Contributions of Convective and Orographic Gravity Waves to the Brewer–Dobson Circulation Estimated from NCEP CFSR

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ABSTRACT

Contributions of convective gravity waves (CGWs) and orographic gravity waves (OGWs) to the Brewer–Dobson circulation (BDC) are examined and compared to those from resolved waves. OGWD drag (OGWD) is provided by NCEP Climate Forecast System Reanalysis (CFSR), while CGW drag (CGWD) is obtained from an offline calculation of a physically based CGW parameterization with convective heating and background data provided by CFSR. CGWD contributes to the shallow branch of the BDC regardless of the season, while OGWD contributes to both the shallow and deep branches except for the summertime, when OGWs hardly propagate into the stratosphere. At 70 hPa, the annual-mean tropical upward mass fluxes from Eliassen–Palm flux divergence (EPD), OGWD, and CGWD are 68%, 7%, and 4% of the total mass flux, respectively. The tropical upward mass flux at 70 hPa shows an increasing trend during the time period from 1979 to 1998, with 28%, 18%, and 6% of the trend driven by EPD, OGWD, and CGWD, respectively. The width of the turnaround latitudes tends to narrow for the streamfunctions induced by OGWD and CGWD but tends to widen for that induced by EPD. The contributions of GWD from MERRA (MERRA-2) to the climatology and long-term trend of the BDC are 7% (7%) and 13% (4%), respectively, somewhat smaller than the contributions of CGWD plus OGWD, which are estimated from CFSR to be 12% and 20%, respectively.

1. Introduction

The Brewer–Dobson circulation (BDC) is a mean stratospheric mass transport that describes air parcels moving upward and poleward from the tropical tropopause and descending in the middle and high latitudes (Andrews et al. 1987; Holton et al. 1995). The BDC includes a residual-mean overturning circulation and a two-way isentropic mixing (e.g., Plumb 2002), and the former can be approximated to the total BDC, except for a surf zone in which mixing is strong (Pendlebury and Shepherd 2003). The BDC is very important to adiabatic warming/cooling in the

stratosphere, the ozone concentration, and the lifetimes of trace gases such as chlorofluorocarbons (CFCs) and greenhouse gases (GHGs) in the stratosphere (Holton et al. 1995; Butchart and Scaife 2001). In addition, the BDC controls water vapor input into the stratosphere (Mote et al. 1994). The BDC is driven by extratropical wave forcing because of the dissipation of Rossby waves and gravity waves (GWs) that propagate from the troposphere (Holton et al. 1995). Westward wave forcing pushes air poleward to conserve angular momentum, hence pumping up tropical air from the tropopause to the stratosphere in a steady-state limit. Chemistry–climate models (CCMs) have shown that the contribution of resolved Rossby waves to the BDC is approximately 70% and that of parameterized GWs is approximately 30% at 70 hPa, yet with a wide spread between the models (e.g., Butchart et al. 2010; Butchart et al. 2011). Uncertainties lie primarily on the realism of the GW parameterization and partly on the different representation of resolved waves in the model.

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To better understand the BDC and its driving mechanism, reanalysis data have been utilized because they provide gridded long-term global atmospheric data that are constrained by observations. [Seviour et al. \(2012\)](#) showed that resolved waves, through Eliassen–Palm flux divergence (EPD), and parameterized orographic GWs provide 70% and 4%, respectively, of the annual-mean upwelling mass flux at 70 hPa in European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-I) data ([Dee et al. 2011](#)) for 1980–2009. They mentioned that the remaining 26% may include effects of underestimated orographic GW forcing and nonorographic GW forcing. [Kim et al. \(2014\)](#) investigated changes in the BDC over 30 years (1980–2009) by using Modern-Era Retrospective Analysis for Research and Applications (MERRA) data ([Rienecker et al. 2011](#)) and found that the resolved wave drag and GW drag (GWD) drive approximately 87% and 9% of the total upward mass flux, respectively, during December–January–February (DJF). [Abalos et al. \(2015\)](#) used three modern reanalysis datasets and showed that resolved waves drive an overall latitudinal structure of residual vertical velocities, whereas GWD slightly increases the magnitudes of residual vertical velocities in the subtropics and extratropics. [Sato and Hirano \(2019\)](#) examined potential GWD contributions to the BDC when the GWD is estimated from the residual of resolved wave drag and zonal wind tendency in the transformed Eulerian-mean (TEM) equation. They emphasized that GWs contribute to the summer-hemispheric part of the winter deep branch, the low-latitude part of shallow branches, and the higher-latitude extension of deep circulation in all seasons except for summer. The large variations in the contributions of GWs and resolved waves to the BDC according to these studies suggest that more investigation is required using various datasets, including physically based GWD data.

Recently, the BDC has been studied intensively in terms of climate change because of evidence of accelerating BDC in a warmer atmosphere ([Butchart et al. 2006](#); [Garcia and Randel 2008](#); [Shepherd and McLandress 2011](#); [Sigmond and Shepherd 2014](#)). [Butchart et al. \(2010\)](#) showed that tropical upward mass fluxes simulated by CCMs increase by around 2% decade^{−1} during the twenty-first century, and parameterized waves drive 59% of the trend ranging from 23% to 95%. This result is also revealed in past records, as reanalysis data show an increasing trend in tropical upwelling of 2%–5% decade^{−1} during 1979–2010, except for ERA-I ([Abalos et al. 2015](#)). Although no evidence of a negative trend in the mean age of air (AoA) has been argued from observations

(i.e., an acceleration trend in BDC) during past decades ([Engel et al. 2009](#); [Stiller et al. 2012](#); [Ray et al. 2014](#)), the increasing trend in the BDC is likely to be robust at least for the lower stratosphere (below ~25 km), where both a negative AoA trend and increasing residual vertical velocity trend exist (e.g., [Diallo et al. 2012](#); [Abalos et al. 2015](#)).

Previous studies investigated the contributions of resolved waves and parameterized GWs to the BDC, but GWs that are generated from individual sources other than orography have not been investigated comprehensively. Some recent studies estimated BDC contributions from each source of GWD by using climate models (e.g., [Chun et al. 2011](#); [Palmeiro et al. 2014](#)). [Chun et al. \(2011\)](#) performed climatological simulations by using the Whole Atmosphere Community Climate Model (WACCM), version 1b (WACCM 1b), including three GWD parameterizations (i.e., orographic, convective, and background GWs), and showed that GWs contribute around 19% to the upward mass flux at 70 hPa, with comparable contributions from orographic (9%) and convective (7%) GWs. [Palmeiro et al. \(2014\)](#) used WACCM, version 4 (WACCM4), including three GWD parameterizations (i.e., orographic, convective, and frontal GWs), and showed that the contributions from resolved waves and orographic GWs to both the climatology and trend of the tropical upwelling at 70 hPa are 70% and 30%, respectively, without significant contributions from convective or frontal GWs. The discrepancy between the results from [Chun et al. \(2011\)](#) and [Palmeiro et al. \(2014\)](#) probably occurred because OGWD in WACCM1b is restricted below 30-hPa altitude, which is not the case for WACCM4. In addition, the CGW parameterization in [Chun et al. \(2011\)](#) ([Choi and Chun 2011](#)) induces strong westward CGWD in the lower stratosphere at 30°–50°N/S, where strong source-level CGWs exist in association with winter-hemispheric storm-track regions, which is different from the CGW parameterization in WACCM4 ([Beres et al. 2005](#)). Recently, [Cohen et al. \(2014\)](#) showed that an increase in GW forcing decreases planetary wave forcing and vice versa, which was also shown previously by [McLandress and McFarlane \(1993\)](#), and the contribution of the total wave forcing to BDC is nearly the same regardless of the contributions of each wave. Based on this result, they suggested that considering the contribution of each wave to the BDC may not be meaningful. Although this concept is noteworthy, it is still important to distinguish the contribution of each wave to the BDC using downward-control principle, which will have different spatiotemporal variations.

This is particularly important to understand the contribution of GWs generated by individual sources to the BDC, which have not been analyzed in detail.

In this study, we examine the roles of orographic GWD (OGWD) and convective GWD (CGWD) individually in BDC by using global reanalysis data in terms of climatology and long-term trends; these roles are also compared to those of resolved forcing. The reanalysis dataset used in this study is the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR; Saha et al. 2010), which provides OGWD as a standard output. For CGWD, we use the results from Kang et al. (2018), where the offline calculation of CGW parameterization by Kang et al. (2017) was conducted with convective heating rates and background wind and temperature values provided by CFSR. In addition to CFSR, three reanalysis datasets are used to summarize the potential range of GWD contribution to the BDC, based on either GWD from the reanalysis datasets or estimated GWD from the residual in the TEM equation when GWD is not provided from the reanalysis datasets.

The remainder of this paper is organized as follows. Sections 2 and 3 describe the data and methods used to calculate the BDC, respectively. Section 4 presents the climatology of the BDC in each season, and section 5 examines the long-term trend of the annual-mean BDC and associated trends in zonal wind and wave forcing. Section 6 discusses the possible contribution from the residual term and estimates the role of GWs in driving BDC by using four reanalyses. A summary and conclusions are given in the last section.

2. Data

We use CFSR data for 32 years (1979–2010) provided on a 1° longitude $\times$ 1° latitude regular grid in a 6-h interval. OGWD is a standard output of the CFSR dataset, resulting from the orographic GWD parameterization scheme by Kim and Arakawa (1995) implemented in the NCEP model. The CGWD data used in our study are calculated from the offline CGW parameterization in Kang et al. (2017, 2018) by using CFSR data because no convective GWD parameterization is implemented in the NCEP model to produce CFSR, version 1. Here, we briefly describe the CGW parameterization. At the source level (cloud top) of the CGWs, the analytic solution of the GW momentum flux spectra induced by convective heating is obtained as a function of phase velocity [Eq. (1) of Kang et al. (2017)] by using the deep convective heating rate and wind and temperature data. The source-level

CGW spectrum is determined by the spectral combination of the convective source and the wave-filtering and resonance factor that represents wave filtering by critical levels within the source layer and the resonance between vertical harmonics of the convective heating and natural wave modes (Song and Chun 2005). The momentum flux of CGWs from the cloud top to the upper stratosphere and the CGWD are calculated based on the columnar wave-propagation method by using Lindzen's linear saturation theory (Lindzen 1981). The parameterized CGWs focus on small-scale waves ($\lambda_h < 100$ km and $\lambda_z < 40$ km), which are hardly observed from satellites and rarely simulated from high-resolution general circulation models (GCMs), and the magnitude of the CGW momentum flux is constrained by GW observations from superpressure balloons (SPBs) in the tropical region (Jewtoukoff et al. 2013). Because seasonal and interannual variations in convective activities and background flows are considered in the parameterization, it is useful for investigating the variations in CGW sources and forcings that contribute to atmospheric variabilities (Kang et al. 2018).

In this study, we mainly present the results from CFSR, but comparisons with other reanalysis datasets, such as MERRA (Rienecker et al. 2011), MERRA-2 (Gelaro et al. 2017), and ERA-I (Dee et al. 2011), are also performed. Because GWD data are not provided from ERA-I, the *potential* GWD is estimated from the residual of the terms that are included in the TEM equation, which is similar to the approach by Sato and Hirano (2019). The magnitude of the *potential* GWD is generally larger than that of the model-provided GWD, given that other factors such as the assimilation increment can be included in the residual of the TEM equation. This topic will be discussed further in section 6.

3. Methods

In this study, we use the TEM equation (Andrews et al. 1987) to define the residual-mean circulation:

$$\frac{\partial \bar{u}}{\partial t} = \left[f - \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\bar{u} \cos \phi) \right] \bar{v}^* - \bar{w}^* \frac{\partial \bar{u}}{\partial z} + \frac{1}{\rho_0 a \cos \phi} \nabla \cdot \mathbf{F} + \text{CGWD} + \text{OGWD} + \bar{X}. \quad (1)$$

Here, $(1/\rho_0 a \cos \phi) \nabla \cdot \mathbf{F}$ refers to EPD from the resolved planetary waves and is calculated as follows:

$$\frac{1}{\rho_0 a \cos \phi} \nabla \cdot \mathbf{F} = \frac{1}{\rho_0 a \cos \phi} \left[\frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (F^\phi \cos \phi) + \frac{\partial F^z}{\partial z} \right], \quad (2)$$

where $F^\phi = \rho_0 a \cos\phi(-\overline{u'v'} + \overline{u_z v' \theta' / \bar{\theta}_z})$ and $F^z = \rho_0 a \cos\phi(\hat{f} \overline{v' \theta' / \bar{\theta}_z} - \overline{u' w'})$. Here, the modified Coriolis parameter $\hat{f}$ is defined by $\hat{f} = f - 1/(a \cos\phi) \partial/\partial\phi(\overline{u} \cos\phi)$, where f is the Coriolis parameter. CGWD and OGWD represent convective and orographic GWD, respectively; and $\overline{X}$ represents the residual in the TEM equation [Eq. (1)]. Note that the $\overline{X}$ is a virtual residual term: $\overline{X}$ is the residual of the balance among EPD, CGWD, OGWD, $\partial\overline{u}/\partial t$, and the advection terms in the TEM equation, although the CGWD is not implemented in the NCEP model but instead calculated offline. This approach enables us to distinguish and diagnose the contribution of CGWD in the real residual term in the NCEP model (i.e., CGWD + $\overline{X}$). The residual meridional and vertical velocities are defined following Andrews et al. (1987):

$$\overline{v}^* = \overline{v} - \frac{1}{\rho_0} \frac{\partial}{\partial z} \left(\frac{\rho_0 \overline{v' \theta'}}{d\bar{\theta}/dz} \right), \quad (3)$$

$$\overline{w}^* = \overline{w} + \frac{1}{a \cos\phi} \frac{\partial}{\partial\phi} \left(\cos\phi \frac{\overline{v' \theta'}}{\partial\bar{\theta}/\partial z} \right). \quad (4)$$

We define the mass streamfunction based on a so-called direct method by using Eqs. (3) and (4) and the downward-control principle (Haynes et al. 1991; Randel et al. 2002) by using Eq. (1):

$$\chi_{\text{direct}}(\phi, z) = \cos\phi \int_z^\infty \overline{v}^*(\phi, z') \rho_0 dz', \quad (5)$$

$$\chi_{\text{dc}}(\phi, z) = -\frac{\cos\phi}{\hat{f}} \int_z^\infty \{ \rho_0(z') [\text{WF}(\phi, z') - \overline{u}_t(\phi, z')] \}_{\bar{m}} dz'. \quad (6)$$

The subscript $\bar{m}$ indicates that the vertical integrand in Eq. (6) should be evaluated following constant angular momentum. In practice, however, this value is calculated with constant latitudes, given that the angular momentum lines in the extratropics (poleward of 18°N/S) are approximately parallel to the latitude lines. Therefore, Eq. (6) is calculated at the poleward of 18°N/S (see Fig. 2). WF(ϕ, z) represents wave forcing, including EPD, CGWD, OGWD, and $\overline{X}$ on the right-hand side of Eq. (1). The downward-control principle requires an integration of the wave forcing to a sufficiently higher level (theoretically infinity), so missing wave forcing in the mesosphere because of a limited top height (1 hPa) for CFSR may lead to uncertainties in the mean meridional circulation ($\overline{v}^*$, $\overline{w}^*$). According to Chun et al. (2011), the vertical integration of CGWD up to 70 km

(30 km) is enough to represent 90% of the upwelling at 100 hPa in the tropics and winter hemisphere (summer hemisphere), implying that $\overline{w}^*$ by CGWD in the real atmosphere in the tropics and winter hemisphere would be more than 10% larger than that in the current study. However, for EPD and OGWD that are concentrated on the stratosphere, the difference would be minor.

In addition, the residual vertical velocities from each wave forcing averaged between ϕ_1 and ϕ_2 can be estimated by the downward-control principle (Randel et al. 2008):

$$\begin{aligned} \langle \overline{w}_{\text{dc}}^* \rangle(\phi, z) &= \frac{\chi_{\text{dc}}(\phi_2, z) - \chi_{\text{dc}}(\phi_1, z)}{\rho_0(z) \int_{\phi_1}^{\phi_2} a \cos\phi d\phi} \\ &= \frac{\chi_{\text{dc}}(\phi_2, z) - \chi_{\text{dc}}(\phi_1, z)}{\rho_0(z) a [\sin(\phi_2) - \sin(\phi_1)]}. \end{aligned} \quad (7)$$

Note that $\langle \overline{w}_{\text{dc}}^* \rangle$ is evaluated at the latitude of $\phi = (\phi_1 + \phi_2)/2$ at 5° interval, except in the tropics between $\phi_1 = 18^\circ\text{N}$ and $\phi_2 = 18^\circ\text{S}$ within which $\langle \overline{w}_{\text{dc}}^* \rangle$ calculated using ϕ_1 and ϕ_2 by Eq. (7) is assumed to be constant. This is because χ_{dc} cannot be defined in the tropics [Eq. (6)] due to the proportionality to $1/\hat{f}$. We found that the seasonal variation of $\langle \overline{w}_{\text{dc}}^* \rangle$ averaged between 18°N and 18°S matches that of $\overline{w}^*$ calculated using Eq. (4), consistent with Fig. 2 of Chun et al. (2011).

The strength of the BDC is often quantified by using the net upward mass flux in the tropical region via the mass streamfunction at the turnaround latitudes (Holton 1990; McLandress and Shepherd 2009; Chun et al. 2011; Kim et al. 2014). Following Holton (1990), the area-averaged extratropical vertical mass flux across a pressure surface in the Northern Hemisphere (NH) (F^{NH}) and Southern Hemisphere (SH) (F^{SH}) can be expressed as

$$F^{\text{NH}} = 2\pi a^2 \rho \int_{\phi_{\text{TL}}^{\text{NH}}}^{\pi/2} \overline{w}^* \cos\phi d\phi = -2\pi a \chi(\phi_{\text{TL}}^{\text{NH}}), \quad (8)$$

$$F^{\text{SH}} = 2\pi a^2 \rho \int_{-\pi/2}^{\phi_{\text{TL}}^{\text{SH}}} \overline{w}^* \cos\phi d\phi = 2\pi a \chi(\phi_{\text{TL}}^{\text{SH}}), \quad (9)$$

where $\phi_{\text{TL}}^{\text{NH}}$ and $\phi_{\text{TL}}^{\text{SH}}$ are the turnaround latitudes in the NH and SH, respectively. Note that, in this study, the turnaround latitudes are defined for each day. The net upward mass flux in the tropical region between turnaround latitudes is estimated from the zero-boundary condition of the mass streamfunctions at the poles (i.e., zero global-average mass flux):

FIG. 1. Latitude–height cross sections of the (a) annual-, (b) DJF-, (c) MAM-, (d) JJA-, and (e) SON-mean (top) EPD, (middle) CGWD, and (bottom) OGWD (shading) overlaid with the zonal-mean zonal wind (magenta contours), with a contour interval of 10 m s^{-1} , calculated using the CFSR data averaged from 1979 to 2010. Positive (negative) winds are plotted as solid (dotted) lines and zero winds are plotted as thick solid lines.

$$F^{\text{TR}} = -(F^{\text{NH}} + F^{\text{SH}}). \quad (10)$$

4. Climatology

Figure 1 shows latitude–height cross sections of the annual- and seasonal-mean EPD, CGWD, and OGWD in the stratosphere. The zonal-mean zonal wind is overlaid with magenta contours in each figure. In the annual-mean EPD, negative forcing is dominant in the midlatitudes and upper stratosphere. Negative forcing is strongest in the NH during the boreal wintertime (DJF), with a maximum negative forcing of $-12.6 \text{ m s}^{-1} \text{ day}^{-1}$ (Fig. 1b), and second strongest in the SH during the austral springtime [September–November (SON)], with a maximum negative forcing of $-6.4 \text{ m s}^{-1} \text{ day}^{-1}$ (Fig. 1d). In addition, negative forcing exists above the subtropical jet at 70–100 hPa regardless of the season. The annual-mean CGWD is the strongest in the tropical upper stratosphere with a positive sign and is evident during the

entire year except for DJF, when a negative sign is dominant (Fig. 1b). The reason why negative CGWD exists only in DJF is that the 32-yr-averaged zonal wind profile between 10 and 1 hPa shows negative wind shear in DJF, but not in the other seasons except for JJA between 2 and 1 hPa (Fig. S1 in the online supplemental material). In JJA, strong eastward CGW momentum flux at the source level saturates strongly in the upper stratosphere so the positive CGWD becomes dominant between 2 and 1 hPa (Fig. S2). The strong CGWD in the tropical upper stratosphere possibly contributes to the semiannual oscillation (SAO) in the upper stratosphere, as mentioned in Kang et al. (2018). Another strong negative CGWD forcing exists in the upper stratosphere around 50° – 70° N/S, where a polar night jet exists, and is strongest during SON in the SH, with a maximum negative forcing of $-0.3 \text{ m s}^{-1} \text{ day}^{-1}$. On the other hand, weak positive forcing exists at the equatorward flank of the polar night jet. Negative

CGWD is also noticeable at 30–100 hPa, where the subtropical jet decelerates, and is strongest during DJF in the NH, with a maximum negative forcing of $-0.3 \text{ m s}^{-1} \text{ day}^{-1}$. This strong negative CGWD above the subtropical jet is related to strong westward momentum flux at the source level (i.e., cloud top) that is generated from convective activities in the winter-time storm-track region (Kang et al. 2017). In the OGWD, negative forcing is dominant in the midlatitudes and upper stratosphere, as with the EPD. This negative forcing is strongest during DJF in the NH, with a maximum negative forcing of $-2.2 \text{ m s}^{-1} \text{ day}^{-1}$, and second strongest during June–August (JJA) in the SH. In addition, OGWD exerts negative forcing above the subtropical jet regardless of the season, as with the CGWD, and is stronger in the NH than the SH, especially during DJF. Negative OGWD is evident during all seasons except for the summertime in each hemisphere, when stationary OGWs fail to propagate into the middle atmosphere.

Figure 2 shows latitude–height cross sections of the climatological annual- and seasonal-mean mass streamfunctions calculated by direct method (first row) and downward-control principle (second to seventh rows). In Fig. 2, “Total” means total streamfunction induced by all wave forcing terms, including $\partial \bar{u}/\partial t$, in Eq. (6), and “EPD,” “CGWD,” “OGWD,” “Residual,” and “dudt” mean the streamfunctions that are induced by each forcing. A comparison between the first and second rows reveals that the total streamfunction for the downward-control principle matches the streamfunction for the direct method. Therefore, the downward-control principle can be applied to examine the contribution of each wave forcing to the residual-mean circulation.

The EPD is responsible for the overall two-celled structure, which is consistent with the strongest negative forcing in Fig. 1, but is smaller than the “total” streamfunction in the high latitudes of the upper stratosphere (above 10 hPa) and a portion of the deep branch of the BDC in the summer hemisphere (Figs. 2b,d). The CGWD contributes to the shallow branch of the BDC because of the strong CGWD, where the subtropical jet decelerates (Fig. 1). The streamfunction from CGWD is vertically deepest and strongest during DJF because of the strong CGWD (Fig. 1b) and generally stronger in the winter hemisphere than the summer hemisphere. The streamfunction from OGWD shows strong circulation near the subtropical jet, as with the CGWD, but extends to the upper stratosphere and higher latitudes except in the summer hemisphere, as expected from Fig. 1. The magnitude of the BDC from Residual is comparable

to that from OGWD but with alternating positive and negative values, implying that the Residual could be induced by several factors, which will be discussed in detail in section 6. The BDC from $\partial \bar{u}/\partial t$ is negligible from the annual-mean perspective but non-negligible from the seasonal-mean perspective. The BDC is the strongest during the equinox, when radiative forcing-driven circulation is dominant (Sato and Hirano 2019).

Figure 3 shows the climatological annual- and seasonal-mean streamfunctions from individual forcing terms at 70 hPa. The 70-hPa level is chosen because it is higher than the tropical tropopause, where two-way vertical exchange occurs, but low enough to include most of the BDC (Butchart 2014). In the annual-mean streamfunction (Fig. 3a), the OGWD is stronger than the CGWD in the NH and vice versa in the SH. The streamfunctions from the OGWD and Residual have different signs near 30° – 60°N . This feature is interesting and will be investigated in detail in the future. The streamfunction from $\partial \bar{u}/\partial t$ is nearly zero and negligible, as in Fig. 2a. Similar features are shown in the seasonal-mean streamfunctions (Figs. 3b–e), except that $\partial \bar{u}/\partial t$ is not negligible and its sign depends on the season and hemisphere. Interestingly, the CGWD is always stronger than the OGWD in the SH, while the OGWD is stronger than the CGWD poleward of 25°N except for JJA, when the OGWD and CGWD show comparable magnitudes.

Figure 4 shows a bar chart of the annual and seasonal means of the net tropical upward mass fluxes at 70 hPa (F^{TR}), and their values, including the NH (F^{NH}) and SH (F^{SH}) downward mass fluxes, are also shown in Table 1. The contributions of EPD, CGWD, OGWD, Residual, and $\partial \bar{u}/\partial t$ to the annual-mean upward mass flux are 68%, 4%, 7%, 14%, and 6%, respectively. From a seasonal-mean perspective (Figs. 4b–e), the tropical upward mass flux is strong in the order of DJF, SON, MAM, and JJA, representing a clear annual cycle. It is because EPD, which contributes mostly to the BDC, is strongest in the order of SON, DJF, MAM, and JJA, and both OGWD and CGWD are strongest in DJF and weakest in JJA (Table 1, Fig. 1). The tropical upward mass flux from EPD is strongest in SON but slightly larger than that in DJF because the SH downward mass flux in DJF is very small compared to SON, despite the strong NH downward mass flux in DJF (Table 1). The tropical upward mass fluxes from CGWD and OGWD are strongest in DJF, while those from Residual and $\partial \bar{u}/\partial t$ are strongest in JJA and MAM, respectively. The percentages of the contribution of each forcing to the seasonal-mean total upward mass flux are

FIG. 2. Latitude–height cross sections of the (a) annual-, (b) DJF-, (c) MAM-, (d) JJA-, and (e) SON-mean mass streamfunction from (first row) the direct method and (second to seventh rows) the downward-control principle from the total wave forcing, EPD, CGWD, OGWD, Residual, and $\partial \bar{u} / \partial t$, respectively, averaged from 1979 to 2010.

Streamfunction at 70 hPa

FIG. 3. Latitudinal distribution of the direct mass streamfunction and downward-control mass streamfunctions using the EPD, CGWD, OGWD, Residual, and $\partial\bar{u}/\partial t$ for the (a) annual, (b) DJF, (c) MAM, (d) JJA, and (e) SON means at 70 hPa.

67% (67%), 5% (5%), 10% (3%), 13% (22%), and 5% (3%), respectively, during DJF (JJA) and 58% (80%), 5% (3%), 7% (7%), 14% (9%), and 16% (1%), respectively, during MAM (SON). The smaller OGWD in JJA than DJF is clear from the NH wintertime contribution, as shown in Fig. 1. The increased contribution from the Residual in JJA compared to DJF and that from EPD in SON compared to MAM were mainly caused by the increased SH wintertime contribution. The larger contribution of CGWD than that of OGWD during JJA is consistent with the larger mass streamfunction from CGWD than OGWD, especially in the SH (Fig. 3d).

In Fig. 4, the turnaround latitudes (ϕ_{TL}^{NH} and ϕ_{TL}^{SH}) for calculating the tropical upward mass flux from

each wave forcing are obtained from the direct mass streamfunction. Note that small changes in turnaround latitudes can significantly affect the mass flux calculation, as mentioned in Chun et al. (2011). When the upward mass flux from each wave forcing is calculated by using the turnaround latitudes of the mass streamfunction of each wave forcing (not shown), the flux is stronger than that in Fig. 4. For example, tropical upward mass flux from CGWD using the turnaround latitudes of mass streamfunction of CGWD in DJF is $1.4 \times 10^9 \text{ kg s}^{-1}$, which is about three times larger than that using the direct mass streamfunction. This is because ϕ_{TL}^{NH} and ϕ_{TL}^{SH} from the direct mass streamfunction are 28°N and 46°S , while those by CGWD are 33°N and 22°S , respectively. The upward

FIG. 4. Tropical net upward mass flux estimated from the direct mass streamfunction and downward-control principle using the EPD, CGWD, OGWD, Residual, and $\partial\bar{u}/\partial t$ for the (a) annual, (b) DJF, (c) MAM, (d) JJA, and (e) SON means at 70 hPa.

mass flux from CGWD by additional upwelling at 28° – 33° N and the excluded downwelling at 22° – 46° S increases the tropical upward mass flux. Similar arguments are given in detail in Chun et al. (2011) and Kim et al. (2014).

Figure 5 shows the annual cycle of the tropical upward mass flux at 70 hPa from the direct streamfunction and downward-control principle. Please note that CGWD, OGWD, and Residual are multiplied by 2. The tropical upward mass fluxes estimated by the direct streamfunction and the downward-control principle of the total wave forcing match each other. The annual cycle is dominated by the variation in EPD, as expected. The tropical upward mass fluxes induced by CGWD and OGWD are larger during DJF than JJA by a factor of 1.5 and 4.6, respectively, which enhance the annual cycle. Those induced by the Residual and $\partial\bar{u}/\partial t$ are strong in JJA and MAM, respectively, and the sum of the Residual and $\partial\bar{u}/\partial t$ is nearly flat during the year, increasing the annual-mean upward

mass flux rather than the annual cycle. We further estimate tropical upwelling by averaging the residual vertical velocities (i) between turnaround latitudes and (ii) between 30° N and 30° S (not shown). Still, $\bar{w}_{dc}^*$ from CGWD and OGWD is larger during DJF than JJA by a factor of 1.6 and 5.3 for method (i), respectively, [1.5 and 6 for method (ii)], implying that CGWD and OGWD enhance the annual cycle. Aside from the annual cycle, both CGWD and OGWD increase the annual-mean total upwelling by 5% and 10% for method (i) and method (ii), respectively. Therefore, the annual-mean contributions from CGWD and OGWD to the BDC depend on the estimation and averaging method but range between 4% and 10%.

Figures 4 and 5 show comparable magnitudes of the streamfunctions from OGWD and CGWD, which was also shown in the modeling results by Chun et al. (2011) but is somewhat different from those by Palmeiro et al. (2014), where the contributions from CGWs were negligible in the lower stratosphere. This discrepancy probably occurred because the CGW parameterization in our study, which is similar to that in Chun et al. (2011), exerts strong westward momentum above the subtropical jet (Fig. 1) due to the strong source-level GW momentum flux in the winter-hemisphere storm-track region (Kang et al. 2017).

5. Long-term trend

Stratosphere-resolving GCMs and CCMs have predicted an accelerating BDC in response to climate changes (Butchart and Scaife 2001; Butchart et al. 2006; Garcia and Randel 2008; Li et al. 2008; McLandress and Shepherd 2009; Butchart et al. 2010), although the upwelling branch of the BDC has been predicted to narrow in the lower stratosphere (McLandress and Shepherd 2009; Li et al. 2010; Hardiman et al. 2014). This increasing trend in tropical upwelling is also found in the past records from the observed AoA trend (Aschmann et al. 2014) and vertical residual velocity trend simulated for historical periods (Garfinkel et al. 2017), although exclusively at the lower stratosphere. Abalos et al. (2015) showed a narrowing trend in the width of the turnaround latitude from MERRA for the period of 1979–2012, whereas this trend was not evident from ERA-I and Japanese 55-year Reanalysis (JRA-55). These authors mentioned that this inconsistency might have resulted from artificial discontinuities in the reanalysis data. In this section, the long-term trend in the intensity and width of the BDC and the contribution of each wave forcing during the past decades are examined by using the tropical net upward mass flux and residual vertical velocity.

TABLE 1. NH and SH mass fluxes and their sum (i.e., tropical upward mass flux) at 70 hPa (10^9 kg s^{-1}) calculated from the direct mass streamfunction and downward-control principle by each wave forcing calculated using the CFSR data averaged over the entire year, DJF, MAM, JJA, and SON for 32 years (1979–2010).

	Annual			DJF			MAM			JJA			SON		
	NH	SH	TR	NH	SH	TR	NH	SH	TR	NH	SH	TR	NH	SH	TR
Direct	4.5	4.4	8.9	8.0	2.6	10.6	3.7	5.0	8.7	1.4	5.7	7.1	5.0	4.3	9.3
Total	4.5	4.4	8.9	8.0	2.8	10.8	3.7	5.0	8.7	1.5	5.7	7.2	5.1	4.3	9.4
EPD	2.9	3.2	6.1	5.7	1.5	7.2	2.7	2.4	5.1	0.5	4.3	4.8	2.7	4.8	7.5
CGWD	0.2	0.2	0.4	0.4	0.1	0.5	0.2	0.2	0.4	0.1	0.2	0.3	0.2	0.1	0.3
OGWD	0.5	0.1	0.6	1.0	0.1	1.1	0.5	0.1	0.6	0.1	0.2	0.3	0.5	0.1	0.6
Residual	0.5	0.7	1.2	0.8	0.6	1.4	0.6	0.6	1.2	0.4	1.2	1.6	0.4	0.5	0.9
$\partial \bar{u} / \partial t$	0.4	0.2	0.6	0.1	0.5	0.6	-0.3	1.7	1.4	0.4	-0.2	0.2	1.3	-1.2	0.1

Figure 6 shows the long-term trends in the annual means of the (Fig. 6a) tropical net upward mass flux, (Fig. 6b) residual vertical velocity averaged over the turnaround latitudes, and (Fig. 6c) width of the turnaround latitudes ($\phi_{\text{TL}}^{\text{NH}} - \phi_{\text{TL}}^{\text{SH}}$). Note that convection-related processes in the CFSR systematically changed after 1998, resulting in a spurious increasing trend in the CFSR precipitation, in contrast to no significant trend in the observed precipitation (Li et al. 2015). This change was likely caused by changes in the assimilated radiance from Television and Infrared Observation Satellite (TIROS) Operational Vertical Sounder (TOVS) to Advanced TOVS (ATOVS) (Wang et al. 2011) that increased the cloud depth, which are important factors that determine the source-level CGW momentum flux [cloud-top momentum flux (CTMF)], through changes in the humidity and temperature profiles (English et al. 2000). Because increased cloud depth reduces the magnitude of CTMF through less frequent coincidence of spectral combinations between convective sources and wave-filtering and resonance factors (Kang et al. 2017), the CTMF after 1998 is systematically smaller than before this date, producing an artificial decreasing trend in both the zonal-mean CTMF and CGWD at all latitudes when the analysis period is set to the entire range (1979–2010). Hence, we analyze the data in two periods, namely, P1 (1979–98) and P2 (1999–2010), to avoid the spurious trend in CGWD that is induced by changes in the CFSR assimilation system. Discontinuities in the tropical upward mass flux and residual vertical velocity are less evident in EPD and OGWD (Fig. 6). Stronger discontinuities in GWD than in other forcings are also evident in MERRA and MERRA-2.

In Fig. 6a, the quantification of the factors for the increase in the tropical upward mass flux from each forcing reveals that the increasing trend in the tropical upward mass flux in P1 is contributed by the Residual (47%), EPD (28%), OGWD (18%), CGWD (6%), and $\partial \bar{u} / \partial t$ (1%), with statistically significant trends for

OGWD and CGWD. These contributions are also similar for $\langle \bar{w}^* \rangle$ (Fig. 6b). The large percentage for the Residual implies that unknown processes predominantly strengthen the BDC, which will be discussed later in section 6. One might expect that the increasing trend in the tropical net upward mass flux is affected by the broadening of the tropical upwelling region. However, the width of the upwelling branch is narrowing (Fig. 6c), which is consistent with previous studies (McLandress and Shepherd 2009; Li et al. 2010; Hardiman et al. 2014), whose results indicated that the BDC grew stronger but narrower in the lower stratosphere. This result is confirmed by the increasing trend in the residual vertical velocity averaged between the turnaround latitudes (Fig. 6b). In Fig. 6c, the width of the turnaround latitude from EPD increases, while those from both CGWD and OGWD decrease, implying that GWD significantly contributes to the long-term trend in the turnaround latitudes.

The tropical net upward mass flux (Fig. 6a) and $\langle \bar{w}^* \rangle$ (Fig. 6b) at 70 hPa estimated from both the direct method and downward-control principle show an

FIG. 5. Monthly climatology of the tropical upward mass flux estimated from the direct streamfunction (gray) and downward-control principle from the total wave forcing (black), EPD (blue), CGWD (red), OGWD (green), Residual (light blue), and $\partial \bar{u} / \partial t$ (orange) at 70 hPa.

FIG. 6. Time series of the (a) annual-mean tropical upward mass flux, (b) $\langle \bar{w}^* \rangle$ averaged over the turnaround, and (c) width of the turnaround latitudes at 70 hPa from the direct method and downward-control principle with contributions from each wave forcing. The dashed lines denote the linear trend of each result for P1 (1979–98) and P2 (1999–2010), and the dotted lines denote the linear trend for the total period (1979–2010). The text in the upper-left corner of each panel represents the slope of the linear trend (respective units per year) for P1 and P2, and the trend for the total period is written in parentheses. Statistically significant trends at the 95% confidence level by a t test are marked with an asterisk.

increasing trend over 32 years, although the increasing trend of $\langle \bar{w}^* \rangle$ during P2 is much steeper than that during P1. In addition, the trends of the upward mass flux and $\langle \bar{w}^* \rangle$ estimated from the direct method are statistically significant exclusively in P1 at the 95% confidence level. Considering this situation, we performed a trend analysis for the contribution from each wave forcing during P1 only, and the results are shown in Figs. 7 and 8. The trends in the annual means of the mass streamfunctions calculated based on the direct method and downward-principle induced by EPD, CGWD, and OGWD in the stratosphere are shown in Fig. 7. The trend in the zonal-mean zonal wind is also shown in Fig. 8, alongside the trend in the wave forcing. Statistically significant trends in the zonal-mean zonal wind and wave forcings at the 95% confidence level are stippled, and statistically significant trends in the EP flux vectors at the 90% confidence level are overlaid with magenta. At the bottom of each panel of Fig. 8, the trend and climatology in the zonal mean of the vertical component of the EP flux at 100 hPa, CTMF, and source-level westward OGW

momentum flux are shown. We pay special attention to the streamfunction near the turnaround latitudes, denoted as a thick black line, which is the most influential on the tropical upward mass flux.

Several interesting features are found in Figs. 7 and 8. First, the direct streamfunction shows an increasing trend near the turnaround latitudes from the lower to the upper stratosphere. Second, EPD contributes the most to the trend in the direct streamfunction, except for the low-latitude region in the NH (Fig. 7b). The increasing trend in the EPD-induced streamfunction is explained by the increasing trend in the negative EPD (Fig. 8b), which is statistically significant at 30°–50°N/S above 10 hPa and near 60°S below 10 hPa. At the same location, the zonal-mean zonal wind generally shows a decreasing trend, although it is not statistically significant (Fig. 8a). The vertical component of the EP flux at 100 hPa increases at 30°–60°N/S, possibly because of the acceleration in the zonal-mean zonal wind in the UTLS region, which induces more stationary and transient waves to propagate into the stratosphere (Garcia and

Trend: streamfunction

FIG. 7. Latitude–height cross sections of the linear trend during P1 in the annual-mean mass streamfunction from the (a) direct method and downward-control principle from the (b) EPD, (c), CGWD, and (d) OGWD ($\text{kg m}^{-1} \text{s}^{-1} \text{decade}^{-1}$). Regions with statistically significant trends at the 95% confidence level are stippled. The thick black lines denote the turnaround latitudes averaged in P1. The green contours represent the zonal-mean zonal wind in P1, with a contour interval of 5 m s^{-1} .

Randel 2008; Shepherd and McLandress 2011; Kim et al. 2014). The increase in the EP flux vector and its equatorward propagation near 10 hPa confirm that the increase in the EP flux in the lower stratosphere increases the EP flux divergence above 10 hPa. The strengthening of the BDC in the middle to high latitudes in the lower stratosphere of the SH is caused by the increasing trend in the negative EPD, associated with ozone depletion in the lower stratosphere of SH (Gillett and Thompson 2003; Son et al. 2008) and resultant wind acceleration through the thermal wind relationship and upward shift of the Rossby-wave critical layer, especially during DJF and SON (Kim et al. 2014; Abalos et al. 2015).

Third, CGWD (Fig. 7c) compensates the EPD-induced weakening trend of the BDC in the low latitudes by accelerating the BDC near 30°N/S and decelerates the BDC at latitudes of $40^\circ\text{--}60^\circ\text{N/S}$, especially below 30 hPa. The strengthening of the BDC near 30°N/S is supported by the negative trend in the CGWD above the subtropical jet (Fig. 8c), which implies that the negative CGWD is strengthening based on the negative sign of the mean

CGWD (Fig. 1a). We also emphasize the role of CGWD in changing the streamfunction at 30°S , as the negative wave-forcing trend occurs at the turnaround latitudes. The strengthening of the negative CGWD is caused by both the increase in the westward CTMF (bottom panel of Fig. 8c) and its critical-level filtering by the increased negative wind shear. The increase in the negative wind shear can be inferred from Fig. 8a, which shows westerly acceleration at 50–100 hPa and easterly acceleration above 50 hPa. On the other hand, the positive trend in CGWD at 40°N above ~ 40 km (Fig. 8c) implies that the positive CGWD becomes stronger based on the climatologically positive CGWD (Fig. 1c). Note that the sign of the CGWD at 40°N above ~ 40 -km altitude is determined by the competition between critical-level filtering and wave saturation: the positive drag due to critical-level filtering is generally stronger than the negative drag due to wave saturation, so the CGWD shows a positive sign except during JJA, when negative wind shear exists (Fig. 1). In the bottom panel of Fig. 8c, the eastward CTMF shows a decreasing trend but the westward CTMF

Trend: wind & wave forcing

FIG. 8. Latitude–height cross sections of the linear trend during P1 in the annual-mean and zonal-mean (a) zonal wind, (b) EPD, (c) CGWD, and (d) OGWD (shading). Regions with statistically significant trends at the 95% confidence level are stippled. The thick black lines denote the turnaround latitudes in P1. The green contours represent the zonal-mean zonal wind averaged in P1, with a contour interval of 5 m s^{-1} . The linear trends in the EP flux vectors that are statistically significant at the 90% confidence level are plotted in magenta. The latitudinal distributions of the linear trends in the EP flux $\mathbf{z}$ component at 100 hPa, CTMF (source-level CGW momentum flux), and OGWMF (source-level OGW momentum flux) are plotted at the bottom of (b), (c), and (d), respectively. The eastward (red) and westward components (blue) are shown for CTMF, while only the westward component is shown for OGWMF. Statistically significant trends are plotted with thick lines.

decreases even more, producing a net eastward (i.e., positive) CGWD trend at 40°N above 40-km altitude. In addition, the slightly increased positive wind shear at $\sim 42\text{-km}$ altitude, which can be inferred from the

climatologically positive-wind shear and the weak eastward acceleration trend above 42-km altitude, may have contributed to the positive trend in the CGWD. The trend in the source-level CGWs seem to be the most

influential to the trend in the CGWD, as a statistically significant trend in the CTMF is mostly linked to the significant trend in the CGWD of the stratosphere (30° – 40° S, $\sim 70^{\circ}$ N, and 60° S), although with an exception near 30° – 40° N where significant trend in the CGWD of the upper stratosphere without significant trend in the CTMF.

Fourth, OGWD (Fig. 7d) cancels the weakening trend in the BDC at low latitudes and accelerates the BDC at latitudes of 20° – 60° N/S in the whole stratosphere, except for 30° – 50° N in the lower stratosphere (below 70 hPa). Strengthened and equatorward-shifted westerlies in the upper flank of the subtropical jet enable more OGWs to propagate into the stratosphere, so breaking occurs more frequently in the lower stratosphere (Butchart et al. 2010; Okamoto et al. 2011). On the other hand, less breaking inside the subtropical jet induces a localized decrease in the mass streamfunction (30° – 50° N, below 50 hPa). In addition, OGWs in the stratosphere likely saturate more frequently when the magnitude of the zonal wind decreases, which is shown as a statistically significant OGWD trend at 20° – 60° N above 50 hPa (Fig. 8d). This result is consistent with the fact that OGWs tend to break when the mean wind is weaker (Lindzen 1981; McFarlane 1987). Zonal wind changes seemed to be the most influential to the changes in the OGWD-induced streamfunction, given that OGWD strengthens the BDC, whereas the source-level OGW momentum flux decreases at 20° – 60° N.

To summarize, the increase in the BDC from the EPD is mainly related to the increase in the vertical component of the EP flux at 100 hPa, which likely results from the strengthening of the zonal-mean zonal wind in the UTLS. A statistically significant negative EPD trend is evident where the zonal wind trend is negative, although the negative zonal wind trend is not statistically significant. The increase in the BDC at low latitudes from CGWD greatly depends on the trend in the CTMF. On the other hand, the increase in the BDC from OGWD is controlled primarily by the wave-propagation conditions in both the troposphere and stratosphere. The main driver of the long-term trend in the tropical upward mass flux and tropical upwelling at 70 hPa (Figs. 6a,b) is the increasing negative trends of EPD and OGWD in the stratosphere and the increasing negative trend of CGWD in the lower stratosphere near 30° N/S.

6. Discussion

The above results are based on CFSR data. To confirm these findings, we compare some of the results with three modern reanalysis datasets: MERRA, MERRA-2, and ERA-I. Figure 9 shows latitude–height cross sections of

the annual-mean mass streamfunctions with the direct method and downward-control principle from the total forcing, EPD, Total-(EPD + $\partial\bar{u}/\partial t$), GWD, Residual, and $\partial\bar{u}/\partial t$ in CFSR, MERRA, MERRA-2, and ERA-I averaged over 1980–2010, for which all data are available. However, no GWD exists in ERA-I, so Total-(EPD + $\partial\bar{u}/\partial t$) represents the residual term of the TEM equation, corresponding to the sum of GWD and Residual in the other reanalyses.

The direct mass streamfunctions from the four reanalysis datasets are generally similar, although some differences exist. For example, MERRA and MERRA-2 show stronger circulation in the upper stratosphere compared to CFSR and ERA-I, possibly due to their higher model top (0.1 hPa). A similar pattern is revealed in the total streamfunction and EPD streamfunction (second and third rows of Fig. 9). In the fourth row of Fig. 9, Total-(EPD + $\partial\bar{u}/\partial t$) strengthens the circulation at middle to high latitudes and weakens it at low latitudes in MERRA and MERRA-2. Although a similar pattern is shown in ERA-I above 30-hPa altitude, somewhat stronger circulation than MERRA and MERRA-2 is shown in the lower stratosphere between 60° N and 60° S. CFSR shows similar circulation to ERA-I below 50-hPa altitude but shows counterclockwise circulation at 40° N and clockwise circulation at 40° S, which are not shown in the other reanalysis datasets. The large spread in Total-(EPD + $\partial\bar{u}/\partial t$) implies that the precise quantification of the GWD with this term is not feasible.

The circulations from the GWD (fifth row of Fig. 9) in MERRA and MERRA-2 extend to higher altitudes than those from CGWD+OGWD in CFSR, but the strength is weaker in the lower stratosphere. The Residual (sixth row of Fig. 9) exhibits counterclockwise circulation in the low latitude of NH and clockwise circulation in the low latitudes of SH in MERRA and MERRA-2. In addition, strong circulations with an alternative sign exist in the lower stratosphere. On the other hand, CFSR shows a strong Residual-induced circulation even in the middle to upper stratosphere in the extratropics. The differences in Residual between CFSR and the MERRA series are not surprising because the GWD defined in CFSR is the sum of the CGWD and OGWD, while that in the MERRA series is the sum of the nonorographic GWD and OGWD.

Although the strength of the BDC and its contribution from EPD depend on the reanalysis data, these data seem to have similar magnitudes among the reanalyses in the lower stratosphere. To quantify the similarity, Table 2 shows the annual-mean tropical net upward mass flux at 70 hPa in CFSR, MERRA, MERRA-2, and ERA-I averaged for 1980–2010. The direct mass fluxes from CFSR, MERRA, MERRA-2, and ERA-I are 9.0,

FIG. 9. Latitude–height cross sections of the annual-mean mass streamfunction from (first row) the direct method and (second to seventh rows) downward-control principle from the total wave forcing, EPD, CGWD, OGWD, Residual, and $\partial \bar{\pi} / \partial t$, respectively, averaged from 1980 to 2010 in (a) CFSR, (b) MERRA, (c) MERRA-2, and (d) ERA-I.

TABLE 2. Annual-mean tropical net upward mass flux at 70 hPa (10^9 kg s^{-1}) calculated from the direct mass streamfunction and downward-control principle by each wave forcing averaged for 31 years (1980–2010, left column of each reanalysis) and their long-term trend ($10^9 \text{ kg s}^{-1} \text{ yr}^{-1}$) during the P1 (1980–98, right column of each reanalysis), along with the percentage, by using CFSR, MERRA, MERRA-2, and ERA-I. Note that the percentage of the trend in ERA-I are not calculated because the total trend is nearly zero. Statistically significant trends at the 95% confidence level using a t test are in boldface.

	CFSR			MERRA			MERRA-2			ERA-I		
	Mean (1980–2010)	% (1980–98)	Trend (1980–98)	Mean (1980–2010)	% (1980–98)	Trend (1980–98)	Mean (1980–2010)	% (1980–98)	Trend (1980–98)	Mean (1980–2010)	% (1980–98)	Trend (1980–98)
Direct	9.0		0.054	7.9		0.047	8.2		0.056	8.9		0.002
Total	9.0		0.054	7.9		0.047	8.2		0.054	9.0		0.000
EPD	6.2	68	0.015	6.2	27	0.017	6.3	77	0.014	5.6	62	0.014
GWD	1.0	12	0.011	0.6	20	0.006	0.6	7	0.002	—	—	—
Residual	1.3	14	0.024	0.4	45	0.031	0.5	6	0.031	2.8	31	−0.018
$\partial \bar{u} / \partial t$	0.5	6	0.004	0.7	8	−0.007	0.8	10	0.007	0.6	7	0.005

7.9, 8.2, and $8.9 \times 10^9 \text{ kg s}^{-1}$, respectively. The magnitudes from GWD + Residual, which is the same as that from Total-(EPD + $\partial \bar{u} / \partial t$), are 2.3, 1.0, 1.1, and $2.8 \times 10^9 \text{ kg s}^{-1}$, which correspond to 26%, 12%, 13%, and 31% of the total upward mass flux, respectively. Therefore, the percentage of Total-(EPD + $\partial \bar{u} / \partial t$) in the CFSR ranges between MERRA and ERA-I. The percentage of GWD in both MERRA and MERRA-2 is 7%, while that of CGWD plus OGWD in CFSR is 12%.

We further checked the long-term trend in the tropical upward mass flux at 70 hPa in the three reanalyses during 1980–98 (Table 2). All the datasets show an increasing trend, as with the CFSR: 0.054, 0.047, 0.056, and $0.002 \times 10^9 \text{ kg s}^{-1} \text{ yr}^{-1}$ for CFSR, MERRA, MERRA-2, and ERA-I, respectively, with the differences among the data within 17% except for ERA-I. The relatively flat trend in ERA-I was already reported by Seviour et al. (2012), who suspected that the meteorological variables in ERA-I that are not changed by the assimilated ozone may affect the insignificant long-term trend. The trend in the mass flux from EPD during 1980–98 is 0.015, 0.017, 0.014, and $0.014 \times 10^9 \text{ kg s}^{-1} \text{ yr}^{-1}$ in CFSR, MERRA, MERRA-2, and ERA-I, respectively. This result confirms the positive trend in the EPD, but this value can be either stronger or weaker within 20%. The trend of GWD in MERRA and MERRA-2 is 0.006 and $0.002 \times 10^9 \text{ kg s}^{-1} \text{ yr}^{-1}$, respectively, while that estimated from CFSR is $0.011 \times 10^9 \text{ kg s}^{-1} \text{ decade}^{-1}$, corresponding to 13%, 4%, and 20% of the total trend, respectively. If Total-(EPD + $\partial \bar{u} / \partial t$) is considered as GWD, its contribution to the total trend (GWD + Residual in Table 2) in MERRA, MERRA-2, and CFSR is 79%, 61%, and 65%, respectively, whereas ERA-I shows a decreasing trend.

The current results based on CFSR, as well as recent reanalysis data, show a significant contribution from the residual term of the TEM equation in both the climatology and long-term trend. Possible sources of the residual term are worth considering. First, there are uncertainties in subgrid-scale GWD, given that no global GWD observations are available at the moment. In particular, jet-front-generated GWs (Richter et al. 2010; Plougonven and Zhang 2014) are not considered in this study because of the uncertainties in their generation mechanism and the formulations of the source momentum flux (Chun et al. 2019). However, an additional calculation including frontal GWD (FGWD) parameterization, which is the same as in WACCM4, is performed to assess whether the currently available FGWD parameterization can account for the Residual. Figure 10 shows the 32-yr-averaged mass streamfunction from downward-control principle by FGWD. The contribution of FGWD to the residual-mean circulation is very small, concentrated in the lower stratosphere below

FIG. 10. Latitude–height cross sections of the (a) annual-, (b) DJF-, (c) MAM-, (d) JJA-, and (e) SON-mean mass streamfunction from the downward-control principle from the FGWD calculated using the CFSR data averaged from 1979 to 2010.

~50 hPa, due to the small negative FGWD in the stratosphere (Fig. S3). Therefore, the FGWD contributes 1%–2% of the total upward mass flux at 70 hPa (Fig. S4). This result is consistent with the WACCM4 result that FGWD is strong in the mesosphere while it is very small in the stratosphere (Palmeiro et al. 2014; Richter et al. 2010 for WACCM3.5). One interesting thing in Fig. 10 is, although minor, FGWD acts to slow down the BDC in the middle to upper stratosphere, which partly explains the circulation from midlatitudes to tropics near 30°–50°N/S in Total-(EPD + $\partial\bar{u}/\partial t$) (Fig. 9a). Again, this feature is shown in the WACCM4 simulation by Palmeiro et al. (2014; see their Fig. 6).

Second, intermediate-scale GWs that are neither resolved nor parameterized in the model of CFSR is another source of the Residual. Third, uncertainties in the resolved variables, depending on the model dynamics and physics and the resolution, induce uncertainties in planetary- and synoptic-scale waves. Although these uncertainties in the resolved variables are usually assimilated by using observational data, this assimilation may not be perfect in the stratosphere because of the sparse observation data and larger noise (Fujiwara et al. 2017). Fourth, imbalances in the TEM equation can be caused by the analysis increment during assimilation, especially for the three-dimensional variational data assimilation (3D-VAR) analysis used for CFSR (Sankey et al. 2007; Miyazaki et al. 2016). Fifth, numerical diffusion may not be negligible (Kim et al. 2014), as the model of CFSR has a rather fine horizontal resolution ($\sim 0.3125^\circ$) that requires a high diffusion coefficient to avoid numerical instability. Finally, the uncertainties in $\partial\bar{u}/\partial t$ estimated with 6-hourly data may not be negligible. Further research and discussion are needed to identify the contribution of each source to the Residual, while it is beyond the scope of this paper.

7. Summary and conclusions

In this study, we examined the contributions of CGWD and OGWD to the BDC in terms of the climatology,

seasonal cycle, and long-term trend and compared the results to those of EPD. CGWD contributes to the BDC at low latitudes in the lower stratosphere, while OGWD contributes to the BDC in the lower to upper stratosphere, especially during the wintertime. At 70 hPa, the percentage contributions to the tropical net upward mass flux are 58%–80% for EPD, 3%–10% for OGWD, and 3%–5% for CGWD. CGWD and OGWD increase the seasonal cycle of the BDC by increasing the tropical upward mass flux in DJF more than in JJA. The tropical upward mass flux at 70 hPa increases during P1 (1979–98), and EPD, OGWD, and CGWD contribute to the total increasing trend with values of 28%, 18%, and 6%, respectively. The underlying mechanisms that are responsible for the increasing trend are briefly described in section 5: (i) an increase in the vertical component of the EP flux at 100 hPa, (ii) an increase in the CTMF, and (iii) westerly acceleration in the troposphere and deceleration in the stratosphere, which mainly strengthen the BDC through EPD, CGWD, and OGWD, respectively. The influence of the source-level CGWs on the CGWD trend and resultant BDC trend from the current study suggests that physically based and source-dependent nonorographic GW parameterization is required for understanding/predicting the BDC trend. The current results are compared to three reanalysis datasets in section 6, showing that the percentage from GWD for MERRA and MERRA-2 is 7%, while that from CGWD plus OGWD for CFSR is 12% at 70 hPa. The BDC trends from both the total forcing and EPD increase with differences in the data within 20%, and the BDC from GWD shows an increasing trend for MERRA (13%), MERRA2 (4%), and CFSR (20%) but not ERA-I, implying that our results are within a similar range. Finally, possible sources of the Residual, quite large both in CFSR and in other reanalyses, are discussed.

In this study, we used an offline parameterization because the simulated atmospheric fields are rather realistic in the reanalysis data, and the magnitude of the GW momentum flux in a GCM is usually determined to yield a realistic wind field in the middle atmosphere,

which could be over- or underestimated. Moreover, understanding the past climate by using observation-based data is crucial to better understand future climate projection. Although we used OGWD from reanalysis data and CGWD calculated offline, the best way to understand the past climates and trends of BDC with the downward-control principle is likely through directly using global reanalysis data after implementing source-based GW parameterizations into a GCM, with in situ GW observations from, for instance, satellites.

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