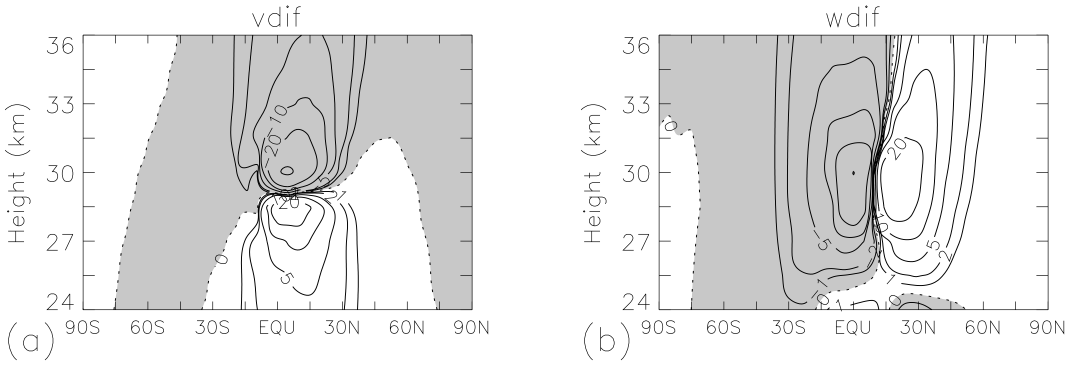

Figure 6: *As figure 2 but for solsticial heating.*

The anomaly in the meridional circulation (figure 6a) is largest at about 5N (with a maximum value about 3 times larger than the maximum in figure 2a) and is also large on the equator (unlike figure 2a, which is zero on the equator). The equatorial anomaly in the vertical velocity (figure 6b) is similar in size to that of the equinoctal run (figure 2b) but the northern anomaly has grown greatly in strength and polewards extent while the southern cell has disappeared and been replaced by an extension of the equatorial arm of the winter cell. Note that the maximum vertical anomaly at 45N is about 5 m/day while at 45S it is less than 1 m/day. At 30N the anomaly is about 10 times that at 30S (and of the opposite sign, unlike figure 2b).

The mechanism behind the large seasonal dependence of the QBO circulation anomaly will be explained in detail in section 5.

4 Impact on a tracer

An inert tracer was included in the model, initialised with a vertical gradient but no horizontal gradient and advected by the modeled circulation. The values on day 60 for the easterly and westerly runs of section 3.2 are shown in figure 7. The shapes are distinctly different, with easterly shear leading to a "nose" projecting into the winter hemisphere above the region of shear, while westerly shear leads to a flattening of the isopleths below the region of shear.

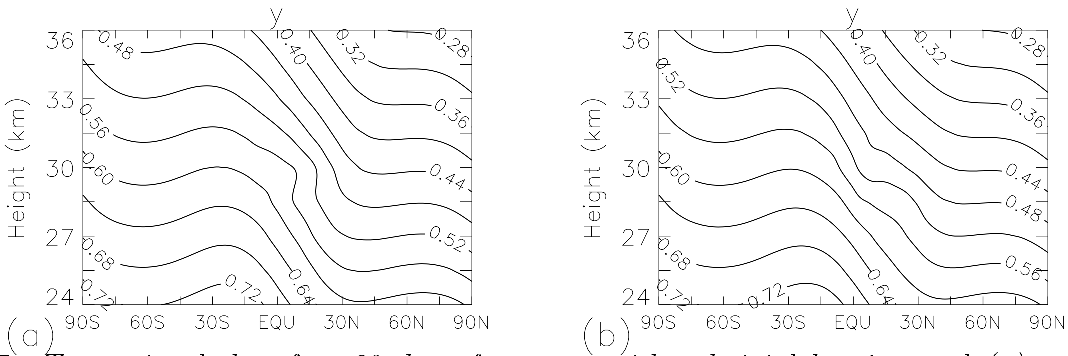

Figure 7: *Tracer isopleths after 60 days from run with solsticial heating and (a) easterly QBO forcing (b) westerly QBO forcing.*

Figure 7b is very reminiscent of Dunkerton and O'Sullivan's (1996) figure 4, with steep gradients near the equator and 25N separated by a region of flat isopleths, which they attributed to horizontal mixing. While there are differences between the simulated and observed isopleths which indicate the need for mixing, it is also very likely that there is a circulation anomaly in the real stratosphere similar to that modeled in section 3.2 which is helping to produce the observed shape of the isopleths. Although in their paper they showed the UKMO winds which had only a weak westerly shear, they pointed out that the Singapore wind at the time of the observation they focus on (March 1992) had a strong westerly shear near 10mb, which would be consistent with the observed shape of the isopleths, given the results presented here. Figure 7a is also very similar to the easterly QBO case shown in Plate 1b of O'Sullivan and Dunkerton (1997).

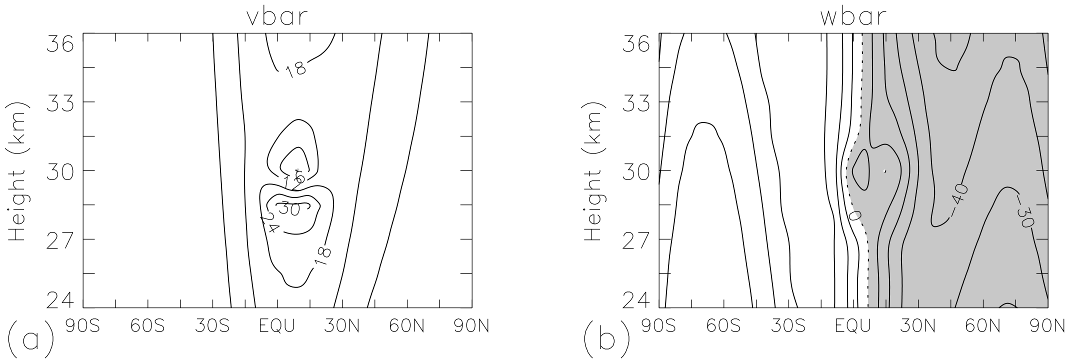

Figure 8: *(a) meridional (in km/day) and (b) vertical (in m/day) velocities from run with solsticial heating and westerly QBO forcing. Negative values shaded.*

The flattening of the isopleths in figure 7b is due both to the enhanced poleward velocity below the level of westerly shear (figure 8a) which stretches the isopleths poleward, and also to the anomaly in the vertical velocity near 15N (see figure 6b), which opposes the downwelling of the mean circulation in winter causing a convergence of the total vertical velocity (figure 8b) which pushes the isopleths together vertically.

This is however just a very simple experiment to demonstrate the effect that the solstice QBO circulation anomaly (which is about 3 times stronger than that at equinox) may have on a tracer distribution.

5 Mechanisms for seasonal asymmetry

Although the model produces a strongly seasonal QBO circulation, we must of course understand the cause of it if we are to decide whether it may represent the behaviour of the real stratosphere. The above seasonality was shown to occur in the model after 60 days when it had reached an almost steady state (in the model, the transition from a state of rest to an almost steady state takes about 40 days in the winter hemisphere, while in the summer hemisphere the easterlies get slowly but progressively stronger since there is no Rayleigh friction to balance the advection of angular momentum). However, the real stratosphere never reaches a steady state due to the seasonal cycle in forcing, and also to the downward movement of the region in which the waves which drive the QBO deposit their momentum. Therefore it is important to investigate whether the asymmetry in the anomalous circulation in the model exists also during the approach to steady state.

5.1 At steady state

At steady state the stream function ψ and hence the horizontal velocity v (using Eq.(6)) are determined by the vertical velocity in isentropic coordinates w (setting $p_t = 0$ in Eq.(7)), which is in turn determined by the temperature (Eqs.(5) and (9)). That these steady-state approximations hold in the model after 60 days is confirmed by the similarity of the temperature anomaly (figure 9a) to the vertical velocity anomaly (figure 6b).

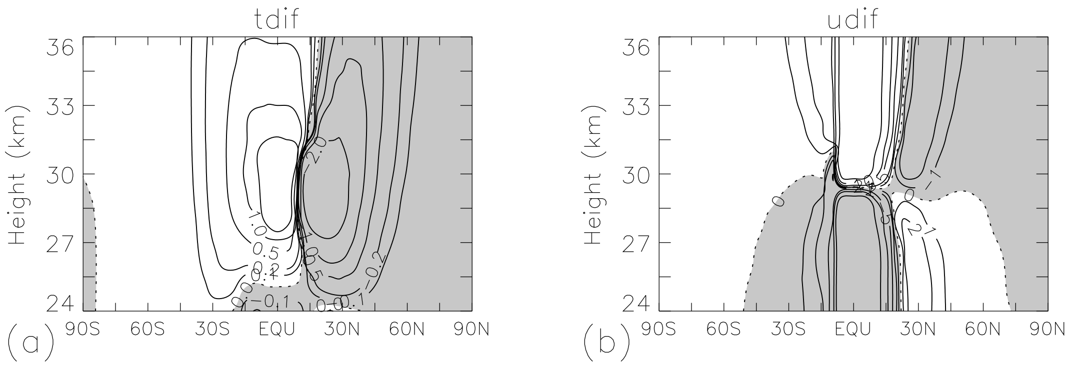

Figure 9: *(a) Temperature anomaly (westerly minus easterly) (contours at 0.1, 0.2, 0.5, 1, 2, 5K) and (b) zonal wind anomaly (contours at 1, 2, 5, 10, 20, 50 m/s) from run with solsticial heating. Negative values shaded.*

However, the temperature anomaly is linked to the zonal wind anomaly (figure 9b), since thermal wind balance (Eqn.(4)) is maintained in the model. Though the two anomalies are linked, it will be argued that the zonal wind anomaly is the major factor determining the stream function anomaly, since there is a constraint placed on zonal wind at steady state (the constraint of zero angular momentum gradient, which will be discussed in this section). Because the temperature gradient is approximately proportional to $sin\phi$ times the zonal wind gradient (from Eq.(4)) the small shift of the zonal wind anomaly into the winter hemisphere by the mean circulation amplifies its effect on the temperature anomaly and hence produces the large temperature anomaly near 30N. Note also that the reversal of the sign of the vertical gradient of the zonal wind anomaly between 15N and 30N is consistent with the polewards decay of the negative temperature anomaly north of 20N, whereas the lack of a sign reversal in the summer hemisphere zonal wind gradient is consistent with the polewards decay of the equatorial positive temperature anomaly in that hemisphere. The sign change in the winter hemisphere and its absence in the summer hemisphere are both due mainly to the advection of planetary angular momentum by the anomaly in the horizontal velocity, as can be seen from figure 6a. For example, above 30km the southward velocity anomaly in figure 6a is the cause of both the negative zonal wind anomaly north of about 25N and the positive anomaly south of 10S in figure 9b. This will be demonstrated in the second experiment below.

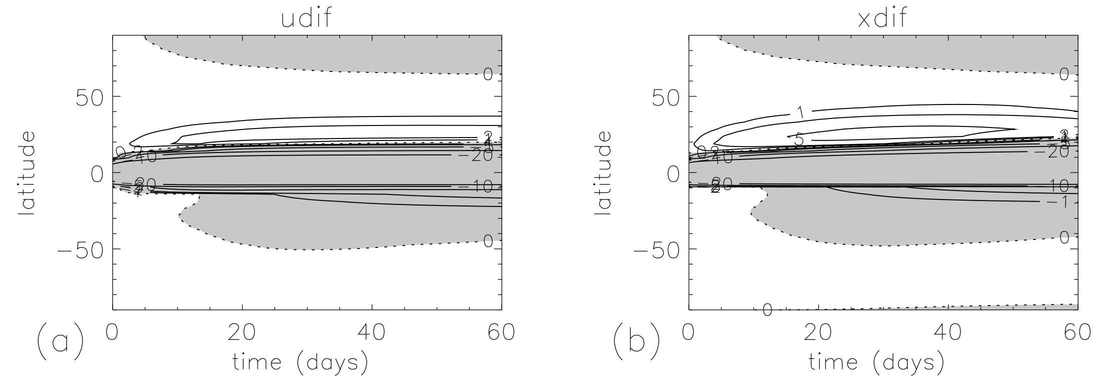

Figure 10: *(a) model zonal wind anomaly (westerly minus easterly) near 27km (contours as figure 9b) and (b) zonal wind anomaly derived from angular-momentum tracer. Negative values shaded.*

To confirm that the zonal wind anomaly is produced by the anomalous advection of angular momentum (and is not due to the inertial stability fluxes or some other cause) a tracer was initialised with the initial value for angular momentum in the model (corresponding to zero zonal wind) and the solstice runs of section 3.2 were performed again. During the course of the runs the tracer was constrained within 5 degrees of the equator (which is where the QBO forcing was applied) to equal the modeled angular momentum there. In figure 10 the anomalies, just below the shear zone, in the modeled zonal wind and in the zonal wind derived from the angular-momentum tracer can be compared. They are very similar within about 15 degrees of the equator in both hemispheres, both showing the northward advection of the negative equatorial anomaly. North of 15N the zonal wind anomaly is due mainly to the poleward advection of planetary angular momentum by the anomalous horizontal velocity. The tracer-derived anomaly extends farther poleward and is larger than the modeled zonal wind anomaly. This difference is due to the frictional damping of the modeled zonal wind anomaly.

The above tracer included the effects of both the anomalous advection of anomalous angular momentum and the anomalous advection of planetary angular momentum. In order to quantify the importance of the anomalous advection of planetary angular momentum, the above runs were performed again, but this time the tracer was constrained at the equator to be equal to the planetary angular momentum there, so that any anomaly in the tracer-derived zonal wind will be due solely to an anomaly in the horizontal velocity. As can be seen from figure 11, below the shear zone this tracer-derived anomaly is positive north of the equator (where, from figure 6a, the anomaly in the horizontal velocity is large and poleward) and negative but smaller south of the equator (where the anomaly in the horizontal velocity is equatorward).

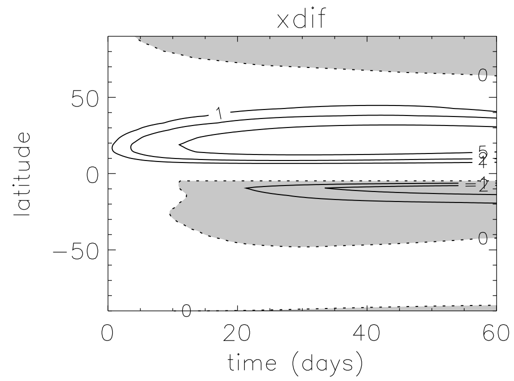

Figure 11: *As figure 10b, but zonal wind anomaly derived from angular-momentum tracer of second experiment.*

To summarise, it appears that the zonal wind anomaly between about 5N and 15N in figure 9 is due to the northward advection of the forced equatorial anomaly, while the anomalies north of 15N and south of 5S are due mainly to the anomalous advection (by the meridional velocity anomaly of figure 6a) of planetary angular momentum.

A steady state is reached in the winter hemisphere in two ways. Where there is friction (which in the runs described above is where the zonal wind is westerly) the zonal wind adjusts itself till friction balances the advection of angular momentum (see Eq.(1)). Where there is no friction (in the above runs, near the equator where the zonal wind is easterly) the constraint on the zonal wind is approximately (ignoring vertical advection) that the horizontal angular momentum gradient, G in Eq.(1), be zero. This is similar to the angular–momentum–conserving circulations studied by Held and Hou (1980), among others, except that here there is a source of angular momentum near the equator (the QBO forcing term) and there is friction on westerly zonal winds. The angular-momentum-conserving circulation is therefore not a closed cell but merely extends the equatorial angular momentum (when the equatorial winds are easterly) into the low latitudes of the winter hemisphere. For the runs of section 3.2 the angular momentum gradient is approximately zero from the equator to 15N (figure 12, solid line) when the equatorial wind is easterly, but is negative north of 5N when the equatorial wind is westerly (dotted line).

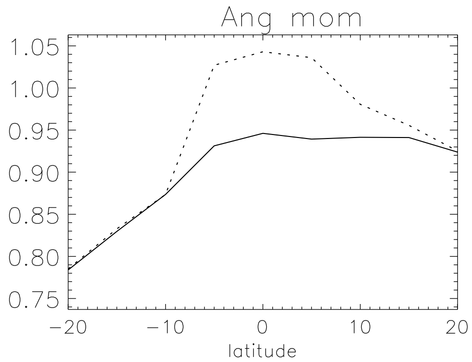

Figure 12: *Angular momentum (normalised by the equatorial planetary angular momentum) near 30km from solstice run with (solid line) easterly forcing and (dotted line) westerly forcing.*

An advantage in studying the case where the equatorial angular momentum is held constant (in this case by relaxing towards the specified QBO profile) is that advection into the winter hemisphere does not create inertially unstable regions (as would be the case if the angular momentum of the summer hemisphere were advected into the winter hemisphere), but merely inertially neutral (with zero angular momentum gradient) regions. The fluxes invoked in the model to maintain inertial stability (see section 2) are therefore redundant when there is a forcing of the QBO. This will be representative of the real atmosphere if the waves driving the equatorial QBO are strong enough to maintain the QBO zonal wind anomaly in the presence of cross-equatorial advection of angular momentum. It is doubtful that this would hold true in the real atmosphere away from the shear zone where the wave driving is presumably weak. However, this does not affect the main conclusions of this paper (see discussion of figure 17).

5.2 During the transition to steady state.

In the previous subsection it was shown that, at steady–state, the temperature anomaly related to the seasonally-asymmetric anomaly in the zonal wind gave rise to a seasonally-asymmetric circulation anomaly. If the asymmetry in the zonal wind anomaly is due to cross-equatorial advection (as will be shown in this subsection) then it is relevant to ask whether there is an asymmetry in the circulation before the zonal wind anomaly has been advected significantly into the winter hemisphere. To address this question the set of runs of section 3.2 were performed again, but the solstice heating was not turned on until day 20 while the QBO forcing was present from day 1 so that on day 21 there were almost symmetric zonal wind and temperature anomalies (figure 13a,b).

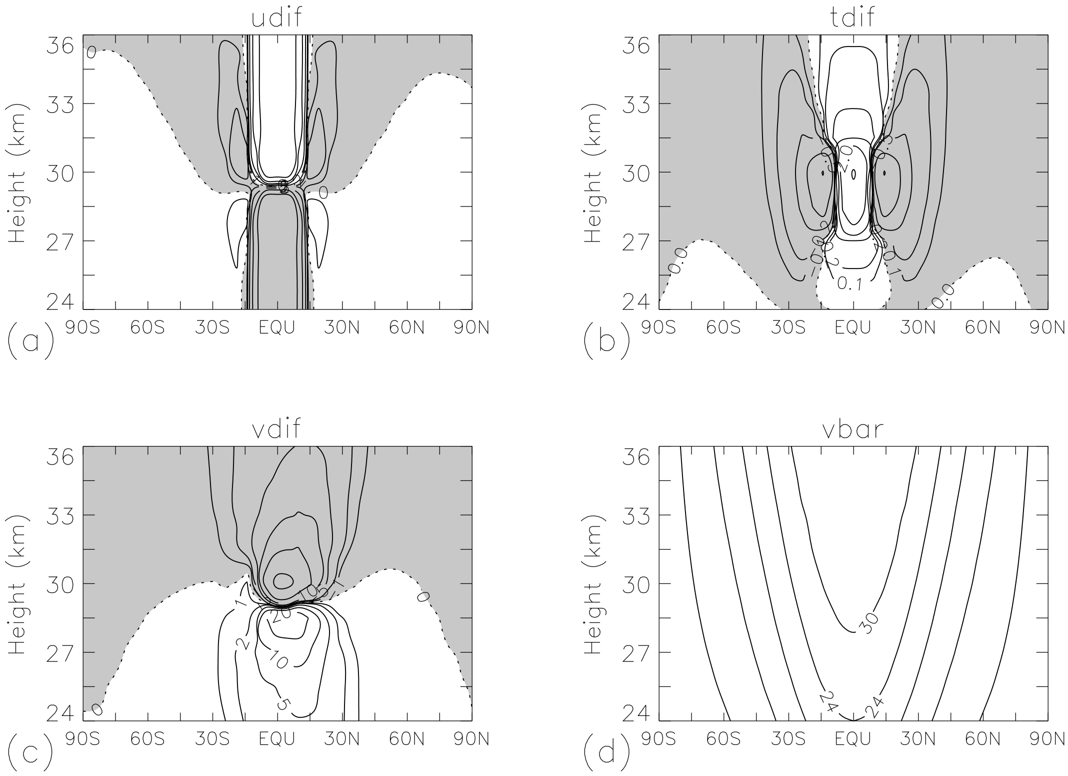

Figure 13: *(a) zonal wind anomaly (westerly minus easterly), contours as figure 9b, (b) temperature anomaly (contours as figure 9a) (c) meridional velocity anomaly (contours at 1, 2, 5, 10, 20, 50 km/day) and (d) meridional velocity with no QBO forcing (contours at 6 km/day intervals) on day 21 of runs described in section 5.2. Negative values shaded.*

If the circulation anomaly were determined by the zonal wind anomaly then it would also be almost symmetric about the equator on day 21, but in fact (figure 13c) it is already strongly asymmetric.

To understand the behaviour of the circulation anomaly on day 21 we must consider the terms in Eq.(7), which may be written using Eq.(8) as

$$w^p = \frac{dz^*}{dz}\left(w - \frac{p_t}{g\rho}\right) . \tag{12}$$

Since the zonal wind anomaly is initially symmetric and evolves slowly towards an asymmetric steady–state distribution, w must behave similarly (since w is directly related to zonal wind through Eqs.(5), (8) and (4)), so that w cannot be the source of the initial asymmetry. However, the last term in Eq.(12), proportional to p_t, will be related to T_t and u_t through Eqs.(3) and (4). Since there is a strong northward circulation (figure 13d) on day 21 and since the zonal wind and temperature anomalies are almost symmetric then u_t and T_t and hence p_t will be asymmetric. This is the source of the asymmetry in the anomalies in w^p and v (figure 13c).

An alternative way of looking at it is to note that advection of the QBO zonal wind anomaly into the winter hemisphere gives an asymmetric u_t which requires an asymmetric T_t to maintain thermal wind balance. This asymmetric T_t must be produced by an asymmetry in w^p and v. As time progresses, the asymmetric T_t produces an asymmetric T and hence an asymmetric w which dominates the right hand side of Eq.(12) as p_t shrinks to zero.

Therefore, initially the asymmetry in w^p is due to the asymmetrical rate-of-change of T while near steady state the asymmetry in w^p is due to the asymmetry in T. Therefore although the asymmetry in the circulation anomaly initially and at steady–state is due to two different mechanisms the source of the asymmetry in both cases is cross-equatorial advection.

6 Effect of friction

From the previous discussion it can be understood that friction at low latitudes will prevent the equatorial angular momentum from being conserved as it is advected polewards and hence will reduce the size of the zonal wind anomaly at low latitudes. However at latitudes poleward of about 15 degrees friction causes a spreading of the QBO circulation anomaly polewards. To demonstrate this, a run with solsticial heating but no friction on the zonal wind (except in the top and bottom layers as described in section 2) was carried out. Without the QBO, this run results in an angular-momentum conserving circulation similar to those studied by Lindzen and Hou (1988) and Dunkerton (1989) and produces the zonal wind of figure 14a after 60 days.

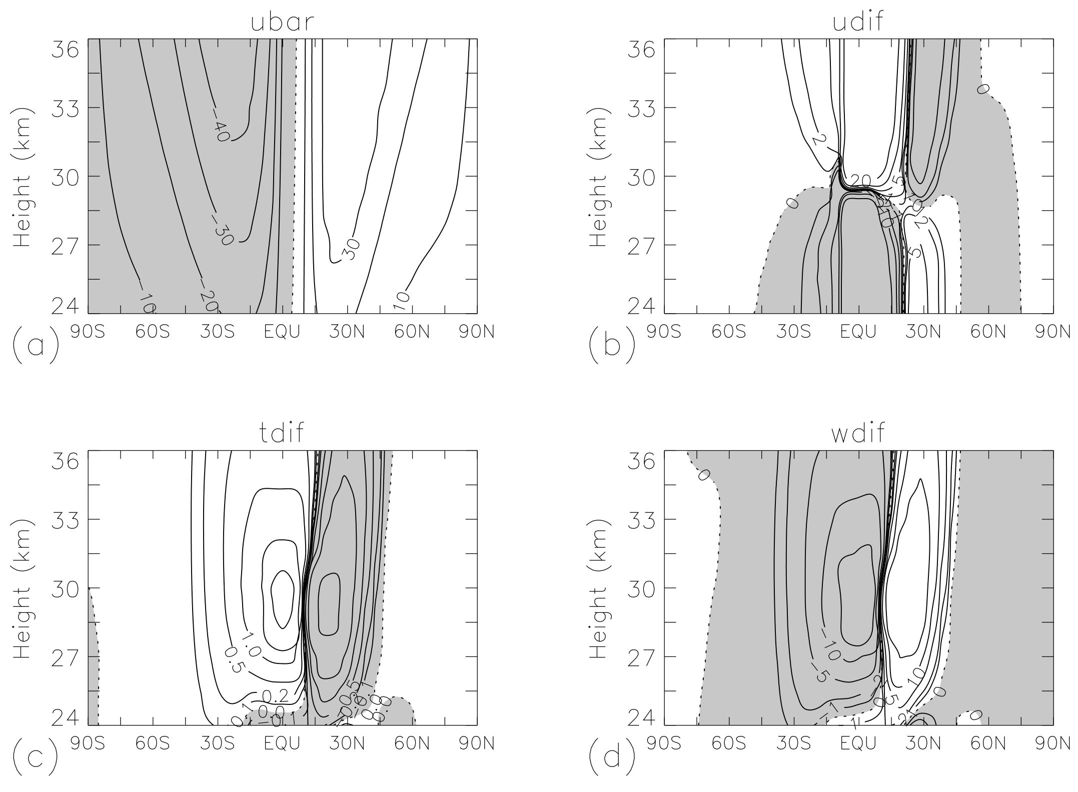

Figure 14: *(a) zonal wind (b) zonal wind anomaly (c) temperature anomaly and (d) vertical velocity anomaly from model with no friction. Negative values shaded.*

The flat angular momentum gradient at low northern latitudes implies in the rapid growth with latitude of westerlies from the equator to about 15N, after which the zonal wind tends towards the value required by thermal equilibrium. The model does not, however, reach a steady state and crashes on day 80 (if allowed to run past day 60). The zonal wind anomaly produced by these no-friction runs (figure 14b) is very similar to that produced with friction (figure 9b) but in the winter hemisphere the anomalies north of about 20N are stronger and narrower than with friction. It seems then that friction damps the anomaly in zonal wind and also in its vertical gradient. Since thermal wind balance is maintained, this implies a weaker horizontal gradient in the temperature anomaly. The temperature anomaly near the equator and at low winter latitudes (figure 14c) is not greatly different from the run with friction (figure 9a) so that the weaker temperature gradient in the winter hemisphere of the frictional run implies a poleward spreading of the temperature anomaly, as is evident from figure 9a. The poleward spreading of the temperature anomaly results in a similar spreading in the heating, vertical velocity (compare figures 14d and 6b), horizontal and hence zonal wind (compare figures 14b and 9b) anomalies.

The other way in which friction leads to a wider circulation anomaly is simply by increasing the meridional flow and hence the advection of the equatorial zonal wind anomaly into the winter hemisphere.

The seasonal asymmetry therefore exists without friction, but is confined to low latitudes. Friction in the winter hemisphere extends the winter anomaly polewards (while hardly affecting the summer anomaly – compare figures 14d and 6b). Note that all that is required of the friction is that it damp out the zonal wind anomaly. If breaking planetary waves or gravity waves are able to do this in the real atmosphere with a time constant of 20 days or less, then this model may be representative of the behaviour of the real atmosphere. However, even if they do not, but also do not damp out the near-equatorial anomaly, then the model predicts a strongly seasonal QBO anomaly within about 30 degrees of equator, with the winter circulation anomaly being much stronger than at equinox. This is, by itself, very different from the "traditional" picture.

In section 3 it was shown that a cross-equatorial mean flow induced by a latitudinal heating gradient gave rise to an asymmetric QBO circulation anomaly, and the mechanism was explained in section 5. However, even with a symmetric equilibrium temperature (such as used for the equinox runs of section 3.1) asymmetric friction on the zonal wind is also able to induce a cross-equatorial mean flow, and hence an asymmetric QBO circulation. To demonstrate this the model was run with the equinox heating of section 3.1 but with the "one–sided" friction applied only in the northern hemisphere. The resulting zonal wind (figure 15a) is, as expected, weaker in the NH than in the SH. The QBO circulation anomaly (figure 15b), although not one–celled as in the solstice experiments, is nevertheless strongly asymmetric, being almost zero south of 30S while still quite large north of 30N. This experiment may be more representative of the stratosphere during equinox when, during the breakup of the winter vortex (in March and October) the spring hemisphere experiences more planetary wave drag than the autumn hemisphere and yet the equilibrium temperature should presumably be the same in both hemispheres.

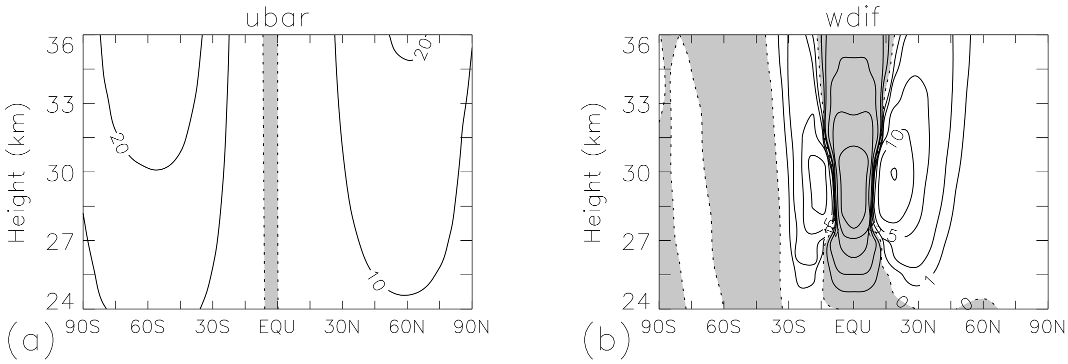

Figure 15: *(a) zonal wind and (b) anomaly in vertical velocity, from run with equinoctal heating and friction in NH only. Negative values shaded.*

Since with the "one-sided" friction used initially easterly winds are not frictionally damped it is worth seeing what effect changing the values of $u_{easterly}$ and $u_{westerly}$ has on the QBO–induced circulation anomaly. Since the westerly QBO zonal wind will be damped at low latitudes while the easterly one will not, the smaller the value of $u_{westerly}$ for a given shear the stronger should be the circulation anomaly. To test this the model was run 25 times more but with $u_{easterly}$ (see section 3) taking the values from -40 to 0 m/s in steps of 10 m/s while the QBO shear ($u_{westerly} - u_{easterly}$) was varied from 10 to 50 m/s in 10 m/s steps. (Note that $u_{westerly}$ is not always westerly, though it is always more westerly than $u_{easterly}$). The graph of the maximum value of the stream-function anomaly (which occurs at about 10N and 30km) plotted against $u_{easterly}$ and the QBO shear (figure 16) shows that when the shear exceeds about 20m/s then the stream-function is more sensitive to variations in $u_{easterly}$ than to variations in shear.

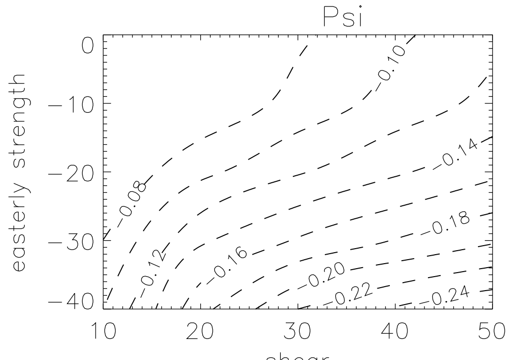

Figure 16: *Maximum value of stream-function anomaly (which occurs near 10N, 30km) as QBO shear and easterly QBO wind strength are varied.*

The usefulness of this result depends strongly of course on the extent to which wave drag in the real atmosphere near the equator is "one-sided" but it does suggest that the vertical shear of the QBO zonal wind may not always be the most important factor in determining the strength of the circulation anomaly (apart, of course, from the strong seasonal dependence demonstrated here).

7 Summary and Discussion.

The model used here has demonstrated that the circulation anomaly induced by the equatorial QBO in zonal wind may depend strongly on season. For a westerly (easterly) QBO shear at equinox there is the usual two-celled circulation anomaly with descent (ascent) over the equator and ascent (descent) in both hemispheres extending about as far as 30 degrees from the equator. For a westerly (easterly) QBO shear at solstice the anomaly is one-celled, with strong ascent (descent) in winter (about 3 times stronger than at equinox) and weak descent (ascent) in summer (about 10 times weaker than in winter). Over the equator the descent (ascent) rate is about the same as at equinox. The experiments in this paper were intended primarily to demonstrate the mechanism by which a seasonal QBO circulation may be generated, rather than to estimate accurately the strength of the circulation. However, the modeled heating rates are generally similar in size to the few estimates derived from observations. For instance, the model produces anomalies in the vertical velocity at 30km of roughly equal and opposite size (about 40 m/day or 1.2 km/month) at the equator and 20 degrees in the winter hemisphere (figure 6b). This compares fairly well with the results of Randel et al. (1998) who found similar equal and opposite anomalies (at the equator and 30 degrees in the winter hemisphere) in the UK Met Office assimilation of about 1.5 km/month at 3mb and about 0.4 km/month at 32mb. Perhaps more interestingly, they found temperature anomalies in the lower stratosphere similar in size and position to those modeled here (about 2K). In addition the equatorial QBO zonal wind anomalies shown in their figure 18 reveal a slight shift towards the winter hemisphere, consistent with the theory outlined here. Note however that an asymmetry in the circulation can exist without there being a large asymmetry in zonal wind or temperature, provided that there is an asymmetric rate of change of temperature and zonal wind, as described in section 5.

The seasonal dependence in the modeled QBO circulation anomaly stems from the cross-equatorial advection of the equatorial QBO anomaly in zonal wind. Since thermal wind balance is maintained in the model a polewards shift of a zonal-wind gradient anomaly implies an increase in the size of the temperature gradient anomaly required to balance it. At steady state the large temperature anomaly in the winter hemisphere leads to a large heating anomaly and hence a cross-equatorial velocity anomaly, which advects the easterly angular momentum anomaly strongly into the winter hemisphere. This leads to a flat gradient of angular momentum in the winter hemisphere near the easterly QBO anomaly, since the equatorial angular momentum is held constant by the forcing of the QBO in the model. This flat angular momentum gradient implies a closeness to inertial instability and perhaps increases the likelyhood of strong horizontal mixing.

However, even before a steady state has been reached and before the equatorial anomaly has been advected far into the winter hemisphere the circulation anomaly is still strongly seasonal because of the asymmetric rate of change of the QBO anomaly. In both cases – at steady state and during the approach to steady state – the seasonal asymmetry in the QBO circulation anomaly is due to cross–equatorial advection by the mean circulation. A cross–equatorial mean circulation (and hence an asymmetric QBO circulation anomaly) can be induced either by asymmetry in the equilibrium temperature (as in section 3.2) or by asymmetry in friction on the zonal wind (as in section 6). Asymmetrical friction may be representative of the wave drag during equinox (if there is more planetary wave breaking in the spring hemisphere) so that the traditional picture of a symmetric QBO circulation anomaly might in fact never occur in the real atmosphere.

Note that the effect of the anomalous QBO circulation at solstice is to advect the easterlies more strongly into the winter hemisphere (figure 8a shows the horizontal velocity for a westerly shear), and the westerlies less strongly. This implies that the equatorial region is more likely to preserve a trace gas mixing ratio (such as that of water vapour) in the region of the QBO where the winds are westerly while in the easterly winds the trace gas will be more strongly blown off the equator. This may have implications for the interpretation of the "tape recorder" signal of water vapour (Mote et al. 1996) in that the cross equatorial velocity may not be negligible when the QBO winds are easterly.

The effect of extra-tropical planetary waves is crudely represented here by the "one–sided" friction (similar to that used by Dunkerton 1989) which acts only on westerly winds. If "one–sided" friction is a fairly realistic description of planetary waves near the equator (though as yet there is little justification for such an assumption) then we would expect the westerly near–equatorial zonal wind QBO anomaly to be more strongly damped than the easterly one, and hence for the QBO circulation anomaly *for a given vertical shear* to be stronger for a weaker westerly phase (as illustrated by figure 16). If observations support the results of figure 16 then this would be a novel way of deducing the dissipation of near-equatorial zonal winds.

Strong seasonality of the QBO circulation exists even in the absence of friction on the zonal wind (see figure 14d), due to the polewards advection of equatorial angular momentum by an angular momentum conserving type of circulation (such as studied by Dunkerton 1989). The anomaly is, however, confined to the tropics (though it is still much stronger in winter than at equinox). If there is moderate (20 day) friction on the winter zonal wind the model demonstrates that the QBO circulation anomaly is extended well into winter middle latitudes (figure 6b). Although such friction may be a parametrisation of the effects of breaking planetary waves or gravity waves in the real stratosphere, it is important to note that this type of planetary-wave interaction is completely distinct from the previously proposed interactions of planetary waves with the equatorial QBO (involving for example critical lines or restricted waveguides), since planetary waves are required here to damp an anomaly rather than to create one.

Where might we expect the strongest seasonal asymmetry in the QBO circulation? Since the degree of seasonal asymmetry in the QBO anomaly depends on the strength of the cross–equatorial mean flow, which in the real stratosphere tends to increase with height due to the tendency of gravity waves to break in the upper stratosphere and mesosphere, the strongest asymmetry may occur in the upper half of the stratosphere. The asymmetry also depends on the strength of the QBO in zonal wind, which is strongest near 10mb (though as suggested by the work of Dunkerton and Delisi (1997) it is possible that the maximum may lie higher up). The equatorial layer that has the largest effect on the asymmetric QBO circulation may therefore lie near 10 or 20 mb, although the winter circulation anomaly itself may extend further down into the stratosphere by a few km (see figure 6b), where it will have a larger effect on the ozone column. It is interesting to note that there is a strong correlation of the winter ozone column between about 15 and 45 degrees in both hemispheres with the Singapore zonal wind near about 15mb (eg. Randel and Wu 1996, Kinnersley and Tung 1998) so it may be that at about 15mb the zonal wind QBO is large enough and the cross–equatorial flow is strong enough for the above mechanism to affect the ozone column in the real stratosphere. Below about 30mb observations of the persistence of the "equatorial tape–recorder" signal (Mote et al. 1996) appear to imply little, if any, cross–equatorial flow. However, since the flow would reverse direction approximately every six months, a weak cross-equatorial flow could exist without disrupting the tape-recorder signal in water vapour, though it would cause it to shift slightly into the winter hemisphere. Such a shift in the QBO zonal wind anomaly could have a significant effect on the anomaly in the meridional circulation anomaly, as has just been shown.

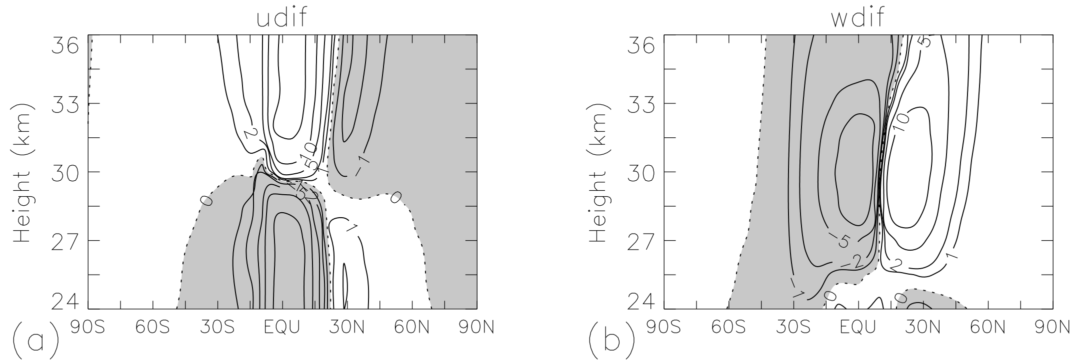

Figure 17: *(a) zonal wind anomaly (westerly minus easterly) and (b) anomaly in vertical velocity, from run with QBO forcing switched off after day 20. Negative values shaded.*

The model used for this study has idealised heating rates and parametrised friction, and a simple mechanistic forcing of the QBO in zonal wind which constrained the equatorial QBO winds over the whole depth of the model stratosphere. The QBO forcing in the real stratosphere is large only in the shear zone. As a simple test of the importance of this difference, the model was run again as in section 3.2, but the QBO forcing was switched off after day 20. The resulting anomalies in zonal wind and vertical velocity (figure 17) are smaller than in that experiment (figures 9b and 6b), although qualitatively very similar. Therefore, although it is very likely that this mechanism exists in the real stratosphere it will no doubt differ in its details from the simple experiments described here.

Acknowledgments. My thanks to K.K. Tung for valuable discussions and comments, and to Tim Dunkerton, Leslie Gray and Bill Randel for helpful reviews. This work was supported by the National Science Foundation through grant ATM-9526136.

References

Angell J.K. and J. Korshover, 1964: Quasi-biennial variations in temperature, total ozone and tropopause height. *J. Atmos. Sci.*, **21**, 479–492.

Baldwin, M.P. and K.K. Tung 1994: Extra-tropical QBO signals in angular momentum and wave forcing. *Geophys. Res. Lett.*, **21**, 2717–2720.

Dunkerton, T.J. 1985: A two-dimensional model of the Quasi-biennial oscillation. *J. Atmos. Sci.*, **42**, 1151–1160.

Dunkerton, T.J. 1989: Nonlinear Hadley circulation driven by asymmetric differential heating. *J. Atmos. Sci.*, **46**, 956–974.

Dunkerton, T.J. and M.P. Baldwin 1991: Quasi-biennial modulation of planetary-wave fluxes in the northern hemisphere winter. *J. Atmos. Sci.*, **48**, 1043–1061.

Dunkerton, T.J. and D.J. O'Sullivan 1996: Mixing zone in the tropical stratosphere above 10mb. *Geophys. Res. Lett.*, **23**, 2497–2500.

Dunkerton, T.J. and D.P. Delisi 1997: Interaction of the quasi-biennial oscillation and stratopause semiannual oscillation. *J. Geophys. Res.*, **102**, 26107–26116.

Gray L.J. and J.A. Pyle 1989: A two-dimensional model of the Quasi-biennial oscillation of ozone. *J. Atmos. Sci.*, **46**, 203–220.

Gray, L.J. and T.J. Dunkerton, 1990: The role of the seasonal cycle in the quasi-biennial oscillation in ozone. *J. Atmos. Sci.*, **47**, 2429–2451.

Gray, L.J. and S. Ruth, 1993: The modeled latitudinal distribution of the ozone quasi-biennial oscillation using observed equatorial winds. *J. Atmos. Sci.*, **50**, 1033–1046.

Hasebe F. 1984: The global structure of the total ozone fluctuations observed on the time scales of two to several years. *Dynamics of the Middle Atmosphere*, J.R. Holton and T. Matsuno, Eds., D.Reidel, 445–464.

Holton J.R. 1989: Influence of the annual cycle in meridional transport on the Quasi-biennial Oscillation in total ozone. *J. Atmos. Sci.*, **46**, 1434–1439.

Holton J.R., and R.S. Lindzen, 1972: An updated theory for the quasi-biennial cycle of the tropical stratosphere. *J. Atmos. Sci.* **36**, 1076–1080.

Holton J.R. and H.-C. Tan 1982: The Quasi-biennial oscillation in the northern hemisphere lower stratosphere. *J. Meteor. Soc. Japan*, **60** 140–147.

Jones, D.B.A., H.R. Schneider and M.B. McElroy 1998: Effects of the Quasi-Biennial Oscillation on the zonally averaged transport of tracers. *J. Geophys. Res., in press.*

Kinnersley, J.S., 1996: The Climatology of the stratospheric 'THIN AIR' model. *Quart. J. R. Meteorol. Soc.*, **122**, 219–252.

Kinnersley, J.S. and K.K. Tung, 1998: Modeling the global inter-annual variability of ozone due to the equatorial QBO and to extra-tropical planetary wave variability. *J. Atmos. Sci.* in press.

Kinnersley, J.S. and K.K. Tung, 1999: Mechanisms for the extra-tropical QBO in circulation and ozone column. Submitted to *J. Geophys. Res.*

Mote, P.W., K.H. Rosenlof, M.E. McIntyre, E.S. Carr, J.C. Gille, J.R. Holton, J.S. Kinnersley, H.C. Pumphrey, J.M. Russel III and J.W. Waters (1996), An atmospheric tape recorder: The imprint of tropical tropopause temperatures on stratospheric water vapor. *J. Geophys. Res.* Vol. 101, pp 3989-4006.

O'Sullivan, D. and T.J. Dunkerton 1997: The influence of the quasi-biennial oscillation on global constituent distributions. *J. Geophys. Res*, **102**, 21731–21743.

Plumb R.A. and R.C. Bell, 1982: A model of the quasi-biennial oscillation on an equatorial beta plane. *Quart. J. Roy. Meteor. Soc.*, **108**, 335–352.

Randel W.J. and J.B. Cobb 1994: Coherent variations of monthly mean total ozone and lower stratospheric temperature. *J. Geophys. Res.*, **99**, 5433–5447.

Randel, W.J. and F. Wu 1996: Isolation of the ozone QBO in SAGE II data by singular–value decomposition. *J. Atmos. Sci.*, **53**, 2546–2559.

Randel, W.J., F. Wu, R. Swinbank, J. Nash and A. O'Neill, 1998: Global QBO circulation derived from UKMO stratospheric analyses. *J. Atmos. Sci., in press.*

Reed, R.J., 1960: The structure and dynamics of the 26-month oscillation. *Fortieth Anniversary Meeting* Amer. Meteor. Soc.

Tung K.K, and H. Yang 1994: Global QBO in circulation and ozone. Part II: A simple mechanistic model. *J. Atmos. Sci.*, **51** 2708–2721.