Influence of Gravity Waves in the Tropical Upwelling: WACCM Simulations

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ABSTRACT

The annual cycle of tropical upwelling and contributions by planetary and gravity waves are investigated from climatological simulations using the Whole Atmosphere Community Climate Model (WACCM) including three gravity wave drag (GWD) parameterizations (orographic, nonstationary background, and convective GWD parameterizations). The tropical upwelling is estimated by the residual mean vertical velocity at 100 hPa averaged over 15°S–15°N. This is well matched with an upwelling estimate from the balance of the zonal momentum and the mass continuity. A clear annual cycle of the tropical upwelling is found, with a Northern Hemisphere (NH) wintertime maximum and NH summertime minimum determined primarily by the Eliassen–Palm flux divergence (EPD), along with a secondary contribution from the zonal wind tendency. Gravity waves increase tropical upwelling throughout the year, and of the three sources the contribution by convective gravity wave drag (CGWD) is largest in most months. The relative contribution by all three GWDs to tropical upwelling is not larger than 5%. However, when tropical upwelling is estimated by net upward mass flux between turnaround latitudes where upwelling changes downwelling, annual mean contribution by all three GWDs is up to 19% at 70 hPa by orographic and convective gravity waves with comparable magnitudes. Effects of CGWD on upwelling are investigated by conducting an additional WACCM simulation without CGWD parameterization. It was found that including CGWD parameterization increases tropical upwelling not only directly by adding CGWD forcing, but also indirectly by modulating EPD and zonal wind tendency terms in the tropics.

1. Introduction

A recent observational study by Thompson and Solomon (2009) revealed that a decreasing temperature trend in the lower stratosphere during 1979–2006, except for the three years following the El Chichón and Mount Pinatubo eruptions, is strongly related to the global overturning circulation. Upwelling equatorward of 50° and downwelling in the polar regions are respectively enhancing and attenuating the ozone-induced cooling trend. This provides additional observational support for acceleration of the stratospheric mean meridional circulation, the so-called Brewer–Dobson (BD) circulation, in the warming atmosphere, as revealed in various recent climate change simulations (e.g., Butchart et al. 2006; Olsen et al. 2007; Fomichev et al. 2007; Li et al. 2008; Garcia and Randel 2008; McLandress and Shepherd 2009; Butchart et al. 2010).

Tropical upwelling, the upward branch of the BD circulation, is important in the stratospheric climate, as it influences the concentration of water vapor entering the stratosphere, which in turn affects radiation and chemistry in the middle atmosphere. Observational studies (e.g., Randel et al. 2003; Kerr-Munslow and Norton 2006) show a clear annual cycle in the tropical upwelling, with a maximum during the Northern Hemisphere (NH) winter and a minimum during the NH summer. It has been recognized that tropical upwelling, as a part of the BD circulation, is driven primarily by extratropical wave forcing due to dissipation of planetary and gravity waves (Holton et al. 1995), and its annual cycle could be related to the annual cycle of planetary-wave forcing. However, Plumb and Eluszkiewicz (1999) showed from their numerical simulations that the extratropical wave forcing is not sufficient to drive upwelling near the equator and that wave drag within 20° of the equator is required. They suggested that vertically propagating equatorial planetary and/or gravity waves may contribute to the angular momentum budget in the tropical region, analogous to the role played by viscosity in their model. Recently, there have been increasing numbers of observational ...
(e.g., Kerr-Munslow and Norton 2006; Randel et al. 2008, hereafter RGW08) and numerical modeling (e.g., Norton 2006; Taguchi 2009 (hereafter T09), 2010; Ryu and Lee 2010] studies that have emphasized the role of the tropical and subtropical wave driving in the annual cycle of tropical upwelling.

Despite recent progress in our understanding of wave driving in the tropical upwelling, it is not clear yet which waves are involved in this process, where they dissipate, and how they interact with each other to drive tropical upwelling. One example that has been explored more recently is the role of gravity waves. Using a high-resolution (horizontal grid spacing of 0.56° and vertical grid spacing of about 300 m) general circulation model (GCM) that explicitly resolves small-scale waves including gravity waves, Miyazaki et al. (2010) showed that the Eliassen–Palm flux divergence (EPD) of the small-scale gravity waves induces mean equatorward flow in the extratropical tropopause region, which partially cancels the poleward flows induced by the planetary and synoptic waves, while it induces mean poleward flow in the subtropical lower stratosphere and mean downward flow in the midlatitude lower stratosphere. From chemistry–climate models that performed twenty-first century reference simulations—the Atmospheric Model with Transport and Chemistry (AMTRAC) by Li et al. (2008), Canadian Middle Atmosphere Model (CMAM) by McLandress and Shepherd (2009), Center for Climate System Research (CCSR)–National Institute for Environmental Studies (NIES) model by Okamoto et al. (2011), and 11 chemistry–climate models including three models by Butchart et al. (2010)—it is shown that the parameterized orographic gravity wave drag (GWD) can contribute to the long-term trend in the net upward mass flux at 70 hPa up to 40%–59% during the NH wintertime. These studies suggested that this rather significant contribution by orographic gravity wave drag is due to an enhanced and upward-shifted gravity wave drag under the stronger subtropical jets in the lower stratosphere as a consequence of climate change, which allows more planetary waves and gravity waves to propagate into the stratosphere.

So far, there have been no proper studies on the contribution of gravity waves generated by individual sources other than orography to the tropical upwelling. One candidate is the gravity waves generated by convective clouds, which can provide significant drag in the tropical lower stratosphere as shown in several GCMs with a convective gravity wave parameterization (e.g., Chun et al. 2004; Song et al. 2007; Jeon et al. 2010; Richter et al. 2010). Also, when the tropical upwelling is estimated by downward mass flux in the extratropical regions, as done by Li et al. (2008), McLandress and Shepherd (2009), and Butchart et al. (2010), gravity waves generated by fronts/jet streams (e.g., Charron and Manzini 2002; Richter et al. 2010) could be important. Because there is no proper observational dataset to derive momentum forcing induced by gravity waves generated by individual sources, using GCM results that include individual GWD parameterizations might be one feasible way. The gravity waves parameterized in GCMs represent subgrid-scale gravity waves with horizontal wavelengths smaller than a few hundred kilometers, and they are the ones that are missed mostly even in current high-resolution GCMs and that are most important for momentum budget in the mesosphere.

In the present study, we examine the annual cycle of tropical upwelling based on a 12-yr simulation of the Whole Atmosphere Community Climate Model (WACCM) including three GWD parameterizations (orographic, nonstationary background, and convective). Contributions of the resolved planetary waves and parameterized gravity waves generated by individual sources are presented. The impact of convective gravity waves, which could be important, especially in the tropical region, is also investigated by comparing results from an additional WACCM simulation without convective GWD parameterization.

2. Experimental design

The climate model used in this study is the WACCM version 1b (WACCM1b) developed at the National Center for Atmospheric Research (NCAR) (Sassi et al. 2002). The WACCM1b is a global spectral model with T63 horizontal resolution at 66 vertical levels from the surface to about 140 km. The model description and physical processes included in WACCM1b can be found in Song et al. (2007). The two 12-yr simulations with and without convective GWD parameterization are performed from an initial condition of 1 July 1978 using the climatological ozone (Wang et al. 1995) and sea surface temperature (SST) (Shea et al. 1992). The constant sea ice thickness is assigned when the SST is less than about −1.8°C (Collins et al. 2004). We refer to the simulations with and without convective GWD parameterization as the GWDC and CTL simulations, respectively. The GWDC and CTL simulations both include the orographic GWD parameterization by McFarlane (1987) and the background GWD parameterization that represents nonstationary gravity waves observable in the atmosphere based on Lindzen (1981). The convective GWD parameterization implemented in the GWDC simulation is the ray-based parameterization to represent the three-dimensional propagation of GWs proposed by Song and Chun (2008). A recent study by Choi et al. (2009) showed that the ray-based parameterization by Song and Chun (2008) represents...
the global temperature variances observed in Microwave Limb Sounder (MLS) on the Upper Atmosphere Research Satellite (UARS) better than the columnar parameterization by Song and Chun (2005). The model climatology for each simulation is made by averaging results over the last 10 yr after a spinup period of 2 yr. The results shown in section 3 are based on the GWDC simulation except in section 3b, where differences between the GWDC and CTL simulations will be presented.

3. Results

a. GWDC simulation

In the present study, the tropical stratospheric upwelling is estimated by the residual mean vertical velocity \( \bar{w}^* \), which is defined following Andrews et al. (1987) as

\[
\bar{w}^* = \bar{w} + \frac{1}{a \cos \varphi} \partial \left( \cos \varphi \frac{\bar{w}' \theta'}{\bar{\theta}' / \partial z} \right),
\]

where \( w \) is the vertical wind, \( v \) is the meridional wind, \( \theta \) is the potential temperature, and \( a \) is the radius of the earth. The overbar denotes the zonal mean, and the departure from the zonal mean is denoted by a prime.

Figure 1 shows the annual cycle in \( \bar{w}^* \) over the tropical region \(( \pm 30^\circ)\) at 100 hPa (Fig. 1a) and its latitudinal average between 10°S and 10°N along with zonal-mean temperature (Fig. 1b). Figure 1a shows a clear annual cycle of the upwelling: maximum during the NH wintertime and minimum during the NH summertime with the latitudes of the maximum being shifted northward. Maximum upwelling is 0.97 mm s\(^{-1}\) during February at 4.2°S; minimum is 0.34 mm s\(^{-1}\) during August at 26.5°N. The maximum value during February is relatively large compared with that from the 50-yr WACCM simulation by T09 of about 0.6 mm s\(^{-1}\) during January. Note that the WACCM simulation in T09 did not include convective GWD parameterization, so it might be similar to the present CTL simulation. Although including the convective GWD parameterization increases the upwelling, especially during February and November, as will be shown later, this magnitude difference is not solely from the convective GWD parameterization, as the difference in the upwelling between the GWDC and CTL simulations is one order of magnitude smaller than that in the CTL simulation. One of reasons may be the different averaging period (50 yr in T09 and 10 yr in the present study), although interannual variation associated with quasi-biennial oscillation (QBO) could not be reproduced in both simulations. In the latitudinal average near the equator (Fig. 1b), seasonal variation of the upwelling is clearer. This is strongly related to the annual cycle of the temperature in the tropical lower stratosphere, as is also revealed in the global reanalysis datasets by Kerr-Munslow and Norton (2006) and in the 50-yr WACCM simulation by T09.

The tropical stratospheric upwelling averaged over the latitudes of \( \varphi_1 \) and \( \varphi_2 \) also can be estimated from the transformed Eulerian mean (TEM) equations (Andrews et al. 1987) of zonal momentum and continuity, following RGW08 as

\[
\langle \bar{w}^* \rangle_\mathcal{M} \langle \varphi, \rho \rangle = \int_{\varphi_1}^{\varphi_2} a \cos \varphi \, d\varphi \left[ -\cos \varphi \int_{\varphi_2}^{\varphi_1} \rho_0(\varphi) \frac{\partial}{\partial \varphi} \left( \bar{w}' \theta' \right) \frac{d\varphi}{\partial \varphi} \right]_{\varphi_1}^{\varphi_2}.
\]

FIG. 1. Annual cycle in \( \bar{w}^* \) defined in (1) (a) over the tropical region \(( \pm 30^\circ)\) at 100 hPa and (b) its latitudinal average over 10°S–10°N (solid) along with zonal-mean temperature at 100 hPa (dotted) from the GWDC simulation. Contour interval of (a) is 0.1 mm s\(^{-1}\) and negative values are shaded.
where \( \rho_0 \) is the basic-state density that decreases exponentially with height, \( \mathbf{W}(\phi, z) \) represents wave forcing by planetary and gravity waves, \( \mathbf{n}(\phi, z) \) is the zonal wind tendency, and \( f = f - \frac{1}{a \cos \phi} \left( \frac{\partial}{\partial \phi} \mathbf{\bar{u}} \cos \phi \right) \). The momentum-balance estimation by (2) generalizes the calculated from gravity wave drag from the parameterization. The EPD is downward control (Haynes et al. 1991) by including momentum-balance estimation by (2) generalizes the zonal wind tendency. The subscript \( m \) in \( \langle \mathbf{w}_m^* \rangle \) denotes momentum. The wave forcing consists of Eliassen–Palm averaging over 15\(^\circ\) latitude band poleward of 70\(^\circ\) (Fig. 2a) and the annual cycle of the tropical upwelling at 100 hPa estimated from (1) and (2) and the annual cycle of the tropical upwelling averaged over 15\(^\circ\)S–15\(^\circ\)N (Fig. 2b). Note that the estimate of \( \langle \mathbf{w}_m^* \rangle \) from (2) is problematic near the equator because of the proportionality to \( \frac{1}{f} \). Based on RGW08, climatological calculations of \( \langle \mathbf{w}_m^* \rangle \) from (2) are reasonably well behaved for latitudes higher than 15\(^\circ\). Therefore, in (2), \( \langle \mathbf{w}_m^* \rangle \) is calculated for a latitude band of 15\(^\circ\)S–15\(^\circ\)N in the tropics and for each individual 5\(^\circ\) latitude band poleward of 15\(^\circ\). Figure 2 shows clearly that the two upwelling estimates are well matched in most latitudes except in the NH polar region where \( \langle \mathbf{w}_m^* \rangle \) is larger than \( \mathbf{w}_m^* \). The degree of matching the two upwelling estimates equatorward of 70\(^\circ\) is much higher than that in the previous study by RGW08 calculated using reanalysis data (Fig. 3 of RGW08) and WACCM simulations (Fig. 4b of RGW08). Figure 2b, which shows upwelling estimates in the tropical region averaged over 15\(^\circ\)S–15\(^\circ\)N, demonstrates a good match between the two estimates throughout the year. This result implies that the tropical upwelling estimate from the balance of the zonal momentum and
the continuity by (2) can be used to understand the annual cycle of tropical upwelling through detailed analyses of individual forcing terms.

Figure 3 shows latitude–height cross sections of (top) the zonal-mean zonal wind, and the EPD, BGWD, CGWD, and OGWD forcing terms during (middle) January and (bottom) July. The zonal-mean zonal wind shown in Fig. 3 exhibits the characteristic features of the zonal wind observed in solstice seasons such as the hemispheric difference in the strength of the polar night jets, zonal wind reversal near the mesopause level, and the cross-equatorial westerly flow from the winter mesosphere to the summer thermosphere. However, compared with Upper Atmosphere Research Satellite (UARS) Reference Atmosphere Project (URAP) (Swinbank and Ortland 2003), the zonal-mean zonal wind above $z = 90$ km in the present simulation is significantly small, as also shown in Song and Chun (2008). Although this discrepancy may induce unrealistic wave forcings in the upper mesosphere and lower thermosphere, this is unlikely to influence in the present upwelling calculation significantly, given that the wave forcing below $z = 50$ km is effective to the upwelling estimation at 100 hPa, as will be shown in Fig. 4. In EPD, negative forcing is dominant in the troposphere and most of the winter middle atmosphere. During January, the maximum negative

![Figure 3](image-url)
forcing is \(-19\) m s\(^{-1}\) day\(^{-1}\) near \(z = 60\) km at \(45^\circ\)N and maximum positive forcing is \(10\) m s\(^{-1}\) day\(^{-1}\) in the NH polar stratopause. During July, their magnitudes increase to \(-21\) and \(12\) m s\(^{-1}\) day\(^{-1}\), respectively, both at similar heights and latitudes in the Southern Hemisphere (SH). The magnitude of BGWD is largest among the four wave forcing terms, with maximum positive forcing during January of \(127\) m s\(^{-1}\) day\(^{-1}\) in the SH midlatitudes near \(z = 80\) km and the maximum negative forcing during January of \(-60\) m s\(^{-1}\) day\(^{-1}\) in the NH midlatitudes near \(z = 60\) km. During July, the areas of negative and positive BGWD forcing are reversed in the different hemispheres, with a slightly less positive forcing maximum (\(124\) m s\(^{-1}\) day\(^{-1}\)) and significantly larger negative forcing maximum (\(-88\) m s\(^{-1}\) day\(^{-1}\)).

The CGWD forcing also shows clear seasonal variation, with negative forcing in the winter mesosphere subtropics and midlatitudes and positive forcing in the tropics and most of the summer hemisphere above \(z = 60\) km. The maximum positive and negative forcing during January are \(31\) and \(-7.6\) m s\(^{-1}\) day\(^{-1}\), respectively, while they are much larger during July at \(41\) and \(-49\) m s\(^{-1}\) day\(^{-1}\), respectively. The stronger negative forcing during the SH midlatitude wintertime compared with the NH wintertime is due to the larger SH wintertime convective forcing and resultant cloud-top momentum flux in the storm-track region in which the main convective sources exist during the winter. The OGWD is restricted below \(30\) hPa (\(z \approx 24.5\) km) in the model, and its magnitude is much less than the other three forcing terms. As will be shown below, however, its contribution can be significant when the density is weighted to the forcing terms.

Note in (2) that the wave forcing contributes to the upwelling with a weighting of density. Therefore, the forcing terms shown in Fig. 3 are not the ones that directly contribute to the vertical integration. To understand the forcing that actually contributes to the upwelling calculation in (2) at \(100\) hPa, effective wave forcing multiplied by \(\rho(z)/\rho(100\) hPa\) is shown in Fig. 4. The negative (positive) values are dashed (solid) lines, while areas with values less (larger) than \(-0.01\) m s\(^{-1}\) day\(^{-1}\) (0.01 m s\(^{-1}\) day\(^{-1}\)) are shaded in blue (orange). The thick black solid lines denote the height at which a 90% contribution is made for each individual forcing term in the vertical integration of (2). The figure clearly shows that effective

![Fig. 4. Latitude–height cross sections of the effective (a) EPD, (b) BGWD, (c) CGWD, and (d) OGWD forcing terms that are multiplied by \(\rho(z)/\rho(100\) hPa\) during (top) January and (bottom) July. The thick line in each plot is the height at which 90% of contribution is made in the vertical integration of (2). Contours are 0, 0.005, 0.01, 0.02, 0.05, 0.1, 0.2, and 0.5 m s\(^{-1}\) day\(^{-1}\), and negative values are dashed. Values less than \(-0.01\) m s\(^{-1}\) day\(^{-1}\) are shaded in blue and those greater than 0.01 m s\(^{-1}\) day\(^{-1}\) are shaded in orange.](image-url)
EPD forcing predominates, with mostly negative values except in the winter hemisphere polar region. To represent 90% of the upwelling at 100 hPa, vertical integration of EPD up to 50 km is required in the tropics and winter hemisphere midlatitudes, whereas vertical integration of EPD below 30 km is enough in the summer hemisphere. Among the three effective GWD forcing terms, effective OGWD forcing is largest in the winter hemisphere subtropics and midlatitudes below $z = 25$ km. The magnitudes of the effective BGWD and CGWD forcing terms are comparable, although effective BGWD forcing is distributed widely in latitude. For the effective BGWD and CGWD forcing terms, integration up to 70 km is required in most regions, except in the summer hemisphere midlatitudes where integration of CGWD forcing below 30 km is enough. This result implies that integration of the effective forcing in a sufficiently deep vertical layer is required to estimate balanced upwelling accurately, especially for nonorographic GWD forcing and even for the EPD forcing in the tropical region.

Figure 5 shows latitudinal distributions of $\langle \pi_{m}^* \rangle$ contributed by individual forcing terms of (left) EPD and OGWD and (right) BGWD and CGWD during (a) January and (b) July. Note that the $y$-axis scaling is different in the left and right panels.

EPD forcing predominates, which is positive equatorward of about $20^\circ$ and poleward of $75^\circ$N and negative elsewhere, with the maximum magnitude in the winter hemisphere midto high latitudes. Among the three GWD parameterizations, the contribution by OGWD is largest, although it is restricted to the subtropics and midlatitudes where mountains exist. During January, the magnitude of the OGWD subtropical positive spike in $\langle \pi_{m}^* \rangle$ near $26^\circ$N is larger than that of the midlatitudinal negative spike near $46^\circ$N. Downwelling by EPD near $20^\circ$–$30^\circ$N is largely cancelled by upwelling by OGWD, while that near $37^\circ$–$60^\circ$N is enhanced by OGWD there. The contributions of BGWD and CGWD are one order of magnitude smaller than those of EPD and OGWD. BGWD provides upwelling equatorward of about $50^\circ$ while upwelling by CGWD is restricted to equatorward of about $30^\circ$. Contributions of CGWD and BGWD are opposite in $20^\circ$–$46^\circ$N and $30^\circ$–$43^\circ$S, with a larger contribution by CGWD, while those poleward of about $50^\circ$ are in the same sign, with a larger contribution by BGWD. The features shown during January appear mostly during July with hemispherical shift according to the seasonal change of circulation. However, the magnitudes of $\langle \pi_{m}^* \rangle$ during July by EPD and OGWD are much smaller and by
CGWD and BGWD are mostly larger than those during January.

Figure 6 shows the annual cycle of $\langle \overline{w'} \rangle$ at 100 hPa averaged over 15°S–15°N contributed by each forcing term in (2): (a) by two major forcing terms (EPD and $dU/dt$) and total values, and (b) by gravity wave forcing (CGWD, OGWD, and BGWD) terms.

![Graph showing annual cycle of $\langle \overline{w'} \rangle$ at 100 hPa](image)

**Fig. 6.** Annual cycle of $\langle \overline{w'} \rangle$ at 100 hPa averaged over 15°S–15°N contributed by each forcing term in (2): (a) by two major forcing terms (EPD and $dU/dt$) and total values, and (b) by gravity wave forcing (CGWD, OGWD, and BGWD) terms.

Given that EPD forcing is the dominant component of tropical upwelling, we calculated $\langle \overline{w'} \rangle$ contributed by each component of EPD. Figure 7 shows the annual cycle of $\langle \overline{w'} \rangle$ contributed by each EPD component along with the total EPD contribution, averaged over 15°S–15°N. It shows that the EPD-Y1 contribution is dominant and explains about half of the total EPD contribution, whereas contributions from EPD-Z1 and EPD-Z2 explain the remaining part. The contribution from EPD-Y2 is negligible. The dominance of EPD-Y1 has been reported in several previous studies (e.g., RGW08 and T09). However, compared with upwelling estimates using the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and the National Centers for Environmental Predition (NCEP) reanalysis dataset by RGW08, the relative contribution of EPD-Y1 is smaller in the present study, and the contributions from EPD-Z1 and EPD-Z2 are larger. The predominant contribution of EPD-Z2 associated with equatorial Rossby waves was shown by Kerr-Münslow and Norton (2006) using ERA-40 data, and a significant contribution from EPD-Z2 was also shown in a WACCM simulation by T09.

In the present study, the contribution of meridional heat flux $\partial F_1/\partial z$ to upwelling is comparable to that from the vertical flux of zonal momentum $\partial F_2/\partial z$. ERA-40...
analysis by RGW08 showed a similar result. In Kerr-Munslow and Norton (2006) and T09, the contribution of $\frac{\partial F_1^z}{\partial z}$ was smaller than that of $\frac{\partial F_2^z}{\partial z}$. Somehow, the relative contribution of each EPD component to upwelling differs between the present study and each of the previous studies by Kerr-Munslow and Norton (2006), RGW08, and T09, likely due to two factors. First, in Kerr-Munslow and Norton (2006) and T09, the contribution of each EPD component to upwelling is considered by wave forcing at a single level (90 hPa) and a thin layer in the lower stratosphere (50–100 hPa), respectively, rather than vertical integration of wave forcing throughout a sufficiently deep layer, as was done by RGW08 and the present study. To confirm this point, we calculated (not shown) the annual cycle of each EPD forcing component averaged over $15^\circ S$–$15^\circ N$ at 100 hPa. The result shows that EPD-Y1 should explain more than 80% of the total EPD forcing most of the time, except during the NH winter when EPD-Y1 and EPD-Z2 are comparable, and that EPD-Z2 should be much larger than EPD-Z1 throughout the year. This is significantly different from the actual contribution of each EPD component to upwelling shown in Fig. 7, implying that the EPD (or any wave) forcing at a single pressure level in the lower stratosphere cannot sufficiently explain its contribution to tropical upwelling. Second, GWD forcing, which cannot be represented in the global reanalysis data, is likely to increase the contribution of the vertical components of EPD forcing to upwelling, as revealed in RGW08 and the present study. In addition to these two factors, there may be several other factors that make a difference in the contribution of each wave forcing to upwelling, such as the length of data (or simulation period) and representation of convective clouds in the tropics and associated equatorial waves, as pointed out by T09. Further investigations on this topic are required to form a robust conclusion.

b. Impact of CGWD on tropical upwelling

To understand effects of convective gravity waves on tropical upwelling, results from GWDC and CTL simulations are compared. Figure 8 shows differences between the GWDC and CTL simulations in the annual cycle of $\overline{w^*}$ over the tropical region ($\pm 30^\circ$) at 100 hPa and (b) the latitudinal averages of $\overline{w^*}$ (solid) and $\langle \overline{w^*_m} \rangle$ (dotted) over $15^\circ S$–$15^\circ N$. The contour interval of (a) is 0.02 mm s$^{-1}$ and negative values are dashed. Light and dark shading in (a) denote the 90% and 95% confidence levels, respectively.

![Figure 8](image_url)

**Fig. 8.** Differences in (a) the annual cycle of $\overline{w^*}$ between the GWDC and CTL simulations over the tropical region ($\pm 30^\circ$) at 100 hPa and (b) the latitudinal averages of $\overline{w^*}$ (solid) and $\langle \overline{w^*_m} \rangle$ (dotted) over $15^\circ S$–$15^\circ N$. The contour interval of (a) is 0.02 mm s$^{-1}$ and negative values are dashed. Light and dark shading in (a) denote the 90% and 95% confidence levels, respectively.

as near 10$^\circ S$ during January, near 20$^\circ N$ during February, near 30$^\circ S$ during October, and near 25$^\circ S$ during November, a decrease of $\overline{w^*}$ by including the CGWD parameterization also appears. When averaged over $15^\circ S$ to $15^\circ N$ (Fig. 8b), both $\overline{w^*}$ and $\langle \overline{w^*_m} \rangle$ are increased by including CGWD during most months, except during January, with similar seasonal variations. The magnitude of CGWD’s impact (GWDC – CTL) is about 1/20 of the total upwelling value in the GWDC simulation. However, it is larger than $\langle \overline{w^*_m} \rangle$ contributed by CGWD forcing alone for all months except January, as shown in Fig. 6b, and is even larger than the sum of all three GWD forcing contributions during November and December. This implies that including CGWD in the GWDC simulation enhances upwelling not only by direct CGWD forcing but also by changing the other processes that determine upwelling.
Figure 9 shows the difference between the GWDC and CTL simulations for each of the forcing terms determining $h_{wm}^*$ (Fig. 9a) and the difference in each component of EPD between the GWDC and CTL simulations (Fig. 9b). Several features can be found in Fig. 9. First, decreased upwelling in the GWDC simulation during January is mainly due to the EPD forcing change, which is greater than 10% of the EPD contribution to upwelling in the GWDC simulation (Fig. 6a). However, the significant increase of $dU/dt$ and the slight increases in CGWD and BGWD partially offset the negative spike in EPD during January. Second, the EPD change during January is mainly due to the change of EPD-Y1, which is out of phase with EPD-Z1 change most of the time (Fig. 9b). The change of EPD-Z2 is comparable to the EPD-Z1 change with nearly opposite sign. Third, the EPD and $dU/dt$ changes are largely out of phase with each other, although with a relatively larger magnitude for $dU/dt$, except during the NH wintertime (November–January). Figure 9 demonstrates that including CGWD forcing increases upwelling in the tropical region (15°S–15°N) not only directly by adding the CGWD forcing term, but also indirectly by modulating the other forcing terms such as EPD and $dU/dt$. Although the indirect influence has the greater effect most of the time, direct CGWD forcing also significantly influences upwelling change when cancellations between the two major indirect processes occur, especially during March–June.

c. Upwelling estimation from net upward mass flux

In the previous figures, upwelling averaged over 15°S and 15°N and seasonal contribution of each forcing terms in those latitudes are considered exclusively. This latitude band is selected because $w^*$ at 100 hPa is positive for all months only inside that band, as shown in Fig. 1a. However, Fig. 1a also shows that upwelling extends up to 30°N and 30°S from the equator during the NH summer- and wintertimes, respectively. Given that the wave forcing terms determining upwelling have latitudinal distribution as shown in Figs. 4 and 5, their contribution to upwelling can be changed when a wider latitude band including the subtropics is considered. We found (not shown) that upwelling is reduced by about 30% when the latitude band is extended poleward to include the subtropics (±25°), mainly due to the compensating downwelling from EPD in the winter hemisphere subtropics (Fig. 5a). Magnitudes of upwelling from all three GWD forcing terms increase in the extended latitude band (±25°), especially by OGWD forcing, and consequently the relative contribution of GWD forcing to the upwelling increases (~10%). In addition, the upwelling averaged over 15°S and 15°N at 70 hPa (not shown) is generally similar to that at 100 hPa, except with much smaller magnitude due mainly to the decreases in upwelling by the EPD and OGWD forcings.

To take into account the contribution of each wave forcing term in a wider latitudinal band in which tropical upwelling actually occurs, we estimate the net upward mass flux between turnaround latitudes where tropical upwelling changes extratropical downwelling, from the mass streamfunctions obtained from the simulated
residual mean meridional \( \bar{v} \) and vertical \( \bar{w} \) velocities, denoted by “direct,” and from the downward control principle. Details on the derivation of the net upward mass flux are given in the appendix. Estimation of the BD circulation based on the net upward mass flux at 70 hPa described in the present study has been conducted for several previous studies (e.g., Li et al. 2008; McLandress and Shepherd 2009; Butchart et al. 2010; Okamoto et al. 2011).

Figure 10a shows vertical profiles of annual-mean net upward mass flux calculated using the direct mass streamfunction and downward-control mass streamfunction contributed by individual forcing terms in the CGWD simulation. The annual cycles of the upward mass flux at 70 hPa contributed by the three GWD forcings are shown in Fig. 10b, and downward-control mass streamfunctions for December–February (DJF) at 70 hPa induced by each forcing terms are shown in Fig. 10c along with the direct mass streamfunction. The annual and seasonal means of the net upward mass fluxes at 70 hPa are shown in Fig. 10d as color bars with their values in Table 1.

In Fig. 10a, net upward mass fluxes generally decrease with height except for those by CGWD and BGWD that increase below about 70 and about 50 hPa, respectively. Among the three GWD forcings, contribution by OGWD is largest during January, February, March, and November, while that by CGWD is largest during the rest of months except December, where contribution by BGWD is largest (Fig. 10b). Compared with 11 chemistry–climate model results reported by Butchart et al. (2010, see their Fig. 11), the upward mass flux by OGWD in the present study is much less, especially during DJF. One possible reason for relatively small contribution by OGWD in the present study, especially

![FIG. 10. Net upward mass flux calculated from the direct mass streamfunction and from downward-control mass streamfunctions using the EPD and three GWD forcing terms in the CGWD simulation: (a) vertical profiles of the annual mean net upward mass flux, (b) the annual cycle of the net upward mass flux at 70 hPa from the three GWD forcing terms, (c) mass streamfunctions for DJF at 70 hPa, and (d) annual, DJF, and June–August (JJA) means of net upward mass flux at 70 hPa. The mass flux is computed between the turnaround latitudes (see text for details).]

<table>
<thead>
<tr>
<th>Direct</th>
<th>EPD+GWDs</th>
<th>EPD</th>
<th>OGWD</th>
<th>BGWD</th>
<th>CGWD</th>
</tr>
</thead>
<tbody>
<tr>
<td>Annual</td>
<td>5.45</td>
<td>5.29</td>
<td>4.29</td>
<td>0.45</td>
<td>0.16</td>
</tr>
<tr>
<td>DJF</td>
<td>6.61</td>
<td>6.76</td>
<td>5.98</td>
<td>0.33</td>
<td>0.27</td>
</tr>
<tr>
<td>JJA</td>
<td>5.01</td>
<td>5.02</td>
<td>4.33</td>
<td>0.18</td>
<td>0.20</td>
</tr>
</tbody>
</table>
for DJF, is that the turnaround latitude in the NH side of the direct mass streamfunction (26°N), which is used for mass flux calculation, is located equatorward of that of downward-control mass streamfunction induced by the OGWD forcing (37°N) (Fig. 10c). Consequently, downwelling by the OGWD forcing between 26° and 37°N cancels the upwelling poleward of 37°N by the OGWD forcing. Compared with McLandress and Shepherd (2009), the turnaround latitude in the NH side in the present DJF mean shifts somewhat equatorward (30°N vs 26°N). When we calculate net upward mass flux using the turnaround latitudes derived from the downward-control mass streamfunction including all forcing terms (EPD+GWDs), which is 29°N during DJF (Fig. 10c), the mass flux by OGWD during DJF becomes almost twice as large as the present result (0.61 vs 0.33 × 10^6 kg s⁻¹). This implies that small changes in turnaround latitudes are likely to make significant differences in the mass flux calculation.

Contributions by the EPD, OGWD, CGWD, and BGWD forcings to upward mass flux are 81%, 9%, 7%, and 3%, respectively, in the annual mean (Fig. 10d). The contribution by CGWD, which could not be considered in the previous studies, is comparable to that by OGWD. This implies that CGWD can contribute to tropical upwelling significantly, as much as by OGWD, and proper treatment of convective gravity waves in GCMs through CGWD parameterization may be required for realistic climate and climate change simulations. Compared with previous studies based on chemistry–climate models (e.g., Butchart et al. 2010), the relative contribution by the present EPD (OGWD) forcing is generally larger (smaller). Considering that the present results are based on a 10-yr annual simulation using climatological boundary conditions without chemical processes, it is not straightforward to directly compare the present results with those from the previous chemistry–climate modeling for more than 100 yr.

4. Summary and conclusions

Seasonal variations of tropical stratospheric upwelling are investigated from the results of the 10-yr WACCM simulations including mountain, convection, and background GWD parameterizations. The tropical upwelling is estimated by the residual mean vertical velocity \( \mathbf{w} \) at 100 hPa averaged between 15°S and 15°N. This is well matched with the upwelling estimate from the balance of the zonal momentum and continuity (\( \nabla_z \mathbf{u} \)) throughout the year, so that the annual cycle of the tropical upwelling can be understood from the detailed analyses of individual forcing terms that contribute to the balance. Contributions by resolved planetary waves through the Eliassen–Palm flux divergence (EPD) and by gravity waves through the three GWD parameterizations (OGWD, CGWD, and BGWD) are investigated. OGWD and CGWD denote GWD generated by orographic and convective sources, respectively, while BGWD, so-called background GWD, largely represents jet stream–related gravity waves. Among the four effective wave forcing terms weighted by density, EPD forcing predominates below \( z = 60 \text{ km} \) in the winter hemisphere, whereas OGWD is comparable to EPD below \( z = 25 \text{ km} \) exclusively in the 20°–50° latitudes in the winter hemisphere. Effective CGWD and BGWD forcing are one order of magnitude smaller than effective EPD forcing below \( z = 60 \text{ km} \) in the winter hemisphere, but comparable or larger in the summer stratosphere and mesosphere. The upwelling at 100 hPa is determined by effective wave forcing terms integrated from 100 hPa to model top, and vertical integration at least up to \( z = 50 \text{ km} \) is required for accurate calculation of tropical upwelling at 100 hPa from the EPD, CGWD, and BGWD forcing terms. The relative contributions of each wave forcing term and their annual cycles are found to be different when vertical integration is conducted within a shallow layer.

A clear annual cycle of tropical upwelling is found, with a maximum during the NH wintertime and a minimum during the NH summertime, and it is determined primarily by the EPD contribution along with the secondary contribution from \( \partial \mathbf{U}/\partial t \). Among the four components consisting of EPD, contribution by the horizontal momentum flux (EPD-Y1) is dominant, while smaller contributions from vertical momentum flux (EPD-Z2) and meridional heat flux (EPD-Z1) are similar to each other. Compared with some previous studies, however, the contribution from EPD-Y1 is relatively small while that of EPD-Z1 is larger in the present study, likely due to using deeper vertical integration of forcing terms in calculating upwelling in the present study. Gravity waves mostly increase tropical upwelling throughout the year, and among the three sources the contribution of convective gravity waves is largest, especially during NH springtime. However, the total contribution of all three gravity waves to tropical upwelling is not larger than 5%. When upwelling is extended to the subtropics (25°S–25°N), the relative contribution from all three gravity wave forcings is larger than 10% during the NH wintertime, mainly due to the increase from OGWD.

To investigate effects of convective gravity waves, an additional WACCM simulation without CGWD parameterization (CTL simulation) is conducted, and differences between the GWDC and CTL simulations are examined for each wave contribution as well as the annual cycle of upwelling. The analysis shows that including CGWD parameterization increases tropical upwelling...
during most months by about 5%, which is larger than the single contribution by CGWD. The increase of tropical upwelling caused by including CGWD is due not only to the direct CGWD forcing but also to the changes in EPD and dU/dt forcing terms that are modulated by including CGWD parameterization.

In the present study, upwelling averaged over 15°S and 15°N at 100 hPa and seasonal contribution of each forcing terms in those latitudes are considered mostly. To take into account the contribution of each wave forcing terms in a wider latitudinal band in which upwelling actually occurs, we also estimate net upward mass flux between turnaround latitudes where tropical upwelling changes extratropical downwelling. It shows that contributions by the EPD, OGWD, CGWD, and BGWD forcings to the annual mean of the net upward mass flux at 70 hPa are 81%, 9%, 7%, and 3%, respectively. The relative contribution by the OGWD forcing in the present study is much less, especially during DJF, while that by the EPD forcing is much larger than the previous studies based on chemistry–climate models. The contribution by CGWD, which was not considered in the previous studies, is comparable to that by the OGWD forcing, implying that proper treatment of convective gravity waves in GCMs through CGWD parameterization may be required for realistic estimation of BD circulation and its trend associated with climate change.

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APPENDIX

Derivation of the Net Upward Mass Flux

Following Holton (1990), the area-averaged extratropical vertical mass flux across a pressure surface in the NH \( F^{NH} \) and SH \( F^{SH} \) can be expressed as

\[
F^{NH} = 2\pi a^2 \int_{\phi_{TL}^{NH}}^{\pi/2} \psi^* \cos \phi \, d\phi,
\]

\[
F^{SH} = 2\pi a^2 \int_{-\pi/2}^{\phi_{TL}^{SH}} \psi^* \cos \phi \, d\phi. \quad (A1)
\]

Here, \( \phi_{TL}^{NH} \) and \( \phi_{TL}^{SH} \) are the turnaround latitudes in NH and SH, respectively, which are located at the minimum and maximum of the mass streamfunction \( \Psi \), respectively. From the continuity equation of the residual mean meridional \( \psi^* \) and vertical \( \psi^* \) velocities,

\[
\frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\psi^* \cos \phi) + \frac{1}{\rho_0} \frac{\partial}{\partial z} (\rho_0 \psi^*) = 0, \quad (A2)
\]

the mass streamfunction is defined as

\[
\psi^* = -\frac{1}{\rho_0 \cos \phi} \frac{\partial \Psi}{\partial z}, \quad \psi^* = \frac{1}{\rho_0 a \cos \phi} \frac{\partial \Psi}{\partial \phi}. \quad (A3)
\]

Using (A3), the vertical mass fluxes in (A1) can be obtained as

\[
F^{NH} = -2\pi a \Psi(\phi_{TL}^{NH}), \quad F^{SH} = 2\pi a \Psi(\phi_{TL}^{SH}). \quad (A4)
\]

Here, we applied zero boundary condition of mass streamfunction at the poles. From the constraint of zero global mean mass flux, the net upward mass flux in the tropical region \( F^{TR} \) between \( \phi_{TL}^{NH} \) and \( \phi_{TL}^{SH} \) can be estimated by

\[
F^{TR} = -(F^{NH} + F^{SH}) = 2\pi a [\Psi(\phi_{TL}^{NH}) - \Psi(\phi_{TL}^{SH})]. \quad (A5)
\]

The net upward mass flux using the downward control principle \( F^{DC}_{DC} \) is also obtained by \( 2\pi a [\Psi(\phi_{TL}^{NH}) - \Psi(\phi_{TL}^{SH})] \), where \( \Psi_{DC} \) is the mass streamfunction that derived from the downward control principle (Haynes et al. 1991), which can be expressed as

\[
\Psi_{DC} = -\cos \phi \int_{z_0}^{\infty} \frac{WF(\phi, z^*)}{\tilde{f}(\phi, z^*)} \, dz^*, \quad (A6)
\]

Here, \( WF(\phi, z) \) and \( \tilde{f}(\phi, z) \) represent the wave forcing and modified Coriolis parameter, respectively, that are defined in (2). In the calculation of \( F^{DC}_{DC} \), \( \Psi_{DC} \) is evaluated at the turnaround latitudes obtained from \( \Psi \), which are almost the same as those obtained from \( \Psi_{DC} \) including EPD and all GWD forcing terms. Given that the turnaround latitudes obtained from \( \Psi \) generally increase with height, this approach can only be applied to levels higher than about 85 hPa where the turnaround latitudes exist in extratropics that are safely escaped from singular latitudes near the equator in the denominator of (A6).

REFERENCES


