Modeling the Quasi-biennial Oscillation's Effect on the Winter Stratospheric Circulation

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ABSTRACT
The influence of the equatorial quasi-biennial oscillation (QBO) on the winter middle atmosphere is modeled with a mechanistic global primitive equation model. The model's polar vortex evolution is sensitive to the lower stratosphere's tropical winds, with the polar vortex becoming more (less) disturbed as the lower stratospheric winds are more easterly (westerly). This agrees with the observed relationship between wintertime polar circulation strength and the phase of the QBO in the lower stratosphere. In these experiments it is the extratropical planetary Rossby waves that provide the tropical-extratropical coupling mechanism. More easterly tropical winds in the lower stratosphere act to confine the extratropical Rossby waves farther north and closer to the vortex at the QBO altitudes, weakening the vortex relative to the case of westerly QBO phase. While the QBO winds occur in the lower stratosphere, the anomaly in the polar vortex strength is strongest at higher levels.

1. Introduction
The Northern Hemisphere winter stratospheric polar vortex demonstrates considerable interannual variability in its time-mean circulation strength, with a significant fraction of its interannual variability correlated with the phase of the equatorial stratosphere's quasi-biennial oscillation (QBO). As noted by Holton and Tan (1980, 1982), there is a strong correlation between the wintertime average polar vortex strength and the phase of the QBO at 50 mb. By compositing winters according to the 50-mb QBO phase, they showed that during winters when the QBO's phase was westerly at 50 mb the polar vortex was stronger and zonal wavenumber one weaker than was the case in easterly QBO phase winters. The monthly or seasonally averaged extratropical QBO composite anomaly (westerly years minus easterly years) shows a meridional dipole pattern in the zonal wind field, having a westerly wind anomaly near 65°N with a weaker easterly anomaly around 30°N. Labitzke (1982) analyzed polar temperature data at the 30- mb level and pointed out a tendency for polar warmings (both early winter “Canadian warmings” and midwinter “major warmings”) to occur most often when the QBO was easterly at 50 mb. This tendency is consistent with the time-average extratropical anomaly seen by Holton and Tan, since a weaker vortex implies a warmer polar temperature in the stratosphere by thermal wind balance. More recently, Dunkerton and Baldwin (1990) analyzed NMC data from 1964 to 1988 and confirmed the continued existence of the Holton and Tan extratropical dipole anomaly pattern in the 1980s. They found a strong QBO modulation of the EP flux of wave activity whose pattern was consistent with the observed wind anomaly. They also showed the strengthening of the QBO dipole anomaly as winter progressed until early January, suggesting the cumulative nature of the extratropical coupling mechanism.

In this study we examine the possibility that the winter polar vortex's QBO-related interannual variability arises from the modification of the Rossby wave propagation in the stratosphere by the equatorial QBO winds. This mechanism was suggested by Holton and Tan (1980, 1982) and McIntyre (1982). We do not examine tropospheric sources (e.g., the Southern Oscillation) for interannual variability of the winter stratosphere here.

Depending on the QBO's phase, stratospheric equatorial winds may range from strong zonal easterlies (~30 m s⁻¹) to westerlies (~15 m s⁻¹). The presence of a deep layer of easterlies or westerlies may modify the propagation of Rossby waves refracting equatorwards from the extratropics, so it is possible that the extratropical Rossby wave field is modified according to the QBO phase. This would be most likely to occur for wave amplitudes sufficiently strong to develop a nonlinear critical layer. Such nonlinear critical layers are regions of efficient quasi-horizontal stirring of air.

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lying to the north of the zero wind line's location, and that widen with increasing wave forcing. Thus, the development of an extratropical anomaly via modulation of Rossby wave activity by the QBO depends on (i) the easterly tropical winds confining the Rossby waves northwards, closer to the vortex than is the case when the tropical winds are westerly, that is, the critical layer forms farther north during easterly phase winters than during westerly phase winters; and (ii) the waves having sufficient amplitude to develop a nonlinear critical layer of significant width so that the wave fields at latitudes significantly to the north of the critical wind line are influenced.

Our results support the importance of nonlinear wave dynamics, since in the case of dissipative linear wave propagation (with absorbing critical lines), Rossby wave activity propagates from the troposphere into the stratosphere and refracts equatorwards where it is absorbed at the critical surface. Dissipative linear wave theory predicts that the region where the QBO can influence wave driving of the mean flow is confined to the low latitudes where the QBO determines the location of the zero wind surface. As will be seen, no QBO modulation of linear wave driving occurs north of \(\sim 40^\circ\)N. For this to occur nonlinear wave dynamics is required.

Observations supporting QBO modulation of extratropical Rossby waves were presented by Dunkerton and Baldwin (1990) and Baldwin and Dunkerton (1991), who noted quasi-biennial modulation of extratropical planetary wave Eliassen–Palm fluxes consistent with the extratropical QBO in zonal-mean wind. Relative to the QBO's phase at 40 mb they noted that the EP flux convergence was stronger for more easterly rather than westerly QBO phase winters. Furthermore, Hamilton (1989) used station observations of total ozone in the subtropics to infer that Rossby wave transport in the northern subtropics (15° to 25°N) is modulated by the QBO's phase.

Extratropical Rossby waves also play a major role in determining the global ozone distribution. Thus, QBO modulation of Rossby wave transport should generate an extratropical QBO in ozone during wintertime, as amplified meridional exchange of air will lead to a stronger poleward flux of ozone when the QBO phase is easterly in the lower stratosphere. An extratropical QBO in ozone has been observed (Hasebe 1983; Garcia and Solomon 1987; Bowman 1989). Lait et al. (1989) found that the rate of decline of ozone in the Antarctic ozone hole was correlated with the QBO phase, and furthermore, that eddy activity between 40° and 70°S correlated well with the QBO phase.

The work reported in this paper is an extension of an earlier study of such tropical–extratropical coupling (O'Sullivan and Salby 1990) that used an equivalent barotropic framework to model the vertically averaged circulation of the lower stratosphere above 30 mb. In that study it was found that the equatorial QBO did strongly influence the polar vortex strength in agreement with the sense of the observed tropical–extratropical coupling, with stronger (weaker) polar warmings occurring if the QBO phase was easterly (westerly). Such a framework is appropriate for studying the tropical–extratropical coupling by Rossby wave transport since it successfully represents quasi-horizontal nonlinear critical layer development at low latitudes and transport by eddies. The modulation of the time-mean polar vortex strength is a result of the QBO's modulation of eddy transport of potential vorticity.

Here we present results of 3D simulations of the extratropical winter stratosphere's dependence on the QBO using a global primitive equation stratospheric model. As in O'Sullivan and Salby, we use the technique of matched pairs of integrations, differing only by the sign of the tropical QBO winds, to determine the effect of the QBO on the winter extratropical circulation. We again find that the polar vortex is more disturbed when the QBO wind is easterly rather than westerly and that the tropical–extratropical coupling mechanism involves a QBO modulation of Rossby wave propagation, providing the forcing is of realistic strength. We examine both the coupling mechanism and the resulting extratropical anomaly in detail.

Previous modeling studies of the QBO's extratropical influence have been along similar lines. These include Bridger (1984), Damiris and Ebel (1990), Holton and Austin (1991), and Kodera et al. (1991). Except for the latter study these also used mechanistic stratospheric models. Kodera et al. used a general circulation model including the troposphere. Excepting Bridger and Kodera et al., these studies indicate a tendency for polar warmings to occur when the lower-stratospheric QBO phase is easterly. Bridger's integrations may have been of too short duration and have had excessive Rossby wave forcing, similar to experiment in this study. Kodera et al. examined the extratropical QBO dependence on the solar cycle. While they did not find the extratropical anomaly for average solar heating, they did find an extratropical anomaly consistent with Holton and Tan when they reduced the radiative heating due to ozone absorption of solar UV by \(\sim 20\%\) or more, taken to represent the minimum extreme of the 11-year solar cycle. The purpose of this study is to investigate the extratropical anomaly and, in particular, the tropical–extratropical coupling mechanism responsible. We do not address solar cycle influences here.

In section 2 we describe the numerical model used for this study. The results from a series of matched-pair experiments are presented in section 3. These experiments examine the modulation of extratropical Rossby wave propagation by the QBO at several Rossby wave forcing strengths. We also examine the sensitivity of the extratropical anomaly to the width and altitude of the specified QBO wind. The results are discussed in section 4.
2. Model description

The numerical simulations were performed with a global primitive equation model developed at NASA/Ames Research Center (Young and Villere 1985) modified to a mechanistic middle atmospheric model. The model is a spectral transform model using a log-pressure grid vertically. For horizontal resolution, we retain spherical harmonics with zonal wavenumber \( m \leq 21 \) and total wavenumber \( n \) such that \( m \leq n \leq 42 \). Vertically, there are 20 levels with a grid spacing of 3 km. The lower boundary is set to the 250-mb level and the uppermost level is \( z = 0.03 \) mb (vertical range spans the 10–70-km range approximately). A semi-implicit time integration scheme is used with a 30-minute time step.

The initial atmospheric state used consisted of a global mean equal to the U.S. Standard Atmosphere on which was superimposed an idealized December zonal-mean flow as shown in Fig. 1. (The analytic expression for this flow is given in O’Sullivan and Hitchman 1992.) The atmospheric QBO’s vertical structure consists of layers of westerly and easterly wind that descend with time. Thus, the QBO often consists of easterlies overlying westerlies, or vice versa, and at other phases consists of deep easterly or westerly layers. To simplify interpretation, we specify the QBO wind as a single layer of easterlies or westerlies, approximately 15 km deep with a Gaussian profile in latitude. Thus,

\[
u_{QBO} = \begin{cases} 
\pm 20 \exp \left\{ -\left( \phi/\Delta \phi_{QBO} \right)^2 \right\}, \\
23.5 \leq z \leq 35.5 \text{ km}; \\
0, \quad 38.5 \leq z, \text{ or } z \leq 20.5 \text{ km}.
\end{cases}
\]

where \( \phi \) is latitude and \( \Delta \phi = 15^\circ \). The corresponding geostrophically balanced temperature anomaly is also included in the initial temperature field.

The model atmosphere is integrated forward from this initial state subject to specified Rossby wave forcing at the lower boundary, diabatic heating, and dissipation. At the upper boundary the log-pressure vertical velocity is set to zero. At the lower boundary the geopotential height of the 250-mb surface is specified; the zonal-mean geopotential is in gradient balance with the specified zonal-mean wind and constant with time. To represent Rossby wave forcing the lower boundary geopotential height is specified to be

\[
\Phi_B = \begin{cases} 
 h \cos^2(3\phi) \cos(\lambda) \left\{ 1 - \exp(-t/2.5) \right\}, \\
30^\circ \leq \phi \leq 90^\circ \\
0, \text{ otherwise,}
\end{cases}
\]

where \( h \) is the wave amplitude in meters, \( \lambda \) is longitude, and \( t \) is time in days. Only zonal wavenumber one is used for wave forcing. The absorption of Rossby waves in the middle atmosphere weakens the westerly flow.

This effect is counterbalanced by differential radiative heating acting to cool the polar atmosphere and strengthen the westerly flow. We simplify radiative forcing by assuming Newtonian cooling toward a radiative equilibrium temperature that is in gradient balance with a zonal wind identical to the initial zonal wind but whose extratropical jets are 50% stronger than the initial wind. The Newtonian cooling time scale ranges from approximately 20 days at the tropopause to approximately five days in the upper mesosphere. Since the radiative equilibrium temperature is identical to the initial temperature field in the tropics, no QBO-driven mean meridional circulations are generated in this model. Thus, any QBO tropical-extratropical coupling must be due to wave processes.

A Rayleigh friction sponge layer is also included with
an approximately two-day damping time scale that is effective at the uppermost levels and is applied only to zonal asymmetries. The sponge layer is intended to reduce reflection of wave activity from the model's upper-boundary rigid lid. To represent horizontal mixing due to subgrid-scale processes (gravity waves, small-scale threedimensional turbulence, etc.) a second-order diffusion was applied to the vorticity, divergence, and temperature fields of the form $\nu_2 \nabla^2$ where $\nu_2 = 10^5$ m$^2$ s$^{-1}$. A scale-dependent diffusion of the form $\nu_6 \nabla^6$ was also used to prevent accumulation of enstrophy at smallest scales; $\nu_6$ was chosen to give an e-folding time of 12 minutes at the smallest resolvable scale. The sixth-order diffusion is essential for stability. The second-order diffusion was included to better represent physical processes, though it is of minor importance in the simulations discussed. In integrations having westerly QBO phase specified a westerly, equatorially confined, zonal restoring force was applied to maintain the QBO westerly anomaly against the effects of horizontal diffusion and Rossby wave absorption. Diffusive effects on the specified QBO wind anomaly were most noticeable in the west QBO case because of the strong shear, $\partial u/\partial y$, which develops on the southern flank of the QBO as the southern summer easterlies strengthen (e.g., see experiment a (section 3a)). Comparisons with integrations without this restoring force showed that it did not significantly alter the extratropical wind anomaly.

Six experiments will be discussed here. The first "standard" experiment involves a matched pair of simulations using the basic state and forcings just described. The evolution of both the easterly and westerly QBO cases are presented and the QBO's extratropical effects seen by the difference between these simulations. The Eliassen–Palm (EP) flux, $F$, and its divergence, $\nabla \cdot F$, in this experiment are then compared with that found for linear wave propagation, with dissipative critical lines. To understand the nature of the QBO–extratropical coupling, two sensitivity experiments are discussed: 1) the latitudinal width of the QBO wind anomaly is widened and 2) the altitude of the QBO wind anomaly is elevated. Finally, the sensitivity of the extratropical response to the Rossby wave forcing strength is examined in two further experiments using either stronger or weaker wave forcing levels.

3. QBO experiments

a. Standard experiment

In the standard experiment we use a pair of integrations to determine if the QBO can influence the winter high-latitude circulation as observations suggest. The model is integrated forward for 80 days, starting from the basic zonal-mean state shown in Fig. 1a (easterly QBO case). This integration is then repeated unchanged except for starting from the zonal-mean flow shown in Fig. 1b (westerly QBO case). Differences that develop between these matched integrations must therefore be due to the QBO.

The forcing amplitude used in this pair is given by (1) and (2) with $h = 215$ m and $\Delta \phi_{\text{QBO}} = 15^\circ$. We will briefly describe the 3D evolutions in both the easterly and westerly QBO cases and then examine the differences that develop between them.

Figure 2 shows the zonal mean for both the easterly and westerly cases on day 60, showing that substantial changes have occurred. In the winter hemisphere a polar warming has occurred in the easterly QBO case with polar night westerlies switching to easterlies from the model top to below the stratopause. The westerly QBO case shows the polar jet shifted closer to the pole but no zonal easterlies have developed.

![Fig. 2. Zonal-mean wind on day 60 for easterly QBO (upper), and westerly QBO phase (lower), of the standard QBO experiment a. This shows greater disruption of the polar night jet in the easterly case.](image-url)
Horizontal views of the mesospheric polar vortex show that it is distorted in both cases on day 60, but that in the easterly QBO case it has been displaced farther off the pole than in the westerly QBO case. There is no significant difference in the summer hemisphere, where winds have become more easterly in response to the temperature field relaxing toward its radiative equilibrium. The temporal evolution of the zonal-mean wind at 1 mb is shown in Fig. 3. These plots show the summer hemisphere’s easterlies relax to a steady state in the absence of significant wave activity, whereas the northern winter vortex weakens with time after about day 7. At this altitude zonal-mean easterlies appear shortly after day 50 in the easterly case but about two weeks later in the westerly QBO case. The easterly case shows a regular northward shifting of the westerly jet with time as the vortex is deformed from zonal symmetry. The westerly case shows a similar northward shifting of the jet up until day 40 when a minor warming occurs, without the appearance of mean easterlies. In the two weeks following this minor warming the westerlies recover strength until a second warming episode completes the zonal-mean westerly jet breakdown with easterlies appearing after day 60. The appearance of easterlies at high latitudes is accompanied by westerly acceleration in the subtropics and cooling (not shown) from about 50°N to 30°S.

To simplify the description we focus on the final 30 days of these integrations time averaged, and in particular the time-average difference between the easterly and westerly QBO cases. Figure 4 shows the zonal-mean winds time averaged from day 50 to day 80 showing a weaker polar night jet resulting in the easterly QBO case. Figure 5 shows the difference in time-averaged zonal-mean wind, temperature, and EP flux divergence between the simulations with westerly and easterly QBO winds. The zonal-mean wind difference (Fig. 5a) shows the specified QBO winds in the tropics with its peak value slightly reduced from 40 m s⁻¹ initially due to the effects of horizontal diffusion. Figure 5a shows a large westerly wind anomaly occurring at high winter hemisphere latitudes and extending through the mesosphere and most of the stratosphere. At low latitudes (south of 30°–40°) there is also a deep easterly wind anomaly. The corresponding temperature anomaly (Fig. 5b) is in approximate thermal wind balance with the wind anomaly away from the pole. The tropical anomalies represent the westerly and easterly wind-shear layers bounding the QBO anomaly. The maximum anomaly, however, occurs at the North Pole in the lower stratosphere (~ −11 K). Both above and to the south of this cold polar anomaly is a much weaker positive temperature anomaly.

Comparing the westerly QBO minus easterly QBO case’s wind and temperature anomalies with observations shows general agreement in the extratropics (cf. Figs. 6 and 7 of Dunkerton and Baldwin 1991; note that they show three-month averages, whereas the model figures show 30-day averages). The observed zonal-mean wind anomaly has an unequal dipole structure meridionally, with a positive westerly maximum near 60°N, flanked to the south by a weaker negative anomaly. This dipole anomaly is strongest in the upper stratosphere. The observed QBO (west minus east phase) zonal-mean temperature anomaly is maximum negative at the North Pole at the altitude the QBO composited is based on (40-mb level in Dunkerton and Baldwin). A weaker positive temperature anomaly occurs in the subtropics at a slightly higher level than the polar anomaly. The model shows a similar temperature anomaly structure (Fig. 5b).

In comparing model with observations, the exact altitude of extratropical anomalies should not be expected to match closely since the model’s QBO was arbitrarily centered at 20 mb. Given the simplicity of this mechanistic model we do not expect to be able to accurately reproduce the observed anomaly details. The similarity of the extratropical anomaly found in the model with observations is evidence that the model contains the QBO tropical–extratropical coupling mechanism. Figure 5c shows the westerly QBO minus
easterly QBO EP flux divergence displayed following the graphical convention of Dunkerton et al. (1981). (The quantity contoured is $\Delta = 2\pi a^2 \cos(\phi) \nabla \cdot \mathbf{F}$, and the contour units are to be multiplied by $2\pi a^2 \rho_0 10^{-7}$ m s$^{-2}$, where $\rho_0$ is the U.S. Standard Atmosphere 250-mb density.) Figure 5c shows a negative anomaly in the tropics trailing to the subtropics at higher levels, and a positive anomaly in the extratropics. The negative anomaly represents greater wave propagation into the tropics during the westerly QBO simulation. The stronger positive anomaly indicates a tendency for more wave activity to be injected into the stratosphere during the easterly QBO simulation, as discussed later.

Having obtained an extratropical anomaly in a 3D model that approximately agrees with observations, we next examine the nature of the tropical QBO–extratropical coupling mechanism responsible.

b. Broader QBO latitudinal profile

If the extratropical anomaly develops from the nonlinear interaction of extratropical Rossby waves with low-latitude easterly winds we expect an enhanced extratropical anomaly if the QBO wind profile is widened. In this situation the zonal-mean zero wind line is moved farther north in the westerly QBO case and farther south in the easterly QBO case than in the standard experiment. To examine this possibility the standard QBO experiment is repeated with a broader latitudinal profile by changing $\Delta \phi_{QBO}$ from 15° to 20° in (1). The resulting anomaly averaged from day 50 to day 80 is shown in Fig. 6. This broad QBO experiment develops an extratropical anomaly whose zonal-mean pattern is very similar to the original experiment, except that it is much stronger, almost by a factor of two. The EP flux divergence westerly QBO minus easterly QBO difference has also strengthened, consistent with the Rossby wave coupling interpretation of the extratrop-

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**Fig. 4.** As in Fig. 2 but averaged from day 50 to day 80.

**Fig. 5.** Zonal-mean difference in (a) zonal wind, (b) temperature, (c) EP flux divergence between easterly and westerly QBO integrations, averaged between day 50 and day 80. In plotting the EP flux and its divergence we follow the graphical convention of Dunkerton et al. (1981); see text for details. Contour interval in (c) = 0.1.
tical zonal-mean anomaly. We found that the extratropical anomaly strengthened further if the QBO was made still broader meridionally ($\Delta \phi_{\text{QBO}} = 25^\circ$, not shown), or if the vertical extent of the QBO anomaly was increased from that used in this paper.

c. Linear wave propagation

Is it possible that modification of the zonal-mean wind by the QBO changes the refractive index in the subtropics sufficiently to alter the Rossby wave activity at higher latitudes? We now investigate the effect of the QBO on linear wave propagation by repeating the last integration pair, $b$, but with very small wave forcing, $h = 1$ m. In order to avoid the thermal relaxation to a very strong westerly jet in the absence of significant easterly drag due to Rossby wave absorption, which would complicate comparison with experiment $b$, we change the radiative equilibrium wind from $1.5 \times u_{\text{initial}}$ to $u_{\text{initial}}$. Due to the lack of wave–mean flow transience, the weakly forced wave is already close to steady state by day 20. We therefore continue the integration to day 50 and average over the last 30 days. The day 20 to 50 average of $\mathbf{F}$ and $\nabla \cdot \mathbf{F}$ are shown in Fig. 7 for (a) easterly QBO, (b) westerly QBO, and (c) westerly QBO minus easterly QBO difference of $\nabla \cdot \mathbf{F}$. These show that the wave driving is concentrated near the critical line and there is essentially no extratropical effect of the QBO on wave driving of the mean flow. This agrees with WKB theory, which would suggest that wave activity propagates along ray paths, which mostly terminate at the subtropical zero wind line. Apart from weak thermal damping in the extratropics, the wave activity is completely absorbed at the critical line. Since

FIG. 7. EP flux and its divergence for experiment $c$, $h = 1$ m in Eq. (2), showing linear wave propagation. (a) easterly QBO phase, (b) westerly QBO phase, (c) westerly minus easterly difference in divergence, showing confinement of westerly minus easterly difference in linear wave driving to the tropics. The zonal mean zero wind line is highlighted in (a) and (b). Contour interval in (c) = $10^{-5}$. 
wave activity flows only southwards there is no means of communicating information on the phase of the QBO back to higher latitudes in the linear wave propagation experiment, if the critical lines are absorptive for Rossby waves.

We should mention that the QBO imposed in this model is tropically confined and does not have or generate mean meridional circulations extending to high latitudes as discussed in section 2. Thus, this model is not exactly comparable with the linear model of Chen and Robinson (personal communication). They argued that the small changes in the high-latitude zonal-mean flow accompanying the QBO-driven mean meridional circulations have an important influence on extratropical Rossby wave propagation.

d. QBO at higher altitude

Observations show major warmings are more likely when tropical winds in the lower stratosphere are easterly rather than westerly. The QBO consists of descending easterly and westerly wind regimes, however, so easterly QBO winds in the lower stratosphere are usually accompanied by QBO westerlies in the upper stratosphere and vice versa (Dunkerton and Baldwin 1990). It is unclear from observations, therefore, whether the occurrence of major warmings is encouraged by the lower easterlies or the upper westerlies. In our experiments we simplified the QBO’s vertical structure by specifying only the westerly or easterly regime of \( \sim 15 \)-km depth. Having so far shown the response to QBO winds in the lower stratosphere, we next examine the response to QBO winds at a higher altitude by repeating the standard QBO experiment but having the QBO wind layer raised by 9 km in altitude. The Rossby wave coupling mechanism should couple the tropics and extratropics in a similar manner regardless of the altitude that the QBO is specified at, except that the QBO at lower altitudes may be more important for the extratropical winter stratosphere because of the density weighting effect. It is also possible that there is a stronger flux of Rossby radiation incident at the tropics in the lower stratosphere than at higher altitudes. The results shown in Fig. 8 indicate that while the extratropical anomaly is broadly similar to the standard experiment (cf. Fig. 5), having colder time-average polar temperature in the westerly QBO case, the extratropical anomaly is weaker than before. It also appears to be shifted upwards relative to the standard experiment. Comparing this experiment with the two earlier experiments where the QBO was situated in the lower stratosphere suggests, therefore, that it is primarily the QBO phase at the lower levels that determines whether major warmings are likely or not. [This is consistent with Smith (1992), who found that the lower-stratospheric wind distribution is much more important than upper-stratospheric winds in determining whether or not the polar vortex is preconditioned for a warming event.]

e. Stronger wave forcing

We next discuss simulation pairs at other Rossby wave forcing strengths. Repeating the standard experiment but with larger Rossby wave forcing accelerates the vortex evolution and breakdown, while reduced forcing does the opposite. For example, when the zonal wave-one forcing amplitude, \( h \), is increased to 300 m the zonal-mean westerlies at 1 mb reverse to easterlies near day 30 for either easterly or westerly QBO phase, rather than day 50 (easterly) or day 60 (westerly) in the standard experiment where \( h = 215 \) m. Compared with the winter stratosphere such vortex evolution is unrealistically fast. We nevertheless compare the westerly and easterly QBO integrations averaged from day 20 to day 50 (rather than day 50 to day 80 because of

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**Fig. 8.** As in Fig. 5 but for experiment \( d \), where the QBO is elevated by 9 km. The resulting extratropical wind and temperature anomaly is similar to Fig. 5 but is weaker and occurs at a higher altitude. Contour interval in (c) = 0.1.
the earlier vortex breakdown). The resulting westerly minus easterly QBO anomaly, shown in Fig. 9, is quite small in the extratropics where the $\overline{u}$ anomaly maximum is $\sim -4$ m s$^{-1}$. The tendency for reduced extratropical QBO coupling in very strongly forced integrations was also seen by Holton and Austin (1991), and may explain the lack of any extratropical QBO anomaly in Bridger (1984). This is not too surprising considering that a strong Rossby wave will quickly disrupt the polar vortex into a major warming, and the QBO-influenced low-latitude “boundary condition” to wave propagation will be relatively unimportant. The westerly minus easterly difference in $\nabla \cdot \mathbf{F}$ (Fig. 9c) still shows greater wave propagation into the tropics during the westerly QBO case than in the easterly QBO case, but this QBO effect does not have a large impact in the extratropics, as Fig. 9a or 9b shows.

$f. \text{ Weaker wave forcing}$

Finally we examine forcing reduced below that of the standard experiment, to $h = 190$ m. At this forcing level the evolution of the polar vortex is retarded and the development of polar warmings is delayed by about a month compared to that of the standard experiment. Thus, the day 50 to day 80 time-average west-minus-east phase difference (Fig. 10) shows that the extratropical anomaly in zonal-mean wind and temperature is mainly confined to the mesosphere, with very weak stratospheric values. We may explain this result by considering that in this model much of the extratropical QBO anomaly is attributable to the later development of polar warming events in the westerly QBO phase simulations compared to the easterly phase simulations. Thus, the prewarming period should not show a significant extratropical anomaly. The dipole anomaly in zonal-mean wind represents the earliest stages of a polar warming event, which is developing slightly earlier (later) when the QBO is easterly (westerly).

We extended this pair of integrations to day 110 and found that the day 80 to 110 time-average extratropical anomaly did descend to lower altitudes and resembled that of the standard experiment. This was the longest integration we performed, though presumably the development of polar warmings and an extratropical QBO anomaly would be further retarded by reducing wave forcing again. Uncertainty concerning the optimal length of the integration period is removed in recent work where we have used a version of this model incorporating annually varying radiative and wave forcing to perform winter-long (September through March) simulations. For wave forcing that generates a polar warming in January or February we find a December–February average extratropical QBO anomaly qualitatively identical to that discussed here.

$4. \text{ Discussion}$

We have shown that the presence of a QBO-like easterly or westerly wind layer in the tropical stratosphere influences Rossby wave propagation in the winter stratosphere and hence modulates the polar vortex strength. With forcing levels that give realistic polar vortex evolution, the vortex becomes more disturbed and polar warmings occur sooner when the QBO is easterly rather than westerly. The QBO’s influence on the extratropical circulation was examined at different Rossby wave forcing strengths, $h = 1$ m for linear wave propagation, and $h = 190, 215, 300$ m representative of weak to excessively strong forcing. The results are consistent with the idea that more easterly QBO winds confine Rossby waves farther north of the equator and, with the development of a broad critical layer, act to communicate the QBO phase to the polar vortex circulation.

![Fig. 9](image-url)

**Fig. 9.** As in Fig. 5 but for experiment $e$, with excessively strong forcing, $h = 300$ m. The averaging period is from day 20 to day 50. There is only a very weak extratropical QBO anomaly in wind or temperature. Contour interval in (c) = 0.1.
With "infinitesimal" Rossby wave forcing the linear wave driving is confined to the vicinity of the low-latitude zero wind line, and there is therefore no sensitivity of the extratropical flow to the phase of the QBO when wave propagation is linear. At finite, but weak, Rossby wave forcing strengths, no polar warming occurs during the integration period irrespective of the phase of the QBO and, consequently, the extratropical west minus east anomaly is weak. Stronger Rossby wave forcing generates polar warmings, most readily in the east QBO phase simulation, so the day 50–80 average shows a significant extratropical west minus east anomaly. Very strong Rossby wave forcing causes major warmings to occur quickly for either QBO phase and the 30-day average west minus east difference is small. In the stratosphere the Holton–Tan extratropical QBO anomaly is largely contributed to by the occasional occurrence of major stratospheric warming events. Similarly, in this model the most appropriate Rossby wave forcing strength to use is that which leads to minor or major polar warming events after two or three months.

The model's tropical–extratropical coupling is interpreted in terms of the propagation of Rossby wave activity toward the low-latitude easterlies, and the development of a nonlinear critical layer north of these easterlies. Westerly QBO winds permit some Rossby wave propagation into the summer (Southern) hemisphere, whereas easterly QBO winds confine the wave activity farther north into the winter hemisphere. This interpretation is supported by the results of experiment $b$, where the meridionally wider QBO implies a greater difference in the latitude of the zero wind line between easterly and westerly QBO cases. In this matched-pair experiment the resulting extratropical anomaly increased relative to the standard width QBO profile, experiment $a$, as would be expected if the extratropical Rossby waves were sensitive to confinement by tropical easterlies. Shifting of the low-latitude critical line is not sufficient to cause an extratropical QBO effect, as the linear results show. It is necessary for a broad nonlinear critical layer to develop north of the easterlies, which in effect creates a QBO modulation of the midlatitude Rossby wave activity. That is, if the tropical critical line is dissipative to Rossby wave activity then the extratropical QBO anomaly does not arise from the QBO's influencing linear Rossby wave propagation. At the opposite extreme, unrealistically strong Rossby wave forcing is ineffective at communicating the QBO polewards because very strong Rossby wave forcing causes waves to quickly amplify and break, generating a major warming with very little sensitivity to the low-latitude winds. With moderate Rossby wave forcing strength (standard experiment), the polar vortex breaks down after approximately day 60, though the timing of the breakdown is sensitive to the QBO phase.

The shifting of the critical layer northwards when the QBO is easterly relative to its latitude during westerly QBO phase was also seen in O'Sullivan and Salby (1990; cf. their Fig. 11, which was using the broad QBO profile, $\Delta \phi_{QBO} = 20^\circ$). The extratropical anomaly seen in that equivalent barotropic model was stronger than that seen here partly because the two dimensionality (horizontal) of the Rossby wave propagation may have enhanced tropical–extratropical coupling relative to the three-dimensional simulation discussed here.

The low-latitude nonlinear critical layers that form in both the 2D equivalent barotropic model and the present 3D model do cause some wave reflection northwards, as is evident from localized episodes of meridionally folding of isobaric potential vorticity contours (where potential vorticity is locally advected northwards) during the wave absorption process. We do not quantify the magnitude of wave reflection by the critical layer here because of the complexity of the flow. At all times, however, the net zonal-mean flux of
wave activity at low latitudes is southwards into the critical layer.

Our use of a stratospheric circulation model with a lower boundary at 250 mb was motivated by the hypothesis that the QBO's tropical–extratropical coupling is an in situ stratospheric phenomenon. It should therefore arise in a model where the Rossby wave activity radiating up from the troposphere is identical between easterly and westerly QBO simulations. We attempted to create such identical forcing conditions by imposing a specified zonal wind and zonal wave–one geopotential height perturbation at the lower boundary. Although the geopotential height forcing is identical between QBO simulations, the evolution of the injected flux of Rossby wave activity ($F_z$, proportional to the heat flux at the lower boundary) is coupled to the evolution of the stratospheric circulation. Thus, as time evolves, the flux of Rossby wave activity injected into the stratosphere becomes different between easterly and westerly QBO cases. Figure 11 shows the time evolution of $F_z$ at the lower boundary for the easterly and westerly QBO simulations of the standard experiment. This shows that the upward EP flux, $F_z$, in both cases, has an initial rapid growth for ~8 days, as forcing is turned on, followed by a brief relaxation period. Subsequent sustained growth from day ~40 (easterly QBO case), day ~50 (westerly QBO case), onwards leads to a stratospheric warming event. In this situation $F_z$ at the lower boundary is almost always greater during the easterly QBO case. This is typical of all the moderately forced QBO simulation pairs examined. The linear experiment had almost identical $F_z$ (~1% difference) at any time. The difference in $F_z$ for the moderately forced experiments may be viewed as the faster development of a polar warming when the QBO phase is easterly rather than westerly. During the finite winter season period this would imply that polar warmings are more likely to occur during winters with easterly QBO phase rather than westerly.

It is unclear how $F_z$ at the tropopause would differ between QBO phases if the troposphere was also modeled. Since $F_z$ at the tropopause would be less constrained, it is possible that the east–west differences would be greater than we have found. These results imply that the stratospheric QBO tropical–extratropical coupling will also communicate QBO variability to the extratropical winter troposphere, though with reduced amplitude. Such an effect is consistent with the results of Boville (1985), who showed that changes in the stratospheric winter circulation lead to tropospheric circulation changes. There is also observational evidence (Dunkerton and Baldwin 1991) that $F_z$ at the tropopause may be coupled to the QBO phase. Dunkerton and Baldwin showed that $F_z$ at the tropopause is slightly stronger during winters when the 40-mb QBO phase was easterly rather than westerly. We cannot discount the possibility suggested by such observations of a tropospheric origin for the extratropical stratospheric quasi-biennial variability, although no tropospheric mechanism capable of generating QBO-related interannual variability is known. We have shown, however, that the tropopause $F_z$ observations may be explained by coupling with the stratospheric QBO.

It has been observed that the extratropical QBO anomaly in zonal-mean wind evolves through the winter season (Holton and Tan 1980; Dunkerton and Baldwin 1991). At 10 mb the dipole in the extratropical zonal-mean wind anomaly strengthens and shifts northwards from November through January. Such behavior is consistent with the cumulative effect of vortex erosion by Rossby waves whose strength is modulated by the QBO's phase. The Rossby wave coupling should cause the polar vortices' westerly QBO minus easterly QBO anomaly to increase as the winter season advances, at least until shortly after the climatological vortex strength maximizes (the climatological vortex is most intense in early January in the Northern Hemisphere and mid-August in the Southern Hemisphere; Randel 1987). The westerly QBO minus easterly QBO difference seen in the model [e.g., standard

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![Figure 11](image-url)  
**Fig. 11.** Time evolution of $F_z$ at 198 mb, showing differences developing in the upward flux of activity at the lower boundary between the easterly (upper) and westerly (lower) QBO integrations.
experiment) also shows the extratropical anomaly increasing with time until the vortex breakdown occurs (equivalent to January/February in the Northern Hemisphere).

While we have shown that the QBO phase influences the polar vortex evolution in this model, there are numerous other possible sources of interannual variability. In particular, the effects of the Southern Oscillation, internal atmospheric variability, and solar influences must be clarified through the use of more realistic numerical simulations, along with analysis of the increasing observational record.

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