Momentum budget of the migrating diurnal tide in the Whole Atmosphere Community Climate Model at vernal equinox

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[1] The momentum budget of the migrating diurnal tide (DW1) at the vernal equinox is studied using the Whole Atmosphere Community Climate Model, version 4 (WACCM4). Classical tidal theory provides an appropriate first-order prediction of the DW1 structure, while gravity wave (GW) forcing and advection are the two most dominant terms in the momentum equation that account for the discrepancies between classical tidal theory and the calculation based on the full primitive equations. It differs from the conclusion by McLandress (2002a) that the parameterized GW effect is substantially weaker than advection terms based on the Canadian Middle Atmosphere Model (CMAM). In the region where DW1 maintains a large amplitude, GW forcing in the wave breaking region always damps DW1 and advances its phase. The linear advection largely determined by the latitudinal shear of the zonal mean zonal wind makes a dominant contribution to the phase change of DW1 in the zonal wind compared to the GW forcing and nonlinear advection. However, nonlinear advection is more important than GW forcing and linear advection in modulating the amplitude and phase of DW1 in the meridional wind. The DW1 amplitudes in temperature and winds are smaller than the TIMED observations, suggesting that GW forcing is overestimated in the WACCM4 and results in a large damping of DW1.


I. Introduction

[2] Atmospheric thermal tides are mainly generated by the daily periodic heating due to the absorption of infrared solar radiation and latent heat release in the troposphere, ultraviolet solar radiation in the lower stratosphere and extreme ultraviolet solar radiation in the thermosphere. Migrating thermal tides follow the apparent motion of the Sun, propagating westward with periods that are harmonics of a solar day [Hagan and Roble, 2001]. Atmospheric tides propagate upward to the mesosphere and lower thermosphere (MLT), where they can reach large amplitudes and have significant impacts by causing ionospheric variabilities and coupling the whole atmosphere vertically [Forbes et al., 2000; Rishbeth and Mendillo, 2001; Immel et al., 2006; Liu et al., 2010a]; introducing large temperature gradient and/or wind shear and resulting in instabilities [Hecht et al., 1997; Liu and Gardner, 2004; Li et al., 2009; Yue et al., 2010]; affecting the propagation of short-period gravity waves (GWs) [Walterscheid, 1981; Liu and Hagan, 1998; Lu et al., 2009] and modulating GW momentum fluxes [Fritts and Vincent, 1987; Espy et al., 2004] by affecting its breaking and dissipation; affecting the vertical transport of the chemical composition and thus the total solar heating in the upper part of the middle atmosphere (70–105 km) [Smith et al., 2003], etc.

[3] Classical tidal theory which involves a set of linearized primitive equations on the global-scale tidal oscillations was proposed by Chapman and Lindzen [1970] to analytically study atmospheric tides in an idealized atmosphere; that is, the background temperature only depends on altitude, and the background atmosphere is motionless and without dissipation. These assumptions lead to a separation of Laplace’s tidal equation, which determines the latitudinal structure of tides (Hough mode), from the vertical structure equation. The migrating diurnal tide (DW1) is the most dominant at low latitudes and the (1, 1) Hough mode is the gravest mode of DW1 in the MLT region [Forbes, 1995]. At equinox, the prediction of the (1, 1) mode by classical tidal theory is in good agreement with the global-scale satellite measurement, reaching a maximum temperature amplitude near the equator and horizontal wind near 20° latitude [Wu et al., 2008; Mukhtarov et al., 2009]. At solstice, contributions from higher-order modes increase, resulting in an asymmetric tidal structure [Hays et al., 1994; McLandress et al., 1996a].

[4] Mean winds are important for modulating the amplitude and phase structure of DW1. Forbes and Vincent [1989] found that mean winds and dissipation are responsible for the amplitude and phase asymmetries about the equator and a broadening of DW1 oscillations to higher latitudes.
At solstices, mean winds introduce significant hemispheric asymmetries in the structure of DW1 amplitude and phase via a mode coupling into the antisymmetric (1, 2) and (1, −1) modes [Forbes and Hagan, 1988]. Ortland [2005] found that the asymmetry of the tidal structure arises as the result of strong vertical shear in the mean winds from the reversal of the sign of the summer and winter jets in the upper mesosphere. Mean winds are also responsible for the seasonal variation of both amplitude and phase of DW1, and the latitudinal shear of the zonal mean easterlies in the summer mesosphere plays a key role in the phase modulation [McLandress, 2002b].

[5] The effects of GW forcing on the tide have been recognized for decades on the basis of model simulations [Miyahara and Forbes, 1991; Forbes et al., 1991; Mayr et al., 1998; McLandress, 1998; Watanabe and Miyahara, 2009], although the conclusions about GW effects are still controversial. The GW momentum deposition has been observed to modulate the tidal amplitude and phase [Fritts and Vincent, 1987] and simulated to shorten the tidal vertical wavelength by several kilometers [Ortland and Alexander, 2006]. The subgrid GW forcing effects are parameterized in general circulation models (GCMs) due to insufficient observations and difficulties in quantifying them. The source spectra for the GW parameterizations are not physically constrained but normally tuned to obtain the realistic mean winds and background temperature in the middle atmosphere. Since DW1 is a salient and persistent wave in the MLT region, besides adjustment of mean winds and temperature, a reasonable simulation of DW1 is crucial for validating and optimizing GCMs.

[6] The examination of DW1 momentum budget in different GCMs is of great value because it can help to diagnose the important forces affecting the DW1 characteristics, which is difficult to complete from observations, and to identify the possible reasons causing discrepancies from observations. The momentum budget of DW1 has been systematically studied by McLandress [2002a] on the basis of the Canadian Middle Atmosphere Model (CMAM). Like CMAM, the Whole Atmosphere Community Climate Model, version 4 (WACCM4), also extends from the ground to the lower thermosphere. The diurnal tide in the WACCM4 compares favorably with the observations [Lu et al., 2011], and it is worthwhile assessing the momentum budget of DW1, in particular with the new physics-based schemes for GW sources. In this paper, we follow the methodology of McLandress [2002a] and examine the discrepancies from classical tidal theory caused by mean winds and GW forcing and their effects on the tide. We compare the WACCM results with the CMAM in order to substantiate the same conclusions from the models while appreciating the differences. The comparisons with the observations [Lieberman et al., 2010] are also discussed. This study is focused on March, when the DW1 amplitude reaches the maximum value. The seasonal variations of the mean wind and GW forcing effects on DW1 will be investigated in a follow-up work.

2. Model and Method

[7] The WACCM is a GCM extending from the Earth’s surface through the lower thermosphere, incorporating most of the important physical and chemical processes [Garcia et al., 2007]. It is a superset of the NCAR Community Atmosphere Model (CAM) and one of the atmospheric components of the Community Earth System Model (CESM). It has 66 vertical levels from the ground to ~145 km. The vertical resolution is about 1.1 km in the troposphere, 1.1–1.4 km in the lower stratosphere, 1.75 km at the stratopause and 3.5 km above 65 km. The horizontal resolution of WACCM4 used for this study is $2.5^\circ \times 1.9^\circ$ (longitude $\times$ latitude). Atmospheric tides are generated self-consistently in WACCM and subject to interactions with resolved PWs and parameterized GWs.

[s] The GW drag and vertical diffusion parameterizations were modified in WACCM3.5 and are also used in WACCM4. Frontal systems and convection are used to determine the wave sources for nonorographic GWs. Instead of being an arbitrarily specified GW source spectrum, the new wave source schemes are physically parameterized [Richter et al., 2010]. For convection, the parameterization scheme developed by Beres et al. [2005] is used. In this scheme, convective heating depth is an important factor for determining the dominant GW horizontal phase speed and the basic shape of the momentum flux spectrum in phase speed. Deep heating ($\geq 10$ km) tends to generate GWs with long vertical wavelengths and large horizontal phase speeds while shallow heating ($\leq 5$ km) generates GWs with short vertical wavelengths and small horizontal phase speeds [Beres et al., 2005; Richter et al., 2010]. For the frontal GW source parameterization, GWs are launched from a source level of 600 hPa only if a frontogenesis threshold is reached; this typically occurs in a region characterized by strong wind shear and temperature gradient. The intermittency in the wave source partly depends on the status of the simulated atmosphere, rather than an artificial intermittency parameter [Richter et al., 2010]. The GW source by orography is the same as in WACCM3 [Garcia et al., 2007]. For both orographic and nonorographic GWs, the parameterization is based on Lindzen’s scheme [Lindzen, 1981]. By using the source-oriented GW parameterization in the WACCM4, an attempt is to make the wave generation more realistic and self-consistent.

[8] We extract global horizontal winds and GW forcing from the model every 3 h. In order to obtain the amplitude and phase of DW1, a 2-D Fourier transform in longitude and time is applied to the wind components to get the spectrum density in the domain of wave frequency and zonal wave number. On the basis of the spectrum density corresponding to DW1, an inverse Fourier transform is used to derive its amplitude and phase. We also calculate the DW1 component of GW forcing and advection terms by applying the same Fourier transform method. More details about calculating the DW1 component of each forcing are discussed in section 3.2.

3. Model Results

3.1. Climatology of Mean Winds, DW1 Amplitudes and GW Forcing

[10] Figure 1 shows the seasonally averaged zonal mean zonal winds in March, April and May obtained from WACCM4. Log pressure altitude is used for the vertical coordinate, and it is calculated as $Z = H \ln(P_r/P)$, with the surface pressure $P_r = 1000$ hPa and the scale height $H = 7$ km. Compared with the zonal winds based on the Upper Atmosphere Research Satellite (UARS) Reference Atmosphere Project (URAP) [Swinbank and Ortland, 2003; Garcia et al.,...
2007; Richter et al., 2008, 2010], WACCM4 captures the salient features of the observed zonal mean zonal wind climatology. For instance, it generates the tropospheric westerly jets in both hemispheres and the stronger westerly stratospheric zonal mean winds in the Southern Hemisphere (SH) than the Northern Hemisphere (NH) at middle and high latitudes. At the equator, the weak easterly zonal winds from 20 to 50 km are also consistent with URAP observations. For the URAP zonal wind climatology, the reader is referred to Figure 1b of Richter et al. [2008] for details. Overall, the simulation of the mean winds is in good agreement with observations and provides a reasonable background for DW1.

Differences between URAP zonal mean winds and the WACCM4 simulations still exist. The simulated stratospheric westerly jets in the SH tend to be stronger by ~20 m s⁻¹ than observations. In the mesosphere over the equatorial region, the zonal mean winds in the MLT are relatively weaker in the model than the observations, which may be associated with the unresolved mesospheric semi-annual oscillation (MSAO) in the WACCM. Richter and Garcia [2006] showed that near the equinoxes, all forcing terms from both GWs and the resolved waves are weak, suggesting that the easterly phase of the MSAO is driven by a process not well represented in the WACCM. A wind reversal is observed in the NH from the weak westerly in the stratosphere to the easterly around 80 km, which is not captured by the model. Note that although the differences may partly originate from the year-to-year variabilities because a 1 year simulation is used for the WACCM4, it is believed to not play a major role for the discrepancies.

The amplitude of DW1 has a strong semiannual oscillation with stronger amplitudes at equinoxes and weaker at solstices, and the strongest amplitude of DW1 occurs at the vernal equinox [Hays et al., 1994; McLandress et al., 1996b; Wu et al., 2008; Xu et al., 2009], which are captured by the WACCM4 [Lu et al., 2011]. Figure 2 shows the mean amplitudes of DW1 in the zonal wind, meridional wind and temperature in March retrieved from WACCM4. The zonal and meridional winds reach maximum tidal amplitudes at ~20° in both hemispheres. The temperature tide is the strongest at the equator and has a secondary peak near ~30°. The latitudinal structure of DW1 largely resembles that of the (1, 1) Hough mode, except that the zonal wind amplitude peaks 5° equatorward of the prediction by classical tidal theory, especially in the SH.

The tidal structure is not strictly symmetric about the equator, indicating a superposition of antisymmetric Hough modes, which can be generated by asymmetric tidal heating and/or mean winds via mode coupling. At low and middle latitudes, DW1 is strong between 70 and 110 km and greatly dissipates above 110 km due to the increased molecular diffusion. In this region, DW1 is weaker in the NH than in the SH for the wind component and the opposite for the secondary peak of temperature. As will be shown in sections 3.2 and 3.3, this is partly caused by the asymmetric forcing and dissipation in different hemispheres. Above 110 km, an enhancement of DW1 persists in the polar lower
thermosphere region which likely originates from the auroral heating. It is also asymmetric with a stronger amplitude near the Arctic region than near the Antarctic region, similar to the WACCM3 simulation [Chang et al., 2008].

The general structure of DW1 in Figure 2 is in good agreement with satellite observations while the magnitudes are underestimated in the model, as also found by Liu et al. [2010b]. In WACCM4, the monthly mean amplitude of the tidal temperature has two peaks located at 90 and 110 km with magnitudes of ~15 K. The Thermosphere Ionosphere Mesosphere Energetics and Dynamics/Sounding of the Atmosphere using Broadband Emission Radiometry (TIMED/SABER) shows that the 6 year average amplitude for temperature is 18 K in February-March and reaches double peaks at 85 and 95 km [Mukhtarov et al., 2009], which is slightly larger than the maximum temperature amplitude in WACCM4. From TIMED Doppler Interferometer (TIDI) observations, the maximum amplitudes for the zonal and meridional winds are of the order of 50 and 60 m s$^{-1}$ within a 60 day window centered in March [Wu et al., 2008] whereas in WACCM4 the corresponding amplitudes are 30 and 50 ms$^{-1}$, respectively. However, the amplitudes of the diurnal tide from WACCM4 are more comparable with the observations of the horizontal winds measured by the ground-based meteor radar in Maui, Hawaii (21°N) [Lu et al., 2011]. More comparative studies based on both ground-based and spaceborne observations are needed in order to evaluate the degree to which the DW1 amplitude is underestimated in WACCM4. Since it is an important tidal source, an underestimation of the diurnal cycle in the parameterized convection is one of the possible causes for the underestimation of DW1 (A. Smith, private communication, 2011). Another possible cause is GW forcing that is too strong and introduces a large damping on DW1. The impact of damping by GW is discussed further in section 3.3. In addition, the vertical resolution can affect the amplitude and phase of DW1 in the MLT region; an insufficient vertical resolution is in part responsible for a systematically smaller tidal amplitude [Liu et al., 2010b].

Figure 3 shows the projection of GW forcing on DW1 in terms of the time tendency projecting onto DW1 in March. The unit is m s$^{-1}$ d$^{-1}$. [15] Figure 3 shows the projection of GW forcing on DW1 in terms of the time tendency in the zonal momentum, a new finding in the WACCM that summarizes the distribution of GW forcing from the three different wave sources. For the convective and frontal GW forcing, GWs start to play an important role above 70 km as their amplitudes grow large enough, depositing considerable momentum into the mean flow and affecting the tide, whereas for the orographic GW forcing, it can be present with a same magnitude at much lower altitude (20 km) and decreases above 100 km. The contribution from convection is confined to the equatorial region where the strongest convective activities occur most frequently. Frontal system and orography excite GWs mainly at middle and high latitudes; the magnitude of the latter GW forcing is much smaller than the former one. The
maximum amplitude of the time tendency for DW1 in the zonal wind caused by frontogenesis GWs is \(~80\text{ m s}^{-1}\text{ d}^{-1}\), by convection GWs is \(~40\text{ m s}^{-1}\text{ d}^{-1}\) and by orography GWs is \(~1\text{ m s}^{-1}\text{ d}^{-1}\). These magnitudes are comparable to the GW drag effect on the mean flow \([\text{Richter et al., 2010}]\). A large projection of GW forcing onto the DW1 component indicates that the modulation of GW forcing by DW1 \([\text{Walterscheid, 1981; Fritts and Vincent, 1987}]\) is captured in the WACCM4. Compared with the zonal momentum, the latitudinal distribution of the amplitude of GW forcing in the meridional momentum is similar, but its magnitude is 2 times smaller (not shown here).

3.2. Comparative Magnitudes of Classical Terms, GW Forcing and Advection

[16] On the basis of Figure 2, we confine our analysis to the latitude region between \(\pm 50^\circ\) and \(\pm 50^\circ\), where DW1 attains a large amplitude. Equations (1) and (2) are the momentum equations \([\text{e.g., Andrews et al., 1987}]\), where \(u, v\) and \(\Phi\) are the zonal wind, meridional wind and geopotential, respectively. The variable \(f\) is the Coriolis parameter, and \(a\) is the Earth radius; \(\phi\) and \(\lambda\) are latitude and longitude, respectively. In classical tidal theory, only the DW1 components of the first two terms, the Coriolis force (CF) and pressure gradient force (PGF) determine the time tendencies of the horizontal tidal winds, which are referred to as the classical terms \([\text{Chapman and Lindzen, 1970; McLandress, 2002a; Lieberman et al., 2010; Chang et al., 2011}]\). The other terms are nonclassical terms, including advection (due to both mean and perturbation winds), curvature, GW forcing and dissipation:

\[
\frac{\partial u}{\partial t} = fu - \frac{1}{a \cos \phi} \frac{\partial \Phi}{\partial \lambda} - \frac{\Phi}{a} \cdot \nabla u + \frac{uv}{a} \tan \phi + F_{GW,x} + X
\]

\[
\frac{\partial v}{\partial t} = -fu - \frac{1}{a \cos \phi} \frac{\partial \Phi}{\partial \lambda} - \frac{\Phi}{a} \cdot \nabla v - \frac{\nabla^2 u}{a} \tan \phi + F_{GW,y} + Y
\]

\(F_{GW,x}, F_{GW,y}\) are GW forcings in the zonal and meridional directions. \(X, Y\) represent the other nonconservative mechanical forcings. In WACCM4, they mainly include ion drag and eddy and molecular diffusion. Below 100 km, GW forcing is more significant than ion drag and the molecular diffusion terms, while above 100 km, molecular diffusion becomes dominant \([\text{McLandress, 2002a}]\).

[17] In order to evaluate the accuracy of classical tidal theory and to assess the contribution from the nonclassical terms, the amplitude of \(\partial u/\partial t\) based on the tidal wind (left-hand side of the momentum equation) and the time tendency caused by CF and PGF are calculated. They are shown in Figure 4 for the zonal wind and Figure 5 for the meridional wind. By writing the tidal wind as \(u' = \tilde{u} e^{i(\omega t - \phi \lambda)}\), the amplitude of the time tendency can be simply written as \(\partial u'/\partial t = \omega|\tilde{u}|\), where \(\tilde{u}\) is the complex amplitude in the zonal wind and \(\omega\) and \(s\) are frequency and zonal wave number of DW1. Similarly, for CF in both the zonal and meridional winds and PGF in the zonal wind, the solutions can be obtained on the basis of the complex amplitudes of horizontal winds \((\tilde{u}, \tilde{v})\) for CF and the complex amplitude of the geopotential \((\tilde{\Phi})\) for PGF. The same analyses are performed by \(\text{McLandress [2002a]}\) in evaluating the momentum budget of DW1 in the CMAM and by \(\text{Chang et al. [2011]}\) in studying the effect of the nonlinear interaction between DW1 and quasi-2 day wave on the short-term variability in DW1. The differences in time tendencies calculated on the basis of the classical terms and full primitive equations are also shown in Figures 4c and 5c, which represent the contributions from the nonclassical terms.

[18] Comparison of Figures 4a and 4b illustrates that the calculation based on the classical terms provides a reasonable first-order prediction of the structure and magnitude of DW1 in the zonal wind component. The discrepancy from the calculation based on the classical terms is about 2–3 times smaller than the first-order prediction. In the zonal wind, the amplitude of the time tendency in the presence of only CF and PGF for DW1 is up to 200 m s\(^{-1}\) d\(^{-1}\), and the difference caused by the nonclassical terms is up to 90 m s\(^{-1}\) d\(^{-1}\). A significant difference between them is found at latitudes 20°N–40°N, where the amplitude of the time tendency becomes smaller after including the nonclassical
terms. As will be shown in section 3.3, the decrease of the DW1 amplitude is caused by GW drag.

[19] Figure 5 shows that the amplitude of the time tendency in the meridional tidal wave is larger than the zonal wind component by a factor of ~1.4, which is consistent with a larger meridional amplitude of DW1 (Figure 2). In the meridional wind, the amplitude of the time tendency in the presence of only CF and PGF for DW1 is up to 270 m s⁻¹ d⁻¹, and the difference caused by the nonclassical terms is up to 90 m s⁻¹ d⁻¹, comparable to that in the zonal wind. Compared with the observations from the TIMED satellite that the momentum residuals range between 50 and 150 m s⁻¹ d⁻¹ for the zonal component and 100 and 250 m s⁻¹ d⁻¹ for the meridional component [Lieberman et al., 2010], the discrepancy caused by nonclassical terms in the meridional wind is significantly underestimated in the WACCM4.

[20] Opposite to the zonal wind component for which GW drag damps DW1 in the NH, the time tendency is due to CF and PGF is smaller than the total tendency in the meridional wind, implying that the nonclassical terms tend to increase the DW1 amplitude. According to Figure 5, another notable difference is the variation of the vertical structure after the nonclassical forcings are introduced. On the basis of the calculation including CF and PGF, the double-amplitude peaks are located near 90 and 105 km, which are replaced by a single peak around 100 km when the nonclassical terms are considered.

[21] As described in equations (1) and (2), the nonclassical terms that may contribute to the discrepancies from the prediction of classical tidal theory are advection, curvature, GW forcing and dissipation. In order to study the relative significance of these terms and to identify the most dominant terms in the momentum budget of DW1, we calculated each term and projected it onto the DW1 component. GW forcing is obtained from the model output directly. The advection terms are calculated as

\[ F_{\text{advect},x} = -\mathbf{\nabla} \cdot \mathbf{u} = - \left( \frac{u}{\cos \phi} \frac{\partial u}{\partial \lambda} + \frac{v}{\cos \phi} \frac{\partial u}{\partial \phi} + \frac{w}{\cos \phi} \frac{\partial u}{\partial z} \right) \]  

\[ F_{\text{advect},y} = -\mathbf{\nabla} \cdot \mathbf{v} = - \left( \frac{u}{\cos \phi} \frac{\partial v}{\partial \lambda} + \frac{v}{\cos \phi} \frac{\partial v}{\partial \phi} + \frac{w}{\cos \phi} \frac{\partial v}{\partial z} \right) \]  

[22] After writing the mean wind as \( u = \bar{u} + u', \ v = \bar{v} + v' \), we divide the advection terms into linear and nonlinear advection. Linear advection involves the advection of mean winds by wind perturbations as well as the advection of perturbations by mean winds. Nonlinear advection involves the advection of wind perturbations themselves. The linear advection terms are as follows:

\[ F_{\text{Linadvect},x} = - \left( \frac{\bar{u}}{\cos \phi} \frac{\partial \bar{u}'}{\partial \lambda} + \frac{\bar{v}}{\cos \phi} \frac{\partial \bar{u}'}{\partial \phi} + \frac{\bar{w}}{\cos \phi} \frac{\partial \bar{u}'}{\partial z} \right) \]  

\[ F_{\text{Linadvect},y} = - \left( \frac{\bar{u}}{\cos \phi} \frac{\partial \bar{v}'}{\partial \lambda} + \frac{\bar{v}}{\cos \phi} \frac{\partial \bar{v}'}{\partial \phi} + \frac{\bar{w}}{\cos \phi} \frac{\partial \bar{v}'}{\partial z} \right) \]  

where the mean winds \((\bar{u}, \bar{v})\) are computed by averaging the wind fields in time and longitude.

[23] When calculating the time tendency of linear advection, for the terms having derivatives with respect to longitude (e.g., \( \partial u' / \partial \lambda \)), the DW1 component with an explicit form of \( u' = \bar{u} \exp(i(\omega t - s\lambda)) \) is used. The \( \omega = 2\pi/24 \) h is the frequency and \( s = -1 \) is the zonal wave number of DW1. By substituting this form into equations (5) and (6), we have

\[ F_{\text{Linadvect},x} = i s \bar{u} \frac{\partial \bar{u}'}{\partial \phi} = i s \bar{v} \frac{\partial \bar{v}'}{\partial \phi} = \frac{\bar{w}}{\cos \phi} \frac{\partial \bar{u}'}{\partial z} \]  

\[ F_{\text{Linadvect},y} = i s \bar{v} \frac{\partial \bar{v}'}{\partial \phi} = i s \bar{v} \frac{\partial \bar{v}'}{\partial \phi} = \frac{\bar{w}}{\cos \phi} \frac{\partial \bar{v}'}{\partial z} \]

For other terms such as GW forcing, a 2-D Fourier transform with respect to time and longitude is applied at each altitude and latitude to obtain the projection onto the DW1 component. After the linear advection terms for DW1 are derived, the nonlinear advection terms for DW1 are calculated by subtracting these linear advection terms from the DW1 component of total advection. The total advection is calculated on the basis of the advection of the total wind fields from equations (3) and (4). Just as for GW forcing, it is projected onto the DW1 component by applying the 2-D Fourier transform. The 2-D Fourier transform method and the method of using the explicit form of DW1 may result in
some differences in the derived time tendencies since the
finite length data used in the Fourier transform may cause
some spectrum leakage and lead to smaller amplitudes. The
differences are not significant for the large horizontal scales
and will not change the global structure of time tendencies
[McLandress, 2002a].

[24] Figure 6 shows the time tendencies of DW1 from the
total advection (Figure 6a), GW forcing (Figure 6b) and the
sum of these two terms (Figure 6c) in the zonal direction. It
is obvious that the effects of the total advection and GW
forcing are comparable. GW forcing is more important
above 80 km and at latitudes >20°N. Below 80 km, advec-
tion terms are dominant at the equator between 70 and 80 km
and in the region near 20°S and 60 km. The zonal forcing
due to advection and GW forcing is stronger in the NH than
in the SH. It is partly caused by the stronger GW drag gen-
errated by frontal systems in the NH in all seasons (not
shown). The distribution of convection-generated GW forc-
ing is determined by the location of deep convective systems
which changes with season, thus not contributing to the
stronger GW forcing in the NH in March.

[25] Figures 4c and 6c are similar in both the distribution
and magnitude of time tendencies, which indicates that
advection and GW forcing are the two dominant factors
to account for the momentum budget of DW1 besides
the classical terms (i.e., CF and PGF). This is different from
the conclusion of McLandress [2002a], who found that the
direct effect of parameterized GWs is substantially weaker
than wave-wave and wave-mean flow interactions. In the
CMAM [McLandress, 1998, 2002a], the effects of nonoro-
graphic GWs are parameterized on the basis of the Hines'
Doppler spread parameterization (DSP) [Hines, 1997a,
1997b]. Instead, Lindzen’s GW parameterization scheme is
used in the WACCM [Lindzen, 1981], and it incorporates a
physically based GW source parameterization [Richter et al.,
2010]. A different GW parameterization is an important
factor that leads to different GW-tidal interaction results.

[26] The similar structures in Figures 5c and 7c show the
same dominances of advection and GW forcing in the
meridional momentum. However, in contrast to forcing of
the zonal wind tide, GW forcing has a magnitude about 2–
3 times smaller and is less important than the advection

Figure 6. Time tendency amplitudes in the DW1 component of (a) total advection forcing, (b) total GW
forcing and (c) the sum of them in the zonal momentum. The unit is m s⁻¹ d⁻¹.

Figure 7. Same as Figure 6 except for the meridional momentum.
terms. Especially in the equatorial region, the contribution by GW forcing is very weak. Compared to the numerical modeling study by McLandress [2002a], the effect on the meridional component of DW1 of GW forcing is stronger and that of advection is weaker in the WACCM. As discussed by Richter et al. [2010], a low frontogenesis threshold is used in order to produce enough GWs to reverse the stratospheric jets and to cool the mesopause to observed temperatures. It is likely that there is an overestimation of GW generation by fronts, which in turn affects DW1. Since the advection terms involve wave perturbations, the underestimation of the tidal amplitudes is a possible explanation for the weak advection effect.

In order to further elucidate which advection term is more important, we show the effects from linear advection and nonlinear advection separately in Figure 8. For the zonal wind, linear advection is about 2–3 times stronger than nonlinear advection, and it is the opposite for the meridional wind. By comparing Figure 8 with Figures 6a and 7a showing the total advection in the zonal and meridional winds, respectively, it is clear that linear advection is dominant for the zonal wind and nonlinear advection is dominant for the meridional wind, which are in agreement with the findings of McLandress [2002a]. This is because linear advection is essentially a product of mean wind and wave perturbation and the zonal mean zonal wind is much stronger than the zonal mean meridional wind. Nonlinear advection is the product of wave perturbations, so the meridional nonlinear advection is larger due to the stronger tidal amplitude in the meridional wind for DW1.

3.3. Effects of GW Forcing and Advection

In section 3.2, the relative importance of GW forcing, linear and nonlinear advection in the momentum budget of DW1 is investigated in terms of comparative magnitudes in a form of time tendency. However, it does not provide the information of the phase relationship between these forcings and DW1. In order to study whether the amplitude of DW1 is increased or decreased and whether the phase is advanced or delayed, we introduce the coefficient of equivalent Rayleigh friction (ERF) [Forbes et al., 1991; McLandress, 2002a; Chang et al., 2011] defined as

\[ \gamma_{u} = \frac{\bar{F}_u}{\bar{u}} \]

where \( \bar{F}_u \) and \( \bar{u} \) are the complex amplitudes of forcing and DW1, respectively. The real part of the ERF represents the effect on changing tidal amplitude. If Re(\( \gamma_{u} \)) > 0, it damps the tide; if Re(\( \gamma_{u} \)) < 0, it enhances the tide. The imaginary part of the ERF corresponds to the change of the phase. Im(\( \gamma_{u} \)) < 0 corresponds to an advance in phase and a local shortening of the vertical wavelength for the tide with upward energy propagation. Conversely, the phase is
delayed and the vertical wavelength is increased locally when \( \text{Im}(\gamma_a) > 0 \).

Figure 9 shows the real part of the ERF calculated for GW forcing. The regions with tidal amplitudes too small are excluded because the ERF is unreasonably large after dividing by a small complex amplitude of the tide. Above 70 km, \( \text{Re}(\gamma_a) \) is mostly positive, which means that GW drag damps DW1 in most regions. The largest value of \( \text{Re}(\gamma_a) \) is around \( 8 \times 10^{-5} \) s\(^{-1}\) and is found where the GW drag is large and the tidal amplitude is small, i.e., around latitudes 50\(^\circ\)–60\(^\circ\). For the meridional wind, the largest value is \( 3 \times 10^{-5} \) s\(^{-1}\), about ~2.5 times smaller than the zonal wind component. The vertical and latitudinal distribution of the ERF in the WACCM4 is similar to the effective Rayleigh friction in the Global-Scale Wave Model (GSWM) [Hagan et al., 1995, 1999], while the magnitude is of 1 order larger than the maximum effective Rayleigh friction in the GSWM. Contrary to the WACCM4, GW drag from momentum deposition acts to amplify the tide in the CMAM [McLandress, 2002a], and the magnitude is also 1 order smaller than that in the WACCM. It is noticed that the DW1 maximum amplitudes simulated in the CMAM reach 45 and 70 m s\(^{-1}\) for the zonal and meridional winds, respectively, which are larger than those in the WACCM4. It suggests that the GW damping...
effect may be overestimated in the WACCM and result in the underestimation of the DW1 amplitude.

[30] A significant difference for the imaginary part of the ERF from GW forcing is the changing sign as it goes from low to high latitudes. As shown by Figure 10, in the tropical and subtropical regions, Im(\(g_u\)) and Im(\(g_v\)) are negative and become positive toward higher latitudes, meaning that GW drag tends to advance the tidal phase at low latitudes where DW1 is strong and delay it at middle and high latitudes. The latitude dependence of the GW effect on the tidal phase has not been reported before to our knowledge. Similar to the real part of the ERF, the largest values are found at latitudes where GW forcing is strong and the tide is weak. In these regions, GWs are basically excited by the frontal systems, indicating that weather systems in the lower atmosphere have a notable impact on DW1 in the MLT region through GW excitation and momentum deposition.

[31] Figures 11 and 12 show both the real and imaginary parts of the ERF by linear advection and nonlinear advection, respectively, in the region between \(-50^\circ\) and \(+50^\circ\). Figure 13 displays the total effect by adding the ERF of GW drag, linear and nonlinear advection together. Unlike GW drag with more uniformly positive values for Re(\(g_u\)) and negative values for Im(\(g_u\)) in the region of interest, linear advection and nonlinear advection have finer vertical and latitudinal structures. For instance (Figure 11), Re(\(g_u\)) of linear advection resembles a wave structure near the equator, which is positive near 70 km and negative near 90 km and becomes positive again near 100 km. On the basis of equations (7) and (9), a scale analysis for the linear advection reveals that the largest term contributing to Re(\(g_u\)) is caused by the vertical advection of the zonal mean zonal wind. It is expected that Re(\(g_u\)) has the largest value at the equator where the tidal amplitude of the vertical wind and the vertical shear of the zonal mean zonal wind are both strong.

[32] Im(\(g_u\)) of linear advection is larger than that of either GW drag or nonlinear advection; therefore, the linear advection term makes a dominant contribution to the phase

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**Figure 11.** Same as Figures 9 and 10 but for linear advection and confined within \(\pm 50^\circ\) in latitude.
change of DW1 in the zonal wind. The negative area corresponding to an advance of tidal phase is centered in the equatorial region below 90 km. It becomes narrower as the altitude increases from 50 to 90 km. At a latitude of 20° N, the positive \( \text{Im}(\gamma_u) \) up to \( 5 \times 10^{-5} \text{ s}^{-1} \) is 1 order of magnitude larger than those originating from GW drag and nonlinear advection, indicating a strong delay of the tidal phase caused by linear advection. A scale analysis for linear advection (equation (7)) indicates that the meridional advection of the zonal mean zonal wind (\( \partial (\bar{v} / a) \cdot \partial \bar{u} / \partial \phi \)) is the largest factor determining \( \text{Im}(\gamma_u) \). Since the latitudinal shear of the zonal mean zonal wind is stronger than that of the mean meridional wind, \( \text{Im}(\gamma_v) \) is much larger than \( \text{Im}(\gamma_u) \). Owing to the importance of linear advection, it is expected that the phase change of DW1 is very sensitive to the latitudinal structure of the zonal mean zonal wind. This is supported in the numerical modeling by McLandress [2002b] that attributes the strong modulation of the DW1 phase by the mean winds to latitudinal shears in the zonal mean easterlies in the summer mesosphere.

By comparing the real part of the ERF in the zonal momentum in Figures 9–13, it can be seen that linear advection is important at the equator and GW drag is important at latitudes higher than 20°, while nonlinear advection is less significant. For the phase modulation, however, linear advection makes a major contribution while GW drag and nonlinear advection also play a role in decreasing or even offsetting the phase delay caused by linear advection, such as the region between latitude 0° and 10° S and altitudes from 70 to 90 km and in the latitude range 20° S–50° S and altitudes 100–110 km.

In contrast to the case for the zonal momentum, nonlinear advection is important for the meridional momentum. A comparison of ERFs in the meridional momentum in Figures 9–12 shows that nonlinear advection, in addition to the classical terms, largely determines the momentum budget of DW1 in the meridional wind. This is due to the fact that the zonal mean meridional wind is weak which means that linear

![Figure 12](image_url)
advection is also weak. In addition, the GW drag effect on the meridional momentum of DW1 is not as strong as that on the zonal component. Since nonlinear advection involves the products of wave perturbations, after being divided by the tidal wave as in the definition of the ERF, $\text{Re}(\gamma)$ and $\text{Im}(\gamma)$ have wave structures with the same vertical scales as the tide. A scale analysis shows that for nonlinear advection, the meridional and vertical advections of the meridional wind perturbations \( \left( \frac{v'}{C_0} \cdot \frac{\partial v'}{\partial f} - \frac{w'}{C_0} \cdot \frac{\partial v'}{\partial z} \right) \) are the most important terms and as latitude increases to 40°, the zonal advection increases and becomes comparable (not shown).

4. Conclusion and Discussion

[35] The momentum budget of DW1 contributed by GW forcing and advection terms is investigated by using the WACCM4, which is capable of simulating both mean background winds and DW1 reasonably well. Classical tidal theory provides a good first-order prediction of the magnitude and structure of DW1. The discrepancies mainly result from GW drag and advection terms. The GW drag on DW1 from the three wave sources in the WACCM4, i.e., orography, convection and frontal systems, is shown for the first time in this paper.

[36] GW drag is found to be important in the zonal momentum balance of DW1, and it partly causes the asymmetric tidal structure in the zonal wind, which is a new finding different from the conclusions based on the CMAM and NCAR Thermosphere Ionosphere Mesosphere Electrodynamics General Circulation Model (TIME GCM) results. In the CMAM, McLandress [2002a] found that the GW momentum deposition and eddy diffusion are much smaller than the advection terms. The GW forcing in the WACCM always damps the tide, while it amplifies the tide in the CMAM [McLandress, 2002a], indicating that different GW parameterization schemes can lead to different GW effects on the tide. Chang et al. [2011] found that the GW momentum forcing is 1 order smaller than the linear and nonlinear

Figure 13. Same as Figure 11 but for the total ERF (GW forcing plus linear and nonlinear advection).
advection in the TIME GCM, which is similar to the CMAM result but different from the WACCM.

[37] The projection of GW drag onto the DW1 component reaches a substantial magnitude (Figure 6) and a maximum real ERF reaches $8 \times 10^{-5} \text{s}^{-1}$ (Figure 9) in the zonal momentum, implying that the tide also has a significant modulation on GWS. The modulation previously reported by Bergman and Salby [1994] shows that convectively generated GWS have a substantial diurnal component. At the equator, linear advection mainly originates from the advection of the zonal mean zonal wind by the vertical tidal wind, which is also important for the momentum budget of DW1. Unlike GW drag, which always damps the tide in the zonal wind direction, the effect of linear advection depends on altitude because the vertical shear of the zonal mean zonal wind changes its sign with altitude. The contribution from nonlinear advection in the momentum budget of the zonal wind, however, is insignificant compared to GW drag and linear advection.

[38] In the area where DW1 is strong, GW drag tends to advance the tidal phase and thus to shorten its vertical wavelength locally, which has also been reported by previous studies [Orlant and Alexander, 2006; Watanabe and Miyahara, 2009]. Nevertheless, GW drag is not the most important factor in changing the tidal phase. The major contribution for the phase change of DW1 is from linear advection, and it is largely determined by the latitudinal shear of the zonal mean zonal wind, which is supported by studies of McLandress [2002a, 2002b]. Note that linear advection also affects the tidal amplitude according to the study by McLandress [2002b]. Compared to the zonal wind, GW drag and linear advection are less important for the meridional wind. Instead, the changes of amplitude and phase are attributed to nonlinear advection.

[39] Lieberman et al. [2010] studied the momentum budget of DW1 and calculated the “wave drag” upon the tide inferred as a residual of the classical and linear advective terms in the MLT region, based on TIMED/SABER and TIMED/TIDI observations. Some similarities can be found between the WACCM and observations while differences are also present. In their study, the classical terms and linear advection of the mean winds by tidal perturbations are inferred from measurements and referred to as the momentum budget terms. The residual terms are the sums of linear advection of the tide by the mean wind, frictional terms and nonlinear advection. Similar to the WACCM, for the zonal momentum budget, the meridional advection of the zonally averaged momentum $-(v/a) \cdot \hat{z}$ is found to be the most important nonclassical term, approaching 70–80 m s$^{-1}$ d$^{-1}$. In the WACCM, this term also makes a significant contribution to linear advection, and it reaches 60 m s$^{-1}$ d$^{-1}$, which is slightly smaller than observations, probably due to the underestimation of the tidal amplitude in the meridional wind. The difference is that from the observations, the net residual zonal force is generally in quadrature with the zonal wind, which means that it advances the tidal phase but has small impact on the tidal amplitude, while on the basis of the WACCM, the linear advection delays the tidal phase and GW drag can significantly damp the tidal amplitude in the region where DW1 is strong. The finding of Lieberman et al. [2010] that the meridional momentum residual is in antiphase with the meridional wind is also different from the results in the WACCM.

[40] The WACCM4 is capable of resolving global structure and seasonal variation of DW1, while the magnitude is slightly underestimated. The possibility that the large GW drag on DW1 causes a large damping effect needs a further study. As pointed out by Richter et al. [2010], other extratropical GW sources should be included in the model to account for the momentum budget from GWs, instead of applying a relatively low frontogenesis threshold which may result in launching more GWs than the reality. A more realistic GW parameterization is expected in the future. It should be noted that observational studies on the momentum budget of the tide are still rare and need further investigations.

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