Influence of the quasi-biennial oscillation on isentropic transport and mixing in the tropics and subtropics

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Abstract. The influence of the quasi-biennial oscillation (QBO) on isentropic transport and mixing in the tropical and subtropical stratosphere is investigated over a period of 6 years. The transport and mixing is quantified by the equivalent length diagnostic, calculated from tracers simulated in chemical transport models using European Centre for Medium-Range Weather Forecasting (ECMWF) analyzed winds. A new procedure for calculating equivalent length from tracers, such as nitrous oxide (N₂O), with a tropical maximum or minimum is devised. Results from different tracers and different chemical transport models demonstrate the robustness of the equivalent length diagnostic. Equivalent length calculated both from an artificial tracer and from simulated N₂O indicates that, when the QBO winds are easterly, mixing is inhibited in the tropics throughout the broad region occupied by the easterlies and that, when the QBO winds are westerly, mixing is strongly inhibited within the narrow region occupied by the westerlies themselves but is enhanced in the subtropics. Examination of absolute vorticity gradients and horizontal Eliassen-Palm (EP) fluxes (broken down into contributions from different zonal wavenumbers) suggests that, in the ECMWF analyses, barotropic shear instability of the westerly jet, as well as propagation of planetary waves from the extratropics, drives the subtropical mixing seen in the westerly phase.

1. Introduction

The zonal wind structure in the tropical lower and middle stratosphere is characterized by descending layers of alternately easterly (up to ~30 m s⁻¹) and westerly (up to ~15 m s⁻¹) winds with a somewhat irregular period of between 24 and 32 months. This is known as the quasi-biennial oscillation (QBO). The QBO is thought to be primarily driven by the convergence of zonal momentum fluxes transported by various upward propagating equatorial waves (reviewed most recently by Dunkerton [1997] and Balwin et al. [2000].)

The QBO modulates the transport of chemical tracers in two distinct ways. Firstly, the QBO modulates diabatic transport. There is a local meridional diabatic circulation in the tropics and subtropics associated with the equatorial QBO winds [Plumb and Bell, 1982]. In periods with descending easterly QBO shear, there is equatorward motion at low levels, relative upwelling at the equator at the level of maximum shear, and poleward motion at upper levels. In periods with descending westerly QBO shear, the pattern of the meridional circulation is reversed with relative downwelling at the equator. These patterns of meridional circulation induce corresponding QBO-associated patterns in chemical tracer distributions [Gray and Pyle, 1989]. Secondly, it is thought that the QBO winds modulates isentropic mixing through their influence on extratropical planetary-wave propagation and breaking. It has been suggested that westerly QBO winds permit some planetary-wave propagation from the winter extratropics into the tropics, and perhaps even into the summer hemisphere, whereas easterly QBO winds confine the planetary-wave activity in the winter extratropics [O’Sullivan and Young, 1992; Orland, 1997; Hamilton, 1998]. The effects on planetary-wave breaking have been investigated in a modeling study by O’Sullivan and Chen [1996]. There is direct observational evidence for QBO modulation of tracer fields from volcanic aerosol [Trepte and Hitchman, 1992; Trepte et al., 1993; Hitchman...
et al., 1994; Grant et al., 1996], and chemical tracers, such as nitrous oxide, methane, and water vapor [O'Sullivan, 1997; Randel et al., 1998]. The current state of knowledge of the effects of the QBO on tracer fields in the tropics is summarized by Balwin et al. [2000], section 5.

Haynes and Shuckburgh [2000a] investigated isentropic transport and mixing in the stratosphere by calculating the diagnostic quantity “effective diffusivity” [Nakamura, 1996] from a tracer advected by analyzed winds from the 18-month period December 1996 to May 1998. The effective diffusivity diagnostic has been shown to have considerable advantages over diagnostics based on particle stretching rates, such as contour advection [Haynes and Shuckburgh, 2000a]. In particular, from one long-timescale integration of a tracer using analyzed winds the effective diffusivity can straightforwardly be calculated globally on many isentropic levels, allowing comprehensive comparison of transport and mixing between different spatial regions and different time periods. The results of Haynes and Shuckburgh [2000a] showed generally very weak mixing in the tropics/subtropics, with little transport into/out of a “tropical reservoir” region, but there was considerable variability over the 18-month period. Haynes and Shuckburgh [2000a] noted that some variability appeared to be associated with a seasonal cycle, with a shift of the entire tropical reservoir region in the direction of the summer hemisphere. They also noted that much of the remaining variability was consistent with the potential influence of the QBO on planetary-wave propagation and breaking. Their results show a broad region of weak mixing in the tropics associated with easterly QBO winds. With westerly QBO winds, their results show a narrow band of very weak mixing associated with the westerlies themselves, adjacent to a region in the summer hemisphere subtropics of enhanced mixing. However, since Haynes and Shuckburgh [2000a] only considered 18 months (less than one QBO period), any suggestion of an influence of the QBO on isentropic transport and mixing had to be speculative.

In this paper we investigate further the possible QBO modulation of transport and mixing over a longer time period, spanning three QBO periods. We quantify the isentropic transport and mixing in the stratospheric tropics and subtropics (between 30°N and 30°S) using an “equivalent length” diagnostic which is closely related to the effective diffusivity. The details of the method used are described in section 2.2. Results are presented of equivalent length calculated from artificial (section 2.3) and realistic (N2O in section 2.4) tracers advected with European Centre for Medium-Range Weather Forecasting (ECMWF) analyzed winds for the period February 1992 to February 1998. From these results the influence of the QBO on isentropic transport and mixing is inferred. Then, in section 3 we analyze further the QBO-associated mixing using various dynamical diagnostics. We suggest that the signature of mixing in the westerly case is due both to planetary waves propagating from the extratropics, and to barotropic shear instability of the westerly jet.

2. Transport and Mixing Diagnosed by Equivalent Length

2.1. Derivation of Equivalent Length

Haynes and Shuckburgh [2000a, 2000b] investigated the isentropic transport and mixing structure of the stratosphere using the effective diffusivity diagnostic [Nakamura, 1996; Nakamura and Ma, 1997]. The method used was to solve the advection-diffusion equation for an artificial tracer with concentration \( c(x, t) \),

\[
\frac{\partial c}{\partial t} + \mathbf{u} \cdot \nabla c = \nabla \cdot (\kappa \nabla c),
\]

where \( \mathbf{u}(x, t) \) is a nondivergent, two-dimensional, analyzed wind field, and \( \kappa \) is a constant diffusivity. The advection-diffusion equation (1) may be written in the form of a diffusion-only equation with an “effective diffusivity” \( \kappa_{\text{eff}} \), by using “equivalent latitude” \( \phi_e \) as an independent variable (the equivalent latitude is the latitude corresponding to an area-preserving, zonally symmetric, rearrangement of a tracer contour). Thus

\[
\frac{\partial C(\phi_e, t)}{\partial t} = \frac{1}{r^2 \cos \phi_e} \frac{\partial}{\partial \phi_e} \left[ \kappa_{\text{eff}}(\phi_e, t) \cos \phi_e \frac{\partial C(\phi_e, t)}{\partial \phi_e} \right],
\]

with

\[
\kappa_{\text{eff}}(\phi_e, t) = \frac{\kappa r^2}{(\partial C/\partial \phi_e)^2} \frac{\int [\nabla c]^2 \frac{dt}{\sqrt{\nabla c}}}{\int \frac{dt}{\sqrt{\nabla c}}} = \frac{\kappa L_{\text{eq}}^2(\phi_e, t)}{(2\pi r \cos \phi_e)^2},
\]

where \( r \) is the radius of Earth, and the integrals are around a tracer contour \( c(x, t) = C \). Equation (3) also defines the related diagnostic, the “equivalent length” \( L_{\text{eq}} \). The tracer was advected on isentropes by the nondivergent component of analyzed winds using a single-layer spectral chemical transport model (SLS) with a constant diffusivity \( \kappa \), and \( L_{\text{eq}}(\phi_e, t) \) was calculated from the tracer as it evolved. From the diffusion-only equation (2) it is clear that \( \kappa_{\text{eff}} \) is a measure of transport and mixing, and Haynes and Shuckburgh [2000a, 2000b] showed that the spatial structure of \( \kappa_{\text{eff}} \) is largely independent of the chosen \( \kappa \). Hence \( \kappa_{\text{eff}} = \kappa L_{\text{eq}}^2/(2\pi r \cos \phi_e)^2 \) was used to compare the relative strengths of mixing regions and barriers to transport in the upper troposphere and the stratosphere.

When using tracers simulated in chemical transport models that do not use advection-diffusion equations (such as models using the Prather advection scheme [Prather, 1986], which conserves second-order moments), or observed tracers, \( \kappa \) is either unknown or difficult to determine, and so \( \kappa_{\text{eff}} = \kappa L_{\text{eq}}^2/(2\pi r \cos \phi_e)^2 \) cannot be defined. However, it is clear that \( L_{\text{eq}} \) and \( \kappa_{\text{eff}} \) contain the same information about the tracer field. Both can be regarded as measures of the complexity of the geometric structure of the tracer, since the equivalent length at a particular equivalent latitude is always greater than or equal to the actual length of the cor-
responding tracer contour [Haynes and Shuckburgh, 2000a, section 2.2]. Thus, both $L_{eq}$ and $\kappa_{eq}$ will be large in regions of strong mixing (where the geometric structure will be complex), and both will be small in regions of weak mixing (where the geometric structure will be simple). Therefore, even in circumstances where $\kappa_{eq}$ cannot be defined, one might hope that $L_{eq}$ can be used as an alternative transport and mixing diagnostic. For example, $L_{eq}$ calculated from various simulated tracers [Lee et al., 2000; Allen and Nakamura, 2000] and from satellite tracer data [Nakamura and Ma, 1997] has used to analyze isentropic transport and mixing in the stratosphere.

2.2. Methodology for Equivalent Length Calculations

We have calculated equivalent length from artificial and realistic tracers in 6-year integrations of two chemical transport models, with isentropic velocity fields taken from 6-hourly ECMWF data (reanalysis up to February 1994 and operational analyses thereafter) with linear, time interpolation. It should be noted that there is not much observed tropical wind data inserted into the ECMWF analyses in the stratosphere, and so the accuracy of the winds in this region may be questionable. However, Rogers et al. [1999] showed that with an earlier version of the ECMWF analyses the winds could produce a reasonable simulation of the Mount Pinatubo aerosol cloud over the timescale of 1 month. The equivalent length results are presented in the form of the nondimensionalized quantity $L_{eq} = L_{eq}^2 / (2 \pi \cos \phi_e)^2$, which takes the value 1 when a tracer contour is geometrically equivalent to a latitude circle. The quantity $L_{eq}$ can be calculated directly from a tracer field using the method described by Haynes and Shuckburgh [2000a], section 2.2.

To demonstrate whether the seasonal/interannual variations in isentropic transport and mixing in the tropical stratosphere noted by Haynes and Shuckburgh [2000a] for December 1996 to May 1998 do indeed reflect a seasonal cycle and the influence of the QBO, we have calculated equivalent length using the same approach (i.e., from an artificial tracer, simulated on various isentropic surfaces, in the chemical transport model SLS which is based on an advection-diffusion equation) for the longer period February 1992 to February 1998. Ten isentropic levels were used, the lowest level having potential temperature $\Theta = 380$ K ($\sim 100$ hPa) and the highest level having $\Theta = 800$ K ($\sim 10$ hPa), with intermediate levels equally spaced in $\ln(\Theta)$. The results presented here are from one of these levels, $\Theta = 624$ K ($\sim 30$ hPa).

To demonstrate whether $L_{eq}$ calculated from tracers simulated in a chemical transport model that does not use an advection-diffusion equation (for which $\kappa_{eq}$ cannot be defined) can indeed be used as a transport and mixing diagnostic, we compare $L_{eq}$ calculated from the SLS tracer with $L_{eq}$ calculated from an artificial tracer simulated in the SLIMCAT model, which uses the Prather advection scheme. (SLIMCAT is an off-line chemical transport model and has been extensively used in simulations of stratospheric chemistry [e.g., Chipperfield, 1999; Hansen and Chipperfield, 1999].)

Equivalent length has been calculated previously from satellite observations of $N_2O$ [Nakamura and Ma, 1997]. These results may be influenced by two important factors that do not apply to our artificial tracer. The first is that $N_2O$ has a vertical gradient and hence will be influenced by diabatic transport. The second is that $N_2O$ is nonmonotonic in latitude, which causes problems when applying the equivalent latitude approach. We have developed a new approach for calculating $L_{eq}$ which circumvents these problems by dividing the domain into latitude bands within which the zonal mean of the tracer is locally monotonic, and calculating $L_{eq}$ separately in each band. Since the $L_{eq}$ is related to the length of tracer contours, it is expected that at the boundaries of the bands, where tracer contours will generally be cut off, $L_{eq}$ will be anomalously low. Therefore, the $L_{eq}$ corresponding to the first and last few equivalent latitudes of each band are automatically rejected. To minimize the rejected area, the tracer is first interpolated onto a higher-resolution grid, and correspondingly, a larger number of bins is used in the equivalent length calculation (for details of binning procedure see Haynes and Shuckburgh [2000a], section 2.2.) Note that this approach can be applied to almost all realistic, long-lived chemical tracers in the lower stratosphere which have either a maximum in the tropics, e.g., tropospheric source gases such as $N_2O$, or a minimum, e.g., upper-stratospheric source gases such as $O_3$. To relate our results to those of Nakamura and Ma, we have used this approach to calculate $L_{eq}$ from $N_2O$ simulated in diabatic SLIMCAT integration, where the cross-isentropic motion is provided by the heating rates calculated from the MIDRAD radiation scheme [Shine, 1987].

2.3. Equivalent Length Calculated From an Artificial Tracer

In Haynes and Shuckburgh [2000a, 2000b], $L_{eq}$ was calculated from a set of 4-month SLS integrations, each with artificial tracer initialized as sine latitude ($\sin\phi$) and using results only from the last 3 months of each integration to allow for spin-up time. An alternative approach is to include a weak source term in the tracer equation. This is advantageous for longer integrations since it avoids the need to re-initialize, but ensures that useful contrast remains in the tracer field. In this subsection we present results from a 6-year integration (February 1992 to February 1998) where such a source term, a relaxation back to the initial $\sin\phi$ profile, was included. Various tests were performed with relaxation timescales varying from 30 days to 2 years. We concluded that a relaxation timescale of 120 days is optimal, such a forcing being weak enough not to have any significant effect on the spatial structure seen in $L_{eq}$, but strong enough to prevent the tracer being homogenized. The calculations were performed on each isentropic surface at T85 spectral resolution (roughly corresponding to 2.1° lat-
itude by 2.1° longitude), using the nondivergent component of ECMWF winds, and using a constant diffusivity \( \kappa = 0.81 \times 10^2 \text{ m}^2 \text{s}^{-1} \).

We also present the results from a corresponding 6-year adiabatic SLIMCAT integration using the nondivergent component of the ECMWF winds, also for February 1992 to February 1998. The tracer was initialized and relaxed back to a \( \sin \phi \) profile on each isentropic surface, as for the SLS integration. (Separate calculations with the tracer initialized and relaxed in the same way, one including the divergent part of the winds and another using a diabatic integration gave only minor differences in the values of \( \Lambda_{\text{eq}} \).) A horizontal grid of approximately 2.8° latitude by 2.8° longitude was used. Since the Prather advection scheme used in SLIMCAT allows for much better maintenance of steep tracer gradients and filamentary structures than the diffusion scheme used in SLS, the effective resolution of the SLIMCAT integration is probably at least comparable to, if not greater than, the SLS integration.

In Plate 1 we present \( \Lambda_{\text{eq}} \) as a function of equivalent latitude and time for the 624 K isentropic surface. Of the 10 isentropic levels investigated, the \( \Lambda_{\text{eq}} \) for 624 K was found to have the strongest modulation by the QBO. The \( \Lambda_{\text{eq}} \) calculated from the artificial tracers are shown in Plate 1a (SLS) and Plate 1b (SLIMCAT). Only the tropics and subtropics (30°S < \( \phi_e \) < 30°N) are plotted. The zonal-mean zonal winds are overlaid, plotted against \( \phi \) rather than \( \phi_e \), to indicate the phase of the QBO (solid contours are westerly, dashed are easterly). The similarity of \( \Lambda_{\text{eq}} \) between the two models (which continues throughout the extratropical region not plotted) gives us confidence that it is reasonable to interpret \( \Lambda_{\text{eq}} \) calculated from SLIMCAT tracers in the same manner as that calculated from SLS tracers. Furthermore, the presence of many of the same detailed features of interannual variability in the tropics in the two plates suggests that these are real, meaningful features. There are, however, some differences, the most obvious of which is that the values of \( \Lambda_{\text{eq}} \) calculated from SLS are lower than those calculated from SLIMCAT in the subtropics of the summer hemisphere (particularly the Northern Hemisphere in July 1994). The values of \( \Lambda_{\text{eq}} \) calculated from SLS are also slightly lower in the tropics when the QBO winds are westerly. It was noted by Haynes and Shuckburgh [2000a], section 2.1, that the geometric structure of a tracer at any instant will depend on the history of the flow (weighted toward the recent past, defined by some “memory time”). This memory time is shorter in the SLS integration, which uses a substantial numerical diffusivity \( \kappa \), than in the SLIMCAT integration, which uses the Prather advection scheme. This means, for example, that structure created in the tracer during the strong mixing in the subtropics in winter would persist longer in the weakly mixed summer subtropics in the SLIMCAT integration than in the SLS integration. Correspondingly, \( \Lambda_{\text{eq}} \) would be, as is observed, lower for SLS than for SLIMCAT.

There is a strong annual cycle in the transport and mixing structure, with high values of \( \Lambda_{\text{eq}} \) in the winter hemisphere subtropics (indicating strong mixing in the surf zone). The tropical reservoir, which is represented by low values of \( \Lambda_{\text{eq}} \), exhibits a seasonal shift in position toward the summer pole.

There is also a strong tropical interannual variability in \( \Lambda_{\text{eq}} \), which appears to be closely associated with the QBO winds. There is a broad region of weak mixing (0.5 < \( \ln \Lambda_{\text{eq}} \) < 1) in the tropics and summer subtropics, directly associated with QBO easterly winds (pre-October 1992, June 1993 to October 1994, October 1995 to January 1997, post-October 1997). There is a narrow region in the tropics of very weak mixing (\( \ln \Lambda_{\text{eq}} \) < 0.5), directly associated with strongest (i.e., \( > 10 \text{ m s}^{-1} \)) westerly QBO winds (January 1993, January to July 1995, July 1997). During westerly QBO winds, the surf zone in the winter hemisphere extends into the tropics, the strongest mixing tending to follow the zero-wind line. In the summer hemisphere between about \( \phi_e = 5° \) and \( \phi_e = 15° \), there is a region of strong mixing (\( \ln \Lambda_{\text{eq}} > 2 \)) reminiscent of the “second surf zone” noted in the shallow water model simulations of Polvani et al. [1995]. This description of the QBO modulation of the isentropic transport and mixing is consistent with the modeling study of O’Sullivan and Chen [1996] (although the region of mixing they found in the summer hemisphere was farther from the equator at 10°–30°).

### 2.4. Equivalent Length Calculated From SLIMCAT N\textsubscript{2}O

A diabatic SLIMCAT integration was performed to simulate N\textsubscript{2}O (with initial conditions and simplified photolysis chemistry taken from a two-dimensional chemical transport model) for the same 6-year period. The \( \Lambda_{\text{eq}} \) was calculated following our new method outlined in section 2.2, with the N\textsubscript{2}O simulated at approximately 2.8° latitude by 2.8° longitude resolution and then interpolated onto a 0.7° latitude by 0.7° longitude grid. The first and last 7 equivalent latitudes of each band (corresponding to about 5°) were considered anomalous and have been plotted as missing data. The \( \Lambda_{\text{eq}} \) calculated from the N\textsubscript{2}O for the 624 K isentropic surface is shown in Plate 1c. There is remarkably good agreement with the \( \Lambda_{\text{eq}} \) calculated from the artificial tracer (Plate 1b). This demonstrates that, by use of our new method, accurate \( \Lambda_{\text{eq}} \) results can be obtained from realistic tracers which are nonmonotonic in latitude. There is some indication in the tropics of regions of weak mixing associated with the strongest westerly winds, although there is much missing data. The only region of significant difference is in the Northern Hemisphere between January and July 1997 when the values of \( \Lambda_{\text{eq}} \) calculated from N\textsubscript{2}O are larger than those calculated from the artificial tracer. This may be due to diabatic effects creating small-scale features in the N\textsubscript{2}O.

The \( \Lambda_{\text{eq}} \) calculated from SLIMCAT N\textsubscript{2}O can be compared with the results of Nakamura and Ma [1997] who presented \( \Lambda_{\text{eq}} \) calculated from N\textsubscript{2}O measured by the Cryogenic Limb Array Etalon Spectrometer instrument on the Upper Atmosphere Research Satellite (UARS) for the period January 1992 to February 1993. Their results concerning transport and mixing associated with the QBO appear inconsistent with our results in some respects. When there were east-
Plate 1. The $\Lambda_{eq}(\phi_0, t)$ distribution for 624 K, February 1992 to February 1998, calculated from an artificial tracer (initialized with a $\sin \phi$ profile and relaxed back to that profile over 120 days) simulated with (a) SLS and (b) SLIMCAT, and also calculated from (c) SLIMCAT-simulated $N_2O$. Contours of zonal-mean zonal wind, as a function of $\phi$ and $t$, are superimposed, with negative contours dashed and positive contours solid. A 21-day running mean has been applied to the daily $\Lambda_{eq}$ and wind data.
Plate 2. Time-latitude plots from February 1992 to February 1998 at 30 hPa, derived from monthly-mean data, of (a) the horizontal component of the Eliassen-Palm (EP) flux, (b) the divergence of the horizontal component of EP flux, and (c) latitudinal gradient in the absolute vorticity ($\beta^*$). Contours of zonal-mean zonal winds are superimposed (negative contours dashed and positive contours solid), with a contour interval of 10 m s$^{-1}$. 
erly QBO winds (January and February 1992), Nakamura and Ma calculated quite high values of $\Lambda_{\text{eq}}$ in the middle stratosphere in the tropics (between about 20°S and 20°N) and well-defined barriers in the sub tropics (at around 20°–30°) in both hemispheres. In contrast, our calculations (for February 1992 and other easterly phases) give a broad region of generally low values of $\Lambda_{\text{eq}}$ in the tropics and summer hemisphere sub tropics. When there were westerly QBO winds (January and February 1993), they calculated, in the winter hemisphere, low values of $\Lambda_{\text{eq}}$ in the sub tropics (30°–10°), with slightly higher values in the tropics. In contrast, our results give quite high values in the winter subtropics and very low values in the tropics. However, in the summer hemisphere the results are more similar, both showing high $\Lambda_{\text{eq}}$ just off the equator and a barrier in the sub tropics.

There are a number of possible explanations for these differences. Some have already been noted by Haynes and Shuckburgh [2000a], section 5, where it was suggested that diabatic effects may have been important in producing $\Lambda_{\text{eq}}$ calculated from UARS N$_2$O. In light of the similarity in the tropics and sub tropics between $\Lambda_{\text{eq}}$ calculated here from the artificial tracer with no diabatic advection, and $\Lambda_{\text{eq}}$ calculated from SLIMCAT N$_2$O, this explanation seems unlikely. Some differences may be expected as a consequence of the different methods used to calculate $\Lambda_{\text{eq}}$ since, to allow for the nonmonotonicity of N$_2$O, we calculated $\Lambda_{\text{eq}}$ using the method described in section 2.2, whereas Nakamura and Ma calculated $\Lambda_{\text{eq}}$ from the pole of one hemisphere to 10° in the other hemisphere, and then plotted the data only between the pole and the equator (N. Nakamura, personal communication, 1999). However, this cannot account for all the differences since it will only affect those in the region closest to the equator, i.e., 10°N–10°S. It is possible that the results from SLIMCAT are misleading because of errors in the ECMWF winds in the tropics, or that the results from UARS data are misleading because of the intrinsic limitations of satellite data such as poor temporal and spatial resolution. However, it seems that the most likely explanation is that in the lower and middle stratosphere the UARS N$_2$O observations were contaminated, particularly in early 1992, by the presence of aerosol from the June 1991 eruption of Mount Pinatubo, as acknowledged by Nakamura and Ma.

3. Dynamical Diagnostics

In this section we examine various dynamical diagnostics in order to investigate the processes that give rise to the modulation by the QBO of isentropic transport and mixing in the tropics and sub tropics seen in Plate 1.

3.1. Horizontal Eliassen-Palm Fluxes

Plate 2a shows a time series of the horizontal component of the Eliassen-Palm (EP) flux [Andrews et al., 1987, page 128], at 30 hPa (~624 K), between 30°N and 30°S, for the same period as the equivalent length calculation, i.e., February 1992 to February 1998. The zonal-mean zonal winds have been superimposed to indicate the phase of the QBO. Plate 2 shows a strong annual cycle in the subtropics with equatorward propagation of planetary-wave activity in the winter hemisphere (negative values indicate southward propagation in the Northern Hemisphere winter, and positive values indicate northward propagation in the Southern Hemisphere winter). During the easterly phase of the QBO, planetary-wave activity does not penetrate into the tropics, remaining poleward of about 20°. However, during the westerly phase of the QBO (January 1993, January to July 1995, and July 1997), strong wave activity extends from the subtropics of the winter hemisphere across the equator to about 5° in the summer hemisphere. Note that during these periods there is often a local minimum in EP flux at about 15° in the winter hemisphere.

Dissipation of planetary waves, for example by breaking, is indicated by the convergence of the horizontal component of the EP flux (negative values in Plate 2b). Breaking planetary waves will give rise to mixing, and hence regions of negative EP flux divergence are expected to be correlated with regions of high $\Lambda_{\text{eq}}$. The negative values (< 0.3 m s$^{-1}$ d$^{-1}$) in the winter subtropics poleward of 20° do indeed correlate with much of the regions of high $\Lambda_{\text{eq}}$ in Plate 1. However, it is noticeable that whereas the regions of high $\Lambda_{\text{eq}}$ appear to extend farther equatorward in the westerly QBO phase, the regions of strong EP flux convergence do not.

In Plate 2b there are also regions of negative EP flux divergence in the summer hemisphere in the tropics (≈5°) during the westerly phase of the QBO (in particular November 1992, November 1994 to February 1995, July 1995, August 1997). Again, these regions appear correlated with regions of high $\Lambda_{\text{eq}}$, which can be seen in Plate 1 between 5° and 15° in the summer hemisphere when there are westerly QBO winds. This may indicate that extratropical planetary waves are able to propagate from the winter hemisphere through the tropical westerlies, and break causing mixing in the summer hemisphere. Such a scenario was suggested by Polvani et al. [1995].

Plate 2b has the striking feature that there are extensive regions of EP flux divergence (positive values), with particularly large values around 15°N and 15°S. The regions of strongest divergence (> 0.15 m s$^{-1}$ d$^{-1}$) are associated with QBO westerlies and form a dipole structure with the region of strong EP flux convergence near the equator noted above. These regions of EP flux divergence would not be expected if planetary-wave propagation into the tropics from the extratropics were the sole dynamical mechanism.

3.2. Barotropic Instability

To investigate other possible dynamical mechanisms, we calculated the latitudinal gradient of absolute vorticity $\beta^* = \frac{2\Omega}{\cos \phi} - \frac{1}{\cos \phi} \frac{d}{d\phi} \left( \frac{u \cos \phi}{\cos \phi} \right)$ from monthly-mean ECMWF analyzed data for the 6-year period, and we present the results in Plate 2c. It can be seen that $\beta^*$ is strongly modulated by the QBO. When there are westerly QBO winds, there are regions of high positive $\beta^*$ (> 4 × 10$^{-11}$ m$^{-1}$ s$^{-1}$) at the equator associated with the strongest westerly winds (> 10 m s$^{-1}$). These regions are flanked at about 15° by regions of very low and often negative $\beta^*$ (note that Plate 2c
Figure 1. Artificial sin φ tracer on 624 K isentrope on (a) December 19, 1993 (easterly QBO winds), and (b) December 19, 1994 (westerly QBO winds).

shows monthly averages and so β* may be more strongly negative for shorter periods). This pattern is what would be expected from a sufficiently narrow westerly jet. The negative β* regions tend to occur at and immediately after the transition to westerly winds at the equator. Neighboring regions of negative and positive β* fulfill the necessary (Rayleigh-Kuo) condition for barotropic instability [Gill, 1982, section 13.6]. Furthermore, the patterns both of EP flux divergence (with positive values associated with negative β*), and negative values associated with positive β*), and of high Δeq values (between 5° and 15° in both hemispheres), are precisely consistent with the expected structure of the barotropic instability. The possibility of barotropic instability associated with the QBO has been discussed before, for example, by Hamilton [1984] who noted negative β* in association with both the transition to westerlies and to strong easterlies.

To investigate this possibility further, we consider synoptic plots of the artificial sin φ tracer on the 624 K isentrope that was used to calculate Δeq for Plate 1b. In Figure 1 we present a representative day from a period when there were easterly QBO winds, December 19, 1993 (Figure 1a), and a representative day from a period when there were westerly QBO winds and regions of negative β* north and south of the equator, December 19, 1994 (Figure 1b). There are a number of distinct differences between the generic features of the tracer in the easterly and the westerly case. Firstly, there is a region of steep gradients near the equator in the westerly case (Figure 1b, all longitudes, but especially close to 240°E), as is consistent with low values of Δeq. (Note that this tracer has been advected only by the nondivergent component of the winds, so the steep gradients can have formed only through the action of isentropic eddy mixing in the regions indicated by high Δeq values.) Secondly, while there is some weak folding of tracer contours in the easterly case (such as at 240°E, 10°S in Figure 1a), there is strong wrapping up of tracer contours just off the equator in the summer hemisphere in the westerly case (in particular near 300°E, 10°S in Figure 1b). This indicates, in the westerly case, eddy activity on much smaller longitudinal scales than could be expected from breaking large-scale planetary waves which have propagated from the winter hemisphere (but on a scale
consistent with barotropic instability). Furthermore, the locations of eddy features such as that near 300°E, 10°S in Figure 1b correspond to regions of EP flux divergence and negative $\beta^*$. This strongly suggests that these tracer features result from eddies that have been generated through barotropic instability. In the easterly case, the weak folding of tracer contours in Figure 1a is probably due to the action of upward propagating Rossby-gravity waves in the tropics (spectral analyses of the meridional wind, not shown, reveal significant Rossby-gravity wave activity).

3.3. EP Flux Zonal Wavenumber Spectra

The results in section 3.2 indicate that the modulation by the QBO in the ECMWF analyzed winds of isentropic transport and mixing in the tropics and subtropics is a consequence of both breaking planetary waves and barotropic instability.

To determine the relative importance of the two processes, the horizontal EP flux and its divergence at 30 hPa have been decomposed into the contributions from each zonal wavenumber. Figure 2 shows the result for January 1993 (QBO westerly), December 1993 (QBO easterly) and December 1994 (QBO westerly). Figures 2a, 2c, and 2e show the horizontal EP flux; Figures 2b, 2d, and 2f shows the horizontal EP flux divergence. Immediately striking is the difference between the easterly case, December 1993 (Figure 2c), when the EP flux is strongly confined north of 20°N, and the westerly cases, January 1993 and December 1994 (Figures 2a and 2e), when the flux extends right through the tropics. In January 1993, when the westerlies are well developed, the flux in the tropics is dominated by wavenumbers 1–4. The EP flux divergence associated with these wavenumbers shows negative values in localized regions between 5°S and 20°S. Such a pattern of flux and divergence is consistent with planetary-wave propagation out of the winter hemisphere and low-latitude breaking. In December 1994, immediately after the westerly transition, the flux in the tropics has a component at wavenumber 1. However, there is also a strong contribution to the EP flux at wavenumbers 6–10, and there are corresponding dipole structures in the EP flux divergence, centered on 10°S and 5°N. This pattern of flux and divergence is consistent with barotropic instability (as are the eddies in Figure 1b).

4. Discussion

In this paper we have used the equivalent length diagnostic to demonstrate that the QBO in the ECMWF analyzed winds modulates isentropic mixing in the tropics and subtropics. We have introduced a new procedure for calculating equivalent length from realistic tracers, such as N$_2$O, with a tropical maximum or minimum. Results of calculations of equivalent length from different tracers and different chemical transport models have demonstrated the robustness of the equivalent length as a diagnostic of isentropic mixing and transport. Our calculations have shown that, when there are QBO easterly winds, there is weak mixing throughout the tropics and subtropics. When there are QBO westerlies there is an extended surf zone in the winter hemisphere, very weak mixing at the equator associated with the strongest westerly winds, and a “second surf zone” between 5° and 15° in the summer hemisphere. These results depend on the ECMWF analyses, but there is prospect that equivalent length, for example calculated using the 6 years of ozone data from the Microwave Limb Sounder on UARS (using the new method for calculating equivalent length from realistic tracers presented in this paper), may verify the modulation by the QBO of the isentropic transport and mixing structure in the real atmosphere.

Diagnostics using horizontal EP fluxes show that, in QBO easterlies, planetary waves from the winter hemisphere are unable to propagate into the tropics. In QBO westerlies, the winter hemisphere surf zone extends farther into the tropics, and some planetary waves appear to propagate into the tropics and break on the summer hemisphere side of the QBO westerlies. However, the analysis shows that part of the enhanced mixing in the westerly phase, particularly immediately after the westerly transition, is associated with eddies with zonal wavenumbers 6-10 centered at about 10° from the equator. On the basis of absolute vorticity and EP flux diagnostics we have identified the cause of these eddies as barotropic instability that arises due to the strong curvature of the narrow westerly jet.

The pattern of tracer contours seen in Figure 1b suggests that mixing due to instability occurs on either side of the westerly jet axis, which is centered at the equator. The jet itself acts as a transport barrier, as is familiar in other fluid-dynamical contexts [e.g., del Castillo-Negrete and Morrison, 1993], in particular at the vortex edge in the extratropical winter stratosphere [e.g., Bowman, 1996]. This pattern of two mixing regions separated by a barrier occurs naturally on unstable westerly jets; see Rogerson et al. [1999] for another example. The mixing enhances tracer gradients at the equator, as seen in Figure 1b, and there is likely to be a corresponding enhancement of potential vorticity gradients, which would reinforce the barrier effect.

There is a convincing signature of barotropic instability in the ECMWF analyzed winds. Whether or not the instability occurs in the real atmosphere remains to be determined. There is relatively little wind information in the tropics and it seems unlikely that those observations would resolve the instability. What is more likely is that a QBO with barotropically unstable westerlies is being forced in the ECMWF model (which underlies the analyses) by the observations, and the model then simulates the unstable disturbances. The QBO in the ECMWF analyses used here is somewhat weak compared with Singapore radiosonde winds [Pawson and Fiorino, 1998]. With QBO winds of a more realistic amplitude, it is possible that the barotropic instability will be more important.

The new 60-level ECMWF analyses have a much better representation of the stratosphere and a stronger QBO signal in the equatorial winds. Preliminary calculations of \( \Lambda_{qt} \), based on these new analyses, for the period December 1999
to March 2000 indeed show, in association with a narrow QBO westerly jet at 500 K (~50 hPa), a similar structure to that described above. This is particularly significant since, as noted, when using the old 31-level ECMWF analyses we only found strong modulation by the QBO at around the 624 K (~30 hPa) level.

We suggest that the barotropic instability may be an important mechanism for transport and mixing in the tropics of the real atmosphere. If the instability is indeed present, then it is likely to broaden the QBO westerlies, thereby acting against the well-known narrowing effect of advection by the meridional circulation. (Note from Plate 2b that the mean
force associated with the barotropic instability may be as much as 0.2 m s\(^{-1}\) d\(^{-1}\).

It seems unlikely that in the near future there will be sufficient wind information in the tropics to assess whether barotropic instability is a feature of the real tropical stratosphere. Furthermore, the horizontal resolution of the instruments on UARS is probably too coarse in the tropics (spacings in longitude between measurements is approximately 20\(^\circ\)) to detect the signature of barotropic instability on tracer fields. However, the High Resolution Dynamics Limb Sounder (HIRDLS), planned to be launched in 2002, will have a horizontal resolution of approximately 4\(^\circ\) and should be able to show the signature of barotropic instability if it exists.

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