Momentum budget of the migrating diurnal tide in the mesosphere and lower thermosphere

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[1] The leading terms of the momentum budget of the migrating diurnal tide are diagnosed from temperatures and geopotential height provided by Thermosphere-Ionosphere-Mesosphere Energetics and Dynamics (TIMED)/Sounding of the Atmosphere using Broadband Emission Radiometry (SABER) and the horizontal wind vector provided by TIMED Doppler Interferometer (TIDI). The “wave drag” upon the tide is inferred as a residual of the classical and linear advective terms in the zonal and meridional momentum equations. Between 85 and 100 km, the migrating diurnal tide is generally far from the classical description of Chapman and Lindzen (1970). The leading nonclassical terms are the meridional advection of zonally averaged momentum and wave drag effects encapsulated in the momentum residuals. The magnitudes of the momentum residuals range between 50 and 150 m s\(^{-1}\) d\(^{-1}\) for the zonal component and 100–250 m s\(^{-1}\) d\(^{-1}\) for the meridional component. The zonal momentum residual is generally in quadrature with the meridional wind above 95 km. Winds and temperatures exhibit maximum amplitudes during the vernal equinox season, but only a minor reduction in wave damping is observed. Our results suggest that Rayleigh damping is inadequate to describe the relationship between the tide and the forces acting upon it.


1. Introduction

[2] Atmospheric tides are among the most prominent motions in the mesosphere and lower thermosphere (MLT), often dominating the meridional wind field at low latitudes [Hays et al., 1994]. Migrating diurnal tides exhibit seasonal and interannual variations in amplitude, that correlate with equatorial zonally averaged zonal mean winds [Burrrage et al., 1995; Wu et al., 2008; Xu et al., 2009a]. This variability is thought to be caused by turbulent diffusion, small-scale wave drag, large-scale nonlinear wave interactions [McLandress, 2002a], and advective effects [Hagan et al., 1999a; McLandress, 2002b]. Although seasonal and year-to-year tidal variations can be reproduced in models with tunable dissipation parameters [Yudin et al., 1997; Hagan et al., 1999b], the underlying physical causes remain unclear. Moreover, no observational confirmation of the magnitude or seasonality of the dissipation upon the tides has ever been carried out.

[3] The Thermosphere-Ionosphere-Mesosphere Energetics and Dynamics (TIMED) satellite has been monitoring the mesosphere and lower thermosphere (MLT) since late 2001. The goal of this study is to apply simultaneous measurements of TIMED winds, temperatures and geopotential height to diagnose the momentum budget of the diurnal tides between 80 and 100 km. We compute the migrating diurnal winds and geopotential, examine the relative contributions of each of the terms in the momentum equations, and determine their seasonal variations. The momentum residual is interpreted as the net forcing upon the tide, which we shall refer to as the “wave drag.” Meridional wave drag generally acts to damp the motion above 14 scale heights. The zonal wave drag both accelerates and decelerates the motion, and generally leads the zonal wind by one-quarter cycle.

2. Data

[4] The TIMED satellite was launched in December 2001 with the mission of studying the influences of the sun and climate change on the MLT and the ionosphere. One of the instruments on board is the Sounding of the Atmosphere using Broadband Emission Radiometry (SABER), a 10-channel infrared (1.27 to 16.9 \(\mu\)m) limb radiometer that measures radiatively active species in both local thermodynamic equilibrium (LTE) and non-LTE environments. SABER LTE retrievals of temperature and geopotential height are obtained from infrared emissions of CO\(_2\) at both
15 and 4.3 \( \mu \text{m} \), from cloud tops to about 120 km. The vertical resolution is about 2 km. This study uses Version 1.07 temperatures retrieved assuming both LTE and non-LTE processes.

[5] The SABER instrument provides information about the spacecraft position, pointing, and the vertical spacing between radiance samples [Rong et al., 2009]. A relative altitude profile is estimated, and then adjusted to match that reported by the National Centers for Environmental Prediction (NCEP) on the 10 hPa surface [Remsberg et al., 2008]. Subsequently, the vertically varying gravity \( (g(z)) \) is integrated with respect to the adjusted altitudes to produce the geopotential \( (\Phi) \). The quantity actually reported in the SABER files is the geopotential height \( Z \), defined as \( \Phi/g_0 \), with the value of \( g_0 \) given by 9.8 m s\(^{-2}\). Remsberg et al. [2008] have presented a thorough evaluation of version 7 temperatures retrieved on pressure surfaces, and concluded that these show good agreement with airglow- and sodium lidar-based temperatures in the MLT. V1.07 retrievals are archived at http://saber.larc.nasa.gov.

[6] TIMED orbits the Earth about 15 times per day, viewing on the ascending and descending portions of the orbit. Ascending and descending node measurements at a given latitude are nearly fixed with respect to local time, drifting backward by about 12 min on subsequent days. Over the course of one TIMED orbit, SABER observes between 52\(^\circ\)S and 83\(^\circ\)N during the northward-looking mode, switching after 60 days to an analogous southward-viewing mode. On an individual day, SABER makes measurements covering 15 longitude bands. The difference between ascending and descending node local times at the equator ranges from approximately 8.7–9.3 h at equatorial latitudes, depending on whether SABER was in a northward viewing or southward viewing yaw period.

[7] Another instrument aboard TIMED, the TIMED Doppler Interferometer (TIDI) performs high spectral resolution limb scans with 2.5 km vertical resolution through the terrestrial airglow layers in the MLT [Killeen et al., 2006]. The horizontal wind vector is derived from the Doppler shifts and line shapes of the spectral features of the airglow emissions. TIDI obtains these scans simultaneously in four orthogonal directions: two at angles 45° forward, but on either side of the satellite’s velocity vector, and two at 45° rearward of the satellite. These four views provide the measurements necessary to construct the horizontally resolved vector winds as a function of altitude along two parallel tracks corresponding to the “warm” (sunward) and the “cold” (anti-sunward) sides of the spacecraft. Between 52°S and 52°N TIDI views to both the warm and the cold sides. Poleward of these latitudes, TIDI views to only one side (warm in one hemisphere, cold in the other) The difference between the ascending and descending local times (denoted \( t_{asc} \) and \( t_{desc} \), respectively) varies from 8 to 12 h between 20°N and 20°S. The warm- and cold-side winds are separated on average by about 3 h in local time.

[8] Wind retrievals are reported every 2.5 km, although the effective vertical resolution is half a scale height. Useful daytime winds are reported between 70 and 110 km, while nighttime winds are retrieved only between 85 and 100 km. Although the nominal horizontal spacing between profiles is approximately 750 km along the orbit track, the effective meridional resolution of vector winds is about 6°. Details about instrument design, performance, wind retrievals, and product hierarchy appear in the work of Skinner et al. [2003], Killeen et al. [2006] and Niciejewski et al. [2006]. We have opted to analyze inverted line-of-sight winds from the “PRF” files, since these data can be easily clustered by viewing angle, and by ascending and descending nodes. This study uses version 10 PRF files distributed by the University of Michigan (http://tidi.engin.umich.edu).

3. Tidal Analyses

[9] The momentum budgets of the diurnal tide are given by the equations of horizontal motion linearized with respect to a three-dimensional basic state: \( (\mathbf{U}, \mathbf{v}, \mathbf{w}) \).

\[
\begin{align*}
\frac{D\mathbf{v}}{Dt} + \left[(a \cos \phi)^{-1} (\mathbf{U} \cos \phi)_{oz} - f\right] \mathbf{v} &= \mathbf{U} + \mathbf{w} \\
+ (a \cos \phi)^{-1} \mathbf{v} &= \mathbf{X}_{res} \\
\end{align*}
\]

(1)

\[
\begin{align*}
\frac{D\mathbf{w}}{Dt} + (f + 2\Omega a^{-1} \tan \phi) \mathbf{w} + a^{-1} \mathbf{v} &= \mathbf{Y}_{res} \\
\end{align*}
\]

(2)

with

\[
\begin{align*}
\mathbf{X}_{res} &= X - a^{-1} \mathbf{v} - \mathbf{w} - a(\cos \phi)^{-1} \mathbf{u} - a^{-1} \mathbf{v} - \mathbf{w} \\
\end{align*}
\]

(3)

\[
\begin{align*}
\mathbf{Y}_{res} &= Y - a^{-1} \mathbf{v} - \mathbf{w} - a^{-1} \mathbf{v} - \mathbf{w} - a(\cos \phi)^{-1} \mathbf{u} - a^{-1} \mathbf{v} - \mathbf{w} \\
\end{align*}
\]

(4)

Aside from the designations of \( X_{res} \) and \( Y_{res} \) (1) and (2) are identical to equations (3.4.2a) and (3.4.2b) by Andrews et al. [1987]. \( u', v', w' \) and \( \Phi' \) represent diurnal zonal, meridional, vertical wind, and geopotential, respectively. \( \mathbf{D} \) is defined as the time tendency plus the zonal advection by \( \mathbf{U} \). \( f \) is the Coriolis parameter, \( a \) is the radius of Earth, and \( \phi \) is latitude. The vertical coordinate \( z \) is scaled log-pressure. \( X_{res} \) and \( Y_{res} \) hereafter referred to as the zonal and meridional momentum residuals, are the sums of frictional terms \( X' \) and \( Y' \), tidal advection by the mean meridional circulation, and nonlinear eddy forcing. The terms on the left-hand side of (1) and (2) can be computed or inferred from measurements; these are hereafter collectively referred to as the “momentum budget” terms. \( X_{res} \) and \( Y_{res} \) are unknown, and estimated as residuals of the momentum budget.

[10] The first step in defining the left-hand sides of (1) and (2) is to obtain the migrating diurnal harmonics of \( u' \), \( v' \), \( w' \), and \( \Phi' \). TIMED samples 24 h over nearly 60 days, and many satellite-based tidal studies are based upon fitting the data to 24 h in local time [Huang et al., 2006; Wu et al., 2008]. Our approach here is slightly different. Tidal analysis begins with the formation of differences between daily ascending and descending node zonally averaged winds and geopotential height on log-pressure surfaces. (SABER data are reported on pressure surfaces, but because TIDI winds are reported on altitude surfaces, these must be referred to pressure surfaces using the corresponding daily SABER altitude and pressure coordinates.) The difference approach is feasible because the ascending-descending node observation times are separated by 9 h for SABER, and nearly 12 h at equatorial latitudes in the case of TIDI. The analyses therefore emphasize the diurnal harmonic, while mitigating
the effects of semidiurnal tides, and daily variability of background conditions [Wu et al., 1998].

If we express the migrating diurnal tide at ascending and descending node observation times as

\[ V_{\text{asc}} = V \cos(\Omega t_{\text{asc}} - \phi) \]

\[ V_{\text{desc}} = V \cos(\Omega t_{\text{desc}} - \phi) \]

then forming the ascending-descending node difference (denoted \( \Delta V \)) yields

\[ \frac{\Delta V}{2 \sin \Omega \Delta t} = -V \sin(\Omega \Delta t - \phi), \]

where \( V \) is the diurnal amplitude and \( \phi \) is the phase.

\[ \Delta t = \frac{1}{2} (t_{\text{asc}} - t_{\text{desc}}), \quad \overline{t} = \frac{1}{2} (t_{\text{asc}} + t_{\text{desc}}). \]

We note that aliasing of (7) by migrating semidiurnal tides can occur at latitudes where the local time difference departs from 12 h. These effects are examined and discussed in Appendix A; they turn out to be relatively minor between 30°S and 30°N.

Daily SABER and TIDI measurements are gridded into 6° × 30° latitude-longitude bins. Ascending and descending node differences are formed only from measurements made on the same day, in order to minimize temperature and geopotential variations arising from non-tidal, day-to-day variability. The tidal analysis interval is 60 days, approximately equal to one yaw period during which nearly 24 h of local time coverage is obtained. We apply 60 day fitting windows that contain 2 full calendar months, slid forward by 30 days. Analyses are carried out using data between 2002 and 2009. Thus, results labeled “March 2004” reflect data gathered between 1 February to 1 April 2004.

Figure 1 shows an example of the retrieval of TIDI diurnal winds, and SABER diurnal temperature and geopotential height for March. The uncertainties associated with the tidal determinations (also listed in Table 1) are 4.3 m for winds, 0.3K for temperature, and 1.8 m for geopotential height. The error estimates were obtained by iterating the analysis method 100 times upon a known tidal field perturbed by random fluctuations scaled by the measurement uncertainties associated with individual profiles (21 m s\(^{-1}\) for wind components, 4K for temperature, and 20 m for geopotential height).

All of the derived tidal fields are qualitatively consistent with the classical theory of tides. Meridional winds are strongly antisymmetric about the equator, and temperatures, zonal winds and geopotential height are largely symmetric about the equator. The temperature, defined as the vertical gradient of geopotential, is in quadrature with the vertical wind component.

Table 1. Uncertainties of Tidal Variables

<table>
<thead>
<tr>
<th>( \phi' ) (m s(^{-1}))</th>
<th>( \psi' ) (m s(^{-1}))</th>
<th>( \tau' ) (K)</th>
<th>( \psi' ) (cm s(^{-1}))</th>
<th>( \Phi' ) (m² s(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.3</td>
<td>4.4</td>
<td>.3</td>
<td>2.5</td>
<td>1.8</td>
</tr>
</tbody>
</table>

*Defined as one standard deviation (“one-σ”). Values are averaged between 12.5 and 15 scale heights and 30°S and 30°N.
the geopotential height. A dominant vertical wavelength of slightly more than 3 scale heights (approximately 22 km) is inferred from the temperature. Retrievals of diurnal winds are very similar to those obtained by Wu et al. [2008] using conventional Fourier methods.

Validation of TIDI zonal mean zonal winds is presented in Appendix B. For this study, zonal mean winds are computed from 60 days of daily averages of ascending and descending node winds. Measurements of both cold-side and warm-side winds (offset by approximately 3 h) are used. The resulting patterns of zonal mean winds are very similar to TIDI winds averaged over 24 h, examples of which are shown in Figure B1c. Between 35°S and 35°N, winds at all times are directed eastward at midlatitudes and westward at equatorial latitudes. The ascending-descending node averaging strategy was adopted in order to minimize contamination of the daily zonal wind by the migrating diurnal tide. However, this approach does not remove semidiurnal effects. Estimates of their effects upon the mean winds and pertinent momentum budget terms are shown in Appendix A.

Diurnal vertical velocity is calculated by integrating the continuity equation:

\[
\left( a \cos \phi \right)^{-1} \left[ u'_x + (v' \cos \phi)_y \right] = \rho_0^{-1} (\rho_0 w').
\]  

\( \rho_0 \) is the background density, computed from the daily sums of SABER zonally averaged ascending and descending node temperatures. \( u' \) and \( v' \) are the diurnal wind harmonics, computed as described above. Thus, the longitudinal derivative of \( u' \) is obtained from the quarter-cycle shift of \( u' \), while the meridional derivative of \( v' \) is computed as a centered difference about the analysis data grid points. At the top boundary (generally situated at 15 scale heights) \( w' \) is assumed to be a monochromatic wave that transmits energy upward ("radiation" condition). To implement this condition, the vertical wavelength and amplitude structure of \( w' \) must be known. The vertical wavelength is given by the phase gradient of the horizontal divergence (the left-hand side of equation (4)). The amplitude of \( w' \) is assumed to vary linearly at the top boundary.

4. Momentum Budget

Figure 2 presents the terms on the left side of equation (1) referenced to 00LT. Uncertainty estimates for zonal and meridional momentum budget terms are tabulated in Tables 2 and 3. These are computed as standard deviations of momentum budget terms from the 100 sets of tidal variables used for the calculation of the one-\( \sigma \) tidal uncertainties, described in section 3. "Classical" terms are shown in Figures 2a–2c. Away from the equator, the tendency \( (\partial u' / \partial t) \) and Coriolis \( (fv) \) terms dominate, with the same order of magnitude and opposite signs. Near the equator, where the
Uncertainties of Momentum Budget Terms during equinox.

and exp WU et al. with D20105 and Re − Y over much of the height (m s K as ð ð = w exp ′ due to nonlinear, small-scale advective effects. Uncertainty estimates for momentum residuals are given in Table 3. Xx res generally ranges between 50 and 125 m s −1, with an uncertainty of about 30 m s −1 d −1. At 00 LT there is net eastward acceleration below 14.5 scale heights, and westward acceleration above that level.

[18] The classical terms of the meridional momentum budget referenced to 00LT are shown in Figures 3a–3c, for March. The dominant terms are the tendency and the pressure gradient force. Yx res is large, with peak values exceeding 240 m s −1 d −1. Comparison with Figure 1 shows that this residual is in opposite phase to v ′ over much of the height range. However, some caution must be exercised in the interpretation of Yx res. Appendix A shows that this quantity is vulnerable to contamination by the semidiurnal tide. Figure A1 shows that semidiurnal aliasing of Yx res at 00 LT weakly echoes Yx res itself, therefore contributing to the anti-phase relationship with v ′.

[19] Composite monthly variations of tidal variables and momentum terms were examined. Figure 4 displays amplitudes at the equator and at 18°N, where the temporal variations are most prominent. Diurnal temperatures and meridional winds exhibit semiannual variations that are familiar from tidal climatologies derived from UARS data: strongest amplitudes at equinox, and weakest values at solstices [Burrage et al., 1995; Wu et al., 1998]. However, some differences with UARS do exist. Above 14 scale heights, a secondary maximum in the meridional wind amplitude is observed in July. HUANG et al. [2006] have noted that amplitudes of TIDI migrating diurnal winds at 95 km are generally weaker than HRDI winds, possibly owing to the data sets spanning different decades. The seasonal cycle of the zonal wind amplitude differs somewhat from the meridional wind, with maxima in March and August. The seasonality of advection by the meridional wind is similar to the meridional wind itself: maximum amplitude in March, secondary maximum in September, and amplitude minima in January and June. Vertical wind and advection variations are similar to temperature, although the “first” equinox amplitude maxima in w′ and w′eU/ez occur in May, as opposed to March. Both zonal and meridional momentum residuals indicate maximum amplitudes in March and August, with a secondary meridional momentum residual in December.

5. Dissipation

[20] Migrating diurnal amplitudes generally peak near 95 km [HAYS et al., 1994; Wu et al., 2008]. Amplitude growth is checked by molecular viscosity, ion drag, turbulent diffusion due to saturated eddies, and momentum transport by large-scale nonlinear wave interaction [Lindzen, 1981; McLandress, 2002a]. Early modeling experiments concluded that stresses induced by saturated gravity waves had the effect of suppressing tidal amplitudes in the MLT [Miyahara and Forbes, 1991]. Consequently, gravity-wave drag upon the tide is often cast using as equivalent Rayleigh friction, or X′ = −Kg′v′ with Kg positive and real [Khattatov et al., 1997; HAGAN et al., 1995, 1999a; Xu et al., 2009b].

[21] Because we have independent measurements of tidal u′, v′, Φ and inferred w′, Xx res and Yy res, we are able to examine the relationship between zonal and meridional winds and momentum residuals, and see if there is an empirical basis for a Rayleigh formalism. Using the following complex representations

\[ u′ = Re(\dot{u} \exp(\delta t)) \]
\[ v′ = Re(\dot{v} \exp(\delta t)) \]
\[ X′_{res} = Re(\dot{x} \exp(\delta t)) \]
\[ Y′_{res} = Re(\dot{y} \exp(\delta t)) \]

we compute Kx = −\dot{x}/\dot{u} and Ky = −\dot{y}/\dot{v}. In the case of pure damping, Kx = −|\dot{u}|/|\dot{u}| and Ky = −|\dot{v}|/|\dot{v}|. Real (imaginary) components of \dot{u} and \dot{v} would be negatively correlated with the real (imaginary) components of \dot{x} and \dot{y}, respectively. Negative correlations are observed in scatterplots of Re(\dot{y}) and Im(\dot{y}) versus Re(\dot{y}) and Im(\dot{y}) between 14.0 and 15.0 scale heights. These relationships are illustrated in Figure 5 (top). However, no correlations were observed between the corresponding real and imaginary components of \dot{x} versus \dot{u}.

The absence of any linear relationship between them suggests that Xx res cannot be adequately modeled as pure Rayleigh damping.

Table 3. Uncertainties of Momentum Residual and K Coefficients

<table>
<thead>
<tr>
<th>Xx res (m s⁻¹ d⁻¹)</th>
<th>Yy res (m s⁻¹ d⁻¹)</th>
<th>Kx (d⁻¹)</th>
<th>Ky (d⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>28.</td>
<td>31.</td>
<td>0.8</td>
<td>1.9</td>
</tr>
</tbody>
</table>

*Defined as one standard deviation. Values are averaged between 12.5 and 15 scale heights and 30°S and 30°N.

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spread parameterization [Hines, 1997] used to represent the effects of gravity waves in the Canadian middle atmosphere model (CMAM) also resulted in gravity-wave drag that was in quadrature with the tidal winds [McLandress, 2002a].

[23] In light of the quadrature relationship between parameterized gravity-wave breaking and tidal zonal winds, we examined the phase relationships between $u'$ and $X'_{\text{res}}$, and $v'$ and $Y'_{\text{res}}$. Figure 5 (bottom) plots the hour of maximum of $X'_{\text{res}}$, or $t_X$, as a function of the hour of maximum of $u'$ ($t_U$). There is a strong linear relationship between the two parameters, with a least squares fit to the data given by $t_X = 0.8(t_U - 5.7)$. The slope indicates a strong positive phase correlation, while the y-intercept value indicates that the zonal wave drag $X'_{\text{res}}$ generally leads the tidal zonal wind $U'$ by 5.7 h, or close to 1/4 of a cycle. The data empirically support an interpretation of zonal tidal drag induced in part by diurnally modulated gravity waves.

[24] The implications of the phase quadrature between $X'_{\text{res}}$ and $u'$ for Rayleigh friction parameterization can be appreciated by approximating $t_X$ as $t_U - 6$, and recasting the representations in equations (10) and (12) as

\[ u' = |\bar{U}| \cos(\Omega(t - t_U)) \]  

\[ X'_{\text{res}} = -|\bar{X}| \sin(\Omega(t - t_U)) \]
Noting that (17) can be rewritten as \( \dot{x} = i u |\vec{X}|/|\vec{U}| \), which is purely imaginary, and negative. Thus, gravity wave-induced drag upon the zonal wind tide is manifested as an imaginary \( K_x \).

[25] Our calculations show that \( K_x \) and \( K_y \) generally have both real and imaginary components, with the imaginary component of \( K_y \) virtually always negative. Physically, this means that the wave drag affects both the amplitudes and the phases of \( u' \) and \( v' \), and that the zonal wave drag \( X'_{\text{res}} \) nearly universally leads \( u' \). We have elected to present the complex nature of \( K_x \) (Im(\( K_y \))) in terms of the phase offset between \( X'_{\text{res}}(Y'_{\text{res}}) \) and \( u'(v') \), which is the physical manifestation of Im(\( K_x \)) (Im(\( K_y \))). \( K \) coefficients can only be computed at latitudes and altitudes where all the terms on the left-hand side of (1) and (2) exceed their uncertainty thresholds. As a result, global coverage of \( K_x \) or \( K_y \) tends to be somewhat “spotty,” and not conducive to contour plots in altitude versus time or latitude versus time. We therefore present global \( K \) coefficients as distributions.

[26] Figure 6 shows the distributions of zonal wind, momentum residual, and \( \dot{x} \) amplitudes, together with the \( u' - X'_{\text{res}} \) phase offset. Uncertainty estimates for \( K_x \) and \( K_y \) appear in Table 3. The data have been segregated by season, since we wish to explore the connection between seasonal variations in migrating diurnal tides, and in \( K_x \) and \( K_y \). The clearest contrasts in \( u' \) amplitudes are seen between DJF and MAM. DJF amplitudes are sharply peaked at 10 m s\(^{-1}\), compared to the MAM curve, which is more broadly peaked near 20 m s\(^{-1}\). The seasonal characteristics of \( u' \) are echoed in the zonal momentum distribution, with the DJF curve peaked at 60 m s\(^{-1}\) d\(^{-1}\), and the MAM curve peaked more broadly about 100 m s\(^{-1}\) d\(^{-1}\). The phase offsets between \( X'_{\text{res}} \) and \( u' \) are very sharply peaked at \(-6\) h during JJA. During the remaining seasons,
phase offsets have a broader distribution, with the largest occurrences of 12 h offsets (pure damping) in SON. 

[27] A leading objective of this study is to provide observational support for Rayleigh friction estimates in tuned mechanistic models [Khattatov et al., 1997; Yudin et al., 1997; Hagan et al., 1999a]. Hence, we examine the statistical behavior of $K$ coefficients as a function of season. The real part of $K_x$ is distributed symmetrically about the zero value, denoting nearly equal degrees of damping and amplification. Most of the real values range between $-3$ and $3$ d$^{-1}$. The complex part of $K_x$ is almost universally negative. A clear contrast is observed between curves representing equinox (red and green) and solstice amplitudes (black and blue). Solstice amplitude distributions are more sharply peaked at 5 d$^{-1}$, while equinox distributions are more broadly peaked about 4 d$^{-1}$. These differences arise primarily through the imaginary component, and the real component during JJA. At equinoxes, $\text{Im}(K_x)$ is centered about $-4$ d$^{-1}$, with a broad dropout toward zero and weakly positive values. During solstice, $\text{Im}(K_x)$ is more sharply peaked at $-5$ d$^{-1}$. The real and imaginary variations translate into a $u' - X_{\text{res}}$ phase offset that peaks very sharply about $-6$ h during JJA. Thus, the emerging picture is one of a zonal Rayleigh friction with a distribution whose peak value is 20% stronger during equinox than solstice, but acting primarily to advance the tide by one-quarter cycle rather than to damp it.

[28] The meridional counterparts are shown in Figure 7. Both $\nu'$ and $Y_{\text{res}}$ show prominent seasonal variations, with highest amplitudes and broader distributions at equinox. The phase offsets between $\nu'$ and $\nu$ have a primary peak near $-7$ h during solstice and closer to $-10$ h at equinox. A secondary maximum at $\pm 12$ h, indicative of pure damping, is also observed during all seasons. Another way to display the essential damping nature of $K_y$ is by viewing the real and the imaginary components, and noting that the distribution of $\text{Re}(K_y)$ is centered about 2 d$^{-1}$ during all seasons. This value is very consistent with that obtained by Khattatov et al. [1997] in a numerical model that was tuned to HRDI winds. Distributions of $K_y$ amplitudes do show a clear seasonal contrast. Peak equinox values are about 20% lower than those at solstice, and the solstice curves are broader. The real part of $K_y$ has a slightly broader distribution at solstices, indicating a higher incidence of damping, while $\nu' - Y_{\text{res}}$ phase offsets are closer to $-6$ h during solstices. Thus, while the meridional wave drag acts to damp the tide, it behaves similarly to the zonal wave drag during solstice when it advances the phase of the meridional wind tide.

6. Summary and Conclusions

[29] We have obtained 60 day averages of diurnal winds, temperatures, and geopotential using 8 years (2002–2009) of TIMED measurements. These estimates have been used to evaluate the tidal momentum budgets given by equations (1) and (2). A leading objective of this study is to provide observational support for diurnal Rayleigh friction coefficients obtained from tuned mechanistic models [Khattatov et al., 1997; Yudin et al., 1997; Hagan et al., 1999a]. We therefore converted momentum residuals (3) and (4) to Rayleigh friction, and examined the variability of the global distributions of the Rayleigh coefficients.

[30] The leading term in (1) in addition to the classical terms is the meridional advection of zonal momentum. This term maximizes during equinox periods, when it attains values of about 80 m s$^{-1}$ d$^{-1}$. The enhanced advection at equinox is due in part to stronger equinox $\nu'$, and the increased meridional shear of $\overline{U}$ during equinox (see Figure B1c). These observations are consistent with numerical experiments of McLandress [2002b]. Whereas McLandress [2002b] was able to isolate the effects of meridional wind shear on tidal propagation and amplitude evolution, we are not able to determine from the available observations whether enhanced advection is a cause or a consequence of the tidal enhancement at equinox.

[31] The net zonal force $X_{\text{res}}$ has an order of magnitude of about 100 m s$^{-1}$ d$^{-1}$, and is generally in quadrature with the zonal wind. This is consistent with theoretical scenarios of momentum deposition by gravity waves that are modulated by the tide. The net forcing upon the diurnal meridional wind, or $Y_{\text{res}}$ is generally stronger than $X_{\text{res}}$ by about a factor of two. In contrast to the zonal momentum residual, $Y_{\text{res}}$ is to a significant degree in antiphase with the meridional wind. Thus the diagnosed meridional wave drag on balance tends...
to damp the tide, while the diagnosed zonal wave drag advances the phase of the tide.

[32] We have examined seasonal variations in the diurnal parameters. Tidal amplitudes are strongest during vernal equinox, and exceed solstice amplitudes by up to a factor of two. Advection of zonal momentum by the diurnal meridional wind also maximizes at equinox, and has minima at solstice. However, we are unable to determine whether the seasonal behavior of meridional advection is a cause or a consequence of the tidal seasonal variations. Zonal and meridional momentum residuals are also stronger during equinox, exceeding solstice values by up to a factor of two. Amplitudes of the corresponding Rayleigh friction coefficients are weaker at equinox, and stronger at solstice. However, the variation in the peak values is about 20%, in contrast to 100% variations (and greater) required to tune mechanistic models to reproduce observed tidal amplitudes. Moreover, the observed amplitude variations arise largely through changes in tidal phasing, rather than damping.

[33] Comparisons of our calculations of $K_x$ and $K_y$ with previous studies show both similarities and differences. Khattatov et al. [1997] solved for $K_R$ in a linearized tidal model whose meridional winds field consisted of HRDI diurnal winds. $K_R$ was constrained to be real, positive, and identical for both zonal and meridional winds. Magnitudes ranged between 1 and 3 d$^{-1}$, which are very similar to those we obtain for Re($K_x$). More recently, Xu et al. [2009b] computed diurnal winds from SABER diurnal temperatures using a linear tidal model with nonmolecular diffusive processes represented by equivalent Rayleigh damping. The frictional parameters (e.g., altitude of maximum, Gaussian half-width) were determined so as to minimize the differ-

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**Figure 6.** Histograms of (a) zonal wind amplitude, (b) $X_{res}$ amplitude, (c) phase offset between $u'$ and $X_{res}$, (d) $K_x$ amplitude, (e) Re($K_x$), and (f) Im($K_x$). Y axis is the fraction of occurrences (or ratio of occurrences to total number of data points). Only data points between 30°S and 30°N and 12.5 and 15 scale heights are used. Portions of the curve within dotted lines are within the uncertainty range. Black, December–January–February (DJF) values; green, March–April–May (MAM); blue, June–July–August (JJA); red, September–October–November (SON).
ences between the model winds, and TIDI diurnal winds. The seasonal variations we reported in $X'_{\text{res}}$ and $Y'_{\text{res}}$ are similar to those calculated by Xu et al. [2009b] (maxima at equinox and minima at solstice), and the range of real $K_y$ displayed in Figure 7 is highly consistent with the $K_{\text{ERF}}$ values of Xu et al. [2009b]. However, our calculations of momentum residuals were carried out independently for the zonal and the meridional components, and were not constrained to represent pure damping. This has resulted in complex values of $K_x$, and in $X'_{\text{res}}$ universally leading $u'$ by one-quarter cycle. The quadrature relationship differs very substantially from that obtained in the more constrained models of Khattatov et al. [1997] and Xu et al. [2009b].

It is important to bear in mind that equations (3) and (4) express $X'_{\text{res}}$ and $Y'_{\text{res}}$ as the combined effects several mechanisms: nonlinear wave interactions, advection by the mean meridional circulation, and small-scale turbulence and eddy diffusion ($X'$ and $Y'$). Thus, the diagnosed momentum residuals and associated $K_x$ and $K_y$ are describing the net effect of multiple processes that can act in phase, antiphase or in quadrature with each other. For example, McLandress [2002a] explicitly calculated the projection of resolved, nonlinear eddy advection in CMAM onto the migrating diurnal component. This mechanism maximized at equinox, but was associated with a net damping of the tide, so did not explain equinox wind maxima in CMAM. An additional caveat in the interpretation of $Y'_{\text{res}}$ is the possibility of aliasing by the semidiurnal tide, that may exaggerate the antiphase relationship with $v'$ (see Appendix A).

The quadrature relationship between $X'_{\text{res}}$ and $u'$ (and to a lesser extent, between $Y'_{\text{res}}$ and $v'$) suggests turbulent and diffusive processes associated with gravity waves breaking. However, other mechanisms can induce similar phase offsets. Preliminary experiments using a nonlinear model of Ortland and Alexander [2006] (to be reported in a future publication) indicate that tidal dissipation forces a mean circulation that advects zonal momentum vertically and meridionally. During the vernal equinox, total advection was found to be as large as eddy diffusion, and precedes $u'$ and $v'$ by 90° in phase. A fuller understanding of the rela-

Figure 7. As in Figure 6, but for the meridional component.
tion between tidal variability and wave drag requires that all the effects encapsulated by the momentum residual be quantified.

Appendix A: Impact of Semidiurnal Tide on Analyses of Diurnal Tides and Mean Fields

This appendix presents quantitative estimates of aliasing by semidiurnal tides of TIMED ascending-descending node averages and differences. Estimates of semidiurnal winds, temperature and geopotential are provided from a modified version of the model of Ortland and Alexander [2006]. In this version, eddy diffusion is given by the vertical profile from McLandress [2002b], and Newtonian cooling is replaced by the heating code of Zhu et al. [1999]. The heating algorithm uses O₃, O, and CO₂ taken from the NRL Mass Spectrometer and Incoherent Scatter radar (MSISE-00) empirical model [Picone et al., 2002], O₃ from UARS Halogen Occultation Experiment (HALOE), and H₂O from the NASA Water Vapor Project (NVAP) [Lieberman et al., 2003a]. The tides in the model are forced by daily heating of H₂O and O₃. These modifications result in a semidiurnal tide that closely resembles that produced by the Global Scale Wave Model (GSWM) [Hagan et al., 1995]. We apply TIDI and SABER sampling during the 15 March to 15 May yaw period to the semidiurnal tides. Diurnal tides and zonal mean winds and temperatures are then computed in a manner identical to that reported in section 3.

Figure A1. Aliasing by the semidiurnal tide. (top) Zonal mean zonal wind (m s⁻¹). (middle) The 00LT X’ₚₑₛ (m s⁻¹ d⁻¹). (bottom) The 00LT Y’ₚₑₛ (m s⁻¹ d⁻¹).
The effects of the semidiurnal tide on zonal mean temperature are less than 1K. Semidiurnal aliasing of the diurnal winds is relatively mild between 30°S and 30°N (not shown), but increases very steeply poleward of these latitudes. Aliasing of the zonal mean zonal wind does occur, and is shown in Figure A1 (top). The effects of the aliased zonal wind upon the advective terms in equation (1) do not exceed 10 m s$^{-1}$ d$^{-1}$. Errors in $X_{res\prime}$ are shown in Figure A1 (middle). For the most part, these range between 10 and 20 m s$^{-1}$ d$^{-1}$ in the subtropical latitudes, but increase sharply poleward of 30°. These considerations have led us to retain only values between 30°S and 30°N for analyses of seasonal variations in the momentum budget and residuals.

Semidiurnal geopotential aliases the retrieval of diurnal geopotential at low latitudes. This effect leads to spurious meridional gradients in geopotential, and is the primary cause of the errors in $Y_{res\prime}$, shown in Figure A1 (bottom). These maximize at about 10° on either side of the equator, with values between 60 and 70 m s$^{-1}$ d$^{-1}$ (about 25% of the $Y_{res\prime}$ maxima shown in Figure 3). Comparison of aliased $Y_{res\prime}$ pattern with the 00 LT diurnal $v'$ in Figure 1 indicates an antiphase relationship between these 2 variables. Thus, it should be kept in mind that the negative correlations observed between these 2 variables can be exaggerated by the effects of the semidiurnal tide.

Appendix B: TIDI Zonal Mean Zonal Winds

TIDI “zero wind” spectral line positions (required for the absolute vector wind) are telescope dependent, and subject to both random and systematic fluctuations that contribute to TIDI’s wind uncertainties [Niciejewski et al., 2006]. The purpose of this appendix is to argue the validity of TIDI zonal mean zonal winds. We compare TIDI with UARS High Resolution Doppler Imager (HRDI) winds, and with the radar-oriented Global Empirical Wind Model (GEWM) [Portnyagin et al., 2004].

Figure B1. (a) HRDI daytime averaged zonal mean zonal winds (m s$^{-1}$), (b) TIDI daytime averaged zonal mean zonal winds (m s$^{-1}$), and (c) TIDI 24 h averaged zonal mean zonal winds (m s$^{-1}$) centered on (left) 1 March and (right) 1 December. (d) GEWM zonal winds (m s$^{-1}$) averaged over (left) February–March and (right) November–December.
HRDI data were collected between 1991 and 1998, while TIDI winds were gathered between 2002 and 2009. We focus on the 35°S–35°N range where the bulk of the momentum budget calculations are made. Because HRDI winds are measured only during daylight hours over most of the MLT, we first focus on the comparison of HRDI with TIDI daytime-only winds. These are shown in Figures B1a and B1b. Using values sampled only between 5 and 20 LST, winds from each instrument are averaged over 60 days. (Thus, results labeled “March” are obtained from measurements gathered between 1 February to 1 April.) TIDI and HRDI daytime zonal mean zonal winds are morphologically similar, both showing westward winds at equatorial latitudes during equinox, and eastward winds at midlatitudes. HRDI generally exhibits stronger values at midlatitudes than TIDI. The reason for this behavior is not known, but may be related to HRDI and TIDI viewing winds during different decades [Huang et al., 2006]. Uncertainties in the TIDI’s zero wind line position may also be a factor.

[41] Because TIDI measures nighttime winds between 80 and 100 km, it is illuminating to consider the differences in zonal wind averages formed from daytime-only data (Figure B1b), and those formed from 24 h of data (Figure B1c). Comparing plots in Figures B1b and B1c indicates that when 24 h of TIDI data are used for averaging, the following features are observed: (1) Vertically continuous westward winds over the equator during equinox. (2) Reduction of westward flow over the equator by 10 m s\(^{-1}\) (or 50%) during solstice. (3) Reduction of eastward flow by 10 m s\(^{-1}\) (or 50%) at 30°. These comparisons suggest that averages formed from daytime-only winds may contain localized enhancements most likely due to incomplete removal of the diurnal tide.

[42] Finally, we show GEWM winds for the corresponding months in Figure B1d. The GEWM data set is derived from 24 h averages of HRDI winds near ~95 km, radar (mainly medium frequency) measurements, and UARS wind imaging interferometer (WINDI) empirical winds in the 90–110 km range [Wang et al., 1997] that have been divided by 2. We note immediately that GEWM winds are significantly weaker than HRDI winds, which is due to the speed bias of spaced receiver radar measurements [Jacobi et al., 2009]. However, their latitude–height structure appears quite consistent with TIDI winds, apart from winter when the GEWM zero wind line is found at greater altitudes than measured by TIDI. This may be due in part to incomplete correction of medium frequency radar heights, which are overestimated due to group retardation in the ionospheric D region [Nambottirhi et al., 1993].

[43] In summary, Figure B1 (and similar analyses for the remaining months, not shown) demonstrates the following: (1) TIDI zonal mean zonal winds agree morphologically with HRDI winds. (2) TIDI 24 h averaged winds are in very good agreement with the GEWM climatology that is derived from and tuned to radar winds. (3) HRDI zonal mean zonal winds often exceed TIDI values by as much as 50%, especially at midlatitudes. (4) Zonal averages formed from daytime-only TIDI winds are locally enhanced relative to 24 h averages. (5)
TIDI daytime zonal mean winds lie in the intermediate range of magnitude between HRDI and radar winds.

[44] Since both radar- and HRDI-based winds have been used extensively for diagnostic analysis [Meek et al., 1996; Vincent et al., 1998; Fritts et al., 1999; Lieberman, 1999; Lieberman et al., 2003b], TIDI zonally averaged zonal winds would appear to be a reasonable data set for the calculations of tidal advection in section 4. Nevertheless, insofar as TIDI-HRDI zonal mean differences remain unexplained, we have opted to view these as an additional source of uncertainty. We therefore perform a sensitivity study of the tidal momentum budget to the zonal mean zonal winds. The calculations described in section 4 have been repeated using TIDI “monthly” zonal mean zonal wind patterns multiplied by a factor of 1.5 so as to bring them into better quantitative agreement with HRDI.

[45] Figure B2 illustrates the impact upon the vertical and meridional advection and the zonal momentum residuals. The quantities in the plot are the differences between values obtained at 00LT with TIDI winds (shown in Figure 2), and TIDI winds enhanced by 1.5. Results are displayed for March and May, when the effects upon the zonal momentum residual were most prominent. (The effects of artificially enhanced TIDI zonal winds upon the meridional budget do not exceed 10 m s⁻¹ d⁻¹.) Enhanced vertical advection in May (Figure B2d) increases the zonal momentum residual by about 20 m s⁻¹ d⁻¹ over the equator at 14 scale heights. This corresponds to a nearly 50% increase at this location, compared to calculations with unmodified TIDI winds. Enhanced meridional advection leads to an increases to the zonal momentum budget ranging between 40 and 60 m s⁻¹ d⁻¹ centered at 25° in each hemisphere. The phase of the momentum residual (not shown) is affected so as to center its offset from ut more sharply about −6 h.

[46] We conclude that the uncertainties in TIDI’s zonal mean zonal winds do not affect our conclusion that the zonal momentum residual leads the zonal wind tide by approximately one-quarter cycle. The estimates for the advective terms and the zonal momentum residual presented in sections 4 and 5 represent a lower limit of the range of likely values.

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