The Roles of the Hadley Circulation and Downward Control in Tropical Upwelling

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ABSTRACT

Several idealized models of tropical upwelling are presented in order to clarify the roles of the nonlinear Hadley circulation and extratropical wave driving. In particular, it is shown that the Hadley circulation and wave-driven circulation interact to determine the nature of tropical upwelling. The authors explain several observed features such as maximum upwelling in the summer subtropics and the annual variation of the upwelling.

1. Introduction

Since the pioneering work of Brewer (1949) and Dobson (1956), based on the distributions of stratospheric water vapor and ozone, it has been recognized that tropospheric air generally ascends into and within the tropical stratosphere, and spreads poleward and downward from there into the winter hemisphere. This global-scale exchange between stratosphere and troposphere, called the Brewer–Dobson circulation, is an important aspect of dynamic, chemical, and radiative coupling in the atmosphere, which is essential for prediction of global change (Holton et al. 1995).

Tropical upwelling brings greenhouse gases, and sources of reactive gases (CO₂, CH₄, N₂O, CFCs) into the stratosphere, profoundly modifying stratospheric radiation and chemistry, for example, by ozone depletion (WMO/UNEP 2003). The magnitude and configuration of tropical upwelling of tropospheric air into the stratosphere plays an important role in determining tropical tropopause properties and in stratosphere–troposphere exchange. It is likely of great importance in understanding observed variations in stratospheric water vapor.

Mechanisms for tropical upwelling are not yet satisfactorily explained. Dickinson (1971), Held and Hou (1980), WMO (1985), Haynes and McIntyre (1987), McIntyre (1990), and particularly Haynes et al. (1991) have shown that global-scale axisymmetric meridional circulations entail some kind of drag acting on the flow. Holton (1990) and Rosenlof and Holton (1993), exploited the principle of downward control (Haynes and McIntyre 1987; McIntyre 1990; Haynes et al. 1991) to derive seasonal mean upward mass fluxes from the troposphere to the stratosphere in the Tropics via mass continuity and an area-averaged tropical upwelling velocity at 100 hPa by calculating seasonal mean downward mass fluxes across the 100-hPa level in the extratropics of both hemispheres. Since the net trace species exchange depends on the latitudinal and temporal correlations between the tracer concentrations and the mass flow (Ko et al. 1985; Holton 1990), both spatial and temporal variations of tropical upwelling velocity must be addressed.

The principle of downward control fails to account for two important observed features. First, the atmospheric tape recorder (Mote et al. 1996) points to persistent equatorial upwelling throughout the year, whereas the meridional circulation depicted by downward control cannot extend outside the latitudes of the surf zone, which terminates a finite distance from the equator (McIntyre and Palmer 1983; Holton et al. 1995; Plumb and Eluszkiewicz 1999). Second, downward con-
trol is unable to account for the observations that indicate the latitudinal band of upward velocities is shifted toward the summer hemisphere and the latitude of maximum upwelling migrates toward the summer subtropics at both solstices (Yang et al. 1990; Rosenlof 1995; Randel et al. 1998; Plumb and Eluszkiewicz 1999), the effect of which is manifested in the observed seasonal distribution of aerosol (Hitchman et al. 1994) and the Halogen Occultation Experiment (HALOE) water vapor and methane (Rosenlof et al. 1997, Randel et al. 1998).

Plumb and Eluszkiewicz (1999) show that in a model with viscosity, equatorial tropical upwelling can indeed persist in the lower stratosphere when the wave drag extends to within 20° of the equator, but the tropical upwelling maximum is still located at the low-latitude edge of the wave-drag region. In addition, Plumb and Eluszkiewicz (1999) demonstrate that stratospheric thermal forcing results in a small increase in the maximum tropical upwelling and, what is more, a modest bias of the upwelling toward the summer side of the equator.

Semeniuk and Shepherd’s (2001b) results indicate that the observed annual mean tropical upwelling necessitates the nonlinear Hadley circulation (Dunkerton 1989) in the upper stratosphere and wave drag, tropical and subtropical, in the lower stratosphere; but the latitude of maximum upwelling shifts to the vernal subtropics during equinox seasons and the equatorial upwelling velocity is almost zero in the lower stratosphere.

Tung and Kinnersley (2001) argue that the influence of the winter extratropical wave drag is transmitted across the equator in the upper stratosphere so that it can produce a summer upwelling in the stratosphere. It appears that the summer upwelling in the lower stratosphere is substantially weaker in their model than that inferred by observations though.

Using a wave-mean flow model (Scott and Haynes 1998, 2002), in which the zonal momentum forcing arises self-consistently from wave dissipation, Scott (2002) showed that tropical upwelling increased with increasing forcing amplitude and there exist two qualitatively distinct stratospheric equilibrium states. Employing a zonally symmetric model, in which the zonal momentum forcing was prescribed externally, Scott (2002) found that the interaction between seasonally varying wave driving and seasonally varying tilting of the angular momentum contours associated with seasonal cycle in thermal forcing leads to a 25% increase in annual mean tropical upwelling as compared with that forced by annual mean wave drag and annual mean thermal forcing. Scott (2002) also pointed out that wave-driven tropical upwelling increasingly spread across the Tropics and into the opposite hemisphere as the equatorward boundary of wave drag shifted gradually from 25° to 5°N and cross-equatorial circulation was caused by the nonlinearity in the upper stratosphere. This latter point was also revealed by Tung and Kinnersley (2001). Finally, Scott (2002) suggested that maximum upwelling in the summer hemisphere might be explained by wave drag in the summer hemisphere.

In this paper we investigate how the Hadley and wave-driven circulations interact to determine the nature of the tropical upwelling in the lower stratosphere, thereby explaining its magnitude and configuration to a large extent. We construct a mechanistic model of the residual circulation that can be forced with specified distributions of thermal forcing and/or wave drag. This is applied first in somewhat idealized experiments, but we will also test our results by seeing how well the mechanistic model can reproduce the rather complicated spatial and seasonal variation of the residual circulation that appears in a realistic comprehensive general circulation model (GCM) simulation. In section 2, a zonally symmetric nonlinear balance model is introduced (Plumb and Eluszkiewicz 1999; Semeniuk and Shepherd 2001a). Annual variation of tropical upwelling is examined in section 3. Conclusions and discussions are presented in section 4.

2. Model description and solution

The numerical model in this paper is based on the balanced transformed Eulerian mean equations (Andrews et al. 1987; Plumb and Eluszkiewicz 1999). The numerical approach follows closely that in Semeniuk and Shepherd (2001a).

\[
\frac{\partial \bar u}{\partial t} + \bar v^* \left( \frac{1}{a} \frac{\partial \bar u}{\partial \varphi} - f \frac{\tan \varphi}{a} \right) + \bar w^* \frac{\partial \bar u}{\partial z} = d + V(\bar T),
\]

(1)

\[
\frac{\partial \bar T}{\partial t} + \bar v^* \left( \frac{1}{a} \frac{\partial \bar T}{\partial \varphi} \right) = \frac{\bar w^*}{a} \left( S \frac{\partial \bar T}{\partial z} + \frac{\delta T}{\delta \varphi} \right) = -\alpha_d \left( \bar T - T_e \right) + V(\bar T),
\]

(2)

\[
\left( f + \frac{2 \bar u \tan \varphi}{a} \right) \frac{\partial \bar u}{\partial z} = -\frac{R}{a H} \frac{\partial \bar T}{\partial \varphi},
\]

(3)

\[
\frac{1}{a \cos \varphi} \frac{\partial (\bar v^* \cos \varphi)}{\partial \varphi} + \frac{1}{\rho} \frac{\partial (\rho \bar w^*)}{\partial z} = 0,
\]

(4)
with boundary conditions

\begin{align}
\text{at } z &= 60 \text{ km}: \nabla^* = 0; \psi = 0; \frac{\partial \pi}{\partial z} = 0; T = 0, \\
\text{at } z &= 0: \nabla^* = 0; \psi = 0; \frac{\partial T}{\partial z} = 0; K_{zz} \frac{\partial \nabla}{\partial z} = c \nabla,
\end{align}

and

\begin{align}
\text{at } \varphi &= \pm \frac{\pi}{2}; \nabla = \nabla^* = 0; \psi = 0; \frac{\partial T}{\partial \varphi} = 0, \tag{6}
\end{align}

where \( \nabla, (\nabla^*, \nabla^*) \) represent the zonal wind and residual circulation, respectively. Here, \( \psi \) is the mass streamfunction for the residual circulation, that is \( \partial \psi / \partial \varphi = a \cos \varphi \nabla^* \), and \( \partial \psi / \partial z = -a \cos \varphi \nabla^* \). \( T \) is the departure of temperature from a horizontally uniform reference profile \( T_0(z) \), which has a corresponding static stability \( S = dT_0/dz + g/c_p \). The corresponding buoyancy frequency \( N = \sqrt{gR/H} \) is specified to be \( 2 \times 10^{-2} \) \( \text{ s}^{-1} \) above the tropopause at 15-km altitude, and \( 1.22 \times 10^{-2} \) \( \text{ s}^{-1} \) below. Here, \( H = 7 \) km is the constant density-scale height, \( \varphi \) is latitude, \( z = -H \ln(p/p_0) \) is the reference profile, toward which temperature is relaxed at a rate \( \alpha_T \). The thermal relaxation time \( \alpha_T^{-1} = 10 \) days is spatially uniform.

Here, \( V(\varphi) = (K_{xy}(a^2 \cos^2 \varphi)(\partial \varphi / \partial \varphi)[\cos^2 \varphi \partial^2 \varphi / \partial \varphi^2]) + (K_{zz}(p)(\partial \varphi / \partial z)[p(\partial \varphi / \partial z)]) \) are the diffusion terms for the zonal wind, where this form for the horizontal diffusion insures that the mean angular momentum on a log-pressure surface is conserved (Holton 1979). Also, \( V(\varphi) = (K_{yy}(\partial \varphi / \partial \varphi)^2 + K_{zz}(\partial \varphi / \partial z)^2 \) are the diffusion terms for \( T, V(\varphi) \) and \( V(\varphi) \) are employed to dissipate two-grid waves; \( K_{yy} \) and \( K_{zz} \) are set to 2500 \( \text{ m}^2 \text{ s}^{-1} \) and 0.1 \( \text{ m}^2 \text{ s}^{-1} \), respectively. The surface friction scheme is based on the stress condition of Held and Hou (1980) with a drag coefficient of \( c = 0.005 \text{ m s}^{-1} \). Combining the above \( K_{zz}, c \), and (6) leads to

\begin{equation}
\bar{u} |_{z=0} = \frac{1}{1 + 0.05 \Delta z} \bar{u} |_{z=\Delta z}, \tag{8}
\end{equation}

where \( \Delta z \) is the vertical grid interval of the model.

Introducing \( M = (m/a)^2 \), where \( m = a \cos \varphi (\pi + a \Omega \cos \varphi) \) is the absolute angular momentum per unit mass, we can transform (1) and (3) into (1)’ and (3)’, respectively:

\begin{equation}
\frac{\partial M}{\partial t} + \frac{\nabla^* \partial M}{a} + \frac{\nabla^*}{\nabla^*} = 2[d + V(\pi)] \sqrt{M} \cos \varphi, \tag{1}’
\end{equation}

\begin{equation}
\tan \varphi \frac{\partial M}{\partial z} = -\frac{R}{H} \cos^2 \varphi \frac{\partial T}{\partial \varphi}. \tag{3}’
\end{equation}

Using (3)’ and \( \psi, (1)’ \) and (2) can be combined into:

\begin{equation}
L[\psi] = A \frac{\partial^2 \psi}{\partial \varphi^2} + B \frac{\partial^2 \psi}{\partial \varphi \partial z} + C \frac{\partial^2 \psi}{\partial z^2} + D \frac{\partial \psi}{\partial \varphi} \tag{9}
\end{equation}

\begin{equation}
+ E \frac{\partial \psi}{\partial z} = F, \tag{9}
\end{equation}

where

\begin{align}
A &= R \cos^2 \varphi \left( S + \frac{\partial T}{\partial z} \right), \\
B &= \frac{2 \tan \varphi \partial M}{a} \tag{10}, \\
C &= -\tan \varphi \frac{\partial M}{\partial z} \tag{11}, \\
D &= \left[ \frac{\partial M}{\partial z} + R \cos^2 \varphi \left( S + \frac{\partial T}{\partial z} \right) \right] \tan \varphi \frac{\partial M}{Ha}, \tag{13} \\
E &= \frac{1}{a} \left( 1 + 4 \tan^2 \varphi \right) \frac{\partial M}{\partial z} - \tan \varphi \frac{\partial M}{Ha}, \tag{14} \\
F &= \rho \cos \varphi \left[ 2 \sin \varphi \left( \frac{\partial (\sqrt{M}[d + V(\pi)])}{\partial \varphi} + \frac{R \cos^2 \varphi}{Ha} \right) \right. \\
&\left. \times \left[ -\alpha_T \left( \frac{\partial T}{\partial \varphi} - \frac{\partial T}{\partial z} + \frac{\partial (\sqrt{V[\pi])}}{\partial \varphi} \right) \right] \right]. \tag{15}
\end{align}

The model is global, and extends to 60-km altitude, with 257 \( \times \) 257 grid points regularly spaced in latitude and in height. The centered second-order finite-differencing scheme in space is used. The time step \( \Delta t = 1 \) day. The explicit simulated backward difference method (Lilly 1965; Matsuno 1966; Haltiner and Williams 1980) is employed so as to eliminate nonlinear instability.

The MUDPACK multigrid solver developed by John C. Adams at National Center for Atmospheric Research (NCAR) is used to obtain the mass streamfunction at each time step. The MUDPACK software package requires that the operator on the left-hand side of (9), L, be either elliptic or parabolic over the domain. However, cross-equatorial flow leads to symmetric in-
stability (Eliassen 1951; Holton and Wehrbein 1980; Dunkerton 1981, 1989; Stevens 1983; Semeniuk and Shepherd 2001a), generating a hyperbolic region near the equator in the winter hemisphere, where \( B^2 - 4AC > 0 \). Since the Eqs. (1)–(4) address the large-scale circulation rather than small-scale processes, symmetric instability, and the associated adjustment are not captured by the model and have to be parameterized.

The parameterization implemented here is as follows. Cross-equatorial flow associated with the meridional circulation distorts the tropical distribution of absolute angular momentum as the maximum of absolute angular momentum is moved off the equator into the winter hemisphere in the middle stratosphere and above (where the meridional flow is strong) and to a far lesser extent in the lower stratosphere (where the meridional flow is weaker), this maximum is transported across the equator into the summer hemisphere. Consequently, inertial instability occurs, which leads to a latitudinally homogenized and vertically stratified distribution of absolute angular momentum in the tropical stratosphere. The discrete system domain is rectangular and defined by \([i, j]|1 \leq i \leq N_e = 257, 1 \leq j \leq N_l = 257\), in which \( i \) is the latitudinal index and \( j \) is the vertical index. Further, without losing generality, we demonstrate how to remove the hyperbolic regions during northern winter by a conserving redistribution of absolute angular momentum in latitude.

Wherever \( B^2 - 4AC > 0 \), we let \( B^2 - 4AC = 0 \) after the adjustment, that is,
\[
\frac{\partial M}{\partial \varphi} = -H \tan \varphi \left( \frac{\partial M}{\partial z} \right)^2 \left( R \cos^2 \varphi \left( S + \frac{\partial T}{\partial z} \right) \right). \tag{16}
\]

In the middle and upper stratosphere and above, the hyperbolic region is located in the northern Tropics. First, let us remove the hyperbolic region between \( i = 129 \) (i.e., the equator) and \( i = 130 \): if \( M_{i=129} < M_{i=130} \), let \( M_{i=130} = M_{i=129} \). Second, let us remove the hyperbolic region between \( i = 130 \) and \( i = 131 \). For any \( j \), if \( B^2 - 4AC_{i=130} > 0 \), then we calculate \( \partial M/\partial \varphi \)|\(_{i=130} \) via (16). Thus, for every affected \( j \), we get \( M(131, j) = M(130, j) + \Delta \varphi \partial M/\partial \varphi \)|\(_{i=130} \), where \( \Delta \varphi \) is the latitudinal grid interval of the model in radians. Similarly, we can further remove the hyperbolic region between \( i = 131 \) and \( i = 132 \). This algorithm is repeated until the whole hyperbolic region is removed in the northern Tropics. For any \( j \), ranges of \( i \) where the adjustments are performed can be identified at every time step and take the form \([130 \leq i \leq N_l < N_e]\). After calculating the lost absolute angular momentum in the winter hemisphere for any given \( j \)

\[
\Delta m = \sum_{i=130}^{N_l} (m_i - m_i') \cos \varphi, \tag{17}
\]

where \( m_i, m_i' \), \( \cos \varphi \), represent the absolute angular momentum before the adjustment, that after the adjustment, and the latitude, respectively, at a grid point \((i, j)\), we add the same amount of absolute angular momentum, \( \Delta m \), between the equator and \( 20^\circ \) in the summer hemisphere for the same \( j \) with the distribution function \([\sin[\pi(\varphi/20)]]\), where \( \varphi \) is latitude in degrees. The model results are insensitive to the exact value \( 20^\circ \), so long as the result of adjustment is confined to the Tropics where the observed absolute angular momentum distribution in the tropical middle and upper stratosphere has small latitudinal gradients (see Fig. 1 in Haynes et al. 1991). This arbitrary redistribution is ascribed to the barotropic instability near the equator due to the equatorward transfer of angular momentum caused by the inertial instability (Semeniuk and Shepherd 2001a).

In the lower stratosphere, the algorithm of the adjustment is similar to the abovementioned, except that the lost absolute angular momentum in the summer Tropics is projected between the equator and \( 5^\circ \) in the winter hemisphere with the distribution function \([\sin[\pi(\varphi/5)]]\). The model results are also insensitive to the exact value \( 5^\circ \) due to similar considerations discussed above.

In developing the above scheme for adjusting the angular momentum, we have tried to reflect the expected consequences of the instabilities, conserve angular momentum, and make a minimum of ad hoc assumptions (e.g., see appendix B of Semeniuk and Shepherd 2001a).

After the absolute angular momentum is adjusted in the model domain, the temperature field is also adjusted via the thermal wind relation
\[
\tan \varphi \frac{\partial (\Delta M)}{\partial z} = - \frac{R}{H} \cos^2 \varphi \frac{\partial (\Delta T)}{\partial \varphi},
\]

where \( \Delta M \) and \( \Delta T \) are the adjusted amounts of \( M \) and \( T \) at every time step.

3. Annual variation of residual mean meridional circulation

a. Annual variation of nonlinear Hadley circulation

Here, we run the model with seasonally varying stratospheric thermal forcing, but without any wave drag. An idealized radiative equilibrium temperature is specified to be \( T_e = T_{e0} Z(z) Y_T(\varphi, t) \) with \( T_{e0} = -30 \) K. The vertical, latitudinal, and temporal dependences are defined by
\[
Z(z) = \begin{cases} 
0 & z \leq 15 \text{ km} \\
\sin \left( \frac{\pi}{1} \left( \frac{z - 15 \text{ km}}{45 \text{ km}} \right) \right) & z > 15 \text{ km}
\end{cases}
\]

(18)

\[
Y_{\varphi}(\varphi, t) = \begin{cases} 
\sin^2 \left( \frac{\varphi}{2} \left( -\frac{\varphi}{2} - \varphi_0 \right) \right) \left[ 1 - \sin \left( \frac{\pi}{l_1} \right) \right] / 2 - C_0 & \varphi \equiv \varphi(t) \\
\sin^2 \left( \frac{\varphi}{2} \left( \frac{\varphi}{2} - \varphi_0 \right) \right) \left[ 1 + \sin \left( \frac{\pi}{l_1} \right) \right] / 2 - C_0 & \varphi \equiv \varphi(t) \\
\sin^2 \left( \frac{\varphi}{2} \left( \frac{\varphi}{2} - \varphi_0 \right) \right) \left[ 1 - \sin \left( \frac{\pi}{l_2} \right) \right] / 2 - C_0 & \varphi \equiv \varphi(t)
\end{cases}
\]

\[
0 \leq t \leq l_1
\]

\[
t_1 \leq t \leq l_1 + t_2
\]

(19)

where \( \varphi(t) = \begin{cases} 
-20^\circ \sin \left( \frac{\pi}{l_1} \right) & 0 \leq t \leq l_1 \\
20^\circ \sin \left( \frac{\pi}{l_1 + t_2} \right) & l_1 \leq t \leq l_1 + t_2
\end{cases}
\]

(20)

with day \( t = 0 \) corresponding to northern autumnal equinox, day \( t = t_1 = 181 \) northern vernal equinox, day \( t_1 + t_2 = 365 \) next northern autumnal equinox. Since the globally averaged heating vanishes, we have \( \int_{l_1}^{l_1 + t_2} \cos \varphi \, d\varphi = 0 \). That is to say, \( C_0 \) in (19)

\[
C_0 = \begin{cases} 
1 \int_{\varphi}^{\varphi_0} \cos \varphi \sin^2 \left( \frac{\pi}{2} \left( -\frac{\varphi}{2} - \varphi_0 \right) \right) \left[ 1 - \sin \left( \frac{\pi}{l_1} \right) \right] d\varphi & 0 \leq t \leq l_1 \\
+ \int_{\varphi}^{\varphi_0} \cos \varphi \sin^2 \left( \frac{\pi}{2} \left( \frac{\varphi}{2} - \varphi_0 \right) \right) \left[ 1 + \sin \left( \frac{\pi}{l_1} \right) \right] d\varphi & l_1 \leq t \leq l_1 + t_2
\end{cases}
\]

(21)

Figure 1 shows \( T_e \) at northern winter solstice. The maximum latitudinal gradient of the equilibrium temperature in equatorial latitudes and the winter hemisphere occurs at the solstices, but the latitudinal gradient of the equilibrium temperature vanishes between 20° and 90° in the summer hemisphere at the corresponding times. Qualitatively, this is consistent with the calculation of Fels (1985); that is, radiative-convective equilibrium temperatures have a substantial latitudinal gradient across the equator and in the winter hemisphere at the solstices. The thermal wind relation, on the other hand implies that there can be no latitudinal temperature gradient at the equator. This incompatibility gives rise to the nonlinear Hadley circulation in the middle atmosphere (Dunkerton 1989). Quantitatively, the maximum latitudinal gradients of \( T_e \) employed in the model are located about 10 km lower than those of the realistic \( T_e \) in the middle and upper stratosphere for the model are substantially smaller than those for the real atmosphere. Here, the choice is based on giving the correct latitudinal gradients of \( T_e \) in the lower tropical stratosphere.

Figure 2 shows the vertical velocities at 20-km altitude in response to the specified seasonally varying thermal forcing. After the initial several months, the vertical velocities from the mechanistic model behave almost periodically following the annual cycle of thermal forcing. The nonlinear Hadley circulation reaches...
its maximum at both solstices, with maximum upwelling of slightly over 0.09 mm s\(^{-1}\) located around 18.5° in the summer hemisphere and maximum downwelling of about −0.11 mm s\(^{-1}\) located around 20.5° in the winter hemisphere.

Figure 3 illustrates the latitudinal profiles of the vertical velocities at 20-km altitude averaged over January, July, and the whole year, respectively. In the seasonal sense, the role of the nonlinear Hadley circulation (NHC) is striking in determining the latitudinal distribution of tropical upwelling; while in the annual mean, its contribution to tropical upwelling is very modest. In other words, the asymmetric part of the differential radiative heating about the equator is significant in determining the latitudinal profile of vertical velocity in the lower stratosphere while the symmetric part of the differential radiative heating about the equator is comparatively insignificant, though not negligible, which is essentially consistent with the study of the tropospheric NHC by Lindzen and Hou (1988), Plumb and Hou (1992), and Walker and Schneider (2005).

The asymmetric differential heating reaches its maximum near the solstices, and the solstitial fields of model output at northern winter solstice are depicted in Fig. 4. The streamfunction in Fig. 4a clearly shows the Hadley cell related with the asymmetric differential heating. The absolute angular momentum in Fig. 4c is a consequence of the increased nonlinearity with height in the tropical stratosphere. The nearly horizontal orientation of the angular momentum contour in the tropical middle and upper stratosphere results from the latitudinal homogenization of absolute angular momentum associated with inertial instability. The thermal trough and ridge in the middle and upper subtropical stratosphere (Fig. 4b) are closely connected with the upwelling and downwelling branches of the NHC in Fig. 4a. The subtle kinks of contours at the equator in Fig. 4b are coupled to the vertical shears of angular momentum, which are constrained by the thermal wind relationship. Since the atmospheric response is very close to linear in the lower stratosphere (Plumb and Eluszkiewicz 1999), Fig. 4d shows that the tropical upwelling and downwelling velocities are nearly antisymmetric about the equator without appreciable equatorial upwelling. However, it also shows that there is substantial tropical upwelling in the upper stratosphere due to the nonlinearity, which contributes to the annual mean tropical upwelling there (Semeniuk and Shepherd 2001b).

**b. Annual variation of wave-driven circulation**

Now, we run the model with imposed wave drag having the form \(d = d_0 Y(\phi)Z(z)\tau(t)\) without any differential radiative heating. Here the peak value \(d_0 = -2.0 \times 10^{-5} \text{ m s}^{-2}\) from northern fall equinox to northern
spring equinox and halved from northern spring equinox to 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, 1997). Here, $Y(\varphi)$ is given by

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

(22)

from northern fall equinox to northern spring equinox with $\varphi_2 = 45^\circ$, $\varphi_3 = 80^\circ$, $\varphi_1 \geq 0$ representing how close wave drag penetrates to the equator, and is the mirror image of (22) about the equator from northern spring equinox to northern fall equinox. Here, $Z(z)$ is given by (18); $\tau(t)$ is defined by

$$\tau(t) = \begin{cases} \sin\left(\frac{\pi}{\tau_1} \left( \frac{t}{t_1} \right) \right) & 0 \leq t \leq t_1 \\
\sin\left(\frac{\pi}{\tau_1} \left( \frac{t - t_1}{t_2 - t_1} \right) \right) & t_1 \leq t \leq t_1 + t_2 \end{cases},$$

(23)

where $t, t_1, t_2$ are the same as those in (19) and (20).

Figure 4 shows the vertical velocities at 20-km altitude responding to the seasonally varying wave drag with its equatorward edge penetrating to $30^\circ, 15^\circ$, and the equator, respectively, in the winter hemisphere. When the equatorward edge is located at $30^\circ$ (top panel) in the winter hemisphere, there is no appreciable tropical upwelling in the equatorial region. As the equatorward edge is located at $15^\circ$ (middle panel) in the winter hemisphere, there is appreciable equatorial upwelling at solstices and weak equatorial downwelling.
during equinoxes. The equatorial upwelling and downwelling reach their maxima of 0.03 mm s$^{-1}$ at northern winter solstice and $-0.03$ mm s$^{-1}$ at northern vernal equinox, respectively.

The surf zone resulting from planetary waves (McIntyre and Palmer 1983, 1984) seen using coarse resolution data does not exclude the existence of unresolved wave drags. GCMs such as the Canadian Middle Atmosphere Model (CMAM; documented in Beagley et al. 1997) and SKYHI do not have the resolved wave drag confined to the surf zone and there is weak Eliassen–Palm (EP) flux divergence extending to the equator. When the applied wave drag extends into the Tropics (Polvani et al. 1995) and tapers to zero at the equator (bottom panel), there is persistent and substantial upwelling in the equatorial region extending from the winter subtropics into the summer subtropics. The seasonal equatorial upwelling reaches a secondary maximum well above 0.08 mm s$^{-1}$ near the equator at austral winter solstice with the principal maximum being above 0.14 mm s$^{-1}$ 14$^\circ$N at boreal winter solstice. The tropical upwelling is quite uniform in the equatorial region at northern winter solstice.

To explore how wave drag penetrating into tropical regions impacts the morphology of the tropical upwelling arising only from the wave-driven circulation (WDC), we shift the equatorward edge of the wave drag from 30$^\circ$ to the equator in intervals of 2.5$^\circ$ while $\varphi_2$ and $\varphi_3$ are kept invariant as in Eq. (22). Figure 6 shows how the latitudinal profile of the vertical velocities at 20-km altitude varies with the latitude of the equatorward boundary of the wave drag changing from 30$^\circ$ in the winter hemisphere to the equator. The monthly mean equatorial upwelling in January (top panel) increases monotonically as the wave drag penetrates more deeply into the Tropics, particularly when the equatorward edge of wave drag is less than 12.5$^\circ$ away from the equator. That is to say, provided the wave drag extends into the internal tropical boundary layer (Plumb and Eluszkiewicz 1999), upwelling is distributed across the Tropics. The monthly mean tropical upwelling in July (middle panel) behaves similarly as that in January except that upwelling is weaker because of halved wave drag.

The annual mean tropical upwelling (bottom panel) also increases while its latitudinal span shrinks as wave drag penetrates closer to the equator. In order for an annual mean equatorial upwelling to exist, the equatorward edge of wave drag must be less than 15$^\circ$ away from the equator, which is consistent with the result of Sankey (1998) and Semeniuk and Shepherd (2001b); that is, wave drag must penetrate quite close to the equator in order to simulate tropical upwelling in the lower stratosphere.

Since wave drag also reaches its maximum near the winter solstices, the solstitial fields of model output at northern winter solstice are shown in Fig. 7 and Fig. 8 for comparison, where the equatorward edge of wave drag is located at 15$^\circ$ away from the equator in the winter hemisphere (Fig. 7) and the equator (Fig. 8). Although there is no annual mean upwelling at the equator in the bottom panel of Fig. 6 when the equatorward edge of wave drag in the winter hemisphere is located at 15$^\circ$ away from the equator, the wave drag induces a meridional cell of large latitudinal extent (Fig. 7a) reaching into the summer subtropics at boreal winter solstice with its primary maximum upwelling in the subtropics of the northern lower stratosphere and its secondary maximum upwelling in the subtropics of

![Fig. 5. Vertical velocity at 20 km forced by seasonally varying wave drag with its equatorward edge reaching (top) 30°, (middle) 15°, and (bottom) the equator in the winter hemisphere. The contouring is the same as that in Fig. 2.](image-url)
the southern middle stratosphere (Fig. 7d), which correspond to the two cold regions in Fig. 7b. The upwelling in the winter subtropics is approximately governed by the downward control principle (Haynes et al. 1991). The upwelling in the summer subtropics is associated with the zonal flow transience because the streamlines cross angular momentum contours (Figs. 7a,c) in the summer hemisphere. When the equatorward edge of the wave drag in the winter hemisphere reaches the equator, the overall stratosphere–troposphere mass exchange is increased by 22% at northern winter solstice (Figs. 7a, 8a). The tropical upwelling spreads quite symmetrically about the equator, with one weak peak in the subtropics of the winter lower stratosphere and another weak peak in the subtropics of the summer upper stratosphere (Fig. 8d). Figure 8b shows that the temperature distribution fits in with that of vertical velocity. The absolute angular momentum in Fig. 8c is well homogenized in the vicinity of the equator around 40-km height, which partly accounts for the maximum upwelling in the subtropics of the summer upper stratosphere (Tung and Kinnersley 2001).

c. Annual variation of the residual mean meridional circulation by both radiative differential heating and wave drag

Figure 9 shows the vertical velocities at 20-km altitude in response to the specified seasonally varying radiative differential heating in section 3a and wave drag in section 3b with the equatorward edge of the latter penetrating to 30°, 15°, and the equator, respectively, in the winter hemisphere while ϕ2 and ϕ3 are the same as in Eq. (22). After an initial several months, the vertical velocities from the mechanistic model behave nearly periodically following the annual cycle of thermal forcing and wave drag. As the equatorward edge of the drag is shifted from 30° (top panel) to 15° (middle panel) in the winter hemisphere, the equatorial upwelling also increases substantially. The bottom panel of Fig. 9 shows the modeled vertical velocities at 20-km altitude with wave drag extending into Tropics and tapering to zero at the equator. The results are similar to those deduced from tracer observations (Boering et al. 1996; Hall and Waugh 1997; Mote et al. 1996; Niwano et al. 2003) and from radiative calculations (Rosenlof 1995; Eluszkiewicz et al. 1996, 1997; Plumb and Eluszkiewicz 1999) in that

1) the magnitude of our modeled upwelling velocity is similar to that observed with maximum ascent rates of about 0.21 and 0.16 mm s⁻¹ in northern and southern winter, respectively;
2) equatorial upwelling persists throughout the year; and
3) the latitude of maximum upwelling migrates toward the summer subtropics at both solstices.

To explore how wave drag penetrating into tropical region impacts the morphology of tropical upwelling caused by both the NHC and WDC, we shift the equatorward edge of wave drag from 30° to the equator in intervals of 2.5° while ϕ2 and ϕ3 are the same as before. Figure 10 shows how the latitudinal profile of the vertical velocities at 20-km altitude varies with the latitude.

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**Fig. 6.** Latitudinal distribution of vertical velocities at 20 km forced by seasonally varying wave drags in the winter hemisphere whose equatorward edges range from 30° to 0° away from the equator as indicated by abscissa. Contour interval is 0.025 mm s⁻¹. Solid, dash–dotted, and dotted lines denote positive, zero, and negative contours, respectively, for (top) January, (middle) July, and (bottom) annual mean vertical velocity.
of the equatorward boundary of wave drag changing from 30° in the winter hemisphere to the equator.

The monthly mean equatorial upwelling in January (top panel) increases monotonically as wave drag penetrates deeper into the Tropics. When the tropical edge of wave drag is more than 12.5° away from the equator, there is one upwelling band caused by wave drag in the northern subtropics, an upwelling band caused by the NHC in the southern Tropics, and a downwelling band in between. As wave drag penetrates further toward the equator, the downwelling band in between disappears; but the double-peak structure of tropical upwelling is maintained. After the equatorward edge of wave drag is no more than 2.5° away from the equator, tropical upwelling exhibits a single-peak structure consistent with results deduced from observed data. Thus, both single- and double-peak structures in the lower stratospheric upwelling velocity in January can be simulated and may arise from interannual variations in the wave drag.

The monthly mean tropical upwelling in July (middle panel) behaves somewhat differently from January. When the tropical edge of wave drag is more than 7.5° away from the equator, there is one upwelling band caused by wave drag in the southern subtropics, another upwelling band caused by the NHC in the north-
ern Tropics, and a downwelling band in between. As wave drag penetrates further toward the equator, the downwelling band in between disappears and the upwelling no longer exhibits a double-peak structure.

The annual mean tropical upwelling caused by both the NHC and WDC (bottom panel) also increases while its latitudinal span shrinks as wave drag penetrates closer to the equator. When the tropical edge of wave drag is more than 7.5° away from the equator, the annual mean upwelling shows the triple-peak structure: the two peaks in the northern and southern subtropics result from the wave drags in both hemispheres during the winter seasons and the equatorial peak results from the interaction between the NHC and WDC. As wave drag penetrates further toward the equator, the upwelling has only one peak at the equator.

As pointed out by Scott (2002) and Haynes (2005), it is possible, in principle, for the wave forcing to stop abruptly at say, 20°, but this situation is not likely relevant to the real atmosphere where the wave drag required to sustain an upwelling circulation becomes smaller and smaller as the equator is approached. In any model that resolves the large-scale waves, it is very difficult to rule out any penetration of the waves and hence the wave drag to the Tropics (Scott 2002). The realistic behavior of breaking and dissipating gravity waves and, above all, Rossby waves likely leads to small wave forces at low latitudes (Haynes 2005).

Some overall features of the residual circulation may be estimated from observations (e.g., Rosenlof 1995; Niwano et al. 2003), however a more complete picture of the residual circulation may be obtained from realistic comprehensive GCM simulations, and these can provide an interesting comparison for the present

![Fig. 8. Same as Fig. 7 except for the equatorward edge of wave drag located at the equator.](image)
mechanistic model results. Here we examine results from the Geophysical Fluid Dynamics Laboratory (GFDL) SKYHI troposphere–stratosphere–mesosphere GCM (Hamilton et al. 1995, 2001). The SKYHI model is formulated with simple centered differencing on model levels that are isobaric (at pressures less than 321 hPa) and so, as noted by Andrews et al. (1983) and Hamilton and Mahlman (1988), the transformed residual circulation in the middle atmosphere can be deduced in a completely consistent manner with the model dynamics (so that results can be viewed as exact to within machine precision). The wave drag is generated self-consistently via explicitly resolved eddies as well as subgrid-scale heat and momentum transports (which are generally very small relative to the explicitly resolved eddy terms; e.g., Hamilton and Mahlman 1988).

In the case examined in this paper the SKYHI model was run at 2° horizontal resolution and with 80 levels in the vertical from the ground to 0.01 hPa (~80 km). This version produces a fairly realistic zonal-mean temperature and wind field, suggesting that the wave drags generated are also fairly realistic (Hamilton et al. 2001). Figure 11 shows the monthly mean residual vertical velocity for January at 20 km computed from 10 yr of this SKYHI model simulation along with the present mechanistic model result. Here, the wave drag imposed in the mechanistic model reaches 7.5° latitude in the winter hemisphere. The mechanistic model largely mimics the feature of the low-latitude vertical velocity from this SKYHI model run. In January, there exist a maximum upwelling in the summer hemisphere, a secondary maximum upwelling and a minimum tropical upwelling in between in the winter hemisphere for both
models. 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 the imposed radiative differential heating is also smaller than that in the real atmosphere and SKYHI model. Another difference is that the SKYHI model shows conspicuous downwelling in summer hemisphere high latitudes in January. Our mechanistic model does not simulate this feature since no wave drags are specified in the summer hemisphere (Alexander and Rosenlof 1996), which results in downwelling in high latitudes and a shifting poleward of the primary maximum upwelling.

Niwano et al. (2003) also pointed out the lower stratospheric double-peak structure of vertical velocity in the northern winter as inferred from Upper Atmosphere Research Satellite (UARS) HALOE trace gas data (see their Fig. 7). Rosenlof (1995) illustrated the lower stratospheric double-peak structure of vertical velocity in the southern winter derived from radiative calculations (see her Fig. 13).

The comparison of our mechanistic modeling with results from observed data and SKYHI model suggests that both the Hadley circulation and wave-driven circulation are crucial in determining the seasonal cycle of tropical upwelling.

d. Interaction between the NHC and the WDC

We further examine what role the nonlinear interaction between the NHC and the WDC plays in determining the latitudinal morphology of tropical upwelling. Figure 12 shows the difference between the velocities in Fig. 10 and those of the linear superposition of the NHC and WDC at 20-km altitude. The following inferences can be made from these two figures.

1) To a first-order of approximation, the tropical upwelling at 20 km is determined by the linear superposition of NHC and WDC. Figures 2, 3, and 4 indicate that the NHC is significant in determining the latitudinal profile of the vertical velocity in the lower stratosphere, particularly during the solstitial seasons. In the lower stratosphere, the NHC gives rise to tropical upwelling in the summer hemisphere and tropical downwelling in the winter hemisphere. Figures 5, 7, and 8 demonstrate that the WDC results in tropical upwelling and causes the down-
welling in the winter middle and high latitudes. Thus, there is large cancellation of vertical velocities between the NHC and WDC in the Tropics of the winter lower stratosphere, which is responsible for the lower stratospheric double-peak structure of vertical velocity in the northern winter found by Niwano et al. (2003).

2) However, nonlinearity increases the equatorial upwelling and decreases the subtropical upwelling. The role of nonlinearity increases as the equatorward edge of wave drag penetrates closer to the equator. Physically, this is due to the interactions between the WDC and the altered angular momentum distribution that results from the NHC. The asymmetric radiative differential heating across the equator and wave drag maximize around the winter solstice. Along with Figs. 10 and 12, Figs. 4a,c show the cross-equatorial flow of the NHC and nearly horizontal orientation of the angular momentum contours in the winter Tropics. This facilitates the meridional extension of the WDC into the summer hemisphere and results in increased tropical upwelling there. These figures also show that the poleward tilting of angular momentum contours in the subtropics of the winter middle and upper stratosphere and decreased latitudinal gradient of angular

Fig. 13. The model state forced by both radiative differential heating and wave drag at northern winter solstice with the equatorward edge of wave drag located at 15°: (a) mass streamfunction ($10^8$ kg m$^{-1}$ s$^{-1}$); (b) departure of temperature from a horizontally uniform reference profile (K); (c) absolute angular momentum ($10^8$ m$^2$ s$^{-1}$); (d) vertical velocity (mm s$^{-1}$). Contour intervals in (a), (b), and (c) are 4, 1, and 1, respectively. Solid and dotted lines correspond to positive and negative contours while the zero contours are omitted in (a). For clarity, an additional isoline of 29.5 is added in (c). Contour interval in (d) is 0.05, where solid, dash–dotted, and dotted lines denote positive, zero, and negative contours, respectively.
momentum not only increase the in situ meridional flow forced by wave drag, but also help to extend the influence of wave drag in the subtropics of the winter middle and upper stratosphere into the Tropics of the winter lower stratosphere along angular momentum contours. This also gives rise to increased tropical upwelling in the winter lower stratosphere.

In the presence of both radiative differential heating and wave drag, the solstitial fields of model output at northern winter solstice are shown in Fig. 13 and Fig. 14, where the equatorward edge of wave drag is located at $15^\circ$ away from the equator in the northern winter hemisphere (Fig. 13) and the equator (Fig. 14). Figures 13a,c clearly show the lower stratospheric double-peak structure of vertical velocity while Figs. 14a,d illustrate the lower stratospheric single-peak structure of tropical upwelling. These figures also demonstrate the contribution of nonlinearity to persistent tropical upwelling in the upper stratosphere. Figures 7a and 13a indicate that the radiative differential heating does not tangibly affect the overall stratosphere–troposphere mass exchange when the equatorward edge of wave drag is located $15^\circ$ away from the equator in the winter Northern Hemisphere. However, Figs. 8a and 14a reveal that the radiative differential heating increases the overall stratosphere–troposphere mass exchange by 9% due to the nonlinear interaction between NHC and WDC when wave drag in the winter Northern Hemisphere penetrates to the equator. Compared with Fig. 8c, Fig. 14c shows the increased vertical shear of angular momentum in the tropical stratosphere. In comparison with Figs. 13b,c, Figs. 14b,c display increased horizontal curvature of temperature and increased vertical shear.

Figure 14. Same as Fig. 13 except for the equatorward edge of wave drag located at the equator.
of angular momentum as the wave drag penetrates from 15°N to the equator at northern winter solstice.

The vertical velocity averaged between 30°S and 30°N in the lower stratosphere removes most of the asymmetric part of the nonlinear Hadley circulation. Here, the tropical mean upwelling can still be estimated by the downward control principle employed by Holton (1990) and Rosenlof and Holton (1993). Figure 15 shows the agreement between the vertical velocities at 20-km altitude averaged from 30°S to 30°N forced by both thermal forcing and wave drag and those induced only by wave drag. No matter whether the wave drag penetrates to 15° or the equator in the winter hemisphere, the tropical mean upwelling at 20-km altitude forced by both thermal forcing and wave drag is very close to that forced by both wave drag and thermal forcing. This is because angular momentum is horizontally stratified beyond 30° latitude in both hemispheres (see Figs. 7, 8, 13, and 14). Thus, downward control principle is approximately valid.

However, the vertical velocities at 20-km altitude averaged from 15°S to 15°N forced by both thermal forcing and wave drag are substantially larger than those induced only by wave drag (figure is not shown here). This is because angular momentum of the former is not horizontally stratified around 15° latitude in the winter hemisphere. Thus, the downward control principle is no longer valid in this region. The nonlinearity acts to increase the equatorial upwelling.

5. Conclusions and discussions

The nonlinear Hadley circulation, wave-driven circulation, and their interaction determine the latitudinal morphology of tropical upwelling. The nonlinear Hadley circulation plays a striking role in the seasonal variation of tropical upwelling, which accounts for why the latitude of maximum upwelling migrates toward the summer subtropics at both solstices. In the annual mean sense, the nonlinear Hadley circulation per se makes only a modest contribution to tropical upwelling in the lower stratosphere while it significantly contributes to tropical upwelling in the upper stratosphere.

The wave-driven circulation is important not only in the seasonal variation of tropical upwelling but also in the annual mean tropical upwelling. The extent to which the wave drag penetrates toward the equator determines whether tropical upwelling in the lower stratosphere exhibits a single-peak structure or double-peak structure. The existence of both single- and double-peak structures in the lower stratospheric upwelling velocity inferred from observed data may arise from interannual variations in the wave drag.

Nonlinear interaction between the Hadley circulation and wave-driven circulation is important in explaining persistent equatorial upwelling in the lower stratosphere. The simple mechanistic model results are similar to those of SKYHI and the observationally inferred tropical upwelling. Results of both the mechanistic and SKYHI simulations can be interpreted as indicating that both the nonlinear Hadley circulation and a wave-driven (downward control) component interact to determine the configuration of tropical upwelling in the stratosphere.

The results of this paper suggest a possible resolution to the apparent contradiction between the observed increase of water vapor in the stratosphere prior to 1996 (Randel et al. 1999; Rosenlof et al. 2001), only half of which may be explained by methane increases in the troposphere, and the decrease of cold point tropopause (CPT) temperatures (Zhou et al. 2001a). Holton and Gettelman’s (2001) study indicates the importance of the region of coldest tropical tropopause layer (TTL) temperatures in determining stratospheric water vapor mixing ratios. Both the Brewer–Dobson tropical up-
well ing (Yulaeva et al. 1994; Randel et al. 2002) and convection (Highwood and Hoskins 1998; Zhou et al. 2001b, 2004) are important in determining these cold trap temperatures.

In the presence of the cooling minimum CPT temperatures that may have been caused by enhanced convection (as suggested by Zhou et al. 2001a), if the wave drag shifts poleward away from the equatorial region, the equatorial upwelling would in turn decrease while the subtropical upwelling and the latitudinal span of upwelling region would increase. This results in less upward transport of colder and drier air into the stratosphere in the deep Tropics and more upward transport of warmer and moister air into the stratosphere through the subtropical tropopause, which could cause moistening of the stratosphere. Of course, much remains to be done to establish whether this speculation is actually the case.

Our results show that both the Hadley and the wave-driven circulations are crucial in determining the nature of tropical upwelling in the lower stratosphere. A key result of our investigation is the significance of the extent to which extratropical planetary wave drag penetrates toward the equator. It is important to explore this point further in future studies.

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