Morphology of Tropical Upwelling in the Lower Stratosphere

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ABSTRACT

Sensitivity tests of a mechanistic model of the mean meridional circulation driven by specified eddy forcing are conducted to investigate how the morphology of tropical upwelling in the lower stratosphere is related to the structure of the forcing expected to be associated with the stratospheric surf zone. The basic morphology of tropical upwelling is found to be similar among the mechanistic model forced with reasonable eddy fluxes, the Geophysical Fluid Dynamics Laboratory (GFDL) SKYHI GCM, U.K. Met Office (UKMO) analyses, and other climate models, indicating the robustness of the upwelling features. Atmospheric data are analyzed to characterize the interannual variability of wave drag. The influence of such variations on the interannual variability of tropical upwelling in the lower stratosphere is explored, which may help explain the observed interannual variability of stratospheric water vapor.

1. Introduction

The early works of Brewer (1949) and Dobson (1956) indicate that essentially all tropospheric air entering the stratosphere must do so through upwelling in the tropics. Although some tropospheric air can enter the stratosphere through isentropic transport at higher latitudes, this air is subsequently transported down back into the troposphere by the large-scale Brewer–Dobson circulation and, thus, does not reach the chemically active region above 100 hPa (Rosenlof and Holton 1993; Holton et al. 1995).

Variations in the magnitude and meridional structure of tropical upwelling have important implications for stratospheric ozone (Butchart and Scaife 2001; Austin and Wilson 2006), water vapor concentrations, and the age of stratospheric air (Zhou et al. 2001a; Randel et al. 2006; Austin et al. 2007).

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upwelling is particularly important for considerations of stratospheric water vapor. Holton and Gettelman (2001) suggest that it is the transport through the region of coldest lower-stratospheric temperatures, referred to as the dehydration region by Zhou et al. (2004), that determines the water vapor mixing ratio of the air entering the stratosphere. Part of the variation in the temperature of the “dehydration region” can be attributed to changes in tropical upwelling (Yulaeva et al. 1994; Randel et al. 2006). However, some aspects of the variations in tropical tropopause temperatures can be attributed to other factors, such as the intensity and spatial distribution of convection (Highwood and Hoskins 1998; Hartmann et al. 2001; Zhou et al. 2001b, 2004). Thus, decreases in the tropical upwelling mass flux or a latitudinal widening likely would result in a moister stratosphere (Zhou et al. 2001a). These influences on stratospheric water vapor are in addition to the influence from the observed increases in tropospheric methane in recent years (Austin et al. 2007).

Following up on the Zhou et al. (2006, hereafter referred to as ZGH) paper, this paper explores what factors might cause an interannual variation in the magnitude and latitudinal distribution of tropical upwelling in the lower stratosphere, and investigates whether the atmosphere shows variability in these factors. Section 2 briefly reviews the model developed by ZGH and the principal results of their earlier investigation. Section 3 discusses the results of an investigation of the relationship of the morphology of tropical upwelling in the lower stratosphere to the structure of wave forcing using the ZGH model. Section 4 shows how some of the crucial parameters that determine the magnitude and spatial distribution of tropical upwelling might vary in the actual atmosphere, and section 5 discusses some implications of those results.

2. Model used in ZGH and review of some results

The numerical model used by ZGH is based on the balanced transformed Eulerian mean equations (Andrews et al. 1987; Plumb and Eluszkiewicz 1999; Semeniuk and Shepherd 2001a), which are the equation of motion for the mean zonal wind, the thermodynamic equation for the zonal mean temperature, the continuity equation in terms of the residual mean meridional circulation, and the thermal wind relation. These equations are nonlinear and include momentum and heat diffusion. The zonal body force in the model is expressed by the divergence of the Eliassen–Palm flux, and the model radiative damping is represented by Newtonian cooling.

The computational domain extends from the earth’s surface to an altitude of 60 km and from pole to pole, and the boundary conditions are as given by ZGH’s Eqs. (5)–(7). The numerical scheme follows closely that used by Semeniuk and Shepherd (2001a). First, the equations are manipulated to give a single governing partial differential equation for the streamfunction for the residual mean circulation. The corresponding finite-difference equation uses a centered second-order finite-differencing scheme in space, and the Matsuno (1966) time-differencing scheme is used to eliminate nonlinear instability. We use the MUDPACK multigrid solver developed at National Center for Atmospheric Research (NCAR) by J. C. Adams (more information available online at www.cisl.ucar.edu/css/software/mudpack/). This solver requires that the governing partial differential equation be either elliptic or parabolic. Since the partial differential equation becoming hyperbolic corresponds to the condition for symmetric instability (Eliassen 1951; Hoskins 1974; Stevens 1983) of the mean zonal flow and our zonally symmetric model cannot explicitly treat these instabilities, we adjust the zonal wind and temperature fields in a physically reasonable way such that the unstable regions become neutrally stable and the thermal wind relation is maintained. Note that symmetric instability is usually called inertial instability in the middle-atmosphere literature (e.g., Dunkerton 1981).

ZGH were able to reproduce the main features of the annual variation of the tropical upwelling in the stratosphere as well as the nature of the annually averaged tropical upwelling. They found that there was a very important nonlinear interplay between the solstitial Hadley circulation and the wave-driven circulation in that the cross-equatorial flow from the Hadley circulation advected the angular momentum maximum off the equator into the winter hemisphere. The resulting symmetric instability then gave an enlarged region where horizontally oriented angular momentum contours at the equator extend into the winter tropics. Furthermore, it deformed the angular momentum contours toward the winter pole in the winter tropics and subtropics in the middle and upper stratosphere. When wave drag penetrates into the winter subtropics, its influence extends toward the equatorial region via “downward control” (Haynes et al. 1991; Scott 2002), producing significant upwelling there. Thus, the Hadley circulation plays an important role in determining the impact of wave drag on tropical upwelling at lower latitudes (Semeniuk and Shepherd 2001b).

Corroborating the argument about a “frictional boundary layer” near the equator (Plumb and Eluszkiewicz 1999), ZGH also found that the extent to which extratropical planetary wave drag extends toward the
The equator is particularly important for inducing tropical upwelling. ZGH ran a number of numerical experiments where the equatorward edge of the wave drag was varied. They found that when the equatorward edge of wave drag was poleward of about 12° in the winter hemisphere, the annually averaged tropical upwelling maximized around 30°N and 30°S due to the respective wave forcing during each winter period, with weak downwelling near the equator. As the equatorward edge of the wave drag moved equatorward of 12° in the winter hemisphere, however, there was annually averaged upwelling throughout the entire tropics; and as the equatorward edge of the wave drag moved closer to the equator, the latitudinal extent of the tropical upwelling narrowed and the upwelling velocities increased.

3. The relationship of the morphology of tropical upwelling in the lower stratosphere to the structure of wave forcing

The form of the wave drag used by ZGH is

\[ d = d_0 Y(\varphi) Z(z) \tau(t), \quad (1) \]

where \( d_0 = -2.0 \times 10^{-5} \text{ m s}^{-2} \) from the northern fall equinox to the northern spring equinox and is halved from the northern spring equinox to the northern fall equinox, which is similar to that deduced by Rosenlof (1995, see her Fig. 7) and the same as that used by Plumb and Eluszkiewicz (1999), but a factor of 2 smaller than the momentum residual of Eluszkiewicz et al. (1996) and Eluszkiewicz et al. (1997). Here, \( Y(\varphi) \) is given by

\[
Y(\varphi) = \begin{cases} 
\sin^3 \left( \frac{\pi}{2} \left( \frac{\varphi - \varphi_1}{\varphi_2 - \varphi_1} \right) \right) & \varphi_1 \leq \varphi \leq \varphi_2 \\
\sin^3 \left( \frac{\pi}{2} \left( \frac{\varphi - \varphi_3}{\varphi_3 - \varphi_2} \right) \right) & \varphi_2 \leq \varphi \leq \varphi_3 \\
0 & \text{elsewhere}
\end{cases} \quad (2)
\]

in northern winter (from northern fall equinox to northern spring equinox), where \( \varphi_1 \) and \( \varphi_3 \) represent the equatorward and poleward edges of the wave drag, respectively, while \( \varphi_2 \) represents the latitudinal location of the maximum wave drag; in southern winter (from northern spring equinox to northern fall equinox), the mirror image about the equator of (2) applies.

Here, \( Z(z) \) is given by

\[
Z(z) = \begin{cases} 
0 & z \leq 15 \text{ km} \\
\sin \left( \pi \left( \frac{z - 15 \text{ km}}{45 \text{ km}} \right) \right) & z > 15 \text{ km}. 
\end{cases} \quad (3)
\]

The temporal variation in the wave drag is given by

\[
\tau(t) = \begin{cases} 
\sin \left( \pi \left( \frac{t}{t_1} \right) \right) & 0 \leq t \leq t_1 \\
\sin \left( \pi \left( \frac{t - t_1}{t_2} \right) \right) & t_1 \leq t \leq t_1 + t_2 
\end{cases} \quad (4)
\]

with day \( t = 0 \) corresponding to the northern autumnal equinox, day \( t_1 = t_1 = 181 \) to the northern vernal equinox, and day \( t_1 + t_2 = 365 \) to the next northern autumnal equinox.

The spatial distribution of the wave drag applied in the mechanistic model is illustrated in Fig. 1 for the northern winter season. Figure 1 shows that the wave drag maximizes at 37.5 km in the vertical direction, as specified in Eq. (3).

Here, a series of sensitivity tests are conducted to methodically investigate how the morphology of tropical upwelling in the lower stratosphere is related to the structure of wave forcing. These sensitivity tests are motivated by analysis of the interannual variation of the observed convergences of the Eliassen–Palm flux derived from observations (see section 4). The seasonally varying radiative differential heating in this paper is specified as being the same as used in ZGH for all cases.

a. Sensitivity of the tropical upwelling in the lower stratosphere to the magnitudes of wave drag

For this sensitivity test, the control run is conducted by using the wave drag as specified by Eqs. (1)–(4) with \( \varphi_1 = 15^\circ \), \( \varphi_2 = 45^\circ \), and \( \varphi_3 = 80^\circ \) in the winter hemisphere while a sensitivity run is performed with a 50% decrease of the magnitude of wave drag, that is, \( d_0 = -1.0 \times 10^{-5} \text{ m s}^{-2} \) in northern winter and halved in southern winter. Figure 2 shows that the residual ver-
tical velocities averaged over January or July in the summer hemisphere do not differ very much between those two runs, because they are mainly determined by the same radiative differential heating (Zhou et al. 2006). However, ZGH demonstrated that the wave drag in the winter hemisphere enhances the tropical upwelling in the summer hemisphere (also see Tung and Kinnersley 2001). Since the wave drag in the northern winter is two times as large as that in the southern winter, the disparity of tropical upwelling in the southern summer hemisphere between the two model runs is slightly larger than that in the northern summer hemisphere. The upwelling in the northern subtropics in January is larger than that in the southern subtropics in July, which is due to the larger wave drag in the northern winter compared with that in the southern winter. When the magnitude of the wave drag is halved in the sensitivity run, upwelling velocities in the deep tropics are not decreased proportionally because the vertical velocity is approximately equal to the linear combination of the vertical velocities caused by the pure non-linear Hadley circulation and pure wave-driven circulation in the lower stratosphere (Zhou et al. 2006).

Since it is the long-term net transport of gas species from the lower stratosphere into the middle and upper stratosphere that matters radiatively and chemically, great importance is attached to the annual mean tropical upwelling. Figure 3 shows that the annually averaged tropical upwelling approximately scales as the wave drag, consistent with the fact that the lower-stratosphere response is almost linear according to Plumb and Eluszkiewicz (1999). It should be noted that the vertical velocities in the sensitivity run are doubled in Fig. 3 in order to facilitate comparison. As ZGH have shown, there are some nonlinear interactions between the Hadley and wave-driven circulations, which likely accounts for the fact that the linear scaling is not exact. Figure 3 also indicates that the latitudinal width of the annually averaged upwelling region is only slightly larger in the control run than in the sensitivity run, demonstrating a very nearly linear response at 20 km.

Fig. 2. Latitudinal distribution of vertical velocities at 20 km averaged over (a) January and (b) July in the sensitivity run (blue line) with a 50% decrease of the magnitude of the wave drag compared with the control run (red line).

Fig. 3. Latitudinal distribution of annual mean vertical velocities at 20 km in the sensitivity run (blue line) with a 50% decrease of the magnitude of the wave drag against the control run (red line). For the sake of comparison, the vertical velocities in the sensitivity run are multiplied by 2.

Fig. 4. The annual mean residual vertical velocities at 20 km from 10 consecutive individual years (dotted lines) and the 10-yr mean residual vertical velocity at 20 km (solid line).
Figure 4 shows that the annual mean residual vertical velocities at 20 km from 10 consecutive individual years (dotted lines) and the 10-yr mean residual vertical velocity at 20 km (solid line) from the Geophysical Fluid Dynamics Laboratory (GFDL) SKYHI GCM simulation (Hamilton and Mahlman 1988; Hamilton et al. 2001) with 2° latitude × 2.4° longitude resolution and with 80 levels in the vertical. The morphology of tropical upwelling in each year does not differ much from each other or much from that of the 10-yr mean tropical upwelling. Since the sea surface temperatures in the present integrations are prescribed with realistic, seasonally varying, climatological values and the CO₂ volume mixing ratio is set to 330 ppmv uniformly (Hamilton et al. 1995), the SKYHI model does not simulate the response to variations in these external forcings, which are likely responsible for variations in the strength of the Brewer–Dobson circulation (Butchart et al. 2000; Butchart and Scaife 2001; Eichelberger and Hartmann 2005; Austin and Wilson 2006; Austin and Li 2006; Butchart et al. 2006).

Figures 3 and 4 show that our mechanistic model can largely simulate the morphology of annual mean tropical upwelling from the SKYHI model run. Upwelling velocity in the lower stratosphere maximizes in the subtropics of each hemisphere and is relatively weak in the equatorial region. The upwelling velocity in the northern subtropics is larger than that in the southern subtropics due to the stronger planetary waves in the northern winter hemisphere. Obviously, the vertical velocity from the mechanistic model is smaller than that from the SKYHI model. This is because the wave drag imposed in the former is a factor of 2 smaller than the momentum residual of Eluszkiewicz et al. (1996, 1997), and smaller than that in the SKYHI model. In addition, the strong polar downwelling in Fig. 4 simulated by the SKYHI model is absent in Fig. 2, indicating that the idealized wave forcing specified by Eq. (2) is not realistic in that it does not extend sufficiently poleward to reproduce polar downwelling.

The morphology of tropical upwelling in the lower stratosphere from U.K. Met Office (UKMO) model analyses (see Fig. 2 in Butchart et al. 2006) is very similar to the above-mentioned structure, indicating the robustness of the upwelling features. The double-peaked structure of the annual mean tropical upwelling is also simulated by many climate models (see Fig. 2 in Butchart et al. 2006). However, it merits further investigation as to why the upwelling velocity in the southern subtropics is larger than that in the northern subtropics for some climate models such as the Canadian Middle Atmosphere Model and the Freie Universität Berlin Climate Middle Atmosphere Model (see Fig. 1 in Butchart et al. 2006).

b. Sensitivity of the tropical upwelling in the lower stratosphere to the poleward edge of wave drag

In this section, the poleward edge of the wave drag, \( \varphi_p \), is varied between 75°, 80° (the control run), and 85° in the winter hemisphere, with \( \varphi_1 = 15° \) and \( \varphi_2 = 45° \) for all three runs to examine the influence of the poleward edge of the wave drag on the latitudinal distribution of the tropical upwelling in the lower stratosphere. The magnitudes of the wave drag are specified as \( d_\theta = -2.10 \times 10^{-5} - 2.0 \times 10^{-5} \) (the control run), and \( -1.92 \times 10^{-5} \) m s\(^{-2} \), respectively, during the Northern Hemisphere winter half-year and halved during the Southern Hemisphere winter half-year so that the total wave drag integrated over the winter hemisphere is the same for all runs. The results show (figure not shown) that the latitudinal distributions of tropical upwelling in the lower stratosphere averaged over January or July are not sensitive to the locations of the poleward edge of the wave drag even though the latitudinal distributions of the downwelling velocities are influenced by those locations. In Fig. 5, changes in the poleward edge of the wave drag show no influence on the annual mean tropical upwelling in the lower stratosphere while they have strong influence on the annual mean downwelling in the extratropical lower stratosphere.

The net upward mass flux in the tropics (Rosenlof 1995; Butchart et al. 2006) can be expressed as

\[
2\pi a (\psi_{\text{max}} - \psi_{\text{min}}),
\]

where \( a \) is the earth’s mean radius, and \( \psi_{\text{max}} \) and \( \psi_{\text{min}} \) are the maximum and minimum values, respectively,
the mass streamfunction for the annual mean residual circulation.

Using this expression, it is found that at 15 km the net upward mass flux decreases from \(13.07 \times 10^8\) kg s\(^{-1}\) to \(12.47\) kg s\(^{-1}\) and \(11.98\) kg s\(^{-1}\) as the poleward edge of the wave drag shifts poleward from 75\(^\circ\) to 80\(^\circ\) and 85\(^\circ\). These values are directly proportional to the specified magnitudes of the maximum wave drag in these runs.

c. Sensitivity of the tropical upwelling in the lower stratosphere to the latitudinal locations of the maximum wave drag

In these experiments, the latitude of the maximum wave drag, \(\varphi_2\), is varied between 40\(^\circ\), 45\(^\circ\) (the control), and 50\(^\circ\), with \(\varphi_3 = 15^\circ\) and \(\varphi_3 = 80^\circ\) for all three runs. The magnitudes of the wave drag are specified as \(d_0 = -1.93 \times 10^{-5}, -2.0 \times 10^{-5}\) (the control run), and \(-2.08 \times 10^{-5}\) m s\(^{-2}\) during the Northern Hemisphere winter half-year, and halved during the Southern Hemisphere winter half-year so that the total wave drag integrated over the winter hemisphere is the same for all runs. Figure 6 shows the wave drag at 37.5 km, the specified altitude of the maximum wave drag, indicating that the latitudinal gradient of the wave drag is steeper at low latitudes (e.g., \(\varphi \leq 30^\circ\)) when it peaks at a lower latitude. The increased latitudinal gradient of the wave drag at low latitudes in the winter hemisphere leads to increased horizontal mass flux divergence, which is mainly responsible for the enhanced upwelling averaged over January or July in that region shown in Figs. 7a and 7b.

Figures 7a and 7b also show that the residual vertical velocities in January or July in the summer hemisphere do not differ very much among the three runs, because they are mainly determined by the same radiative differential heating. It is clear that a greater near-equatorial wave drag in the winter hemisphere (see Fig. 6) somewhat enhances the tropical upwelling in the summer hemisphere because of the increased cross-equatorial flow (Tung and Kinnersley 2001). Since the wave drag in the northern winter is two times as large as that in the southern winter, the differences of the tropical upwelling in the southern summer hemisphere among the three model runs are slightly larger than those in the northern summer hemisphere. As the latitudinal location of maximum wave drag shifts equatorward, the upwelling and downwelling maxima in the winter hemisphere at solstices also shift toward lower latitudes.

Figure 8 shows that the latitudinal span of the annual mean tropical upwelling becomes somewhat narrower while the magnitude of the annual mean tropical upwelling is slightly increased as the latitudinal location of the maximum wave drag shifts equatorward with its global integral kept constant.

The mass exchange rates between the troposphere and stratosphere are equal to \(13.59 \times 10^8\), \(12.47 \times 10^8\), and \(11.46 \times 10^8\) kg s\(^{-1}\), respectively [refer to Eq. (5)], as the latitudinal location of the maximum wave drag shifts poleward from 40\(^\circ\) to 45\(^\circ\) and 50\(^\circ\). These differences are caused by more wave drag at lower latitudes as the wave drag maximizes at lower latitudes.
Similar results are obtained for these three latitudes of maximum wave drag when the wave drag tapers to zero at the equator (rather than tapering to zero at 15° in the above experiments) except that the latitudinal width of the upwelling region is narrower and there is somewhat less variation in the latitudinal width of the annually averaged tropical upwelling region in this latter situation (figure not shown).

d. Sensitivity of the tropical upwelling in the lower stratosphere to the equatorward edge of wave drag

ZGH discussed the importance of the extent to which the extratropical wave drag penetrates toward the equator in determining the lower-stratosphere tropical upwelling. Their conclusions were consistent with the earlier work of Sankey (1998), Plumb and Eluszkiwicz (1999), Semeniuk and Shepherd (2001b), and Scott (2002). Here, a sensitivity test is conducted to emphasize the importance of the equatorward edge of the wave drag and to highlight the underlying physical mechanism.

In this sensitivity test, the equatorward edge $\phi_1$ of the wave drag region is located at 0°, 7.5°, and 15° in the winter hemisphere for three cases while the latitude of the maximum drag and the poleward extent of the wave drag region are held fixed at $\phi_2 = 45°$ and $\phi_3 = 80°$. The magnitudes of the wave drag for these three cases are $-1.76 \times 10^{-5}$, $-2.0 \times 10^{-5}$, and $-2.31 \times 10^{-5}$ m s$^{-2}$ for $\phi_1 = 0°$, 7.5°, and 15°, respectively, in the northern winter and halved in the southern winter so that the total wave drag integrated over the entire winter hemisphere is equal for all the cases.

Figure 9a shows that the residual vertical velocities averaged over January change significantly as the equatorward edge of wave drag is shifted from 15° latitude in the winter hemisphere to the equator. When the equatorward edge of wave drag is located at 15° latitude in the winter hemisphere, the upwelling velocities in the lower stratosphere exhibit a double-peaked structure, with one peak in the summer hemisphere, mainly explained by the Hadley circulation, and another peak in the winter subtropics caused by the wave drag. The downwelling in the tropics arises from the dominant influence of the Hadley circulation there. A double-peaked structure in the upwelling velocities is also apparent in the case when the wave drag region extends to 7.5°N, but no tropical downwelling is seen since the upward motions induced by the wave drag dominate the downward motions associated with the Hadley circulation there. However, the summer peak velocities are enhanced, and the winter peak velocities are diminished compared with the 15°N case. The increased upwelling in the winter tropics results from the increased wave drag in winter lower latitudes as the wave drag penetrates closer to the equator, which is reflected in Fig. 10a by the shrinking area enclosed by the closed angular momentum contours in the winter tropical stratosphere. The increased upwelling in the...

Fig. 8. Latitudinal distribution of annual mean vertical velocities at 20 km in the control run and the two sensitivity runs. The latitudinal locations of the maximum wave drag in the winter hemisphere are at 40° (red line), 45° (blue line), and 50° (green line), respectively, for those three runs.

Fig. 9. Latitudinal distribution of vertical velocities at 20 km averaged over (a) January and (b) July in the three cases. The equatorward edges of the wave drag in the winter hemisphere are at 15° latitude (green line), 7.5° latitude (blue line), and the equator (red line), respectively, in the winter hemisphere for those three cases.
summer hemisphere arises from the increased cross-equatorial flow and increased nonlinearity as the wave drag penetrates closer to the equator, which is indicated in Fig. 10a by the greater tilt of the angular momentum contours toward the equator in the summer tropics (Tung and Kinnersley 2001). The decreased upwelling in the winter subtropics is due to the fact that both the latitudinal gradient of wave forcing around the latitude of the maximum wave drag and the wave drag in the middle and high latitudes are decreased as the equatorward edge of the wave drag shifts from 15° to 7.5° latitude in the winter hemisphere.

Figure 9b shows that the latitudinal distribution of the residual vertical velocities in July is similar to that in January. Since the wave drag in the winter Southern Hemisphere winter is only half as large as that in the winter Northern Hemisphere, the Hadley circulation is more dominant in July, giving downward motion for both the 15° and the 7.5° cases in the winter tropics and weaker upwelling in the winter subtropics.

When the equatorward edge of the wave drag in the winter hemisphere is located closer to the equator (Polvani et al. 1995; Haynes 2005), the upwelling velocities in the lower stratosphere exhibit a single-peaked rather than a double-peaked structure in the tropical region. Figure 9a shows that the magnitude of the tropical upwelling is further increased while its latitudinal span is narrowed as the equatorward edge of the wave drag in the winter hemisphere shifts from 7.5° latitude to the equator. The increased tropical upwelling is again ascribed to the further increased wave drag in the winter lower latitudes and the consequently increased nonlinearity (see Fig. 10b) as the wave drag penetrates closer to the equator. The increased tropical upwelling is also partly attributed to the fact that the influence of even small values of model dissipation becomes important near the equator, as has been demonstrated by Plumb and Eluszkiwicz (1999). Similarly, the decreased upwelling span in the winter hemisphere is due to the further decreased latitudinal gradient of the wave forcing around the latitude of the maximum wave drag and the further decreased wave drag in the middle and high latitudes in the case when the wave drag reaches the equator.

The relative contributions of the Hadley and wave-driven circulations are a bit more transparent in July. Figure 9b shows that the primary maximum upwelling in the summer tropics associated with the Hadley circulation is strengthened while the primary maximum downwelling at southern midlatitudes, associated with the wave driving, is weakened as the equatorward edge of the wave drag in the Southern Hemisphere is shifted from 7.5° to the equator. Figure 9 also shows a secondary maximum downwelling in the southern subtropics caused by the dominance of the Hadley circulation over the wave-driven circulation there. This is quite different from the feature in the northern subtropics in January (Fig. 9a) due to the stronger wave forcing in the northern winter hemisphere.

In both January and July, the tropical upwelling maximum becomes larger and moves toward the equator as the wave drag penetrates more equatorward. These effects arise from the increased wave drag near the equator, the nonlinearity, and the model viscosity.

Figure 11 shows the latitudinal distribution of the annual mean vertical velocities at 20 km for the three cases. The equatorward edges of the wave drag in the winter hemisphere reach 15° latitude (green line), 7.5° latitude (blue line), and the equator (red line), respectively.
large effect on the latitudinal extent of the tropical upwelling regions. The annually averaged tropical upwelling extends from about 40°S to about 45°N in the \( \varphi_1 = 15° \) case, from about 35°S to about 40°N in the \( \varphi_1 = 7.5° \) case, and from about 22°S to about 28°N in the \( \varphi_1 = 0° \) case.

The mass exchange rate between the troposphere and stratosphere changes from 14.29 \( \times \) 10\(^8\) to 13.39 \( \times \) 10\(^8\) and 13.55 \( \times \) 10\(^8\) kg m\(^{-1}\) s\(^{-1}\) as the equatorward edge of the wave drag in the winter hemisphere shifts from 15° to 7.5° to the equator, respectively. When \( \varphi_1 \) changes from 15° to 7.5° latitude, the effects of the decreased latitudinal gradient of the wave forcing around the latitude of maximum wave drag and the decreased wave drag in the middle and high latitudes dominate over those of the increased wave drag in the winter lower latitudes and increased nonlinearity. Our interpretation of the causes of the slight increase in the exchange rate from 13.39 \( \times \) 10\(^8\) to 13.55 \( \times \) 10\(^8\) kg m\(^{-1}\) s\(^{-1}\) as the equatorward edge of wave drag in the winter hemisphere shifts from 7.5° latitude to the equator is as follows. The effects of the decreased latitudinal gradient of the wave forcing around the latitude of maximum wave drag and of the decreased wave drag in the middle and high latitudes are slightly dominated by those of the increased wave drag in the winter equatorial region and the additional mechanical forcing due to model viscosities.

In \( Z\)GH, we emphasized the important role of the interactions between the Hadley circulation with the wave driving in determining tropical upwelling. The present results complement our earlier paper, by illustrating how changing the strength of the wave driving and its positioning with respect to the Hadley circulation alters the shape and magnitude of the tropical upwelling.

e. Sensitivity of the tropical upwelling in the lower stratosphere to the altitudinal location of the maximum wave drag

In this section, the identical latitudinal structure and temporal variation of the wave drag are specified for each run by Eqs. (2) and (4) with \( \varphi_1 = 15°, \varphi_2 = 45°, \) and \( \varphi_3 = 80° \) in the winter hemisphere. The vertical profile of the wave drag in the control run is the same as that in the previous sensitivity tests, which is specified by Eq. (3). To examine how the altitudinal location of the wave drag influences the latitudinal distribution of the tropical upwelling in the lower stratosphere, the vertical profiles of the wave drag for the two sensitivity runs are specified by the following two equations:

\[
Z(z) = \begin{cases} 
0 & z \leq 15 \text{ km} \\
\sin \left( \frac{\pi (z - 15 \text{ km})}{35 \text{ km}} \right) & 15 \text{ km} < z \leq 32.5 \text{ km} \\
\sin \left( \frac{60 \text{ km} - z}{55 \text{ km}} \right) & 32.5 \text{ km} < z \leq 60 \text{ km}
\end{cases}
\]

(6)

and

\[
Z(z) = \begin{cases} 
0 & z \leq 15 \text{ km} \\
\sin \left( \frac{\pi (z - 15 \text{ km})}{55 \text{ km}} \right) & 15 \text{ km} < z \leq 42.5 \text{ km} \\
\sin \left( \frac{60 \text{ km} - z}{35 \text{ km}} \right) & 42.5 \text{ km} < z \leq 60 \text{ km}
\end{cases}
\]

(7)

The vertical profiles of the wave drag specified by Eqs. (6), (3), and (7) maximize at 32.5-, 37.5- (the control run), and 42.5-km altitudes, respectively, for the three runs. Correspondingly, the magnitudes of the wave drag are specified as \( d_0 = -1.71 \times 10^{-5}, -2.0 \times 10^{-5}, \) and \(-2.31 \times 10^{-5} \) m s\(^{-2}\), respectively, in the northern winter and halved in the southern winter so that the total wave drag integrated over the winter hemisphere is the same for all runs. Figure 12 shows that the residual vertical velocities averaged over January or July in the summer hemisphere are almost the same among those three runs, because they are mainly determined by the same radiative differential heating. However, Fig. 12 shows that the monthly mean upwelling in January or July in the winter hemisphere increases as the wave drag maximizes at higher altitudes. These features of tropical upwelling are similar at 15-km altitude (figure not shown). Since the vertical integral of the wave-induced body force over the altitude range 15–60 km is taken to be identical at all latitudes, the higher altitude of the maximum \( d \) (wave drag) implies that \( d \) is much greater at higher altitudes and slightly less at lower altitudes due to the decreased air density at higher levels. This combined with the resulting angular momentum contours provides an explanation for the Fig. 12 results. In Fig. 13, it is seen that at middle latitudes the angular momentum contours are shifted equatorward for the case where the wave drag maximizes at higher altitudes. This is clearly a consequence of the much enhanced high-altitude wave drag acting on the mean zonal flow. This effect is much less at lower altitudes where the wave drag is only reduced slightly. Figure 13 also shows that the absolute angular momentum contours at higher latitudes, for example, at 70°, tilt more poleward as the wave drag maximizes at higher altitudes, because the positive advection by the stronger residual meridional
velocities at higher levels dominate over the dissipation of the absolute angular momentum by the greater wave drag (i.e., this is a nonlinear effect).

The mass streamfunction $\psi$ can be expressed as

$$\psi = \int_{\zeta}^{60 \text{ km}} \left\{ \frac{\rho a \cos^2 \varphi}{m_\varphi} \right\} \left\{ \frac{m}{m_{-\text{const}}} \right\} \, dz', \tag{8}$$

where the integration is along the lines of the constant angular momentum, $m = (\Omega a \cos \varphi + \pi)a \cos \varphi$, where $\Omega$ is the planetary rotation rate, $a$ is the planetary radius, $\pi$ is the zonal wind, $\varphi$ is latitude, $\rho$ denotes the reference density profile, $d$ denotes the zonal body force per unit mass (the wave drag), the subscript $\varphi$ denotes a latitudinal derivative, and $dz'$ is the vertical projection of a segment of the $m$ contour. Since the wave drag has its maximum values at the middle latitudes, Eq. (8) and Fig. 13 imply that the maximum value of the mass streamfunction increases where the wave drag is largest, at higher altitudes. This is due to increased $d \cos^2 \varphi$ since Fig. 13 shows that the angular momentum contours are located at lower latitudes for the higher-altitude maximum wave drag case, giving rise to the enhanced upwelling (downwelling) in the vicinity of $30^\circ$ ($60^\circ$) in the winter hemisphere.

Figure 14 shows that the magnitude of the annual mean tropical upwelling is slightly increased while there is no obvious influence on the latitudinal span of the annual mean tropical upwelling, as the altitudinal location of the maximum wave drag shifts upward with its global integral kept constant. These features in the annually averaged tropical upwelling and higher-latitude downwelling are quite consistent with the explanation of the Fig. 12 results.

The mass exchange rates between the troposphere and stratosphere are equal to $10.72 \times 10^8$, $12.47 \times 10^8$, and $14.28 \times 10^8$ kg s$^{-1}$, respectively, as the altitudinal location of the maximum wave drag shifts upward from 32.5 to 37.5 and 42.5 km. Again, this is consistent with the interpretation given above. Interestingly enough, our physical interpretation implies that linear dissipation effects are dominating nonlinear effects in this case.

Table 1 gives a quick graphic summary of our sensitivity study results. It should be emphasized, however, that these conclusions apply to our calculations that imposed the condition that the total amount of wave
4. Atmospheric interannual variability

In the previous section, the sensitivity of the tropical upwelling in the lower stratosphere was explored using the idealized model of ZGH. The question naturally arises to what extent do the observations indicate that the winter wave drag associated with extratropical planetary waves varies from year to year? This is not an easy question to answer. On one hand, an interannual seesaw of stratospheric temperatures between low and high latitudes implies that the Brewer–Dobson circulation varies significantly from year to year (e.g., Labitzke 1982). Given our understanding of the stratospheric mean meridional circulation (e.g., Andrews et al. 1987), this would imply a similarly significant variation in the planetary wave drag. On the other hand, the divergence of the Eliassen–Palm flux is a difficult quantity to determine from direct calculations using atmospheric data or reanalyses. Randel et al. (2004) have shown the differences that exist among various stratospheric datasets. Variations are shown in the primary observed variables (i.e., winds and temperatures), and these differences become larger for quadratics of these primary quantities (i.e., heat and momentum fluxes). The Eliassen–Palm fluxes involve these quadratic terms, and the wave drag involves differences in the derivatives of these quadratic fluxes (see Andrews et al. 1987). Furthermore, these differences between the quadratic quantities involve a great deal of cancelation between terms, so the wave drag is very difficult to determine accurately from observations. Nonetheless, the variations of the Eliassen–Palm flux divergence evaluated from the data is largely consistent with those of such quantities as winds and temperatures, so the derived wave drag does contain useful information. In this section, we will investigate whether coherent variations in the wave drag can be identified to give clues to the factors that are likely to cause variations of the tropical upwelling in the lower stratosphere.

The 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) is employed due to its length and availability up to the 1-hPa level. We use the monthly mean dataset (1979–2002) since it is easier to use and earlier studies have shown that stationary waves normally contribute more to the wave drag than the transient waves (Hartmann 1985; Newman and Nash 2000). Monthly mean 100-, 70-, 50-, 30-, 20-, 10-, 7-, 5-, 3-, 2-, and 1-hPa horizontal velocities ($u, v$), vertical velocity $\omega$, and temperature $T$ fields were used with a $2.5^\circ \times 2.5^\circ$ latitude–longitude physical grid spacing. The components of the Eliassen–Palm flux vector $\mathbf{F}$ are defined by

$$F_\phi = \rho a \cos\varphi (\overline{u} \overline{\nabla T} / \partial \phi - \overline{v} u')$$

and

$$F_z = \rho a \cos\varphi \left( f - (a \cos \varphi)^{-1} (\overline{u} \cos \varphi) \overline{\nabla T} / \partial z - \overline{w} u' \right),$$

where the prime denotes the departure from the zonal mean, which is indicated by the overbar, and all other symbols in Eqs. (9) and (10) are as commonly used (Andrews et al. 1987).

The Eliassen–Palm flux divergence is

$$\nabla \cdot \mathbf{F} = (a \cos \varphi)^{-1} \frac{\partial}{\partial \varphi} (F_\phi \cos \varphi) + \frac{\partial F_z}{\partial z}.$$

Sensitivity studies in section 3 show that closer to the equator, small divergences of the Eliassen–Palm (EP) flux become increasingly important. However, $F_z$ associated with vertical eddy fluxes $\overline{w} u'$ in the tropics has a significant contribution from gravity waves, which are likely not well resolved in the ERA-40 reanalysis (e.g., Bergman and Salby 1994; Giorgetta et al. 2002). Thus, the accuracy of the calculated EP flux divergences is further limited in the deep tropics.

To investigate the year-to-year variability and pos-

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**Table 1.** Sensitivity of the annual mean tropical upwelling to the EP flux divergence, where $+$ is positively correlated and $-$ is negatively correlated.

<table>
<thead>
<tr>
<th>Variations in EP flux divergence</th>
<th>Total amount of upwelling mass flux</th>
<th>Latitudinal span</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Correlation</td>
<td>Sensitivity</td>
</tr>
<tr>
<td>Magnitude</td>
<td>$+$</td>
<td>High</td>
</tr>
<tr>
<td>Altitude of the max</td>
<td>$+$</td>
<td>Moderate</td>
</tr>
<tr>
<td>Latitude of the max</td>
<td>$-$</td>
<td>Moderate</td>
</tr>
<tr>
<td>Latitude of the poleward edge</td>
<td>$-$</td>
<td>Low</td>
</tr>
<tr>
<td>Latitude of the equatorward edge</td>
<td>$+/-$</td>
<td>Moderate</td>
</tr>
</tbody>
</table>
sible long-term trends in the atmospheric body forces from planetary waves in the northern winter stratosphere, winter (December–February, DJF) averages of the EP flux divergences from 1979/80 to 2001/02 were calculated by means of Eqs. (9–(11). The computed 23-winter climatological mean EP flux divergence field is shown in Fig. 15 where the winter middle-latitude “surf zone” is clearly indicated. The vertical cellular structure is likely not real and results from the computational inaccuracies discussed earlier.

The winter (DJF) anomalies were constructed by subtracting the climatological mean from the 23 individual winter averages. To identify the spatial structures that characterize the interannual variability in the winter EP flux divergence averages, a standard empirical orthogonal function (EOF) analysis of the DJF anomalies was applied to those computed anomalies in the northern winter stratosphere.

The three largest eigenvalues account for 30.6%, 17.9%, and 10.8% of the total variance. The first EOF is dominant and physically relevant according to North’s rule of thumb (North et al. 1982).

Figure 16 shows the first leading EOF and time series of its coefficient. Broadly speaking, the latitudinal gradient of the wave drag in the northern winter stratosphere vacillates from year to year, which might be due in part to the interannual variations of planetary waves generated in the troposphere and in part due to the stochastic nature of the troposphere–stratosphere system (Holton and Mass 1976; Yoden 1987). The time coefficient has an upward trend at the 90% significance level for the 23 yr since stratospheric satellite data have been included in the analysis. It remains to be determined whether the trend is real, part of the stratospheric natural interannual variability (Butchart et al. 2000), or just an artifact of the way in which these data have been collected and analyzed (von Storch and Zwiers 2001).

Figure 16 also shows that the time coefficients of EOF1 in the winters of 1996/97, 1997/98, and 2000/01 are above the one standard deviation variability of the time series of the coefficient of EOF1 whereas those in the winters of 1980/81, 1989/90, and 1990/91 are below it.

The former three winters are combined to construct the composite of strong wave drag poleward of 45°N, while the latter three winters are merged to build the composite of weak wave drag poleward of 45°N. Their differences reveal a dipole structure (figure not shown) similar to that of EOF1 shown in Fig. 16, that is, negative anomalous EP flux divergences centered near 70°N paired with positive anomalous EP flux divergences in the lower latitudes. Since the climatological mean EP
flux divergences peak around 60°N (see Fig. 15), the dipole structure of the anomalies shown in Fig. 16 indicates that the maximum wave drag in the winters of 1996/97, 1997/98 (during which a very strong El Niño event occurred), and 2000/01 is located at higher latitudes than that in the winters of 1980/81, 1989/90, and 1990/91. Figures 15 and 16 also suggest that the location of the maximum wave drag shifts poleward from 1979 to 2002, which in turn suggests (see Figs. 6–8) that the latitudinal span of tropical upwelling has an increasing trend from 1979 to 2002 while the upwelling velocities in the deep tropics have a decreasing trend and the downwelling velocities in the extratropics have an increasing trend. Note that this conclusion is not necessarily in conflict with Butchart et al. (2006) for the following reasons: (1) EOF1 only accounts for about 30% of the variance; (2) our sensitivity experiments were carried out for fixed integrated wave drag, but this is not necessarily the case in reality; and (3) the discussion is concerned only with the location of the maximum wave drag, but the other sensitivity factors studied in this paper should also play a role.

We have performed a similar EOF analysis using the 6-hourly ERA-40 results to evaluate the effects of both the stationary and transient planetary waves, and a similar picture is seen, except for the increased vertical cellular structure in EOF1 and the larger values of the EP-flux convergence at low latitudes. Very close to the equator in the lower stratosphere, we see a feature in EOF1, when transient waves are included (figure not shown), that is probably consistent with the \( \frac{\partial w}{\partial t} \) contribution from the transient equatorial waves in determining the annual cycles of the tropical upwelling that were noted by Kerr-Munslow and Norton (2006) and Norton (2006). Both analyses of EOF1, using stationary waves only and including transient waves, show vacillations in the lower-latitude body force, so we believe that this can give significant variations in the structure of the tropical upwelling, both in mass flux and latitudinal extent.

5. Conclusions and discussion

The atmospheric data analyses shown in the previous section indicate that the wave drag changes from year to year, and there exist interannual variations in the latitudinal location of the maximum wave drag.

Such variations in the EP flux divergences exert an influence on the interannual variability of the magnitude and latitudinal distribution of the tropical upwelling in the lower stratosphere. The interannual variations in the structure of the wave forcing influence both the seasonal variation and the annual mean of the tropical upwelling in the lower stratosphere. At solstices, the change of the wave forcing induces a variation of the upwelling velocities in the winter hemisphere, and extends its influence beyond the “surf zone” into the summer hemisphere.

This work was meant to illustrate how interannual changes in the wave drag can give rise to changes in both the upward mass flow from the troposphere to the stratosphere in the tropics as well as the latitudinal width of the annually averaged upwelling region. It has been demonstrated that changes in the wave drag magnitude, latitudinal location, altitudinal location, and the extent of its penetration into the tropics can cause significant variations in the annually averaged upwelling, complementing the works of Plumb and Eluszkiewicz (1999) and others. We have not tried to explore all possibilities for wave drag variations in this paper. For instance, variations in the annual cycle for the wave drag in each hemisphere will have an effect. Furthermore, it has been demonstrated that the atmosphere likely shows variations in the wave drag parameters that will cause significant variations in the tropical upwelling.

It is to be expected that variations in the tropical upwelling will significantly affect the stratospheric water vapor. Zhou et al. (2001a) have suggested that a decreased upwelling mass flow would result in a smaller “dryness” flux into the stratosphere, thus resulting in a moister stratosphere. They also suggested that, even with the same mass flux, a wider latitudinal extent of tropical upwelling would result in moister air entering the stratosphere in the flanks, thus resulting in a moister stratosphere. Atmospheric variability in EP flux divergences appears to give rise to such weaker upwelling mass flow and a wider latitudinal span of tropical upwelling. Randel et al. (2006), in their observational analysis, suggested that an increase in the extratropical planetary wave drag caused a corresponding colder tropical tropopause as well as increased mass flux, thus resulting in a drier stratosphere. In a very recent paper, Austin et al. (2007), using a fully coupled chemistry–climate model, have suggested that during the period of maximum ozone decrease, there was an increase in the upwelling tropical mass flux into the stratosphere and that this, together with increasing tropospheric methane concentrations, caused a general increase in the stratospheric water vapor. Also, of course, Butchart et al. (2006) have shown that in most coupled chemistry–climate models, anthropogenically produced changes in atmospheric composition likely have resulted in a strengthened Brewer–Dobson circulation.

Our purpose here is to contribute to understanding the factors that control tropical upwelling in chemistry–
climate models and in the atmosphere. We are encouraged by the agreement between the SKYHI results for tropical upwelling and those of our mechanistic model, and we believe that we have developed a useful tool in interpreting the behavior of tropical upwelling from both the atmosphere and the results from chemistry-climate models.

As a last comment, tropical upwelling through the tropopause is the rising portion of the Brewer–Dobson circulation, but characteristics of tropical upwelling, such as its magnitude and latitudinal distribution, are quite important, particularly for considerations of stratospheric water vapor. This being the case, it is not sufficient to characterize only changes in the magnitude of the Brewer–Dobson circulation for their stratospheric effects. The nature of the tropical upwelling (i.e., its size and latitudinal distribution) is also important.

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REFERENCES


——, ———, and ———, 2001b: Tropical cold point tropopause characteristics derived from ECMWF reanalysis and soundings. J. Climate, 14, 1823–1838.


CORRIGENDUM

The last four references in the reference section of “Morphology of Tropical Upwelling in the Lower Stratosphere,” by Marvin A. Geller, Tiehan Zhou, and Kevin Hamilton, which was published in the July 2008 issue of the *Journal of the Atmospheric Sciences*, contained errors. The corrected references are shown below, as they should have appeared originally. The staff of the *Journal of the Atmospheric Sciences* regrets any inconvenience these errors may have caused.

——, ——, and ——, 2001b: Tropical cold point tropopause characteristics derived from ECMWF reanalysis and soundings. *J. Climate*, 14, 1823–1838.