Momentum Flux Spectrum of Convectively Forced Internal Gravity Waves and Its Application to Gravity Wave Drag Parameterization. Part II: Impacts in a GCM (WACCM)

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(Manuscript received 15 February 2006, in final form 18 October 2006)

ABSTRACT

Impacts of a spectral parameterization of gravity wave drag (GWD) induced by cumulus convection (GWDC) in the NCAR Whole Atmosphere Community Climate Model (WACCM1b) are investigated. In the spectral GWDC parameterization, reference wave momentum flux spectrum is launched at cloud top and analytically calculated based on the physical properties of convection and the large-scale flow. The cloud-top wave momentum flux is strong mainly in the Tropics and midlatitude storm-track regions, and exhibits anisotropy and spatiotemporal variability. The anisotropy and variability are determined by the distributions and variations of convective activities, the moving speed of convection, and horizontal wind and stability in convection regions. Zonal-mean zonal GWDC has a maximum of 13–27 (37–50) m s\(^{-1}\) day\(^{-1}\) in the mesosphere in January (July). Impacts of GWDC on zonal wind appear mainly in the low to midlatitudes of the upper stratosphere and mesosphere. In these regions, biases of zonal wind with respect to observation are reduced more than 50% through the GWDC process. In contrast to zonal wind, impacts of GWDC on temperature occur mainly in the mid- to high latitudes. Through the analysis of forcing terms in the zonal wind and temperature equations, it is found that impacts of GWDC result from interaction among wave forcing terms (resolved wave forcing, parameterized background GWD, and GWDC) and meridional circulations induced by the wave forcing terms. With regard to tropical variability, when GWDC is included, the model produces the stratospheric semiannual oscillation with more realistic amplitude and structure and stronger interannual variabilities in the lower stratosphere. These enhanced variabilities are caused by resolved wave forcing and meridional circulations.

1. Introduction

The middle atmosphere is coupled radiatively, dynamically, and chemically with the lower atmosphere. To elucidate coupling mechanisms and to investigate the climatology and variability of the whole atmosphere, general circulation models (GCMs) have been extended to the middle atmosphere (Pawson et al. 2000). For reasonable middle-atmosphere simulations, several additional physical processes are required. Among the processes, momentum transport by high-frequency gravity waves to the large-scale flow is believed to be one of the most important (Lindzen 1981; Holton 1982; Garcia and Solomon 1985; Fritts and Vincent 1987). Such high-frequency gravity waves may be generated by various sources, but a number of observational studies (e.g., Taylor and Hapgood 1988; Swenson and Espy 1995; Dewan et al. 1998; Vincent et al. 2004) have confirmed convection as one of the most important sources. Numerical modeling studies (e.g., Alexander et al. 1995; Piani et al. 2000; Lane et al. 2001; Beres et al. 2002; Song et al. 2003) also have demonstrated that convective clouds can generate strong vertically propagating high-frequency gravity waves. Given that convective clouds are widespread on the globe, convective gravity waves can have global impacts on the middle atmosphere. In fact, convective gravity waves
waves can significantly contribute to the quasi-biennial and semiannual oscillations (QBO and SAO, respectively) in the Tropics (Garcia et al. 1997; Baldwin et al. 2001) and wind reversal near the high-latitude mesopause (Alexander 1996).

High-frequency convective internal gravity waves usually have horizontal wavelengths less than 100–200 km, too small even for the present-day middle-atmosphere GCMs to resolve. To overcome this problem in middle-atmosphere GCMs, several efforts have been made to parameterize the momentum of convective internal gravity waves and its transport to the large-scale flow (Rind et al. 1988; Kershaw 1995: Chun and Baik 1998, 2002). These parameterizations have been found to improve the realism of middle-atmosphere simulations in some GCMs [e.g., the Goddard Institute for Space Studies (GISS) global climate middle-atmosphere model (Rind et al. 1988), the Yonsei University atmospheric GCM (YONU AGCM; Chun et al. 2001), and National Center for Atmospheric Research (NCAR) Community Climate Model (CCM) version 3 (Chun et al. 2004)]. However, they consider only a part of various spectral components that consist of convective internal gravity waves. This drawback may lead to incomplete representations of the wave–mean flow interaction.

Some theoretical studies have been conducted to explicitly include broad spectra of convective internal gravity waves in parameterizations. Beres et al. (2004) and Beres (2004) derived formulations of wave momentum flux spectra above convection in the two- and three-dimensional frameworks in a uniform wind and stability condition, respectively. Song and Chun (2005, hereafter Part I) analytically formulated the phase speed spectrum of cloud-top wave momentum flux in the two- and three-dimensional frameworks in a shear flow with stability difference between convection region and above. These formulations are validated through the comparison with wave spectra simulated by cloud-resolving models.

In this study, we implement the spectral GWDC (SGWDC) parameterization based on the cloud-top wave momentum flux spectrum formulated in Part I into the Whole Atmosphere Community Climate Model (WACCM) version 1b developed at NCAR. Using WACCM1b with the SGWDC parameterization, we examine impacts of the parameterization on the global climatology of zonal wind and temperature and zonal wind variability in the Tropics. To accomplish this, change in the meridional circulations and wave forcing terms through the parameterization will be quantified and their interrelationship and effects on the model climatology and variability will be examined. Recently, Beres et al. (2005) implemented Beres’ (2004) gravity wave source parameterization into WACCM2, and presented the characteristics of the parameterized spectrum and impact of the spectrum in the SAO. The present study parallels Beres et al. (2005) but with different parameterization and with complete analysis on impacts on climatologies and variabilities. Some differences in the model simulations between two studies will be discussed in section 2.

The paper is organized as follows. Section 2 presents model description and experimental design. In section 3, model results are presented and compared with observations, and several analyses are conducted to examine impacts of the SGWDC parameterization in WACCM1b. Summary and conclusions are given in the last section.

2. Model and experimental design

WACCM1b used in this study is a three-dimensional global spectral model with T63 horizontal resolution. This model simulates the atmosphere from the ground to the lower thermosphere (about 140 km) with 66 levels. Vertical grid size is less than 1.5 km below the lower stratosphere, and increases to 3.5 km in the mesosphere and above. The dynamic equations of WACCM1b are solved using a semi-Lagrangian technique (Williamson and Olson 1994) with a time step of 1800 s.

Physical processes in WACCM1b are mostly based on NCAR CCM3 (Kiehl et al. 1996). A deep convective process, which activates the SGWDC parameterization, is also calculated using the scheme of Zhang and McFarlane (1995) as in CCM3. For reasonable middle-atmosphere simulations, several additional processes are required in WACCM1b (Sassi et al. 2002). Among the additional processes, one of the most essential process is the background GWD (BGWD, hereafter), which is calculated using the Lindzen-type spectral parameterization (Kiehl et al. 1996). In this parameterization, reference wave momentum spectrum is launched at 100 hPa, and is given by the Gaussian-shaped phase speed (c) spectrum with its center at $c = 0 \text{ m s}^{-1}$ and e-folding width of 30 m s$^{-1}$. This reference spectrum is discretized into nine waves that have phase speeds from −40 to 40 m s$^{-1}$ at an interval of 10 m s$^{-1}$. The peak magnitude of the spectrum varies only in the meridional direction, with a minimum value of $6.4 \times 10^{-4} \text{ N m}^{-2}$ throughout the Tropics and a maximum of $3.2 \times 10^{-3} \text{ N m}^{-2}$ poleward of about 70° latitude.

In this study, we conduct two simulations driven by climatological ozone and sea surface temperature to examine impacts of the SGWDC parameterization. The one is the control (CTRL) simulation without SGWDC parameterization, and the other is the SGWDC simu-
lation with SGWDC parameterization. In both simulations, WACCM1b runs for 12 yr starting from an identical initial condition (1 July 1978) obtained from a previous WACCM1b simulation conducted without the SGWDC parameterization. The first 1.5 yr of each simulation is regarded as a spin-up period and 10-yr results after the spin-up are analyzed. In the SGWDC simulation, the momentum deposition to a cloud region is calculated based on momentum conservation, as in Chun et al. (2001, 2004).

In the SGWDC parameterization, reference wave momentum spectrum ($M_r$) is launched at the top of a deep convective cloud and is given by the threedimensional analytic wave momentum flux spectrum based on Part I [see (A1)]. The spectrum is calculated in the phase speed grid ranging from $-100$ to $100$ m s$^{-1}$ at 1 m s$^{-1}$ interval for two propagation directions ($\phi$) of $0^\circ$ and $90^\circ$. The vertical evolution of wave spectrum and wave momentum forcing are calculated using Lindzen-type method based on Kiehl et al. (1996). Details of the SGWDC parameterization are described in the appendix.

Beres et al. (2005) implemented wave spectrum parameterization only for the Tropics, and removed the BGWD there to avoid overlap between the two GWD schemes. However, such an implementation excludes convection in the midlatitudes, and thus makes it difficult to examine global impacts of GWDC. In this study, we do not restrict regions where SGWDC parameterization is implemented, nor do we make any modifications to BGWD parameterization. Therefore in the SGWDC simulation, two GWD schemes can be executed at the same grid point. However, this overlap does not necessarily yield an excessive GWD. In fact, the magnitude of total GWD is overall reduced in the SGWDC simulation compared with the CTRL. This will be demonstrated in section 3b.

For validation of the zonal wind climatology of the model, we employ the Upper Atmosphere Research Satellite (UARS) Reference Atmosphere Project (URAP) zonal-mean zonal wind from 1992 to 1998 (Swinbank and Ortland 2003). The URAP data are averaged over the 7 yr for comparison with the model climatology. For observational temperature data, we use the Stratospheric Processes and Their Role in Climate (SPARC) temperature climatology (Randel et al. 2004).

3. Results
a. Momentum flux of convectively forced internal gravity waves

Figure 1 shows latitudinal distributions of 10-yr averaged zonal-mean zonal and meridional cloud-top wave momentum flux spectra ($M_{ct}$) in the SGWDC simulation in January and July. Wave spectra shown in Fig. 1 demonstrate two essential characteristics of the reference (cloud top) wave spectrum in the SGWDC parameterization: the anisotropy of the spectrum and the spatiotemporal variation of spectrum shape. In the Tropics the eastward momentum flux is slightly stronger than the westward flux, but in the midlatitudes the westward flux is much stronger. This anisotropy in the zonal spectrum is largely due to the zonal speed of convection relative to the cloud-top zonal wind (see Fig. 6 of Part I). In the Tropics, the easterly flow is dominant above convectively active regions, and convection in these regions is found to move slowly eastward relative to the cloud-top wind (not shown). As a result, the eastward momentum flux is slightly larger than the westward flux. A method to estimate the horizontal velocity of convection is described in the appendix. In the midlatitudes, overall, the westerly flow is dominant, and convection is found to move westward relative to the westerly flow (not shown). This results in the stronger westward momentum flux. In contrast to the zonal spectrum, for most latitudes the meridional spectrum is almost symmetric with respect to $c = 0$ m s$^{-1}$ because the meridional wind in the troposphere and the estimated meridional velocity of convection are generally weaker than their zonal counterparts, and neither of them is toward a particular direction.

The momentum flux spectrum is broader in the Tropics than in the midlatitudes for both the zonal and meridional spectra. This latitudinal variation of spectral width is caused by the greater strength and depth of convection and stability in tropical convection regions. For details about the dependency of spectral width on the depth of convection (stability in convection region), see Fig. 4 (Fig. 5) of Part I. Also, broad spectra in the Tropics in January are located south of the equator, but in July they exist north of the equator. This is consistent with the annual variation of deep tropical convection regions.

Figure 2 shows global distributions of 10-yr averaged eastward and westward wave momentum fluxes at cloud top and 100 hPa in January. These global distributions clearly demonstrate that the wave momentum fluxes in the SGWDC simulation exhibit significant spatial inhomogeneity. In the Tropics, a strong eastward momentum flux appears over the ocean near $10^\circ$N and over the ocean and continents near $20^\circ$S, and a strong westward flux appears over the continents in Southern Hemisphere (SH). In the midlatitudes, strong westward momentum flux appears mainly in the Northern Hemisphere (NH) storm-track regions. However,
in the NH midlatitude continents, the magnitude of wave momentum flux is negligibly small owing to weak convective activities in those winter continents. In the SH midlatitudes, weak momentum flux appears only over the Pacific.

At 100 hPa, considerable amounts of waves launched at the cloud top (safely below 100 hPa at all model grids) can be filtered out by wind before the waves propagate to the stratosphere (Fig. 2, bottom). In the midlatitudes, the westerly flow is dominant in the upper troposphere, and as a result eastward momentum fluxes are dissipated significantly below 100 hPa. In the Tropics, eastward momentum flux at 100 hPa is confined only near Africa, South Asia, and South America. This is because there exists the easterly flow from the cloud top to 100 hPa only in those three regions (not shown), and the upward propagation of waves with the eastward momentum flux is allowed only in the three regions. On the other hand, the westward momentum flux is less dissipated below 100 hPa except in those three regions.

The spatial distribution of cloud-top wave momentum flux in July is quite different from that in January (Fig. 3) owing to differences in the horizontal distributions of convection and wind between January and July. In the Tropics, strong eastward momentum flux is localized in some regions over the ocean near 20°N and widely distributed over the Pacific and Indian Oceans between 10° and 30°S. In the SH tropical continents, the magnitude of momentum flux is generally small owing to weak convective activities there, as in the NH continents in January. In the midlatitudes, strong westward momentum fluxes appear in association with zonally aligned SH storm-track regions near 35°S. In the NH midlatitudes, strong momentum fluxes appear mainly in the continents where convective activities are quite strong in summer time. At 100 hPa, eastward momentum fluxes in the midlatitudes have negligibly small magnitudes in both hemispheres owing to filtering by the strong upper tropospheric jet. In the Tropics, the eastward momentum flux regions at 100 hPa are moved northward compared with January because the upper-tropospheric easterly flow regions as well as convectively active regions are displaced to the north of the equator in July (not shown). The westward momentum flux is less dissipated below 100 hPa than the eastward

Fig. 1. Latitudinal distributions of 10-yr averaged zonal-mean (left) zonal and (right) meridional cloud-top wave momentum flux spectra in the SGWDC simulation in (top) January and (bottom) July. Positive (negative) values are contoured with solid (dashed) lines.
flux, as in January, but again, its relatively strong dissipation occurs in the easterly flow regions.

At 100 hPa, the magnitudes of the annual-mean eastward and westward wave momentum fluxes averaged over 30°S–30°N are $3.3 \times 10^{-4}$ and $4.8 \times 10^{-4}$ N m$^{-2}$, respectively. These values are 1.2–2.4 times larger than given in Beres et al. (2005), but are reasonable in light of the results of some previous studies. Ricciardulli and Garcia (2000, hereafter RG2000) estimated momentum fluxes by equatorial waves using the global cloud imagery (GCI) data. They showed that the equatorial Kelvin, Rossby–gravity and inertia–gravity waves, which are believed to drive the QBO, can have eastward or westward momentum fluxes of about $2.2 \times 10^{-3}$ N m$^{-2}$ at the tropopause. Their results suggest that annual-mean equatorial eastward or westward momentum flux in this study only amounts to 13%–18% of the momentum fluxes of the whole waves involved in driving the QBO. These percentages are reasonable when compared with the other estimates (15%–30%) by Alexander and Holton (1997), Piani et al. (2000), and Piani and Durran (2001).

The wave momentum flux estimates by RG2000 are suitable for use in determining tunable parameters in the SGWDC parameterization since the GCI data used in RG2000 covers the entire equatorial region. However, there have been some criticisms (Horinouchi 2002; Horinouchi et al. 2003) that RG2000’s algorithm for retrieving convective heating from the GCI data might overestimate the variability of convective activities; thus the resulting momentum flux estimates might be exaggerated. However, it is not clear how much their estimates are, in fact, larger than the observed. As discussed by RG2000, their estimates for total wave momentum fluxes, where the contribution of Rossby waves is ignored, are comparable to the mean value of momentum flux magnitudes obtained from rawinsonde observations by Sato and Dunkerton (1997). Also, their estimates for total inertia-gravity waves [$\sim 14 \times 10^{-4}$ N m$^{-2}$ at the tropopause level (see Table 2 and relevant discussions in RG2000)] are close to the lower bound of estimates (7.2–43 $\times 10^{-4}$ N m$^{-2}$) obtained from 6-yr rawinsonde observations by Vincent and Alexander (2000).

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1 RG2000 estimated $\overline{u w'}$ ($\sim 1.35 \times 10^{-2}$ m$^2$ s$^{-2}$) at the tropopause level. In this study, for direct comparison, the $\overline{u w'}$ is multiplied by the air density (0.16 kg m$^{-3}$) at 100 hPa.

Fig. 2. Global distributions of 10-yr averaged eastward and westward wave momentum flux at (top) cloud top and (bottom) 100 hPa in the SGWDC simulation in January.
b. Impacts on the whole atmosphere climatologies

1) Zonal-mean zonal wind and temperature

Figure 4 shows latitude–height cross sections of zonal-mean zonal wind climatologies in the URAP data and CTRL simulation and their difference (CTRL – URAP) in January and July. The model’s climatology is made by averaging model results over 10 yr after a spinup period. The structure of the zonal-mean zonal wind in the model is comparable to that of the URAP wind. For example, the equatorward tilt of the winter westerly flow with height and the location of the subtropical jet in the troposphere are reproduced well by the model. However, in the model, the easterly flow in the equatorial upper stratosphere, the polar night jet in the stratosphere and the subtropical jet in the winter mesosphere are too strong. Also, the zonal wind reversal in the winter polar mesosphere appears at lower altitudes (~75 km) in the model than in the URAP.

Figure 5 shows latitude–height cross sections of zonal-mean temperature climatologies in the SGWDC simulation and its difference from the CTRL (SGWDC – CTRL) in January and July. Shading represents regions where wind biases of the CTRL with respect to the URAP are reduced in the SGWDC simulation. Widespread shaded regions indicate that wind biases are reduced in many regions through the GWDC process. Especially, in the low-latitude upper stratosphere and mesosphere, biases in the CTRL simulation (Fig. 4, right) are reduced by more than 50% in the SGWDC simulation.

Figure 6 shows latitude–height cross sections of zonal-mean temperature climatologies in the SPARC data and CTRL simulation and their difference (CTRL – SPARC) in January and July. Below the lower mesosphere (~60 km), the structure of the zonal-mean temperature in the CTRL generally agrees with that of the SPARC temperature. However, model temperatures exhibit warm biases of more than 30 K in the winter polar mesosphere and of about 12 K in the equatorial upper mesosphere and summer midlatitude mesosphere. Besides these warm biases, cold biases (~−10 K) exist in the winter hemisphere midlatitudes (40°) near 65 km in both January and July, and accompany the inversion of zonal-mean temperature above the cold bias regions.

Similar to the case of zonal wind, alleviation of temperature biases in the SGWDC simulation is observed in several regions (Fig. 7). However, temperature differences between the two simulations appear mainly in
the mid- to high-latitude regions in contrast to the zonal wind difference, which are mostly confined within the subtropics. Alleviation of temperature biases is localized compared with that of wind biases, but is quite significant in some regions. In the winter hemisphere high latitudes, temperature biases are reduced locally by up to 100%. It is also interesting that the local cold biases at 65 km in the winter midlatitude (40°) are reduced significantly in the SGWDC simulation.

2) FORCING OF THE MEAN FLOW

Although SGWDC parameterization only gives momentum forcing to the model, it can influence significantly the large-scale thermal structure, as shown in Figs. 6 and 7, because the large-scale momentum and thermal distributions are coupled with each other through the meridional circulations. Therefore, for comprehensive understanding of the impacts of GWDC on large-scale circulations, it is necessary to examine how much GWDC affects the meridional circulations in the model. For this, we take account of the following transformed Eulerian–mean (TEM) equation set:

\[
\frac{\partial \mathbf{u}}{\partial t} = \nabla \cdot \mathbf{F} + \frac{u^*}{a \cos \phi} \frac{\partial (\mathbf{u} \cos \phi)}{\partial \phi} - \frac{w^*}{a \cos \phi} \frac{\partial \mathbf{u}}{\partial z} + \frac{1}{\rho_0 a \cos \phi} \nabla \cdot \mathbf{X}_{GWDC},
\]

\[
\frac{\partial T}{\partial t} = - \frac{\mathbf{v} \cdot \frac{\partial \mathbf{u}}{\partial \phi}}{a \cos \phi} - \frac{w}{H} \left( \frac{H N^2}{R} + \frac{\partial T}{\partial z} \right),
\]

\[
\frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left( \frac{\mathbf{v} \cdot \mathbf{u}}{a} \right) + \frac{1}{\rho_0 a \cos \phi} 0 \frac{\partial}{\partial z} \left( \rho_0 \mathbf{w} \right) = 0,
\]

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

where \( \mathbf{F} \) is the Eliassen–Palm flux vector by the model-resolved waves, \( \mathbf{X}_{BGWD} \) is momentum forcing by the background GWD parameterization, \( \mathbf{X}_{GWDC} \) is mo-

Fig. 4. Latitude–height cross sections of zonal-mean zonal wind in the (left) URAP data, (center) CTRL simulation, and (right) difference between the CTRL and URAP in (top) January and (bottom) July. Contour interval is 10 m s\(^{-1}\). Positive (negative) values are plotted with solid (dashed) lines. Zero lines are plotted with solid gray lines.
mentum forcing by convective internal gravity waves calculated through the SGWDC parameterization, $R$ is the gas constant for dry air, and $H$ is the scale height. In (2), $N$ is the static stability calculated from the horizontally and temporally averaged (reference) temperature $T_0(z)$, $T_1$ is a deviation of $T$ from $T_0$, and term $w^* \kappa T_1 / H$ (where $\kappa = R c_p$ and $c_p$ is the specific heat at constant pressure) is ignored for the energy conservation as in Holton (1975).

In (1)–(3), $\sigma^*$ and $\pi^*$ are the northward and vertical components of the residual mean meridional circulations:

$$\sigma^* = \bar{\sigma} - \frac{1}{\rho_0} \frac{\partial}{\partial z} \left( \frac{\rho_0 \bar{\sigma} T^*}{HN^2/R + \partial T_1/\partial z} \right),$$

and

$$\pi^* = \bar{\pi} - \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left( \frac{\rho_0 \bar{\pi} T^* \cos \phi}{HN^2/R + \partial T_1/\partial z} \right),$$

respectively. Here, $\sigma$ and $\pi$ are the zonal-mean meridional and vertical flow, respectively.

Figure 8 shows forcing terms of the TEM zonal wind equation in the CTRL simulation in January. Each forcing term is calculated using daily averaged variables, and then averaged over 10 yr for each month. In
the NH midlatitude stratosphere, negative EP-flux divergence (EPD) is dominant, and is mostly due to the stationary waves, which correspond to monthly mean perturbations (Fig. 9a). In the mesosphere, however, both positive and negative EPDs appear. Their overall patterns are explained by EPD due to transient waves, which are obtained by removing stationary components from perturbations (Fig. 9b). This transition from stationary waves in the stratosphere to transient waves in the mesosphere is consistent with results revealed from the Sounding of the Atmosphere using Broadband Emission Radiometry (SABER) data (Garcia et al. 2005). These transient waves in the mesosphere can be related to the negative northward gradient of the basic PV (Norton and Thuburn 1996, 1999). In the CTRL, such negative PV gradients exist in the mid- to high latitudes above 60 km (Fig. 9c), and their location and shape almost accord with those of the positive EPD, which is associated with wind reversal due to gravity wave forcing. This correlation has also been observed in the National Meteorological Center [NMC; now known as the National Centers for Environmental Prediction (NCEP)] analyses (Randel 1994) and the SABER data (Garcia et al. 2005). In addition to the negative PV gradient, background gravity wave drag (BGWD) with zonal asymmetry caused by stratospheric filtering may contribute to the generation of planetary waves in the mesosphere (Holton 1984). In fact, the term $q_1 X_{\text{BGWD}}$ (where $q_1$ is the perturbation PV), related to planetary wave forcing due to the BGWD (Andrews et al. 1987), is strong above 50–60 km (Fig. 9d). The perturbation PV and northward gradient of the basic PV are calculated using (3.4.7) and (3.4.8) in Andrews et al. (1987).

The BGWD term gives westward and eastward momentum forcing in the NH and SH mesosphere, respectively, and has its local maximum of $\sim -73$ and $123$ m s$^{-1}$ day$^{-1}$ at 50°N, 65 km and 50°S, 85 km, respectively. In the NH, strong BGWD appears at lower altitudes than in the SH. This asymmetry of the BGWD term is responsible for the zonal wind reversal appearing at the lower altitudes in the winter polar mesosphere (see Fig.

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Fig. 6. Latitude–height cross sections of zonal-mean temperature in the (left) SPARC data, (center) CTRL simulation and (right) difference between the CTRL and SPARC in (top) January and (bottom) July. In temperature (temperature difference), contour interval is 10 (4) K. Contours with values less than 220 (0) K are plotted with dashed lines. Zero lines are plotted with solid gray lines.
4). The meridional flow structure (strong northward flow over all latitudes at 60–100 km and a moderate northward flow from 30°S to the north pole at 40–60 km) is identified clearly in the Coriolis force term (COR). The strong meridional advection term (ADVY) appears over all latitudes near 80 km and near 50 km in the NH subtropics. In both the two regions, there exists a strong northward flow and strong latitudinal gradient of the zonal wind (Fig. 4). The northward flow at 60–100 km accompanies the compensating upward and downward flow in the SH and NH mesosphere, respectively. These vertical motions and the vertical gradient of the zonal wind (Fig. 4) induce the negative vertical advection term (ADVZ) in both hemispheres.

Figure 10 shows the difference in the forcing terms of the TEM zonal wind equation between the SGWDC and CTRL simulations in January. Several interesting features can be found in Fig. 10. First, difference in forcing terms between the two simulations is significant. Especially, differences in the EPD, BGWD, and COR can be larger than the GWDC itself (The eastward and westward GWDC have their maximum values of about 25 and −10 m s⁻¹ day⁻¹ at 30°N, 73 km and
20°S, 80 km, respectively). Second, the impacts of the GWDC on the other terms are globally distributed and extended down even to the troposphere, even though the GWDC is confined in the tropical middle atmosphere. Note that change in the BGWD is large poleward of 70°, and changes in the EPD and COR appear even in the troposphere. Third, in the SGWDC simulation, the BGWD is reduced compared with the CTRL, and the total GWD is actually reduced in most regions above 60 km. Note that change in BGWD off-

**Fig. 8.** Latitude–height cross sections of forcing terms (EPD, BGWD, COR, ADVY, and ADVZ) of the TEM zonal wind equation in the CTRL simulation in January. When the magnitude of values is less than 10, values of 0, ±1, ±2, and ±5 are contoured. When the magnitude is between 10 and 50 (50 and 110) values are contoured every 10 (20). Units are m s\(^{-1}\) day\(^{-1}\). Positive (negative) values are plotted with solid (dashed) lines. Zero lines are plotted with gray solid lines.

**Fig. 9.** Latitude–height cross sections of (a) EPD by stationary waves, (b) EPD by transient waves, (c) the northward gradient of basic PV, and (d) the correlation between the perturbation PV and perturbation X\(_{BGWD}\) term in the CTRL simulation in January. When the magnitude of values is less than 10, values of 0, ±1, ±2, and ±5 are contoured. When the magnitude is between 10 and 50 (50 and 110), values are contoured every 10 (20). Units of variables contoured are m s\(^{-1}\) day\(^{-1}\) in (a), (b), m\(^{-1}\) s\(^{-1}\) in (c), and m\(^{-1}\) s\(^{-3}\) in (d). Variables are multiplied by \(10^{11}\) in (c) and \(10^{13}\) in (d). Positive (negative) values are plotted with solid (dashed) lines. Zero lines are plotted with solid gray lines.
sets the maxima of the BGWD in the CTRL (Fig. 8). This is caused by the modification of the dissipation condition of background waves due to the GWDC.

Fourth, the EPD change is significant near 90 km in the Tropics where the EPD in the CTRL is small. This EPD change in the tropical mesosphere is likely due to direct gravity wave forcing (the GWDC and BGWD change) rather than the instability of the large-scale flow. Fifth, the impacts of the GWDC on the COR are largely out of phase with its impacts on the BGWD above 80 km. Sixth, impacts on the ADVY and ADVZ appear in association with change in zonal and meridional flow structure. For example, positive change in the ADVZ near 25°N, 85 km is due to weakening of the mesospheric jet (Figs. 4 and 5) and downward flow in the SGWDC simulation. Change in meridional circulations will be explained below in detail.

In July, the overall structure of all the TEM zonal wind forcing terms is almost the same as that in January (Fig. 10) except that it is reversed with respect to the equator, and the magnitude of difference in the ADVZ is increased (not shown).

3) Meridional Circulation and Temperature

Analysis of the TEM zonal wind forcing terms clearly demonstrates that change in zonal-mean zonal wind obtained by including SGWDC parameterization in WACCM1b is caused by two factors: directly by the GWDC and indirectly by the change of other forcing terms with multiple positive and negative feedback processes. The contributions of these two factors to change in zonal-mean zonal wind are likely to be comparable to each other. The six forcing terms in TEM equation are categorized into two groups: wave forcing terms (EPD, BGWD, GWDC) and the meridional circulation-related terms (COR, ADVY, ADVZ). In general, these forcing terms are not independent of one another since the meridional circulations are in fact generated by the wave forcing terms (Andrews et al. 1987). Therefore, for completeness, meridional circulations induced

![Figure 10. Latitude–height cross sections of difference in forcing terms (EPD, BGWD, GWDC, COR, ADVY, and ADVZ) in the TEM zonal wind equation between the SGWDC and CTRL simulations in January. When the magnitude of values is less than 10, values of 0, ±0.1, ±1, ±2, and ±5 are contoured, and when the magnitude is between 10 and 50 (50 and 70), values are contoured every 10 (20). Unit is m s⁻¹ day⁻¹. Positive (negative) values are plotted with solid (dashed) lines. Zero lines are plotted with gray solid lines.](image-url)
by wave forcing need to be examined. For this, we consider an elliptic differential equation for the residual mean meridional circulations, which can be derived from (1)–(4) as follows:

\[
C_{yy} \frac{\partial^2 \tilde{X}^*}{\partial y^2} + C_{yz} \frac{\partial^2 \tilde{X}^*}{\partial y \partial z} + C_{zz} \frac{\partial^2 \tilde{X}^*}{\partial z^2} + C_y \frac{\partial \tilde{X}^*}{\partial y} + C_z \frac{\partial \tilde{X}^*}{\partial z} = -\frac{f}{\partial z} \cos \phi, \quad (7)
\]

where \( f \) is the Coriolis parameter, and \( F_w \) is the wave forcing (EPD, BGWD, or GWDC). Coefficients \( (C_{yy}, C_{yz}, \ldots, C_z) \) are the same as those in Garcia and Solomon (1983), and \( \tilde{X}^* \) is the residual mean meridional streamfunction and related to \( \tilde{v}^* \) and \( \tilde{w}^* \) as follows:

\[
\tilde{v}^* = -\frac{1}{\cos \phi} \left( \frac{\partial \tilde{X}^*}{\partial z} - \frac{\chi^*}{H} \right), \quad \text{and} \quad \tilde{w}^* = \frac{1}{\cos \phi} \frac{\partial \tilde{X}^*}{\partial y}. \quad (8)
\]

Equation (7) is solved numerically using the multigrid routine (Adams 1989) on a rectangular domain with resolution of \( \Delta \phi = 3.75^\circ \) and \( \Delta \phi = 1.875 \) km. The domain covers all latitudes and from the ground \( (z_b = 0 \) km) to the top boundary \( (z_f = 120 \) km). The coefficients in (7) are calculated using 10-yr averaged zonal-mean zonal wind and temperature in the CTRL and SGWDC simulations. Boundary conditions for \( \chi^* \) are given as follows. At both poles, \( \chi^* = 0 \), and at \( z = z_b \), \( \frac{\partial \chi^*}{\partial z} = 0 \), as in Garcia and Solomon (1983). At \( z = z_h \), \( \chi^* \) is specified using the linearized steady-state formulation for \( \pi^* \) induced by \( F_w \) (Garcia 1991; Haynes et al. 1991):

\[
\chi^*(y, z_h) = -\frac{1}{\rho_0(z_b)} \int_{z_b}^{z_h} \left[ \frac{\rho_0(z)F_w(y, z) \cos \phi}{2\Omega \sin \phi} \right] dz. \quad (9)
\]

Here, \( \rho_0(z_b) \chi^*(y, z_b) \) is ignored since it is much smaller than the value at \( z = z_b \). Since this bottom boundary condition holds outside of the Tropics, \( \chi^*(y, z_b) \) at \( |\phi| < 15^\circ \) is linearly interpolated using the values at \( |\phi| = \pm 15^\circ \) as in Garcia (1991).

Figure 11 shows differences between the northward and vertical components of the meridional circulations induced by EPD, BGWD, and GWDC between the SGWDC and CTRL simulations in January. For validation of the solutions of (7), the sum of \( \tilde{v}^* \) and \( \tilde{w}^* \) by the three wave forcing terms is compared with \( \tilde{v}^* \) and \( \tilde{w}^* \) calculated using (5) and (6), respectively. The two are comparable to each other in magnitude and structure except in some regions in the equatorial mesosphere and lower thermosphere.

The findings from the meridional circulations by each wave forcing are as follows. First, the GWDC induces the positive \( \pi^* \) throughout the whole mesosphere and negative (positive) \( \pi^* \) in the NH (SH) mid- to high latitudes. Second, change in \( \pi^* \) induced by the EPD accounts for the difference in the COR at 40–60 km in the NH and at 60–70 km over all latitudes (Fig. 10). Difference in \( \pi^* \) induced by the EPD offsets the maxima in the NH (SH) mid- to high latitudes (not shown), and thus indicates that \( \pi^* \) due to the EPD is reduced overall in the SGWDC simulation. Third, change in \( \pi^* \) induced by the BGWD accounts for difference in the COR at 85–110 km (Fig. 10). In the CTRL, the negative (positive) \( \pi^* \) due to the BGWD appear below 90 km in the NH (SH) mid- to high latitudes (not shown) are reduced in the SGWDC simulation except near 70 km in polar regions. Fourth, both the BGWD and GWDC enhance the positive \( \pi^* \) at 75–85 km over most latitudes. This account for large change in the COR and ADV\( \gamma \) near 80 km (Fig. 10), and enhances vertical motions at 70 km in polar regions in the SGWDC simulation.

Change in the meridional circulations caused by GWDC and the other wave forcing terms interacting with GWDC occurs over a broad range of latitudes. This indicates that the remote impacts of GWDC on the zonal-mean temperature (Figs. 6 and 7) can be attributed to the change in the meridional circulations. Using the meridional circulations induced by wave forcing terms and the thermodynamic Eq. (3), the temperature tendency induced by each wave forcing can be calculated. The temperature tendencies by wave forcing terms and their difference between the SGWDC and CTRL simulations are shown in Fig. 12. Since the GWDC is stronger in July (the zonal-mean eastward and westward GWDC have their maximum values of about 30 and \(-50 \) \( \text{m s}^{-1} \) day\(^{-1} \) at 30°N, 78 km and 35°S, 69 km, respectively) and as a result temperature differences due to wave forcing are more clearly identified, results for July are presented in Fig. 12.

As shown in Figs. 6 and 7, local temperature minima in the mid-latitude mesosphere in the CTRL are reduced in the SGWDC simulation (see also the thick contours of 220 K in Fig. 12). As a result, the strength of the mesospheric temperature inversion is reduced in the SGWDC simulation. It is well known that the mesospheric temperature inversion is frequently observed regardless of season and location (e.g., Hauchecorne et al. 1987; Leblanc et al. 1995; Leblanc and Hauchecorne 1997). These observational results might indicate that the GWDC process unrealistically weakens the mesospheric inversion. However, as is seen in the SPARC
temperature data (Fig. 6), such an inversion layer does not appear in the zonal-mean climatology.

From the temperature tendency by BGWD in the CTRL simulation (top-middle panel of Fig. 12), it is clear that the local temperature minimum and accompanying inversion are mainly due to the BGWD term. Given that the horizontal distribution of background wave momentum is specified somewhat arbitrarily (i.e., uniformly in the zonal direction), it is difficult to regard the inversion layer due to the BGWD as being realistic. It is also interesting that the strong EPD appears near the inversion in the CTRL simulation. However, the EPD induces warming in the local temperature minimum region, and thus acts to diminish the inversion (the top-left panel). This result seems contradictory to the mesospheric inversion due to planetary waves (Sassi et al. 2002). However, note that their study is concerned with local inversions by planetary waves, rather than the inversion of zonal-mean temperature. Similar to the EPD, the GWDC also diminish the inversion by inducing warming near the local temperature minimum region. Adiabatic warming or cooling due to wave induced vertical motions accounts for most of these temperature tendencies. The meridional advection term is generally smaller than the adiabatic term.

In the SGWDC simulation, the EPD and BGWD diminish in the middle mesosphere, and therefore both the temperature tendencies by EPD and BGWD are also reduced. The temperature difference by the BGWD and temperature tendency by the GWDC contribute to the positive temperature difference in the SH low- to midlatitudes at 60–80 km and in the NH mid-latitudes at 80–90 km in July (see Fig. 7).

c. Impacts on the tropical middle-atmosphere variabilities

Convection is more persistent and more widespread in the Tropics than in any other region of the globe. Thus, gravity waves induced by tropical convection are expected to have profound effects on variabilities such
as the QBO and SAO in the tropical middle atmosphere. Here, we present model-simulated variabilities in the tropical middle atmosphere, and examine impacts of the SGWDC parameterization on the variabilities.

Figure 13 shows time–height cross sections of zonal-mean zonal wind averaged over 15°S–15°N and their composite seasonal cycles for the CTRL and SGWDC simulations. In both simulations, variabilities in the

![Image of Figure 13](image-url)

**Fig. 12.** Tendencies of zonal-mean temperature due to (left) EPD, (center) BGWD in the (top) CTRL and (middle) SGWDC simulations. (right middle) Temperature tendency due to GWDC in the SGWDC simulation. (bottom) Difference between the two simulations in July. Each wave forcing is shaded behind contours of the temperature tendency because of the wave forcing. Zero BGWD areas below 15 km are stippled. In temperature tendency and its difference, values of 0, ±0.5, ±1.5, and ±10 K day^{-1} are contoured, and positive (negative) values are plotted with solid (dashed) lines. (top), (middle) Thick black lines represent a contour of 220 K between 50 and 100 km in each simulation.
equatorial middle atmosphere are dominated by the SAO. Composite seasonal cycles show that the first cycle (January–June) of the stratospheric SAO (SSAO hereafter) near 50 km is stronger than the second (July–December). This agrees well with observations by Delisi and Dunkerton (1988). The seasonal cycles also demonstrate that the mesospheric SAO (MSAO, hereafter) near 80 km is roughly out of phase with the SSAO, as described in observational studies by Hirota (1978, 1980) and Hamilton (1982). Although some observed characteristics of the SSAO are reproduced, the SSAO in the CTRL simulation has some unrealistic features. For example, the SSAO does not descend below 46 km (about 1.4 hPa), and its easterly phase is too strong compared with the observations. The composite easterly SSAO wind in the first (second) cycle has its maximum of $-74$ ($-53$) m s$^{-1}$ at 51 (53) km. In the SGWDC simulation, meanwhile, the westerly phase of the SSAO descends to 38 km (4.4 hPa), and the intensity of the easterly phase is reduced significantly. The maximum value of the composite easterly SSAO wind at 48 km in the first and second cycles are reduced to $-53$ and $-43$ m s$^{-1}$, respectively. Comparison with observations (Garcia et al. 1997) suggests that in the SGWDC simulation the SSAO becomes more realistic. However, the MSAO in the SGWDC simulation is too weak compared with observations.

Figure 14 shows composite seasonal cycles of equatorial-mean (15°S–15°N) forcing terms (COR + ADV, EPD, BGWD, and GWDC) in the TEM zonal wind equation. Here, COR + ADV represents the sum of the COR, ADVY, and ADVZ terms. Regions where the seasonal cycle of zonal-mean zonal wind tendency is negative are shaded. The seasonal cycle of the wind tendency is constructed using daily mean zonal-mean zonal wind averaged over 15°S–15°N. At the level of the SSAO (≈50 km), in the CTRL simulation, the positive BGWD term exists in the positive wind tendency regions, and the negative EPD and COR + ADV appear in the negative wind tendency regions (Fig. 13). The negative COR + ADV near 50 km is mainly due to ADVY (not shown). These results suggest that the westerly and easterly phases of the SSAO wind in the CTRL are mainly induced by the BGWD term and by the EPD and ADVY terms, respectively. In the SGWDC simulation, the positive GWDC term is added to the positive BGWD term near 50 km, and as a result the westerly SSAO wind descends. At the level of the MSAO (≈80 km), in the CTRL simulation, the positive and negative wind tendencies are produced mainly by
the BGWD term and by the BGWD and EPD terms, respectively. In contrast to the SSAO, the COR + ADV acts to oppose the wind tendency. In the SGWDC simulation, near 80 km, the zonal wind tendency does not exhibit a clear semiannual cycle as a result of the negative tendency in January and June. This negative tendency is due to the enhanced negative EPD term. In addition, in March and September, the positive COR + ADV is greater than in the CTRL simulation, and disrupts the development of the easterly MSAO wind in April and October, respectively (Fig. 13).

Similar to the present study, Beres et al. (2005) reported that WACCM2 with Beres et al. (2004) convective gravity wave parameterization produces quite realistic SSAO, but is not successful in reproducing the realistic easterly phase of the MSAO. To understand the unrealistic feature of the MSAO simulated by Beres et al. (2005), Richter and Garcia (2006) carried out the analysis of the TEM zonal wind forcing terms. For the westerly phase of the MSAO, the EPD, and advection term (identical to the COR + ADV in this study) tend to oppose to the westerly forcing by gravity waves at the solstices. For the easterly phase, the three terms are small compared with those in the westerly phase, and gravity wave forcing and advection terms are nearly out of phase with each other. These results are consistent with those in this study. More realistic MSAO might be produced by improving reality of processes such as inertia–gravity waves and tides in model rather than by improving parameterizations of small-scale gravity waves, as mentioned by Richter and Garcia (2006). However, in case of the MSAO, small-scale gravity waves are still worth consideration, given that, for example, secondary waves induced by breaking of primary waves excited from the tropospheric convection can be effective momentum sources in the mesosphere (Holton and Alexander 1999; Zhou et al. 2002).

The QBO is not produced even in the SGWDC simulation. However, interannual variabilities are enhanced in the lower stratosphere. Figure 15 shows time–height cross sections of the zonal-mean zonal wind with only interannual variability, which is obtained using a low-pass filter with a cutoff frequency of 1 yr. In the CTRL, the interannual variabilities exist only above 38 km but in the SGWDC simulation they appear even at 28 km. To examine how interannual variabilities are enhanced in the lower stratosphere through the GWDC process, spectral analysis is conducted. For this analysis, we cal-

- Fig. 14. Composite seasonal cycles of equatorial-mean (15°S–15°N) forcing terms (COR + ADV, EPD and BGWD) in the TEM zonal wind equation in the (top) CTRL and (bottom) SGWDC simulations. Composite seasonal cycles for the GWDC term in the SGWDC simulation is added at the bottom. The COR + ADV denotes the sum of the COR, ADV Y, and ADV Z terms. Values of 0, ±10⁷, ±2 × 10⁷, and ±5 × 10⁷ m s⁻¹ day⁻¹, where n = −1, 0, or 1, are contoured. Positive (negative) values are plotted with solid black (dashed gray) lines. Shading represents regions where composite seasonal cycle of the tendency of zonal-mean zonal wind averaged over 15°S–15°N is negative. As in Fig. 13, horizontal lines are plotted at 38 and 46 km.
calculate the power spectral densities (PSDs) of the zonal-mean zonal wind and forcing terms in the TEM zonal wind equation, averaged over 15°S–15°N (Fig. 16). Before this calculation, the linear trend in the data is removed, and then the Welch window function (Press et al. 1992) is applied to the detrended data to minimize the leakage of the spectral power.

The PSDs of the zonal-mean zonal wind demonstrate that the interannual variabilities with periods of 20 and 30 months are enhanced at 10 hPa in the SGWDC simulation. The PSDs of the forcing terms suggest that the variability with the period of 20 (30) months is likely due to the ADVY, COR, and ADVZ terms (ADVY, COR, and EPD terms). However, time series of the forcing terms with the period of only 20 or 30 months (Fig. 17) show that actual variabilities arise through the smaller forcing terms. As for the variability with the 30-month period, the ADVY term is canceled by the COR term, and actual wind tendency of the 30-month period arises through constructive interference between the EPD and ADVZ terms. In the case of 20 months, the ADVY term is almost out of phase with the sum of the ADVZ and COR terms, and as a result actual wind tendency is induced by the EPD term.

It is noteworthy that although the EPD term becomes important in relation to the variabilities in the SGWDC simulation, its magnitude does not differ much from that in the CTRL simulation. Also, the EPD term is due not to the upward propagating resolved waves that might be generated by tropical convection, but to the downward propagating waves. In both the simulations, the vertical components of the EP flux are mostly negative in the tropical stratosphere (not shown).

4. Summary and conclusions

A spectral parameterization of GWD induced by cumulus convection based on Part I is implemented into NCAR WACCM1b and its impacts in the large-scale circulations in the middle atmosphere are examined. Reference (cloud top) wave momentum flux spectrum exhibits two important characteristics: anisotropy in wave propagation and spatiotemporal variation in spectrum shape. In the Tropics the eastward momentum flux is slightly stronger than the westward flux, and in the midlatitudes the westward flux is much stronger. This anisotropy is due to the movement of the convection relative to cloud-top wind. Cloud-top wave momentum flux has broader spectra in the Tropics than in the midlatitudes, because the strength and depth of convection and stability in convection regions are larger in the Tropics. Cloud-top wave momentum fluxes also exhibit significant spatial inhomogeneity. In the summer hemisphere, strong momentum fluxes appear mainly in the continents, but in the winter hemisphere, they appear mainly over the ocean. Especially in the winter hemisphere midlatitudes, strong wave momentum fluxes appear in association with the storm-track regions over the ocean. Waves launched at cloud top can be filtered out significantly before they propagate to the stratosphere. As a result, in the midlatitudes the westward momentum flux is dominant at 100 hPa.

Zonal-mean zonal wind and temperature biases of WACCM1b are alleviated in many regions through the GWDC process. As for zonal wind, the easterly flow in the equatorial upper stratosphere, the polar night jet in the stratosphere, and the subtropical jet in the winter mesosphere are improved significantly. In contrast to the zonal wind difference, temperature change through the GWDC process appears mostly in the mid- to high latitudes rather than in the Tropics. Particularly, the temperature inversion in the midlatitude mesosphere is reduced through the addition of GWDC to the model. To examine changes in the wind and temperature in
detail, forcing terms of the TEM zonal wind equation and meridional circulations are analyzed. GWDC generally acts to induce the change in the zonal-mean zonal wind between the SGWDC and CTRL simulations. However, it is found that actual wind change arises as a result of a combination of interactions among wave forcing terms (EPD, BGWD, and GWDC) and meridional circulation-related terms (COR, ADVY, and ADVZ) induced by the wave forcing terms. The BGWD and GWDC terms are major forcing terms to induce change in zonal wind and meridional circulations in the mesosphere, while EPD has relatively large effects on the zonal wind and meridional circulations in the upper stratosphere. The meridional circulations induced by wave forcing terms show that the zonal-mean temperature inversion in the CTRL simulation is primarily induced by the BGWD term, and the EPD term acts to reduce the strength of the inversion. As GWDC is included in the model, circulations induced by the EPD and BGWD terms are generally weakened, and GWDC induces warming in the temperature inversion region. As a result, the temperature inversion is reduced in the SGWDC simulation.

GWDC also improves the structure and magnitude of the SSAO in the model by reducing the excessively strong SSAO easterly wind, but leads to somewhat weak MSAO. The weak MSAO is attributed to the meridional circulations and EPD terms. The QBO is not simulated even in the SGWDC simulation. However, interannual variabilities are enhanced in the lower stratosphere through the GWDC process, largely due to the meridional advection and EPD terms in the lower stratosphere. The SGWDC parameterization can consider explicitly the anisotropy in wave propagation and spatiotemporal inhomogeneity in wave activity, based on the physical properties of convection and large-scale flow. Even though anisotropy and inhomogeneity are the basic characteristics of the atmospheric gravity waves, they have been ignored or specified a priori in most of current spectral GWD parameterizations. Therefore, implementing physically based wave spectra as what is used in the SGWDC parameterization in GCMs will be one of key steps toward the realistic representation of gravity wave processes in GCMs.

Although SGWDC parameterization reduces model biases in many regions, there are still large wind biases in the upper mesosphere and lower thermosphere (UMLT) (>90 km) even in the SGWDC simulation (Fig. 5). Given that the biases in the tropical UMLT regions increase more in equinox seasons and that their

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**Fig. 16.** PSDs of the equatorial-mean (15°S–15°N) (left) zonal-mean zonal wind and (right) forcing terms of the TEM zonal wind equation at 10 hPa (32.2 km) as a function of frequency in the (top) CTRL and (bottom) SGWDC simulations. Periods at which some peaks of the zonal wind PSD appear are shown over the peaks.
pattern is similar to that of the EP flux divergence by the diurnal tides (Miyahara et al. 1993; McLandress et al. 1996) (not shown), the wind biases in those regions are likely due in part to lack of diurnal tide activities in WACCM1b. In mid- to high-latitude UMLT regions, however, obliquely propagating gravity or planetary waves, besides tides, can affect the observed wind structure significantly (Miyahara et al. 1991). Therefore, to understand why biases arise in these regions, detailed analysis of resolved gravity and planetary waves as well as tides will be necessary.

Acknowledgments. This work was supported by the Korea Research Foundation Grant funded by the Korean Government (MOEHRD) (R02-2004-000-10027-0) and by the Ministry of Science and Technology of Korea through the National Research Laboratory Program (M10500000114-O6J0000-11410). The authors wish to thank two anonymous reviewers for their comments and suggestions on the original manuscript. The first (ISS) and second (HYC) authors would also like to acknowledge the support from KISTI (Korea Institute of Science and Technology Information) under The Sixth Strategic Supercomputing Support Program. The use of the computing system of the Supercomputing Center is also greatly appreciated.

APPENDIX

Description of the SGWDC Parameterization

The SGWDC parameterizations are basically composed of two parts: determination of reference-level spectrum and calculation of the vertical profile of wave momentum flux.

a. Reference-level spectrum

In the SGWDC parameterization, reference-level spectrum is given by the three-dimensional cloud-top wave momentum flux spectrum [see also (28) in Part I]:

$$M_{ct}(c, \varphi, z_{ct}) = \text{sgn}[c - U_{ct}(\varphi)]p_{ct} \frac{2|\varphi|}{A_h L_t}$$

$$\times \left( \frac{g}{c_p T_c N_{ct}^2} \right)^2 \frac{N_{ct}}{|c - U_{ct}(\varphi)|} \Theta(c, \varphi),$$

(A1)

where $c$ is the phase speed; $\varphi$ is the wave propagation direction; $z_{ct}$ is the cloud-top height; $U_{ct}(\varphi)$ is the cloud-top horizontal wind parallel to the direction of $\varphi$; $p_{ct}$, $N_{ct}$, and $T_c$ are the air density, stability, and temperature at $z = z_{ct}$, respectively; $N_q$ is the mean stability below $z = z_{ct}$; $A_h$ and $L_t$ are the area and time, respectively, used for averaging.

In (A1), $|X|^2$ is separately calculated for sheared and uniform wind cases, as formulated in Part I. When Ri in a shear layer near the ground $|X|^2 = N_{ct}^2/(U_{ct} - U_{ground})z_{ct}$, where $z_{ct}$ is the height of the shear layer] is larger than $10^4$, $|X|^2$ for uniform wind is used. The $\Theta(c, \varphi)$ is calculated using a three-dimensional forcing [see (30) in Part I] as follows:

$$\Theta(c, \varphi) = q_0 \left( \frac{\delta_h \delta_i}{32\pi^2} \right)^2 \frac{1}{1 + (c - c_{qh})^2 / c_0^2}.$$  

(A2)

where $q_0$ is the magnitude of forcing; $\delta_h$ and $\delta_i$ are the horizontal and time scales of the forcing; $c_{qh}$ is the speed of the forcing in the direction of $\varphi$, and $c_0 = c_{qh}/c_h$.

In (A1) and (A2), $q_0$ is calculated as $c_c f H_{max}$, where

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Fig. 17. Time series of the equatorial-mean (15°S–15°N) TEM zonal wind forcing terms with periods of (top) 30 and (bottom) 20 months in the SGWDC simulation. The thick black curve represents the sum of all the time series plotted in the same panel.
$H_{\text{max}}$ is the maximum deep convective heating rate (K s$^{-1}$) at each grid column, $c_\phi$ is the specific heat, and $c_f$ is a conversion factor. In this study, $c_\phi$ is specified to be 540. The $\delta_t$ and $\delta_l$ are set to 5 km and 20 min, respectively. The $L_t$ is set equal to $\delta_t$, and $A_h$ is set to $\pi(100L_t)^2$ considering the horizontal propagation of waves (see Fig. 6 of Part I and relevant discussions).

Cloud-top and -bottom heights ($z_{ct}$ and $z_{cb}$) are given by $z_{ct} = z_m + D/2$ and $z_{cb} = z_m - D/2$, respectively. Here, $z_m$ and $D$ are calculated as $(\Sigma_k z_k H_k)/(\Sigma_k H_k)$ and $2\sqrt{(\Sigma_k (z_k - z_m)^2 H_k)(\Sigma_k H_k)}$, respectively, where $z_k$ and $H_k$ are the height and deep convective heating rate at the $k$th level from the ground, respectively. The $c_{gh}$ is obtained by projecting the horizontal velocity of convection ($c_\phi$) onto the direction of $\varphi$. The $c_\eta$ is estimated based on empirical formula (Corfidi et al. 1996): $c_\eta = \mathbf{V}_{CT} - \mathbf{V}_{LLJ}$, where $\mathbf{V}_{CT}$ is the horizontal wind vector averaged over $z_{cb} \leq z \leq z_{ct}$, and $\mathbf{V}_{LLJ}$ is the low-level jet and given as the horizontal wind vector with maximum magnitude below $z = z_m$.

b. Vertical profile of wave momentum flux

The vertical profile of wave momentum flux spectrum $M(c, \varphi, z)$ is calculated based on McFarlane’s (1987) Froude number-based formulation of Lindzen’s (1981) linear theory, which is described in Kiehl et al. (1996). The calculation is performed using the momentum flux spectrum for saturated waves $M(c, \varphi, z)$ and wave diffusivity $D_w$.

The $M(c, \varphi, z)$ is given by

$$M(c, \varphi, z) = \rho_0(z) F_z^2 \frac{k_b(c)}{2N^2} [c - U(\varphi, z)]^2. \quad (A3)$$

Here, $k_b(c)$ is the horizontal wavenumber as a function of phase speed, and is derived through $k_b(c) = 2\pi/(\delta_t \sqrt{\pi}) \sqrt{1 + (c - c_{gh})^2 c_b^2}$. The $F_z^2(\leq 1)$ is the critical Froude number squared, and $F_z^2$ is specified to be 0.01 in this study.

The wave diffusivity ($D_w$) is assumed to be nonzero even for unsaturated waves and expressed as $D_w = D_s \min([M]/M_s$, 1). Here, $D_s$ is the saturation diffusivity given by

$$D_s = \frac{[c - U(\varphi, z)]^2}{N^2} \left\{ \frac{F_z^2 k_b(c)}{HN} [c - U(\varphi, z)]^2 - \alpha \right\} - v_m, \quad (A4)$$

where $H$ is the scale height, $\alpha = 10^{-6} \text{ s}^{-1}$ is a rate of radiative dissipation, and $v_m$ is the molecular viscosity given by $3.55 \times 10^{-7} T(z)^{2/3} p(z)$ (Banks and Kockarts 1973).

In grid columns where deep cumulus parameterization is calculated, momentum flux spectrum $M(c, \varphi, z_t)$ is calculated using (A1), and supersaturated parts in the spectrum are made to be saturated using (A3). Between two adjacent levels ($z$ and $z + \Delta z$) above $z = z_{ct}$, waves that reach their critical levels are removed and otherwise diffused as follows: $M(c, \varphi, z) = M(c, \varphi, z) \exp(-2m_\text{D} \Delta z)$, where $m_t$ is the imaginary part of the vertical wavenumber and expressed as $N/[2k_b(c)|c - U(\varphi, z)|^2] [\alpha + N^2[c - U(\varphi, z)]^2(v_m + D_s)]$. If $|M| > M_t$ even after the diffusion, $|M|$ is set equal to $M_t$.

These filtering and diffusion are repeated upward to determine the vertical profile of the $M(c, \varphi, z)$ up to the model top. For given $c_\phi$ and $\varphi$, the basic-state wind tendency by gravity waves is calculated as

$$\frac{\partial U}{\partial t}(c_\phi, \varphi, z) = \text{sgn}(c - U(\varphi, z)) \times \min \left\{ \frac{1}{\rho_0(z) \partial z} \left| \frac{\partial U}{\partial t} \right| \right\}, \quad (A5)$$

where $|\partial U/\partial t|_{\text{theory}}$ is the theoretical wind tendency due to gravity waves and given by $F_z^2 k_b(c)[c - U(\varphi, z)]^2/(2NH)$. Finally, tendencies of large-scale zonal and meridional wind ($\overrightarrow{i}$ and $\overrightarrow{v}$) due to gravity waves are obtained as follows:

$$\frac{\partial \overrightarrow{i}}{\partial t}(z) = \sum_{j=1}^{n_z} \sum_{l=1}^{n_w} \frac{\partial U}{\partial t}(c_\phi, \varphi, z) \cos \varphi_j, \quad (A6)$$

and

$$\frac{\partial \overrightarrow{v}}{\partial t}(z) = \sum_{j=1}^{n_z} \sum_{l=1}^{n_w} \frac{\partial U}{\partial t}(c_\phi, \varphi, z) \sin \varphi_j. \quad (A7)$$

Here, $n_z$ and $n_w$ are numbers of phase speed and propagation direction, respectively.

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