Diagnosing the QBO's Influence on Circumpolar Vortex Variability Using MSU Brightness Temperatures and MSU-Derived Winds

PAUL D. REASOR AND MICHAEL T. MONTGOMERY

Department of Atmospheric Science, Colorado State University, Fort Collins, Colorado

(Manuscript received 21 April 1997, in final form 14 November 1997)

ABSTRACT

Using brightness temperatures from channels 3 and 4 of the Microwave Sounding Unit (MSU) as approximations to mean-layer temperatures, the geostrophic winds at 50 mb can be computed through a "bottom-up" approach. When this method is applied at high latitudes during austral winter and spring, it is found that accurate descriptions of the seasonal evolution and interannual variability of the lower-stratospheric circumpolar vortex are obtained. Variations in early-spring vortex strength from year to year appear to relate well to variations in the timing of the first large late-winter wavenumber one event in the lower stratosphere. Since wave forcing of the mean flow in the lower stratosphere is known to be weak, the variability in vortex strength may result from variations in wave-induced subsidence through the downward control principle.

Previous studies have demonstrated a biennial harmonic in both extratropical wave forcing and the mean flow, suggesting a link with the equatorially confined quasi-biennial oscillation (QBO). This study attempts to find a similar signal in the strength of the lower-stratospheric austral circumpolar vortex. It is first found that during the easterly (westerly) phase of the QBO large-amplitude wavenumber one in MSU channel 4, brightness temperature generally occurs earlier (later) in the season than normal. Subsequently, for most years of the study when the QBO is in its easterly (westerly) phase, the circumpolar vortex is observed to be weaker (stronger) than average.

1. Introduction

The circumpolar vortices of the Northern and Southern Hemispheres are observed to vary interannually, sometimes with dramatic differences in strength over consecutive years. Departures of the vortex from radiative balance depend strongly upon the intensity of planetary wave forcing in the extratropical stratosphere. Topographic and thermal forcing in the troposphere are believed to be the primary sources of planetary waves in the stratosphere, suggesting that this variability might have tropospheric origins. Observations show, however, that a large component of low-frequency variability in the extratropical stratosphere may be tied to the tropical quasi-biennial oscillation (QBO) in zonal winds (e.g., Angell and Korshover 1964; Bowman 1989; Tung and Yang 1994a,b, hereafter TYab).

Holton and Tan (1982) were among the first to propose a plausible mechanism linking the QBO to variability in extratropical dynamical fields (e.g., geopotential heights). They claimed that during the easterly phase of the QBO the low-latitude zero (zonal mean) wind

line, which defines a barrier to propagation for stationary planetary waves, is displaced further poleward than during the westerly phase. Consequently, the response of vertically propagating planetary waves is altered. A more disturbed circumpolar vortex results, consistent with their observations. Recently, TYab have expounded upon this theory, incorporating advances in our understanding of wave breaking and the stratospheric circulation. The essential points of the mechanism are summarized below.

During the easterly phase of the QBO, easterlies descend into the middle and lower stratosphere, centered on the equator with a half-width of about 10° – 12° . Observations of the austral winter zonal-mean wind structure, for example, show that the zero wind line can be displaced as far south as 30° S, which is about 15° from its climatological position. The precise effect of this poleward displacement of the zero wind line on wave propagation and wave forcing is an active area of research. One means by which the circumpolar vortex may become more disturbed is through the displacement of critical layer mixing associated with the zero wind line closer to the vortex edge (O'Sullivan and Young 1992). The vortex may also become more disturbed through enhanced convergence of wave energy in the stratosphere. To understand this latter effect, first consider the wave-activity density, A , valid for quasigeostrophic (QG) disturbances (see Andrews et al. 1987),

Corresponding author address: Paul D. Reasor, Department of Atmospheric Science, Colorado State University, Fort Collins, CO 80523-1371.
E-mail: reasor@eady.atmos.colostate.edu

$$A = \rho_0 \overline{q'^2} / 2\overline{q}_y, \quad (1)$$

where q' is the QG perturbation potential vorticity (PV), $\overline{q}_y$ the meridional derivative of mean-state PV, ρ_0 the mean-state density, and the bar denotes a zonal average. The wave-activity density is a measure of wave amplitude. The amplitude of linear conservative waves is governed by (Andrews and McIntyre 1976)

$$\frac{\partial A}{\partial t} + \mathbf{c}_g \cdot \nabla A = -A \nabla \cdot \mathbf{c}_g, \quad (2)$$

where $\mathbf{c}_g$ is the group velocity in the y - z plane. Planetary waves propagating into the winter stratosphere are generally confined between the polar night jet (PNJ) and the zero wind line or between the jet and the geographical pole. Thus, during an easterly phase the waves confined to the former waveguide are not only redirected further poleward, but are also restricted to a narrower space. The result is an enhancement of wave-energy convergence (i.e., $\nabla \cdot \mathbf{c}_g < 0$) near the vortex edge. According to Eq. (2), wave amplitudes following the group velocity will also be enhanced, leading to a greater tendency for wave breaking at lower levels and higher latitudes.

The impact of wave breaking on the vortex circulation is readily deduced through the transformed Eulerian-mean zonal momentum equation,

$$\frac{\partial \overline{u}}{\partial t} - f_0 \overline{v}^* = \rho_0^{-1} \nabla \cdot \mathbf{F}, \quad (3)$$

where $\overline{u}$ is the zonal-mean wind, $\overline{v}^*$ the residual meridional velocity, $\mathbf{F}$ the Eliassen-Palm (EP) flux vector, and f_0 the Coriolis parameter. The EP flux vector is defined as

$$\mathbf{F} = (-\rho_0 \overline{v' u'}, \rho_0 f_0 \overline{v' \theta'} / \theta_{oz}), \quad (4)$$

where θ is the potential temperature. The right-hand side of Eq. (3) is the wave forcing. One of the defining characteristics of wave breaking at the edge of the circumpolar vortex is its tendency to redistribute PV, taking high values from the vortex edge and mixing it into the midlatitudes. It is well known that

$$\nabla \cdot \mathbf{F} = \rho_0 \overline{v' q'} \quad (5)$$

is less than zero in this case. One can conclude that the effect of planetary wave forcing is to decelerate the mean westerly winds. Although a transverse circulation is induced to counteract this change, it is observed that strong decelerations of the PNJ still occur. Using a 25-yr dataset from the U.S. National Meteorological Center (NMC, now known as the National Centers for Environmental Prediction), Dunkerton and Baldwin (1991) showed that in the Northern Hemisphere during the easterly phase of the QBO the midlatitude upper-stratospheric EP fluxes were more convergent. More recently Baldwin and Tung (1994) demonstrated a QBO signature in both wave forcing and zonal-mean angular mo-

mentum in the Northern and Southern Hemispheres using a spectral analysis of NMC data.

Although much evidence currently exists for a QBO in the dynamical fields of the extratropical upper stratosphere, observations in the lower stratosphere, where wave forcing is considerably weaker, are less numerous. Chemical studies concerned with the lower-stratospheric ozone hole phenomenon, however, have recently drawn attention to a possible QBO modulation of column ozone in the polar regions of the Southern Hemisphere. Some suggest that a large portion of the high-latitude interannual variability in column ozone may result from the same dynamical mechanism responsible for the QBO-related variability in vortex evolution at higher levels [TYa, Kinnersley and Tung (1998)].

In the steady-state limit the residual circulation induced by wave forcing in the middle and upper stratosphere is equivalent to a zonal-mean diabatic circulation. From the QG equations of motion an expression for the vertical component of the residual velocity may be obtained (Haynes et al. 1991; Holton et al. 1995):

$$\overline{w}^*(y, z) = -\rho_0^{-1} f_0^{-1} \frac{\partial}{\partial y} \int_z^\infty (\nabla \cdot \mathbf{F}) dz'. \quad (6)$$

Because the residual circulation ($\overline{v}^*$, $\overline{w}^*$) is determined by the forcing above, the waves are said to exert a downward control. During the easterly phase of the QBO, when the EP flux is more convergent at the edge of the circumpolar vortex (and zero, or slightly divergent, poleward), downward $\overline{w}^*$ will be enhanced in the extratropics. Since ozone has its source in the midstratosphere, this implies a positive column ozone anomaly in the extratropics during an easterly phase and vice versa for a westerly phase. The observational evidence for such a correlation is mounting.

An equation similar to (6) can also be written for departures from radiative equilibrium in the steady state:

$$\delta \overline{T}(y, z) = \frac{N^2 H}{\alpha_r(y, z) \rho_0(z) f_0 R_d} \frac{\partial}{\partial y} \int_z^\infty (\nabla \cdot \mathbf{F}) dz', \quad (7)$$

where α_r is the Newtonian cooling rate, N the Brunt Väisälä frequency, R_d the gas constant for dry air, and H a vertical scale height. According to Eq. (7), it is possible that a QBO in upper-stratospheric wave forcing could, through nonlocal control, show up as a QBO in lower-stratospheric temperatures. Through thermal wind balance, this then implies a QBO modulation of lower-stratospheric dynamical fields such that when the QBO is in its easterly phase the vortex circulation is weak. During a westerly phase, the vortex at upper levels is more isolated from wave activity, leading to diminished departures from radiative equilibrium and a stronger vortex circulation. It should be noted that $\overline{w}^*$ depends inversely on density. Thus, the mean subsidence, and hence departures from radiative equilibrium, will diminish over a few scale heights from the level of wave forcing. The purpose of this paper is to investigate

whether a QBO signal exists in the lower stratosphere in spite of the weaker effects of wave forcing there.

The analysis utilizes brightness temperatures from the Microwave Sounding Unit (MSU) to compute geostrophic wind and vorticity fields. The “bottom-up” method for obtaining lower-stratospheric fields from thicknesses below (here, MSU brightness temperature is used as an approximation to thickness) is explored at high latitudes. Although the scope of the dynamical investigation is somewhat limited by primarily using the MSU data, such a focus allows us to assess the capability of the raw satellite data to represent high-latitude dynamical phenomena, in this study the influence of the QBO on lower-stratospheric circumpolar vortex strength.

2. Data and methodology

This study uses 8 yr (1985–92) of limb-corrected brightness temperature data from channels 2, 3, and 4 of the MSU supplied by the National Oceanic and Atmospheric Administration/Climate Research Division. The data are part of the *TIROS-N* Operational Vertical Sounder Sounding Product. Details of the data are described by Spencer and Christy (1990, 1993). We also use global monthly mean data from the European Center for Medium Range Weather Forecasts (ECMWF), obtained from the National Center for Atmospheric Research.

The MSU constructs its atmospheric temperature soundings by remotely measuring emissions from the strong molecular oxygen band centered about 60 GHz. Emissions of a particular frequency come from an atmospheric layer, so MSU brightness temperature T_B should not be viewed as representing a single level. It is more appropriate to think of it as a weighted mean-layer temperature, defined by the convolution of a layer temperature $\bar{T}$ and weighting function. The weighting functions for channels 2, 3, and 4 for nadir viewing have broad peaks at roughly 700 mb, 300 mb, and 100 mb, respectively.

Because the MSU weighting functions span well-defined deep layers of the atmosphere, we approximate T_B with $\bar{T}$. According to Vroman and Stephens (1989, hereafter VS), MSU channel 2 brightness temperatures (T_{B2}) closely relate to ECMWF 850–300-mb geopotential thicknesses. In addition, we find that MSU channels 3 and 4 (T_{B3} and T_{B4}) resemble 300–200-mb and 200–50-mb ECMWF mean-layer temperatures, respectively. Figure 1 shows examples of T_{B3} and T_{B4} and the corresponding ECMWF $\bar{T}$ for August and October of 1992. Although the temperatures may not agree exactly, the temperature gradients, which are used to estimate the wind and vorticity fields, do show a similarity. Here T_{B3} slightly overestimates the latitudinal gradients but captures the gross features in the near-tropopause 300–200-mb layer, and T_{B4} appears to represent the gradients of 200–50-mb $\bar{T}$ quite well.

To construct dynamical fields from the MSU data, a bottom-up method is used (e.g., Spencer et al. 1995; VS; Reasor and Montgomery 1996). First, the polar stratospheric winds are assumed to be in approximate geostrophic balance,

$$f\mathbf{v} = \mathbf{k} \times \nabla_p \Phi, \quad (8)$$

where $\mathbf{k}$ is the vertical unit vector and Φ is the geopotential. From (8) the thermal wind equation can be obtained:

$$\mathbf{v}(p_1) = \mathbf{v}(p_0) - \frac{R_d}{f} \ln\left(\frac{p_1}{p_0}\right) \mathbf{k} \times \nabla_p \bar{T}, \quad (9)$$

where p is the pressure. Using T_B in Eq. (9), the winds at the top of the approximate layer defined by the MSU weighting function can be computed, given the winds at the bottom of the layer. Assuming zero winds at the surface, one could use T_{B2} to compute the winds at 300 mb. Maps of T_{B2} , however, show anomalously large temperature gradients over Antarctica associated with a surface bias (not shown). This prevents us from using T_{B2} in a dynamical study of the polar regions. ECMWF 300-mb geopotentials are used instead to compute the geostrophic winds at 300 mb. Here T_{B3} and T_{B4} are then used to compute the winds up to 50 mb.

Taking $\mathbf{k} \cdot \nabla_p \times$ (9) gives

$$\zeta(p_1) = \zeta(p_0) - \frac{R_d}{f} \ln\left(\frac{p_1}{p_0}\right) \nabla_p^2 \bar{T}, \quad (10)$$

where ζ is the vertical component of the (geostrophic) relative vorticity. The relative vorticity at 50 mb can be evaluated following the same methodology for determining the geostrophic winds. If one integrates the vorticity over a closed area; that is,

$$\Gamma(p_1) = \int \zeta(p_1) dA, \quad (11)$$

then the circulation is obtained. The circulation is a dynamical index of vortex strength and will be used to investigate the interannual variability of the austral circumpolar vortex.

The use of analyzed data at 300 mb does not compromise our intent to validate the MSU data in a high-latitude dynamical study. We find that in the high vorticity regions of the vortex (i.e., between 60° and 70°S), ζ (300 mb) only accounts for up to 35% of ζ (50 mb), and, as expected in the upper troposphere, the area spanned by the high vorticity at 300 mb is a small fraction of that at 50 mb. Thus, contributions to 50-mb vortex circulation arise largely from the 300–200-mb and 200–50-mb layers.

3. Climatology of the austral polar vortex

a. Seasonal evolution

Figure 2 shows T_{B4} , which closely resembles 100-mb temperature, for June–October (i.e., winter through

FIG. 1. A comparison of MSU channels 3 and 4 brightness temperatures (K) and ECMWF 300–200-mb and 200–50-mb mean-layer temperatures (K), respectively. August and October of 1992 are shown. Contour interval 2 K.

spring) averaged over the years 1985–92. From June to July the meridional temperature gradients increase, implying an intensification of the PNJ. The jet appears to slightly intensify into August, although the central temperatures within the vortex cease to decrease. This is in

stark contrast to the climatological late-winter evolution of the vortex in the upper stratosphere; there, wave forcing has already decelerated the jet and contributed to a warming of roughly $0.3\text{--}0.4 \text{ K day}^{-1}$ (e.g., Mechoso et al. 1985).

FIG. 2. MSU channel 4 brightness temperature (K) depiction of the thermal evolution of the austral circumpolar vortex. The temperatures are averaged over the years 1985 to 1992. Channel 4's weighting function peaks around 100 mb for nadir viewing. Contour interval 2 K.

According to Eq. (7), wave forcing in the upper stratosphere should have an effect on the thermal structure of the entire stratosphere, particularly at high latitudes where the wave-driven subsidence is greatest. Indeed, in the polar lower stratosphere, we observe midwinter T_{B4} to be about 5–15 K greater than radiative equilibrium values. This mean subsidence at high latitudes weakens the radiatively determined strong pole to low-latitude temperature gradient. As wave forcing intensifies at low-

er levels through October, the lower-stratospheric vortex weakens considerably. The warm pool centered near 50°S and 100°E encroaches on the pole, displacing the vortex. Temperatures within the vortex jump some 12–14 K above late-winter values. This transition to the spring break-up of the austral circumpolar vortex occurs each year. The intensity of the vortex and the timing of the break-up, however, often show dramatic interannual variability. Although “internal variability” may ulti-

FIG. 3. MSU 50-mb circumpolar vortex circulation ($\text{m}^2 \text{s}^{-1}$) bounded by the $\zeta = 0.0 \text{ s}^{-1}$ contour line. Each month's circulation is shown separately for June–October (6–10). Along the bottom axis an indication of the QBO phase in the 50–20-mb layer is shown. Here W and E denote westerly and easterly phases, respectively. For example, W/E signifies westerlies over easterlies in the layer. Prior to 1989, too much MSU data was missing in June to create reliable monthly averages. August of 1986 is missing the last third of the month, and likely overpredicts the circulation. September of 1990 is missing the first third of the month and appears to underpredict the circulation.

FIG. 4. As in Fig. 3 except the circulation was computed using ECMWF 50-mb geopotentials. Due to poor monthly mean data quality at high latitudes, June of 1987 and July of 1988 are not plotted.

mately be responsible for the detailed characteristics of the seasonal evolution, we explore the possibility that the tendency for the vortex to be stronger or weaker than average is governed by the QBO-modulation of planetary wave propagation and forcing. The MSU data allow us to examine the net effect of this dynamical connection in the lower stratosphere.

b. Interannual variability

Spurred partly by interest in the frequency of sudden stratospheric warmings and related phenomena, many of the early studies of the extratropical QBO focused on the Northern Hemisphere (e.g., Dunkerton et al. 1988). In the Northern Hemisphere, however, wave activity may saturate the stratosphere and lead to large “internal variability,” obscuring the QBO signal. Although a statistically significant extratropical signal in chemical and dynamical fields has been found there, this study focuses instead on using the MSU to investigate the less scrutinized extratropical QBO in the Southern Hemisphere. There, the stratosphere is less dynamically active overall and the circumpolar vortex is closer to radiative equilibrium.

As discussed in section 1, it is believed that the extratropical QBO arises from the modulation of planetary-wave fluxes by the tropical QBO. Dunkerton and Baldwin (1991) note, however, that one cannot necessarily attribute the observed extratropical QBO *solely* to planetary-wave fluxes. Feedback mechanisms could be operative such that disturbances to the mean flow, unrelated to planetary waves, may be enhanced by plan-

etary-wave forcing. We hope that such details of the dynamical mechanism will be resolved by future theoretical and numerical studies. Our goal here is not to answer the question of how the QBO is transmitted to high latitudes, but rather to determine whether a discernible signal exists in the austral lower stratosphere. Furthermore, we wish to demonstrate the extent to which the MSU, and the technique described in section 2, can be relied upon to reproduce this signal. We focus specifically on the QBO's effect on the circumpolar vortex in an effort to better understand the interannual variability in vortex strength and, indirectly, the natural variability in ozone depletion in the Southern Hemisphere.

Figures 3 and 4 show the 50-mb circumpolar vortex circulation for each month of the study. The circulation in Fig. 3 was computed using the MSU data following the methodology in section 2. For comparison, Fig. 4 shows the circulation using ECMWF 50-mb geopotentials. The agreement between the two methods is quite remarkable given that MSU T_b are only approximations to mean-layer temperatures. From year to year there appears to be a nearly constant discrepancy between MSU and ECMWF for each month (with the exception of 1985). The months of June–August show MSU-derived circulations larger than those from the ECMWF. In September the two methods give similar values, whereas in October the ECMWF-derived circulations are consistently greater than those from the bottom-up method. Although our method slightly misrepresents the month-to-month evolution of the vortex, the observations of interannual variability are reliable. The reader is referred to Fig. 3 for a further discussion of minor discrepancies between MSU and ECMWF.

The most striking feature of Fig. 3 is the large-amplitude interannual variability in October vortex circulation. For example, from 1986 to 1987 the circulation increases by 40% and then decreases by 35% from 1987

FIG. 5. Wavenumber one component of MSU channel 4 brightness temperature (K) as a function of latitude and time (days). Period begins 1 July and extends through 31 October. Gaps in data are confined between the dashed lines. At the top of each plot an indication of the QBO phase in the 50–20-mb layer is shown. Contour interval 2 K.

to 1988. Earlier months show a similar pattern of weak and strong vortex years, but the amplitude of the variation is much less. The variability likely increases with time due to the cumulative effects of wave forcing at upper levels—that is, the wave-driven subsidence warming. Shiotani and Hirota (1985) characterized winter wave activity in the Southern Hemisphere as weak. Following the downward and poleward shifting of the PNJ in late winter, however, they found that wavenumber one tends to be enhanced and continues to be vigorous throughout spring. This suggests that the strength of the lower-stratospheric vortex in October might be intimately tied to the timing of the PNJ shift. We can use the MSU data to verify whether this relationship exists.

Although the MSU does not allow us to “see” the shifting of the PNJ in the upper stratosphere, we can infer the timing by looking at latitude–time plots of wavenumber one in T_{B4} . Figure 5 shows wavenumber one amplitudes at high latitudes for the period July–October.

In all years of this study the first large, latitudinally broad wavenumber one event ($|\hat{T}_1| \geq 5$ K) is followed by enhanced values throughout the period. We therefore identify the time of this first event as the approximate time of the PNJ shift, or at least the time of the lower-stratospheric response to the PNJ shift. In the years 1986, 1988, 1991, and 1992, the PNJ shift appears to occur during or before the first week of September, and in all other years after this time. According to Fig. 3, the aforementioned years of early PNJ shift also correspond to the four weakest vortices in October. Figure 4 might suggest that 1985 also be included, but ECMWF-derived vortex circulation at 10 mb tends to support the MSU’s depiction of a relatively strong vortex in October. Also note in 1987 and 1989, the two largest October circulation years, that the first large wavenumber one event is delayed until early to mid-October. We conclude that the extended period of subsidence warming resulting from prolonged propagation of wavenumber one into the upper stratosphere during years

TABLE 1. Relationship among variability in lower-stratospheric wavenumber one, seasonal circumpolar vortex circulation, and QBO phase for each year of the study.

	Phase of Jul 50–20-mb QBO	MSU $\Gamma'_{50\text{mb}}$ ($\text{m}^2 \text{s}^{-1}$) departure from Aug–Oct average	$ \hat{T}_1 '_{65^\circ\text{S}}$ (K) departure from Jul–Oct average	Timing of first $ \hat{T}_1 \geq 5 \text{ K}$
1985	W	$+3.2 \times 10^7$	−0.7	Mid Sep
1986	E/W	$−6.8 \times 10^7$	+0.5	Early Sep
1987	W/E	$+1.0 \times 10^8$	−1.2	Early Oct
1988	E/W	$−1.2 \times 10^8$	+1.9	Late Jul
1989	E	$+4.0 \times 10^7$	−0.9	Early Oct
1990	W	$+5.6 \times 10^6$	−0.5	Mid Sep
1991	E	$+2.4 \times 10^7$	0.0	Early Sep
1992	E	$−2.3 \times 10^7$	+1.0	Early Sep

in which the PNJ shifts early may account for the observed cases of weak vortex circulation in October. For years in which this shifting is delayed until mid-September to October, the vortex remains strong through October. To understand the variability in vortex strength, then, we must first determine what controls the timing of the PNJ shift.

According to Shiotani et al. (1993), the PNJ in the Southern Hemisphere can be characterized by its position in midwinter. It should be noted that the fall position of the PNJ shows little variability from year to year. Using 10 years of data, they grouped the midwinter PNJ into high-latitude-jet (around 50°S) and low-latitude-jet (around 40°S) categories. Interestingly, they observed larger fall wavenumber one amplitudes during low-latitude-jet years. Thus, they concluded that the characteristics of planetary-wavenumber one during fall may determine how the stratospheric circulation evolves through the rest of the season. They point out that the wave activity observed in the stratosphere can be traced to the troposphere, perhaps suggesting the importance of tropospheric variability. They also note that the high-latitude and low-latitude jet years appear to come in groups, which is uncharacteristic of a QBO signal. This evidence for a link between the timing of wavenumber one in fall and the position of the midwinter PNJ, however, does not necessarily preclude a connection between early spring variability in circumpolar vortex strength and the QBO. For example, they classify 1985 as a high-latitude jet year, which might suggest a weaker vortex, yet we observe the vortex circulation in early spring to be above average, which, as discussed below, is consistent with the presence of westerlies in the tropical stratosphere (see Table 1).

Referring back to Fig. 5, an indication is made above each plot as to whether the QBO is in its easterly (E) or westerly (W) phase. It is presently unclear which levels of the tropical stratospheric winds have the greatest influence on the planetary wave dynamics. Rather than choose a particular level, we characterize the QBO by the winds in the 50–20-mb layer, which is centered about the level of maximum QBO amplitude. We use the midwinter profile, which, for the years studied, is also characteristic of the late-winter and early spring

periods. In 1986 and 1988 the winds are easterly in the upper-half of the layer and westerly below (E/W). The reverse is true in 1987 (W/E). All other years show well-defined winds of a single sign in the layer. The four years mentioned above as having the earliest starts to vigorous wavenumber one activity are all associated with the easterly phase of the QBO. In 1985 and 1990 the first large event is delayed until mid-September, perhaps a consequence of the prevalence of equatorial westerlies at all levels. In 1987 the QBO winds are westerly in the upper portion of the layer and we observe a delay in large-amplitude wavenumber one until mid-October. Wavenumber one is also delayed until mid-October in 1989, but the equatorial winds are easterly. Thus, the timing of large wavenumber one (and perhaps the shifting of the PNJ) appears to have some connection to the tropical QBO in *most* years of the study. Whether this relation also exists in the seasonally averaged circumpolar vortex circulation in the lower stratosphere is discussed next.

Figure 6 shows the circumpolar vortex circulation

FIG. 6. MSU 50-mb, ECMWF 50-mb, and ECMWF 10-mb circulations ($\text{m}^2 \text{s}^{-1}$) bounded by the $\zeta = 0.0 \text{ s}^{-1}$ contour line and averaged over the period, August–October. The respective 8-yr means have been removed. Along the bottom axis an indication of the QBO phase in the 50–20-mb layer is shown.

FIG. 7. As in Fig. 5 except for wavenumber two. Contour interval 1 K.

averaged over the late winter/early spring period. Separately shown are the 10-mb and 50-mb ECMWF circulations and the 50-mb MSU circulation. The respective eight-year means for each time series have been removed to highlight departures from the “climatological” vortex. This also serves to partially remove the bias between MSU and ECMWF 50-mb circulations. In 1985 and 1987 the winds at upper levels are westerly and, consistent with the ideas presented in section 1, the vortex circulation is above average. In 1986 and 1988 easterly winds are observed at upper levels and the circulation is weaker than average. In 1990 the vortex circulation is above average (the MSU underpredicts the vortex circulation in 1990 as discussed in Fig. 3), as expected during the westerly phase of the QBO. The ECMWF 50-mb circulation appears to be near average in 1992, although the MSU clearly shows it to be below average in agreement with the ECMWF 10-mb circulation. Thus, the vortex circulation in 1992 is consistent with the presence of easterlies in the tropical stratosphere. Only in 1989 and 1991 are clear disagreements with theory found. The inconsistency in 1989 has already been noted. In 1991 a weaker than average vortex

circulation might be expected given the earlier start to wavenumber one activity. However, when we make this connection between the duration of wavenumber one and the cumulative effects of wave forcing at upper levels, it is assumed that the average wavenumber one, say over the period July–October, is larger when the PNJ shifts early than when the shift takes place later in spring. Although this is true in most years of the study, it does not hold in 1991 (see Table 1). In 1991 large wavenumber one begins in early September, yet overall the amplitude is about average.

The focus thus far has been on the relationship among the QBO, lower-stratospheric wavenumber one, and the 50-mb circumpolar vortex circulation. Wavenumber two may also play a role in forcing the vortex and in contributing to the residual transverse circulation, even though its amplitude is generally less than that of wavenumber one. Figure 7 shows T_{B4} wavenumber two amplitudes for the late-winter/early spring period. A similar plot is found in Manney et al. (1991) using analyzed geopotential heights at 10 mb. The agreement between the two figures appears to be quite good, suggesting some continuity in wave amplitudes between the lower

and middle stratosphere. Wavenumber two is more active throughout the season, unlike wavenumber one. But like wavenumber one, the timings of the first large-amplitude wavenumber two event ($|\hat{T}_2| \geq 4$ K) can be grouped according to QBO phase. With the exception of 1989, all years for which the QBO is in its westerly (easterly) phase, the first occurrence of large wavenumber two is found after (before) late August. Average wavenumber two amplitudes over July–October, however, do not appear to have any noticeable relation to the QBO.

4. Summary

According to current thought (summarized by *TYb*), the QBO in the Tropics modulates the polar westerly waveguide in the stratosphere, altering planetary wave propagation. During the easterly phase of the QBO the waveguide is narrowed and the low-latitude zero wind line is moved closer to the edge of the circumpolar vortex. The net effect is to enhance wave forcing in the polar stratosphere. During the westerly phase, the barrier to equatorward propagation is reduced, allowing waves propagating from the troposphere into the stratosphere to follow a more equatorward path.

Approximating the MSU brightness temperatures from channels 3 and 4 by mean-layer temperatures and using the bottom-up approach to compute winds and vorticity at 50 mb, we were able to obtain a dynamical estimate of circumpolar vortex strength in the lower stratosphere. A comparison of MSU-derived vortex circulation with that derived from ECMWF 50-mb geopotentials showed a nearly constant bias between the two in each month. Aside from this bias, the representations of interannual variability were quite similar both on a month-by-month basis and in the seasonal average. We also performed a decomposition of the MSU channel 4 brightness temperature into its wavenumber one and two components. The distribution of large-amplitude events was consistent with observations by others using analyzed data at higher levels. We then invoked the downward control principle to relate variability in lower-stratospheric wave amplitudes (assuming some coherence in wave structure between upper and lower levels) to variability in the vortex strength at 50 mb. A connection between this variability and the tropical QBO was then attempted.

We conclude that during the westerly phase of the QBO large-amplitude wavenumber one activity is generally delayed until early spring, resulting in a stronger than average circumpolar vortex. In all three westerly phase cases we observed this type of seasonal evolution. During the easterly phase, the zero wind line is displaced further south and may lead to enhanced planetary wave forcing of the circumpolar vortex, although further theoretical and numerical work is still needed to elucidate the underlying dynamics. We did, however, observe a tendency for wave amplitudes to become large earlier

in the season (i.e., late winter) when the QBO phase was easterly. The increased duration of wave activity led to a weaker vortex circulation in the lower stratosphere in three of the four cases considered. In 1991 average wavenumber one amplitudes over the course of the late winter to early spring period may have resulted in the above-average vortex circulation, in spite of the early onset of wave activity. The year 1989 was inconsistent in every way with the typical (for the 8-yr period of our study) seasonal evolution expected during an easterly QBO phase. Table 1 summarizes the various diagnostics used and clearly shows 1989 to be more reminiscent of 1987 than an easterly phase year like 1988. In 1989 the seasonally averaged 50-mb vortex circulation is above average, wavenumbers one and two are both below average, and large wave amplitudes are delayed until October. Despite our inability to reconcile the seasonal evolution of this year with the QBO, we are encouraged that the MSU did at least show a physically realistic consistency between the simple wave-related diagnostics and vortex circulation. Overall, however, we believe that there does exist some evidence for a QBO modulation of the lower-stratospheric circumpolar vortex circulation, enough so to warrant future studies. It is hoped that as data from the Advanced Microwave Sounding Unit (AMSU) become available, with greater vertical resolution in the form of 12 channels in the 50–60-GHz oxygen band, a more comprehensive study of the connection between the QBO and circumpolar vortex, utilizing both raw satellite data and the methodology outlined in this paper, might be possible.

Acknowledgments. This work was supported by the Cooperative Institute for Research in the Atmosphere (CIRA) at Colorado State University. We wish to thank Professor Graeme Stephens for inspiring us to use the MSU data to explore the connection between the QBO and circumpolar vortex strength, and Professor Wayne Schubert for his useful suggestions. We would especially like to thank Mr. Ian Wittmeyer for his assistance with the MSU data.

REFERENCES

- Andrews, D. G., and M. E. McIntyre, 1976: Planetary waves in horizontal and vertical shear: The generalized Eliassen–Palm relation and the mean zonal acceleration. *J. Atmos. Sci.*, **33**, 2031–2048.
- , J. R. Holton, and C. B. Leovy, 1987: *Middle Atmospheric Dynamics*. Academic Press, 489 pp.
- 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.
- Bowman, K. P., 1989: Global patterns of the quasi-biennial oscillation in total ozone. *J. Atmos. Sci.*, **46**, 3328–3343.
- Dunkerton, T. J., and M. Baldwin, 1991: Quasi-biennial modulation

- of planetary-wave fluxes in the Northern Hemisphere winter. *J. Atmos. Sci.*, **48**, 1043–1061.
- , D. P. Delisi, and M. P. Baldwin, 1988: Distribution of major stratospheric warmings in relation to the quasi-biennial oscillation. *Geophys. Res. Lett.*, **15**, 136–139.
- Haynes, P. H., C. J. Marks, M. E. McIntyre, T. G. Shepherd, and K. P. Shine, 1991: On the “downward control” of extratropical diabatic circulations by eddy-induced mean zonal forces. *J. Atmos. Sci.*, **48**, 651–678.
- Holton, J. R., and H.-C. Tan, 1982: The quasi-biennial oscillation in the Northern Hemisphere lower stratosphere. *J. Meteor. Soc. Japan*, **60**, 140–148.
- , P. H. Haynes, M. E. McIntyre, A. R. Douglass, R. B. Rood, and L. Pfister, 1995: Stratosphere–troposphere exchange. *Rev. Geophys.*, **33**, 403–439.
- Kinnersley, J. S., and K. K. Tung, 1998: Modeling the global interannual variability of ozone due to the equatorial QBO and to extratropical planetary wave variability. *J. Atmos. Sci.*, **55**, 1417–1428.
- Manney, G. L., J. D. Farrara, and C. R. Mechoso, 1991: The behavior of wave 2 in the Southern Hemisphere stratosphere during late winter and early spring. *J. Atmos. Sci.*, **48**, 976–998.
- Mechoso, C. R., D. L. Hartmann, and J. D. Farrara, 1985: Climatology and interannual variability of wave mean-flow interaction in the Southern Hemisphere. *J. Atmos. Sci.*, **42**, 2189–2206.
- O’Sullivan, D., and R. E. Young, 1992: Modeling the quasi-biennial oscillation’s effect on the winter stratospheric circulation. *J. Atmos. Sci.*, **49**, 2437–2448.
- Reasor, P. R., and M. T. Montgomery, 1996: Circumpolar vortex studies using MSU temperature data. Colorado State University Atmospheric Science Paper 619, 102 pp. [Available from P. Reasor, Dept. of Atmospheric Science, Colorado State University, Fort Collins, CO 80523-1371.]
- Shiotani, M., and I. Hirota, 1985: Planetary wave–mean flow interaction in the stratosphere: A comparison between Northern and Southern Hemispheres. *Quart. J. Roy. Meteor. Soc.*, **111**, 309–334.
- , N. Shimoda, and I. Hirota, 1993: Interannual variability of the stratospheric circulation in the Southern Hemisphere. *Quart. J. Roy. Meteor. Soc.*, **119**, 531–546.
- Spencer, R. W., and J. R. Christy, 1990: Precise monitoring of global temperature trends from satellites. *Science*, **247**, 1558–1562.
- , and —, 1993: Precision lower stratospheric temperature monitoring with the MSU: Technique, validation, and results 1979–1991. *J. Climate*, **6**, 1194–1204.
- , W. M. Lapenta, and F. R. Robertson, 1995: Vorticity and vertical motions diagnosed from satellite deep-layer temperatures. *Mon. Wea. Rev.*, **123**, 1800–1810.
- Tung, K. K., and H. Yang, 1994a: Global QBO in circulation and ozone. Part I: Reexamination of observational evidence. *J. Atmos. Sci.*, **51**, 2699–2707.
- , and —, 1994b: Global QBO in circulation and ozone. Part II: A simple mechanistic model. *J. Atmos. Sci.*, **51**, 2708–2721.
- Vroman, T. T., and G. L. Stephens, 1989: Microwave brightness temperature and its relation to atmospheric general circulation features. Colorado State University Atmospheric Science Paper 454, 158 pp. [Available from G. Stephens, Dept. of Atmospheric Science, Colorado State University, Fort Collins, CO 80523-1371.]