RESEARCH ARTICLE

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Key Points:

- A dissipating tidal spectrum modulated by planetary waves (PW) causes the thermosphere to “vacillate” over a range of PW periods
- The same tidal spectrum can amplify penetration of westward propagating PW into the dynamo region, through nonlinear wave-wave interactions
- Zonally symmetric ionospheric oscillations arising from thermospheric vacillation are potentially large

Supporting Information:

- Supporting Information S1

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Zonally Symmetric Oscillations of the Thermosphere at Planetary Wave Periods

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Abstract New mechanisms for imposing planetary wave (PW) variability on the ionosphere-thermosphere system are discovered in numerical experiments conducted with the National Center for Atmospheric Research thermosphere-ionosphere-electrodynamics general circulation model. First, it is demonstrated that a tidal spectrum modulated at PW periods (3–20 days) entering the ionosphere-thermosphere system near 100 km is responsible for producing ± 40 m/s and ± 10 –15 K PW period oscillations between 110 and 150 km at low to middle latitudes. The dominant response is broadband and zonally symmetric (i.e., “S0”) over a range of periods and is attributable to tidal dissipation; essentially, the ionosphere-thermosphere system “vacillates” in response to dissipation of the PW-modulated tidal spectrum. In addition, some specific westward propagating PWs such as the quasi-6-day wave are amplified by the presence of the tidal spectrum; the underlying mechanism is hypothesized to be a second-stage nonlinear interaction. The S0 total neutral mass density (ρ) response at 325 km consists of PW period fluctuations of order ± 3 –4%, roughly equivalent to the day-to-day variability associated with low-level geomagnetic activity. The variability in ρ over short periods ($\sim < 9$ days) correlates with temperature changes, indicating a response of hydrostatic origin. Over longer periods ρ is also controlled by composition and mean molecular mass. While the upper-thermosphere impacts are modest, they do translate to more significant changes in the F region ionosphere.

1. Introduction

During the past decade or so, it has been established that the troposphere-stratosphere region (0–50 km) drives ionosphere-thermosphere-mesosphere (ITM) variability through generation of a spectrum of vertically propagating waves, including gravity waves (GWs), tides, ultrafast Kelvin waves (UFWK), and planetary waves (PWs) (see review by Liu, 2016). GW, tides, and UFWK amplify exponentially as they propagate into the more tenuous upper atmosphere and thus play a dominant role in the meteorology of the mesosphere and lower-thermosphere (MLT) region (~ 80 –150 km). Early work (Lindzen, 1968; 1970; Richmond, 1975; Yanowitch, 1967) recognized that exponential growth of GW and tides (and by extension, UFWK; Forbes, 2000) is curtailed by viscous dissipation between about 100 and 130 km for waves with vertical wavelengths in the range ~ 25 –65 km. As summarized by Forbes and Garrett (1979), the approximate altitude where this occurs, and also where maximum amplitudes are achieved, is where the viscous term in the horizontal momentum and heat equations is of same order as the inertial term, or where

$$\chi \sim \frac{4\pi^2}{\lambda_z^2} \frac{\mu_0}{\rho\sigma} \quad (1)$$

where λ_z is the vertical wavelength; μ_0 is the coefficient of molecular viscosity; ρ is the total mass density, an exponential function of height; and σ is the wave frequency. Note that $\chi \sim 1$ at a higher (lower) altitude during low (high) solar conditions when ρ is lower (higher). The shape of the amplitude profile also depends on the ratio of λ_z to the mean scale height. The net effect is that vertical penetration of waves above ~ 130 km is more (less) effective during low (high) solar activity conditions.

The vertically propagating wave spectrum varies from day to day due to changes in sources associated with tropospheric weather and variable propagation conditions, and due to nonlinear interactions between different parts of the wave spectrum as propagation occurs through the middle atmosphere (20–100 km; Liu,

2016). GW, tides, and UFKW impress their spatial-temporal variability on the ITM through a number of different mechanisms and pathways (Liu, 2016), and when these waves dissipate, they also modify the mean thermal structure and circulation of the thermosphere (Jones, Forbes, & Hagan, 2014; Jones, Forbes, Hagan, & Maute, 2014; Miyoshi & Fujiwara, 2006; Yigit & Medvedev, 2015). This part of the wave spectrum also affects O and O₂ composition, as the result of modified circulation (Yamazaki & Richmond, 2013) and/or the net transport effects demonstrated in Gardner and Liu (2010) and Jones, Forbes, and Hagan (2014). In turn, the modified composition affects electron production and loss rates that translate to changes in electron density (Jones, Forbes, Hagan, & Maute, 2014).

Traveling PWs have periods in the 2 to 20-day range. Prominent among these are the quasi-2-day wave (Q2DW), Q6DW, Q10DW, and Q16DW, which are normal modes (resonant oscillations) of the atmosphere (Forbes, 1995; Salby, 1981). All of these are westward propagating, the Q2DW with zonal wave number $s = 3$; the rest with $s = 1$ (hereafter Q6DW1, Q10DW1, and Q16DW1). Due to their longer periods and greater sensitivity to the mean wind field, PWs penetrate above about 100 km much less efficiently than GW, tides, and UFKW. Recent general circulation model (GCM) simulations by Yue et al. (2013) and Yue and Wang (2014) for the Q2DW and Gan et al. (2017) for Q6DW1 indicate that some direct penetration by PW above 100 km is possible, and Meyer (1999) shows how GW modulation by PW and their subsequent dissipation in the *E* region can facilitate the vertical penetration of the Q2DW and Q16DW1. Pancheva et al. (2009a) and Pancheva et al. (2009b) quantified PW oscillations with periods between 2 and 27 days in Thermosphere Ionosphere Mesosphere Energetics and Dynamics (TIMED)/Sounding of the Atmosphere using Broadband Emission Radiometry (SABER) data extending between 20 and 120 km. In addition to the above westward propagating waves, they found zonally symmetric ($s = 0$) oscillations (hereafter, S0 oscillations) that mainly appeared above 100 km and which they attributed to the mechanism investigated by Meyer (1999).

There is also compelling evidence from ground measurements (e.g., Beard et al., 1999; Kamalabadi et al., 1997; Huang et al., 2013; Pancheva et al., 2000, 2002; Pancheva, 2001; Pancheva & Mitchell, 2004), space observations (e.g., Forbes & Moudeden, 2012; Forbes & Zhang, 2017; Gasperini et al., 2015), theory (Teitelbaum & Vial, 1991), and numerical simulations (e.g., Gan et al., 2017; Gasperini et al., 2017; Palo et al., 1999; Pedatella et al., 2012; Yue et al., 2016) that PW (including UFKW) modulate tides and in fact account for much of their day-to-day variability. If the tidal spectrum can induce net changes in ITM properties through dissipative processes as described above (e.g., Jones, Forbes, & Hagan, 2014; Yamazaki & Richmond, 2013), and the tidal spectrum is modulated by PW, then the next logical question to ask is: Can the PW-modulated tidal spectrum induce significant changes in the zonal mean state at PW periods, that is, S0 oscillations, such as those reported by Pancheva et al. (2009b)? It is the purpose of this paper to demonstrate that this is indeed the case and to note an analogy with the term “vacillation” in the middle atmosphere meteorological literature (e.g., Pancheva et al., 2009a). Additionally, we seek to establish how S0 oscillations forced in the lower thermosphere might manifest themselves at higher altitudes, that is, in neutral densities at satellite altitudes, and to investigate the interplay with S0 oscillations of similar periods that are forced in situ by solar and geomagnetic variability.

The S0 oscillations reported by Pancheva et al. (2009a, 2009b) are in fact a by-product of recent tidal (Truskowski et al., 2014) and PW (Forbes & Zhang, 2015, 2017) analyses of SABER temperatures involving two of the present authors. In those works, residuals from 60-day running means of measurement temperatures on both ascending and descending legs of the TIMED orbit are used to calculate solar and lunar tides. After removal of the tidal signatures, a second set of residuals are used for PW analysis. S0 oscillations are contained in daily longitude-means of these residuals. A few examples for 2009 at 110 km are shown in Figure 1 where we have averaged data from ascending and descending local times due to the similarities in their structures. We note that the temporal variability occurs over periods of roughly 2–20 days and with amplitudes of order ± 10 K. These temperature excursions are of the same order as those attributable to DE3 and UFKW at 110 km (Gasperini et al., 2017) and thus constitute a potentially important component of the meteorology of this atmospheric region.

In order to answer the questions posed above, we employ the National Center for Atmospheric Research thermosphere-ionosphere-mesosphere-electrodynamics and thermosphere-ionosphere-electrodynamics general circulation models (TIME-GCM and TIE-GCM), and National Aeronautics and Space Administration's Modern Era Retrospective-analysis for Research and Applications (MERRA) to perform numerical experiments designed to isolate the effects of tides on the thermosphere. Appendix A provides relevant details about the models and addresses their suitability to address the problem at hand. The numerical experiments

Figure 1. Latitude versus day of month plots of zonal mean and ascending + descending mean temperatures at 110 km from Thermosphere Ionosphere Mesosphere Energetics and Dynamics/Sounding of the Atmosphere using Broadband Emission Radiometry measurements during 2009. (clockwise from upper left) April, June, October, and December. Day zero is the last day of the previous month. Supporting information Figure S1 contains all 12 months.

are described in the following section. Our results are presented in section 3, and section 4 provides some interpretation of results, related discussion, and conclusions.

Throughout this paper, the notation DWs(SWs) or DEs(SEs) is used to denote a westward or eastward propagating diurnal(semidiurnal) tide, respectively, with zonal wave number = s . Zonally symmetric oscillations are denoted by D0(S0).

2. Numerical Experiments

Readers unfamiliar with the TIE-GCM, TIME-GCM, and MERRA are referred to Appendix A for a description of these models as implemented for the current study. In the numerical experiments described below, TIME-GCM (hereafter TIME-GCM/MERRA2009) is forced at its lower boundary near 30 km altitude by output from MERRA for all of 2009 as described in Häusler et al. (2014) and Häusler et al. (2015). Since MERRA has 3-hourly resolution and is constrained by observational data, it is presumed to drive the TIME-GCM with realistic tidal, UFKW and PW forcing, and day-to-day variability at its lower boundary. Additional tides are forced within the TIME-GCM below ~ 100 km by solar radiation absorption by O_3 in the stratosphere and through nonlinear tide-tide interactions (especially for the terdiurnal tide; Moudén & Forbes, 2013; Truskowski et al., 2014) and tidal interactions with stationary PWs (Angelats i Coll & Forbes, 2002; Lieberman et al., 2015; Yamashita et al., 2002). A few examples of the tidal variability to emerge from the model at 100-km altitude are provided in Figure 2 for the largest tidal components in October 2009: SW2, DW1, DE2, and D0. Additional examples are provided in supporting information Figure S1. These examples clearly indicate variability over periods of order 3 to 15 days and, based on prior observational and modeling work summarized in section 1, reflect modulation of the tidal spectrum by PW. The month of October 2009 is chosen for our numerical experiments, since it contains the familiar Q6DW1, Q10DW1, Q16DW1, and UFKW, and the Q6DW1 is particularly robust, extends above 100 km, and is the subject of interest in a number of modeling and observational studies (e.g., Forbes & Zhang, 2017; Gan et al., 2015, 2016, 2017; Gu et al., 2014; Jiang et al., 2008; Liu et al., 2004; Pancheva et al., 2008; Pedatella et al., 2012; Riggin et al., 2006; Talaat et al., 2002).

The numerical experiments that we report on are primarily designed to reveal the nature and origins of S0 thermosphere variability at PW periods due to dissipating tides and consist of the following. First, a *reference* thermosphere-ionosphere simulation is produced by forcing the TIE-GCM at its ~ 97 -km lower boundary with output of TIME-GCM/MERRA2009, and solar and geomagnetic variability based on observed $F_{10.7}$ and Kp indices, respectively. A second TIE-GCM simulation is forced by the daily and zonal means of the ~ 97 -km TIME-GCM/MERRA2009 output at each latitude, thus eliminating all tides and PW. The only TIE-GCM forcing

Figure 2. Latitude versus day of month plots of sample thermosphere-ionosphere-mesosphere-electrodynamics and thermosphere-ionosphere-electrodynamics general circulation model/Modern Era Retrospective-analysis for Research and Applications 2009 tidal temperature amplitudes at 100 km, showing day-to-day variability at planetary wave periods. Clockwise from upper left: SW2, DW1, DE2, and D0. Supporting information Figure S2 contains nine tidal components at 100 and 110 km.

in this *S0 forcing only* simulation is attributable to any PW period *S0* oscillations that might exist at 97 km and to in situ forcing in the thermosphere due to solar flux and geomagnetic variability. Next, we high pass the TIME-GCM/MERRA2009 output with a cutoff frequency of 1/1.3 days (essentially isolating solar tides), and add this to the daily and zonal means at 97-km altitude to generate a *S0 forcing + tides* simulation. The hourly TIE-GCM outputs of all variables at all grid points are differenced between the *S0 forcing + tides* and *S0 forcing only* simulations to produce output fields that we designate *tide contribution*. Note that the reference simulation is the only simulation that includes PW with $s \neq 0$. Also, all *S0* oscillations depicted in this paper are zonal and diurnal means, except where noted (e.g., section 3.2).

3. Results

In this section we present and interpret results from our numerical experiments. The following subsection focuses on the lower thermosphere in the context of potential relevance to atmosphere-ionosphere coupling and to the SABER data in Figure 1. Section 3.2 is devoted to the response in the upper thermosphere and relevance to the interpretation of Challenging Minisatellite Payload accelerometer measurements of total mass density and satellite drag in general.

3.1. Lower-Thermosphere Temperatures and Winds

Figures 3 and 4 provide results from our numerical experiments that focus on PW periodicities (2–20 days) in lower-thermosphere zonal winds at the equator. These figures consist of longitude versus time (day of year [DOY] in October 2009) depictions of daily mean zonal winds, and the corresponding 2-D zonal wave number versus period spectra of these variations, at 97 km (Figure 3) and at 120 km (Figure 4) based on the numerical experiments. Longitude versus time plots for zonal wind and temperature at -40° , -20° , 0° , $+20^\circ$, and $+40^\circ$ latitude are provided in supporting information Figures S3 and S4, respectively, and the corresponding spectra are found in supporting information Figures S5 and S6. The eight panels comprising Figures 3 and 4 are representative results extracted from the total of 80 panels comprising supporting information Figures S3–S6. The altitude of 120 km is chosen since it is centrally located in the dynamo region, and it is a typical altitude where the tides and *S0* oscillations maximize.

Figure 3. Examples of TIME-GCM/MERRA2009 forcing at the lower boundary of the TIE-GCM. (a) Diurnal mean amplitudes of zonal wind over the equator at 97-km altitude as a function of longitude and day of month during October 2009, and (b) wave number versus period spectrum of the data shown in Figure 3a. Similar figures at -40° , -20° , 0° , $+20^\circ$, and $+40^\circ$ latitude for both zonal wind and temperature are contained in supporting information Figures S3–S6.

In order to equitably capture both short- and long-period PW, throughout this paper 2-D spectra with resolution of 0.125 days are formed within windows that are 3 times the wave period (P) and centered on 15 October. Specifically, spectra for waves with periods $2 \text{ days} \leq P \leq 5 \text{ days}$ are constructed based on a window length of 15 days, and for $P > 5 \text{ days}$ the window length is $(3P + 1) \text{ days}$. Therefore, for $P > 10 \text{ days}$, the window length extends into September and November, with a maximum window length of 61 days for the 20-day period. For waves with period $P \leq 3 \text{ days}$ spectra are calculated based on hourly data; otherwise, daily mean data are used. All of this implies that there is not an exact correspondence between the spectra and the daily mean values in longitude versus time plots; however, the spectra do equitably balance the relative amplitudes of the short- and long-period waves, which is the goal.

Figure 4. Examples of responses at 120 km due to different lower boundary forcing of the TIE-GCM. (a–c) Longitude versus day of month depictions of diurnal mean zonal winds over the equator. (d–f) The wave number versus period spectra of the data shown above. S0 forcing only (Figures 4a and 4d). S0 forcing plus tides (Figures 4b and 4e). Tide contribution alone (Figures 4c and 4f). Similar figures at -40° , -20° , 0° , $+20^\circ$, and $+40^\circ$ latitude for both zonal wind and temperature are contained in supporting information Figures S3–S6.

Returning to Figure 3, Figure 3a illustrates the longitude versus day variations in the daily mean zonal winds at the lower boundary of the TIE-GCM, and Figure 3b illustrates the zonal wave number versus period spectra of the zonal winds in Figure 3a. Both figures are complementary ways to illustrate the dominance of $s = 1$ westward propagating oscillations with periods between about 5d and 16 days. The spectrum notably highlights maxima near 5–7 days, 9–10 days, and 12–16 days that are generally interpreted as manifestations of the Q6DW1, Q10DW1, and Q16DW1 normal modes (Forbes, 1995; Salby, 1981). There is also a signature of a weaker UFKW near 3–4 days period and $s = -1$, and amplitudes of order 2–3 m/s at $s = 0$. (Note that the UFKW becomes more prominent in the equatorial temperature field compared to the other waves; see Figures S4 and S6.) Finally, while each wave in Figure 3b spectrum possesses amplitudes of order 9 m/s or less, the aggregate superposition of all waves produces longitude versus time structures (e.g., Figure 3a) that are of order ± 40 m/s. This is typical of such pairs of plots throughout this paper. Similar plots at other latitudes and for the temperature field in Figures S3–S6 are broadly consistent with those described above.

Figure 4 provides results of the numerical experiments for the lower thermosphere (120 km) that complement Figure 3 for 97 km. Figures 4a–4c consist of longitude versus day of month plots of equatorial zonal wind, and Figures 4d–4f illustrate the corresponding 2-D spectra as described above. Figures 4a and 4d depict the response due to daily variability of the zonal and diurnal average fields at 97 km (S0 forcing only) plus in situ forcing in the thermosphere due to day-to-day changes in geomagnetic and solar activity. Note that the spectral amplitudes (< 1.5 m/s) and longitude-time variability ($\sim \pm 4$ m/s) are small compared to those at 97 km in Figure 3. Figures 4b and 4e correspond to the same forcing as in Figures 4a and 4d with the addition of the full spectrum of tides imposed at the lower boundary of the TIE-GCM. Finally, Figures 4c and 4f represent the contributions of the tidal spectrum alone, obtained by differencing the hourly TIE-GCM inputs leading to Figures 4b and 4e from those leading to Figures 4a and 4d and then depicting the results of that differenced output in Figures 4c and 4f. Not surprisingly, given the small amplitudes displayed in Figures 4a and 4d, those in Figures 4c and 4f show little difference from those in Figures 4b and 4e.

The essence of this paper is contained in Figures 4a and 4f, which reveal the following. First, the upward propagating tidal spectrum at 97 km has resulted in production of PW period oscillations at 120 km that are substantial in their aggregate amplitudes (± 40 m/s) and that consist of a broad spectrum of S0 oscillations extending in period from 5 days to 20 days. Our interpretation is that the upward propagating tidal spectrum at 97 km is modulated over a range of PW periods (see Figures 3b and 2, as well as supporting information Figures S2–S6), propagates upward and dissipates in a broad region around 120 km, and deposits momentum into the mean flow over the same range of PW periods. The result is the PW period S0 oscillations evidenced by Figure 4f. Figure 4f also provides evidence that Q6DW1 and Q9DW1 are produced by tides alone. Some further insights into this result are provided below.

Figures 5 and 6 focus solely on the S0 oscillations, revealing their latitude and height structures. Figures 5e–5h show the zonal wind oscillations at 97 km (Figure 5e) and 120 km (Figures 5f–5h) associated with the TIE-GCM numerical experiments as a function of latitude and time. Figures 5a–5d contain the wave number versus periods spectra of these zonal wind oscillations, and Figures 5i–5l contain the S0 oscillations in temperature that correspond to Figures 5e–5h. Supporting information Figure S7 additionally includes the spectra of the S0 temperature oscillations.

Figures 5a and 5e illustrate the S0 oscillations at the lower boundary of the TIE-GCM, which mainly occur at periods near 6, 9, and 10–18 days. These may possibly originate from nonlinear interactions between $s = 1$ traveling PW at these periods, and $s = 1$ stationary PW (Pancheva et al., 2007, 2009a). Figure 5b indicates that for the most part these S0 oscillations do not propagate to 120 km, but instead features appear near $\pm 50^\circ$ latitude, suggesting the equatorward penetration of effects due to in situ forcing at high latitudes, which also project onto the indicated wave periods (see section 3.2). We note that Figure 5f contains large features that project onto periods longer than 20-day period, so that the variability suggested in Figure 5b is masked in Figure 5f. Figures 5c and 5g reflect the S0 response obtained by adding tides to the lower boundary forcing, and Figures 5d and 5h show the S0 oscillations due solely to tidal forcing at the lower boundary. The latter reflect some degree of symmetry with respect to the equator and reflect zonal wind variability on the order of ± 30 m/s. The S0 oscillations in temperature in Figures 5k and 5l are of order ± 6 K, somewhat smaller than the ± 10 K S0 oscillations observed by SABER during October (Figure 1).

Figure 6 provides insights into the height structures of S0 oscillations, while confirming the quasi-symmetry inferred from Figure 5d. The illustrated amplitudes are obtained by averaging fits to S0 oscillations with

Figure 5. This figure provides information on the latitude structures of S0 oscillations from the numerical experiments as described in the text. (e–h) The zonal wind oscillations at 97 km (Figure 5e) and 120 km (Figures 5f–5h) associated with the TIE-GCM numerical experiments as a function of latitude and time. Figures 5a–5d contain the wave number versus periods spectra of these zonal wind oscillations, and Figures 5i–5l contain the S0 oscillations in temperature that correspond to Figures 5e–5h. Supporting information Figure S7 additionally includes the spectra of the S0 temperature oscillations.

periods of 5–9 days (for 5–9 days, $s = 0$) and 10, 12, 14, and 16 days (for 10–16 days, $s = 0$) and thus underestimate the largest amplitude waves within these bands, but do provide information on height structure. We see that the waves begin to achieve significant amplitudes above about 110 km, where molecular dissipation begins to assume importance, but extend to altitudes of at least 150 km.

An unexpected result that emerges from the numerical experiments is provided in Figure 7, which focuses on the height-latitude structures of the Q6DW1 and Q9DW1. Figure 7a shows the zonal wind amplitude of Q6DW1 from the reference TIE-GCM simulation, which is forced near 97 km by the full TIME-GCM/MERRA hourly output and which includes the full PW and tidal spectra. The maximum amplitudes of about 15 m/s at low latitudes and 105 km can be compared with the 20-m/s values measured by TIMED/TIMED Doppler Interferometer during the large Q6DW event of May 2003 (Gu et al., 2014). Moreover, the 10- to 15-m/s values between 110 and 120 km compare favorably with the 10- to 20-m/s values shown by Gan et al. (2016) & Gan et al. (2017) to produce significant effects on the *F* region ionosphere due to the dynamo generation of electric fields. Figure 7b shows the isolated tide contribution to the Q6DW1 zonal wind amplitude. Compar-

Figure 6. Height versus latitude structures of amplitudes of S0 oscillations in zonal wind due to tidal forcing alone. (left) The 5- to 8-day oscillations. (right) The 10- to 16-day oscillations. These were obtained by averaging amplitudes from fits to S0 oscillations with periods of 5–8 days for the 5- to 8-day plot, and periods of 10, 12, 14, and 16 days for the 10- to 16-day plot.

Figure 7. Height versus latitude structures of (a–c) Q6DW1 and (d–f) Q9DW1 wind amplitudes for zonal winds from reference TIE-GCM simulation with full TIME-GCM/MERRA2009 forcing at its lower boundary (Figures 7a and 7d), zonal winds due to tidal contribution alone (Figures 7b and 7e), and meridional winds due to tidal contribution alone (Figures 7c and 7f).

Figure 8. Variation of daily magnetic index, (top) A_p , and 10.7-cm solar flux, (bottom) $F_{10.7}$, during October 2009.

ing Figures 7a and 7b, it is clear that the net effect of the tides is to further extend the Q6DW1 into the 120- to 150-km region, increasing the efficiency with which it couples the atmosphere to the ionosphere through the dynamo generation of electric fields. Figures 7d and 7e show a similar result for the Q9DW1, and Figures 7c and 7f illustrate the tide contributions to the meridional component of wind for Q6DW1 and Q9DW1, respectively.

We furthermore note from Figure 4c that the amplification of Q6DW1 is clearly visible by eye only during the first half of the month. This suggests that the spectrum in Figure 4f and the depictions in Figure 7, which are centered on 15 October, underestimate the impact of the tidal contribution to the generation of Q6DW1. A preliminary theory on the mechanism underlying this interesting result is provided in section 4.

3.2. Upper-Thermosphere Densities

In this subsection we examine the zonally symmetric response of total mass density at 325 km, ρ_{325} , to tidal forcing at the lower boundary of the TIE-GCM. At this altitude local time effects are potentially important. Indeed, we found the PW period S0 response of neutral density to tidal forcing to be measurably dependent on local time. However, as shown below, the PW period S0 density response to geomagnetic and solar in situ forcing is dominant, and it is weakly dependent on local time. Therefore, we focus on midnight results here since the main conclusions are generally applicable to other local times.

Before proceeding further it is necessary to examine the temporal variability of in situ forcing due to magnetic and solar activity, which are parameterized in the TIE-GCM through relationships with K_p and $F_{10.7}$, respectively (see Appendix A). Since we are dealing with daily mean data here, we use the daily magnetic activity index, A_p , instead of K_p . The variability of A_p and $F_{10.7}$ during October 2009 is illustrated in Figure 8. The level of geomagnetic activity is low, with A_p values of order 3 ± 2 during the first 21 days of the month, a sudden increase from 1 to 11 from day 21 to 22, a slow (~ 5 -day) recovery afterward, and then an increase to $A_p = 9$ on day 30. $F_{10.7}$ remains near 69 ± 2 solar flux unit ($1 \text{ sfu} = 10^{-22} \text{ W m}^{-2} \text{ Hz}^{-1}$) during the first 22 days of October, then rises to a peak value of 80 sfu on day 26, and then decreases to $F_{10.7} = 74$ sfu on day 31.

For the present study, it is important to know how the above small variations in geomagnetic and solar activity might project onto PW periodicities. The Morlet wavelet analyses in Figure 9 reveal that $F_{10.7}$ projects mainly onto a broad peak maximizing near 14d during the second half of the month and extending (half-width) from ~ 9 days to > 15 days. The A_p spectrum possesses a peak mainly confined to the second half of the month near 6–7 days with a broad tail extending to longer periods with ~ 8 days half-width. There is also a more localized but weaker peak near 4-day period extending between days 10 and 30.

Figure 9. Wavelets corresponding to data in Figure 3.

Figure 10 compares the latitude versus time response of ρ_{325} to all sources of forcing (Figure 10a, “reference”); to “S0 forcing only” (Figure 10b); and to “S0 forcing + tides.” The depicted density variations represent percent residuals from a 27-day running mean at each latitude, which highlights variability at PW periods. Figure 10b primarily depicts the density response to in situ geomagnetic and solar forcing. One can discern a small positive response ($\sim +5\%$) around day 11 associated with the corresponding $A_p = 7$ peak in Figure 8 and a $\sim +17\%$ relative increase associated with the $A_p = 13$ peak on day 22 in Figure 8. After day 25 elevated densities remain due to increased solar and geomagnetic forcing (refer to Figure 8). Especially if one considers the period prior to day 25 when the effects of increased solar flux are apparent, the addition of tides (Figure 10c) results in a response measurably modified in comparison to the response to in situ forcing alone (Figure 10b) and is necessary to capture many of the details of the reference simulation (Figure 10a).

To gain deeper insights, consider Figure 11 that is based on the same simulations as Figure 10. Figure 11a represents the “tide contributions” to the day-to-day S0 density variability at noon and 325 km depicted in Figure 10c, based on analysis of the difference field between the TIE-GCM simulations corresponding to Figures 10b and 10c. Figure 11d is the spectrum of the density variability in Figure 11a. Figures 11b and 11e and 11c and 11f are the corresponding figures for temperature and atomic oxygen number density [O]. All quantities in Figures 11a–11c are residuals from the 27-day running mean at each latitude. The notable points are as follows. First, the density variability in Figure 11a prior to days 23–25 ranges between about -3% and

Figure 10. This figure shows S0 oscillations in total mass density from the numerical experiments at 325 km at midnight and illustrates the relative importance of in situ versus tidal forcing. (a) The reference simulation. (b) This corresponds mainly to in situ forcing. (c) The same as Figure 10b with the addition of tides at the lower boundary of the TIE-GCM. Note that vertical green areas in Figure 10b correlate with peaks in A_p in Figure 8.

Figure 11. This figure illustrates the “tide contributions” to S0 oscillations in (a) total mass density, (b) temperature, and (c) O density at 325 km. (d–f) The latitude versus period spectra for Figures 11a–11c, respectively. Scales are percent for density and K for temperature.

+4%, which compares with −7% and +2% depicted in Figure 10b for the same period. Therefore, the tidal contribution is roughly equivalent to density variability attributable to Ap variations between 1 and 5, that is, very low-level geomagnetic activity.

A second notable feature is the similarity/correlation between the variability and spectra of total mass density (Figures 11a and 11d) and temperature (Figures 11b and 11e), which suggests a connection between the two. Although not shown here, $[N_2]$ changes are in the range −4% to +7% and in antiphase with those in Figure 11c for $[O]$ and with nearly identical spectrum. Moreover, the $[O]$ and $[N_2]$ changes are smoother, reflecting a longer timescale response to S0 forcing in the lower thermosphere. As elucidated in Thayer et al. (2012) changes in total mass density arise in part due to a direct temperature (hydrostatic) effect and to a composition effect, and $[O]$ and $[N_2]$ densities are known to respond to tidal and PW dissipation through transport of $[O]$ (Gan et al., 2015; Jones, Forbes, & Hagan, 2014; Siskind et al., 2014; Yamazaki & Richmond, 2013; Yue & Wang, 2014). The total mass density response in Figure 11 appears to be consist in part of a temperature effect and a composition effect, with the former response occurring over shorter timescales than the latter.

The Yamazaki and Richmond (2013) mechanism for redistribution of $[O]$ occurs through a poleward residual meridional circulation responding to the westward acceleration of the mean flow by dissipating tides. This mechanism has been elucidated in some detail by Yue and Wang (2014) for the Q2DW and further verified for the Q6DW1 by Gan et al. (2015). The redistribution of $[O]$ through meridional transport by the dissipating DW1 and SW2 tides occurs with a $1/e$ time constant of roughly 9 days at F region altitudes (Yamazaki & Richmond, 2013; inferred from their Figure 3 for total electron content), which is consistent with the longer time constant compositional response noted above. The Jones, Forbes, and Hagan (2014) mechanism entails direct net transport of $[O]$ by the dissipating tides, and these authors show that this mechanism is of equal importance to the advective transport of $[O]$ according to Yamazaki and Richmond (2013). The time constant for the Jones, Forbes, and Hagan (2014) process is not known. In any case, all of the above studies consider either individual or a select set of tidal and PW components. In the present study where a spectrum of eastward and westward propagating tides exists, it is likely that the net compositional response to dissipating tides may be considerably more complex than the responses to individual westward propagating waves.

4. Summary and Conclusions

In this paper we report on a new mechanism for imposing PW variability on the ionosphere-thermosphere system. Numerical experiments with the National Center for Atmospheric Research TIE-GCM are performed to isolate zonally symmetric (S0) oscillations in temperature, wind, total mass density, and composition in the thermosphere due to molecular dissipation of a tidal spectrum that is modulated at PW periods. As the tidal spectrum dissipates in the 100- to 150-km region, it deposits momentum and heat into the zonal mean flow at PW periods, giving rise to oscillations or “vacillations” (see below) of the mean state. The numerical experiments are performed by separately forcing the TIE-GCM at its ~97-km lower boundary by output from TIME-GCM/MERRA2009 and by select dynamical components of this output such as the tidal spectrum, daily diurnal and zonal mean values at each latitude, and the combination of the two. The TIE-GCM outputs and their differences enable isolation of the contributions of the whole tidal spectrum and its variability to the dynamics above 97-km altitude, including S0 oscillations. The present work focuses on October 2009, since it is characterized by the clear presence of PW familiar from observations, in particular the westward propagating quasi-6-day and quasi-9-day waves with zonal wave number $s = 1$ (Q6DW1 and Q9DW1). The Q6DW1 is the focus of several recent studies on atmosphere-ionosphere coupling (e.g., Gan et al., 2015, 2016, 2017; Pancheva et al., 2008) and PW-tide coupling (Forbes & Zhang, 2017; Gan et al., 2017; Pedatella et al., 2012).

The computational results of this study are summarized as follows.

1. Addition of the tidal spectrum at the lower boundary of the TIE-GCM leads to zonal wind and temperature oscillations at 120-km altitude of order ± 40 m/s and ± 10 –15 K at PW periods and over a range of zonal wave numbers. However, the dominant response consists of broadband S0 oscillations that extend from about 110 km to over 150 km in altitude. Within the $\pm 50^\circ$ latitude band, they are equatorially centered with full width at half-maximum of order 15–20° latitude. The corresponding S0 oscillations in temperature are of order ± 6 K, somewhat smaller than the range of values deduced from TIMED/SABER measurements during October 2009 (± 10 K, cf. Figure 1).
2. The PW-modulated tidal spectrum leads to amplification of the Q6DW1 and Q9DW1 in the 110- to 150-km region. A plausible mechanism is discussed below.
3. The S0 total neutral mass density response at 325 km driven by tides consists of PW period fluctuations during October 2009, which span a total of about 7%, roughly equivalent to the day-to-day variability associated with daily Ap values that fluctuate between about 1 and 5, that is, low-level geomagnetic activity. However, implications for the ionosphere are more substantial (see below).

Physical insights into the above computational results are provided below.

Vacillation. One can say that the thermosphere “vacillates” in response to periodic tidal dissipation in much the same way that the middle atmosphere vacillates in response to Eliassen-Palm flux divergences induced by wave-wave interactions. In the middle atmosphere context, vacillation refers to oscillations in the zonal mean atmospheric state induced by wave-wave and wave-mean flow interactions, usually in the context of simple interference between stationary and traveling Rossby waves (Hirota, 1971; Madden, 1975, 1983; Pancheva et al., 2009a; Salby & Garcia, 1987; Smith, 1985). For instance, a 5-day, 10-day, or 16-day normal mode can constructively and destructively interfere with a stationary wave to induce fluctuations in Eliassen-Palm flux divergences that drive the mean state at the period of the traveling PW (e.g., Hirooka, 1986). By the same token, in the MLT region a PW can modulate a tide that then, upon dissipation at a higher level, drives a S0 oscillation at the same period as the modulating PW. In the present context, we have several PW modulating a tidal spectrum, leading to broadband vacillation of the thermosphere over a range of PW periods.

Q6DW1 and Q9DW1 Amplification. The mechanisms underlying the production of $s \neq 0$ PW oscillations as a result of adding the tidal spectrum to lower boundary forcing of the TIE-GCM are difficult to trace. We focus on amplification of Q6DW1 and Q9DW1 in the lower thermosphere, since these are normal-mode oscillations of the atmosphere and, in particular, the Q6DW1 is the subject of recent interest in the literature as noted above. It is hypothesized that the mechanism underlying this result may involve a second-stage nonlinear interaction. Consider the interaction between a PW with zonal wave number m , $\cos(\delta\Omega t + m\lambda)$, and a tide with zonal wave number s , $\cos(n\Omega t + s\lambda)$, where $\Omega = 2\pi \text{ day}^{-1}$, $t = \text{UT (days)}$, $\lambda = \text{longitude}$, $\delta = \text{days}/T$, and T is the nontidal wave period in days. As elucidated by Teitelbaum and Vial (1991) (hereafter TV91), this interaction gives rise to two secondary waves (SW) with frequencies and zonal wave numbers that are the sums (+) and differences (–) of those of the tide and PW: $\text{SW}(+) = A(+)\cos[(n + \delta)\Omega t + (s + m)\lambda]$, $\text{SW}(–) = A(–)\cos[(n – \delta)\Omega t + (s – m)\lambda]$.

As described in TV91, when these SWs combine *linearly* with the tide, the tide beats at the period (T) of the primary wave, causing the types of day-to-day variations in various tidal components that are illustrated in Figures 2 and S2.

Now consider the second-stage interaction where the SW interacts *nonlinearly* with the same tidal component that led to its production. It is simple to show using the same (+) and (−) relationships above, and assuming $s = 1$, that SW(+) produces the (+) and (−) (frequency and zonal wave number) pair $[(2n + \delta)\Omega, 2s + 1]$, $[\delta\Omega, 1]$ and SW(−) produces $[(2n - \delta)\Omega, 2s - 1]$, $[\delta\Omega, 1]$; in other words, both interactions produce $[\delta\Omega, 1]$, the original modulating PW. This leads us to propose the following mechanism as a viable means of how a PW-modulated tidal spectrum can induce a PW oscillation in the thermosphere, similar to that depicted in Figure 7. This mechanism rests on the knowledge that SWs arising from PW-tide interactions or UFKW-tide interactions propagate up into the lower thermosphere as independent oscillations and achieve their maxima in the 110- to 130-km height region (Gasperini et al., 2017; Nguyen et al., 2016; Nystrom et al., 2018; Palo et al., 1999; Pedatella et al., 2012). The potential exists, therefore, for a SW arising from, for example, a Q6DW1-tide interaction to propagate into the region above 100 km, interact nonlinearly with the same tide originally involved in its generation, and produce a Q6DW1 in situ in the thermosphere. A notable attribute of this mechanism is that it is capable of producing Q6DW1 in the thermosphere even if the original Q6DW1 does not penetrate to these altitudes.

An examination of wave number frequency spectra at 120 km indicates that several of the SW(+) and SW(−) do exist within the TIE-GCM simulations reported here. The SW(+) and SW(−) can of course interact in the same way with all the other tidal components in the spectrum to produce a myriad of other waves. Furthermore, a convincing proof of our hypothesis requires confirmation that the required SW are robust in their amplitudes, spatial distribution, and phase coherence; and that the involved waves possess temporal variabilities consistent with evolution of the Q6DW1. This type of analysis places us out of the scope of the present paper and is reserved for future work.

Ionospheric Implications. The S0 total mass density response to introduction of a PW-modulated tidal spectrum consists of a response associated with short-time scale (< 10 days) temperature/hydrostatic changes and a mean molecular mass response with longer time constants. This is analogous to the Yamazaki and Richmond (2013) simulations that show the ionosphere to respond quickly to tidal forcing through dynamo-generated electric fields but with a time constant of order 9 days to the net compositional effects induced by tides. Although not shown here, the S0 electron density variations that accompany the rather modest total mass density, temperature, and [O] density variations in Figure 11 range between -10% and $+25\%$ and are even larger if one considers zonal wave numbers with $s \neq 0$. The respective [O]/[N₂] ratio values range between ± 0.8 and -0.5 to $+1.0$, respectively. The ionospheric impacts are therefore substantial, and that is why we have expended some effort to interpret the neutral response at 325 km. In this connection, the S0 wind oscillations in the *E* region may not be very efficient generators of electric fields (e.g., see Gan et al., 2017), but they will generate the above compositional effects and they will produce currents that are measured as S0 magnetic oscillations on the ground. This may explain the S0 oscillations at PW periods in ground magnetic data reported by Elhawary and Forbes (2016). Although not analyzed in detail here, there are substantial $s \neq 0$ PW period oscillations in *E* region winds that will produce *F* region PW oscillations over a range of periods and zonal wave numbers; the enhanced Q6DW1 discussed above is a notable example. The complexity of the ionospheric response demands that we also address this aspect of the problem more fully in future work.

Appendix A: Model Descriptions and Verification

This Appendix provides some details about the model suite employed for the present study and addresses its suitability to inform us about the generation of S0 oscillations in the actual thermosphere. As noted in the text, TIME-GCM/MERRA2009 (Häusler et al., 2014, 2015) is used as the driver for a series of diagnostic TIE-GCM simulations designed here to isolate the effects of vertically propagating tides from in situ drivers (such as UV/EUV heating and geomagnetic disturbances) in terms of determining the zonal mean state of the thermosphere and its variability.

The TIME-GCM (Richmond et al., 1992; Roble, 1995; Roble & Ridley, 1994) is a global grid point time-dependent model extending from a constant pressure level near 30 km altitude to the upper thermosphere (500–700 km, depending on solar activity). Given a solar irradiance spectrum and parameterizations for energy inputs from the magnetosphere, it solves thermal energy, momentum, continuity, and photochemical equations to arrive at a comprehensive simulation of the temperature, circulation, composition, chemical, and electrodynamic

Figure A1. Sixty-day running mean migrating tidal temperature amplitudes at 100-km altitude as a function of latitude and day of year (DOY) during 2009. (left column to right column) Diurnal, semidiurnal, and terdiurnal. (top row) SABER. (bottom row) TIME-GCM/MERRA2009. Vertical dashed lines identify October 2009.

structures of the upper atmosphere and ionosphere. The TIE-GCM (Qian et al., 2014) is similar to the TIME-GCM except that its lower boundary is at a constant pressure level near 97 km. The TIE-GCM is a documented community model; see <http://www.hao.ucar.edu/modeling/tgcm/>.

MERRA is a National Aeronautics and Space Administration reanalysis product (Rienecker et al., 2016) consisting of a physics-based weather prediction model constrained by global data. This type of model is needed to provide realistic tidal and UFKW spectra and day-to-day variability emerging from the troposphere, since these waves are strongly dependent on the global distributions of water vapor and latent heating and on tropical dynamics. MERRA extends from the surface to about 64-km altitude with 42 vertical levels and has a horizontal resolution of 1.25° . It has been validated in the context of numerous studies and shown to outperform other reanalysis products (Lindsay et al., 2014). In particular, Sakazaki et al. (2012) validated the diurnal tide in MERRA. Compared to other publicly available reanalysis/assimilation products that provide global fields on a 6-hourly cadence, the time resolution of MERRA is 3-hourly, which is essential for defining semidiurnal tides.

TIME-GCM/MERRA2009 is a year-long simulation forced by MERRA at the ~ 30 -km lower boundary and solar and geomagnetic variability based on $F_{10.7}$ and Kp indices, respectively. As described in Häusler et al. (2014) and Häusler et al. (2015), in order to capture the waves of interest TIME-GCM/MERRA2009 employs a $2.5^\circ \times 2.5^\circ$ latitude-longitude grid, four grid points per scale height in the vertical direction, and 60-s time step. The TIE-GCM extends from a constant pressure level at 5×10^7 hPa (nominally ~ 97 km) to 450 to 600 km depending on level of solar activity. It uses the same horizontal and vertical resolutions as TIME-GCM/MERRA2009. Both TIME-GCM and TIE-GCM employ the EUV flux model for aeronomics calculations model (Richards et al., 1994)

Figure A2. Sixty-day running mean nonmigrating tidal temperature amplitudes at 100-km altitude as a function of zonal wave number and day of year (DOY) during 2009. Negative (positive) zonal wave numbers correspond to eastward (westward) propagation. Tides from TIME-GCM/MERRA2009 (SABER) at (top two rows) -27.5° (-25°) latitude and at (bottom two rows) $+27.5^\circ$ ($+25^\circ$).

based on daily $F_{10.7}$ values used to provide spectral irradiances, and Kp-dependent parameterizations are used to specify auroral particle precipitation (Emery et al., 2012; Roble & Ridley, 1987) and electric potential patterns (Weimer, 2005).

Accepting that the TIE-GCM physics is correct, assessment of the above model suite mainly involves verifying the appropriateness of the lower boundary conditions near 97-km altitude provided by TIME-GCM/MERRA2009. For the problem at hand, key points are that the TIME-GCM/MERRA2009 tidal spectrum should reasonably approximate the observed spectrum in terms of tidal components present, their amplitudes, and their levels of variability at PW periods. We use diurnal, semidiurnal, and terdiurnal tidal temperatures derived from measurements made by the SABER instrument on the TIMED spacecraft during 2009 to provide some initial insights. The tidal temperatures are derived in a manner identical to that of Truskowski et al. (2014) and consist of daily values obtained by fitting data within 60-day windows moved forward 1 day at a time between $\pm 50^\circ$ latitude. The 60-day window is necessitated by the local time sampling afforded by the TIMED satellite orbit, and the latitude range conforms to the geographical limits of continuous SABER data from one yaw cycle to the next.

Figure A3. Same as Figure A2 except for semidiurnal tides.

Comparisons between latitude versus (DOY) TIMED/SABER 60-day mean tidal temperature amplitude structures and those from TIME-GCM/MERRA2009 for the migrating components DW1, SW2, and TW3 are provided in Figure A1. Wave number versus DOY amplitude spectra for the nonmigrating tides at several latitudes is provided in supporting information Figures S1 (TIMED/SABER) and S2 (TIME-GCM/MERRA2009); a few representative results from those files are shown here as Figures A2 and A3. Keeping in mind that the displayed tides are 60-day running means, so that the structures are broadened in time and reduced in amplitude compared to daily values (Häusler et al., 2015), inspection of Figures A1, S1, and S2 provides the following information: In terms of the tidal spectrum, the TIMED/SABER data for 100 km during 2009 reveal the presence of the following tidal components: DE1-DE3, D0, DW1-DW4, S0, SE1-SE3, SW1-SW4, TE1-TE3, TW1-TW5, and TW7. TIME-GCM/MERRA2009 contains the same set with the addition of DW5, SW5-SW6, and TE4-TE5; however, these components lie at the tail end of the amplitude spectrum and do not represent a significant departure from that of TIMED/SABER. In terms of amplitudes, the most prominent tides in the TIMED/SABER spectrum at 100 km occur with the following amplitude maxima during the course of 2009: DE3(6-10K), DE1,DE2(4-5K), D0(3K), DW1(10-20K), DW2(4-8K), DW3(4K), SE1-SE3(3K), SW1,SW3(4-6K), SW2(8-18K), and TW3(4-8K), which can be compared with DE3(4-7K), DE1-DE2(3-4K), D0(3K), DW1(15-25K), DW2(7-10K), DW3(2-4K), SE1(3K), SE2(4-6K), SW1,SW3(4-6K), SW2(10-20K), and TW3(3-7K), for TIME-GCM/MERRA2009. The 60-day mean TIME-GCM/MERRA2009 tidal components are thus consistent with those present in the TIMED/SABER spectrum in terms of their periods and zonal wave numbers, and the nominal maximum amplitudes that occur during the course of 2009.

Examination of Figures A1–A3, S1, and S2 reveals that the intraannual variabilities of the above tidal components in TIME-GCM/MERRA2009 share some similarities as well as some differences compared to TIMED/SABER during 2009. However, whether consistency in this regard is crucial depends upon the science question that is posed. In the present context, we simply seek to quantify what effects might be produced by a dissipating tidal spectrum that is modulated at PW periods, and specifically whether S0 oscillations of some significance would result. Above we established that TIME-GCM/MERRA2009 is consistent with observations in terms of the periods and zonal wave numbers of the tides involved, and their overall 60-day mean magnitudes. Figure 2 in the text and Supporting information Figure S1 provides clear evidence that tidal variability at periods of order 2–20 days is a distinct characteristic of TIME-GCM/MERRA2009. The displayed nominal 25–50% level of tidal variability over these timescales (as well as more extreme jumps) is consistent with existing MLT measurements at equatorial to middle latitudes (e.g., Forbes et al., 2017; Guharay & Franke, 2011; Gurubaran et al., 2009; Nguyen & Palo, 2013; Pancheva, 2001; She et al., 2004; Vineeth et al., 2011; Yuan et al., 2013), although ambiguities in terms of identifying zonal wave numbers of individual tidal components on a day-by-day basis usually exist in both space-based and ground-based measurements. With the given available evidence, we conclude that TIME-GCM/MERRA2009 is an appropriate vehicle to pursue our specific stated objectives.

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