Interpretation of the Altitudinal Variation in the Martian Ionosphere Longitudinal Wave-3 Structure

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Abstract Using the electron density profiles from Mars Global Surveyor Radio Occultation observations, the ionospheric longitudinal wave-3 (WN3) structure with different frequencies is investigated. The semidiurnal component is the dominant component for the entire Mars ionosphere profile at northern high latitudes. Both diurnal and semidiurnal components manifest double peaks in altitudinal variations in amplitude, which are 5–15 km lower than the ionospheric M1 and M2 peaks. The phase generally increases with altitude and shows a rapid rate of increase around the M2 peak. To interpret the altitudinal variation in amplitude and phase, we then derive the tidal coupling equation to elucidate the impact of atmospheric tides on the Mars ionosphere. Under the assumption of photochemical equilibrium, the tidal coupling equation indicates that the log density gradient of electrons (k_N) critically modulates the ionospheric response to atmospheric tides. In the observations, a positive correlation between the k_N profile and the amplitude of the WN3 structure at altitude is found. The solar longitude variation in the k_N peaks also agrees with the vertical movement of the amplitude peaks of the WN3 structure along the solar longitude. The rapid phase increase corresponds to the abrupt decrease in k_N around the M2 peak. These observational results agree with the tidal coupling equation prediction, thereby supporting that the interpretation of k_N in modulating the altitudinal variation in the WN3 structure is reasonable.

1. Introduction

Solar heating on the Mars topography produces nonmigrating atmospheric tides that propagate into the upper atmosphere (Forbes & Hagan, 2000; Forbes et al., 2004; Wilson, 2002; Withers et al., 2003), which in turn creates the longitudinal wave structure in the Mars ionosphere (Bougher et al., 2001, 2004; Cahoy et al., 2007; Mendillo et al., 2017). Accelerometer measurements onboard the Mars Global Surveyor (MGS) reveal that the longitudinal neutral density structure near 125 km consists of longitudinal wave-1 (WN1) and wave-2 (WN2) components (Keating et al., 1998). In both the Mars Atmospheric and Volatile Evolution mission (MAVEN) and Mars Climate Database (MCD) results, wave-2 (WN2) and wave-3 (WN3) components in neutral densities of the upper atmosphere are found (England et al., 2016, 2019; Liu et al., 2017; Lo et al., 2015). In the ionosphere, the most prominent examined patterns are wave-1 to WN3 components (Haider & Mahajan, 2014; Haider et al., 2011). The ionospheric WN1 component is considered to be caused by crustal magnetic fields, so it generally occurs more frequently in the Southern Hemisphere and above ~200 km (Flynn et al., 2017). Observations have revealed that dominant contributors to the ionospheric WN2 component are nonmigrating DE1 tides at the equator and S0 tides at the midlatitudes. For the ionospheric WN3 component, DE2 tides at low latitudes and SE1 tides at midlatitudes are more important (Thaller et al., 2020). Forbes et al. (2020) summarized the main features of those nonmigrating tides and stressed their importance on the ionospheric longitudinal structure.

The characteristics of the ionospheric longitudinal structure are worthy of investigation. It is helpful to deeply understand how atmospheric waves propagate into the upper atmosphere and affect the ionosphere. The theory of photochemical equilibrium (known as the Chapman theory) is a widely accepted view describing atmosphere-ionosphere coupling on Mars. Below ~200 km, photoproduction and recombination between ionospheric electrons and ions rapidly achieve balance. Chapman theory motivates researchers to estimate neutral atmospheric density from ionospheric observations (Zou et al., 2011, 2016). On the other hand, the Chapman theory associated with the effects of tidal-induced vertical displacement (Cahoy et al., 2006) can predict that ionospheric density...
perturbations account for approximately 50% of atmospheric density changes (Fang et al., 2020; Thaller et al., 2020). Electron density profiles from MGS radio occultation measurements revealed that the primary component of the ionospheric longitudinal structure, the WN3 structure, has a primary peak at ~125 ± 10 km and a secondary peak in amplitude at ~105 km (Cahoy et al., 2007). The heights of these two peaks are lower (~5–15 km) than the primary peak (M2) and secondary peak (M1) of the Mars ionosphere electron density profiles (Figure 1). The altitudinal variation in the nonmigrating tide at neutral density (Forbes et al., 2020) does not manifest similar features. Cahoy et al. (2006) attempted to explain it using a simplified model of ionospheric vertical displacement, but the setting of displacement height relied on posterior information. Hence, no detailed investigation, to our knowledge, has accurately interpreted such features of altitudinal variation, and it is still a mystery in the present stage.

In this work, we focus on explaining the observed altitudinal variation in the longitudinal WN3 structure in the Mars ionosphere. Using electron density profiles observed by the MGS, the longitudinal WN3 structure is extracted from 90 to 210 km and covers summer and autumn in Mars years 25–27. Based on the governing equation of the ionosphere, we derive a theoretical tidal coupling equation to explain how the neutral atmosphere tide interacts with the Mars ionosphere. The key factor in modulating the altitudinal variation is discussed. A brief summary of the interpretation is given at the end.

2. Data Processing

Electron density profiles measured from the Mars Global Surveyor (MGS) Radio Occultation (RO, Hinson et al., 1999) are used to examine the ionospheric response to nonmigrating atmospheric tides on Mars. Figure 1 illustrates the 5,305 electron density profiles used, which cover Mars years 23–25. The data in Mars year 22 are small, so we do not analyze them in that period. Here, the profiles clearly show a two-layer structure, that is, the M1 peak at ~135 km and M2 peak at ~110 km. Due to the quasisolar synchronous orbit, the measurements perform a rapid scan for a full range of longitude, which is about once every Martain day. At the same time, the data cover half-day local time from early morning to afternoon and a half-year season (summer and autumn, indicated by the Martain solar longitude [Ls]). The data points cover small ranges of large solar zenith angles (SZAs; 70.9°–89.2°) at northern high latitudes (60.6°–85.5°). The limitation of MGS/RO latitudinal coverage causes our analysis to exclude the low- and mid-latitude regions, where Kelvin modes are important (Forbes & Hagan, 2000; Hinson et al., 2008).

To extract ionospheric perturbations responding to nonmigrating atmospheric tides, the electron density is divided into two parts in the first step:

\[ N = N_0 + N_1 \]  

where \( N_0 \) is the running average of \( N \) within a cycle of longitude, and the residual, \( N_1 \), correspondingly indicates the longitudinal fluctuations for the cycle. The running averaged is taken within ±180° around each data point on the nearest Martian day. If the number of data points is less than 10, we will extend the search scope to more Martain days to ensure that number of data points in each cycle is much than 10. The normalized deviation, \( N_1 / N_0 \), clearly shows the longitudinal WN3 structure (top row of Figure 2). The longitudes of the wave crests/valleys change with the \( L_s \), which is coincident with local times (the black lines) and implies a semidiurnal variation. The value of \( N_1 / N_0 \) is then spectrally analyzed in a local time frame by fitting using Equation 2.

\[ N_1 / N_0 = \sum_{f,k} A(f,k)e^{i2\pi(f-L_s)} \]
$f (=0, \pm 1, \pm 2, \ldots)$ and $k (=1, 2, 3, \ldots)$ are the frequency and zonal wavenumber of the spectra in ionospheric perturbations in the local time frame. A positive (negative) frequency denotes that the direction of the apparent wave signature is eastward (westward), which is subject to the eastward procession of MGS’s orbit. $t$ and $\lambda$ are the local time divided by 24 hr and longitudes divided by 360°, respectively. $A_f$ is the complex amplitude of the ionospheric perturbations with a given frequency and zonal wavenumber, which is hereafter called the “ionospheric tide.” In the following, the nomination of “ionospheric tide” will always be in the local time frame, but referring to atmospheric tides, we still use the traditional method to name them in the universal time frame. The spectrum of the ionospheric tide (Figure 3) shows that the WN3 at frequency of 2 cycle per day (2 cpd, $f = 2$, $k = 3$) is the dominant component throughout the M1 layer, M2 layer, and topside ionosphere. Here, the terdiurnal and other higher-frequency components are not given because the local time coverage is only $\sim$12 hr (bottom row of Figure 2). The contributions from 1-cpd WN3 and 0-cpd WN3 (I-SPW3) are secondary contributions with similar strengths. 1-cpd WN3 is slightly stronger in the topside ionosphere and slightly weaker in the M1 and M2 layers than the 0-cpd WN3. Due to intrinsic ambiguities, a single-spacecraft mission in a quasi-sun-synchronous orbit cannot distinguish a given wave with all their potential secondary waves of the nonlinear interactions between the
given wave and all migrating tidal components. An example is the well-known zonal wavenumber-4 pattern in the Earth’s atmosphere and ionosphere. The stationary planetary wave with zonal wavenumber 4 (sPW4) appears as a zonal wavenumber-4 pattern, so do SE2 (=SW2 × sPW4*), DE3 (=DW1 × sPW4*), DW5 (=DW1 × sPW4, see Figure 1 in He et al., 2011). As a single-spacecraft mission, MGS cannot distinguish instantly different tidal contributors to the WN3 structure. However, assuming that the main contributing tides are time-invariant on the time scale on which the spacecraft covers all local time (LT) sectors, one can identify the main contributors according to the local time variation. Combining observations from all local-time sectors allows identifying the frequency of the main contributors in the Mars-fixed coordinate system. The tidal components DE2 and SE1 appear as the 1-cpd and 2-cpd WN3, while the stationary planetary wave with zonal wavenumber 3 (sPW3) appears as the 0-cpd WN3.

3. Tidal Variability

Based on the spectral analysis in Section 2, the altitudinal variation in 2-cpd WN3 for different seasons is displayed in Figure 4, which is for northern high latitudes (∼60°–85°). Double peaks in altitudes are revealed at ∼120–130 km and 95–105 km. The maximum strength of 2-cpd WN3 around peaks is greater than ∼20%. In Mars Year (MY) 25, the upper peak slowly descends from 70°Ls to 100°Ls and then lifts after 120°Ls. A similar uplift from 120°Ls to 150°Ls is displayed in MY26. However, such up-down movement of the upper peak is not clear in MY27. The lower peak is lower in early autumn (∼140°–160°Ls) and slightly higher in summer (∼80°–100°Ls) and late autumn (∼180°–210°Ls). The strength of 2-cpd WN3 around the upper peak decreases from early summer (∼70° Ls) to autumn (∼180°Ls) and deeply depresses between summer and autumn (∼120° Ls). The lower peak is of similar magnitude to the upper peak and manifests as a depression at ∼120°Ls in MY26, although it has some discrepancies. For example, the strength increases during late autumn in MY27 (∼180°Ls–210°Ls). Such seasonal variation is also manifested in the atmospheric SE1 tides at latitudes of ∼60°–70°N (Forbes et al., 2020). The consistency in seasonal variation confirms the tidal coupling to the ionosphere on Mars. In addition, enhancement in the topside Martian ionosphere (>170 km), where the time scale of vertical diffusion becomes much more rapid as the collisional rate decreases (Thaller et al., 2020), is mainly found during the summer and early autumn. This enhancement is consistent with the MAVEN results by Fang et al. (2020).

The seasonal variation presented in Figure 4 should also contain the influence of latitudinal variability, changes in solar zenith angle, and slow processions in local time. Variations in these factors are further examined in Figure 5, which plots the latitudinal variation in 2-cpd WN3 amplitudes with labeled solar zenith angle and local time. The colored scatter of points indicates that the strength of 2-cpd WN3 varies with Ls around the upper peak (∼124 km). In the summer of MY25 and MY26, the amplitude of 2-cpd WN3 decreases slightly with latitudes approaching the northern pole, and the local time changes slowly. Inspection of the period of 150°–180°Ls reveals that the latitude decreases while the 2-cpd WN3 weakens. In the early autumn of MY25 and MY26, the solar zenith angle and latitude do not change rapidly, but the local time changes from 5 to 8. The strength of 2-cpd WN3
increases in MY25 but does not change significantly in MY26. In MY26, 2-cpd WN3 becomes strong at the local times of 8–12, which is consistent with the variation in MY25 for the same $L_S$. This indicates that the main factor affecting the amplitude change in that period is the $L_S$. In late autumn of MY27, the solar zenith angle increases from 76° to 86°, while 2-cpd WN3 apparently does not change. The results suggest that the dependence of local time and solar zenith angle is not clearly shown in this work and could be further investigated by other observations or model simulations. The results of Figure 5 indicate that the variations in local time, solar zenith angle, and latitude contribute little to the seasonal variation in 2-cpd WN3, as shown in Figure 4.

The phase of the extracted 2-cpd WN3 is illustrated in Figure 6, which increases with altitudes of 100–110 km and 120–200 km. At 110–115 km, which is around the M1 peak, the 2-cpd WN3 phase has a small peak. Notably, the 2-cpd WN3 phase has a sudden increase at 130–145 km. With the local time varying from 3 to 14, the 2-cpd WN3 phase increases, but the main feature of altitudinal variation discussed above remains almost the same. The corresponding 2-cpd WN3 amplitudes are plotted in color with the phase variation in height. In the region in which the 2-cpd WN3 phase rapidly increases, the 2-cpd WN3 amplitude is relatively small. The 2-cpd WN3 phase

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**Figure 5.** Amplitudes of the ionospheric longitudinal WN3 semidiurnal component varying with $L_S$ and latitude at 124 km. The solar zenith angle (SZA) and the local times (LT) are correspondingly labeled along the data points. From top to bottom, the panels are MY25–27.

**Figure 6.** Altitudinal variation in the phase of the 2 cpd WN3 for different local times. The strengths of amplitudes are indicated in color, for which the color bar is labeled on the right.
is quite stable at altitudes with large 2-cpd WN3 amplitudes. The increased phase in altitude should correspond to the downward propagation of the atmospheric tide.

4. Interpretation of the Altitudinal Variation

We have discussed the main features of altitudinal variations in the ionospheric longitudinal WN3 structure in Section 3. In this section, we derive the tidal coupling equation first and then use it to interpret the underlying reason for the altitude changes in both amplitude and phase.

Under photochemical equilibrium, variations in electron density are determined by the divergence of the density flux. This can be characterized by the following continuous equation:

$$ \frac{\partial N_i}{\partial t} = -\nabla \cdot N_0 V_e = -U_z (dN_0/dz) - N_0 \nabla \cdot U $$

(3)

where $V_e$ and $U$ are the electron velocity and the neutral wind velocity, respectively. These two velocities are equal in the collision-dominant region. Here, the horizontal convective term, $U_H \cdot (\nabla H N_0)$, is neglected in the first term of the right side of the equation because the horizontal gradient of electron density is much smaller than the vertical gradient. We then derive the following coupling equation from Equation 3 by substituting the spectral analysis in Equation 2,

$$ i2\pi f A_T = -(k_N + ik_P) \cdot A_T $$

(4)

where we have used the definition as follows,

$$ U_z = A_T e^{2\pi(f-k_N)} $$

(5a)

$$ k_P = i\nabla \cdot U / U_z $$

(5b)

$$ k_N = (1/N_0)(dN_0/dz) $$

(5c)

$A_T$ is the complex amplitude of atmospheric tides in vertical wind (Equation 5a). $k_P$ is the polarization relationship between the vertical wind and the horizontal wind, which is defined as Equation 5b. $k_N$ is the term of the log density gradient, expressed as Equation 5c. The amplitude and phase of the ionospheric tide can be expressed as

$$ A_T^2 = \left[ (k_N + \Re (k_P))^2 + (\Im (k_P))^2 \right] |A_T|^2 / 4\pi^2 f^2 $$

(6a)

$$ \varphi = \tan^{-1} \frac{\Im (k_P)}{k_N + \Re (k_P)} - k_z z - \text{Const} $$

(6b)

where $\Re (k_P)$ and $\Im (k_P)$ are the real and imaginary parts of $k_P$, respectively, and $k_z$ is the vertical wavenumber of the atmospheric tide in the vertical wind. The polarization relationship, $k_P$, is relatively small in general because the wind velocity is smaller than the acoustic speed, so the fluid (gas) is approximately incompressible and the divergence of neutral wind is close to zero.

Altitudinal variations in the log density gradient, $k_N$, should effectively modulate the amplitude of the ionospheric tide, as Equation 6a predicts. Examination of the MGS/RO observations shows that the double peaks of $k_N$ are coincident around the peaks of the ionospheric WN3 structure (Figures 7a and 7b) for both 1 and 2 cpd components. This good coincidence indicates the strong modulation of $k_N$ on $A_T$ at different altitude regions: (a) at steep slopes below the M1 peak, a $k_N$ crest leads to the $A_T$ crest; (b) around the M1 peak, a $k_N$ valley produces the $A_T$ valley; and (c) at steep slopes below the M2 peak, a $k_N$ crest brings the $A_T$ crest. The electron density profile is given as a reference in Figure 7c, which obviously shows that the two peaks of the ionospheric WN3 structure are lower than the ionospheric peaks. As the secondary contributor, the 1-cpd WN3 component has almost the same altitudinal variation as 2-cpd WN3, although vertical structures of atmospheric DE2 and SE1 tides on Mars generally have large differences due to their different vertical wavelengths (Forbes et al., 2020). It also indicates the crucial role of $k_N$ on the altitudinal variation in the ionospheric WN3 structure. In addition, the atmospheric tidal amplitudes $A_T$ should also determine the strength of the ionospheric tide. On the one hand, 2-cpd WN3 is stronger than 1-cpd WN3 (Figure 5a), which is related to the atmospheric SE1 tide being larger than the DE2
tide at latitudes of $\sim60^\circ$–$70^\circ$N (Forbes et al., 2020). On the other hand, the lower peak of $A_N$ is larger than the higher peak, but for $A_T$, it is inversely modulated, which should also be modulated by the increased amplitude of $A_T$ with height. The seasonal variability in 2-cpd WN3 discussed in Figure 4 is also consistent with that of the atmospheric SE1 tide.

Furthermore, the correlation between the amplitude of 2-cpd WN3 and $k_N$ is examined in different seasons. As discussed in Section 3, the peak altitude of 2-cpd WN3 has slight vertical movement for different $L_S$. The $L_S$ variation in the $k_N$ profile manifests good agreement with the movement (Figure 8). The upper peak of $k_N$ ($\sim120$–$130$ km) descends slowly from $80^\circ L_S$ to $120^\circ L_S$ and then uplifts gradually from $120^\circ L_S$ to $160^\circ L_S$ in MY25. In MY26, the upper peak of $k_N$ is at a stable altitude before $120^\circ L_S$ and slightly changes in the period of $120^\circ$–$180^\circ L_S$. For MY27, an increase in the $k_N$ upper peak height is found during $120^\circ$–$150^\circ L_S$. These features are all coincident with those in 2-cpd WN3. The variation in $k_N$ at lower peak heights also agrees with the corresponding change in 2-cpd WN3. For example, the uplift of the lower peak during late autumn in MY27 is found in both the $k_N$ and amplitude of 2-cpd WN3.

The log density gradient $k_N$ also modulates the phase change at altitude. Equation 6b suggests that $k_N$ should be inversely proportional to the phase of the ionospheric tide, which is verified by Figures 9a and 9b. In the region of the phase rapidly increasing near the M2 peak, $k_N$ decreases with altitude with a steep slope. When the height of

![Figure 7](image_url)

Figure 7. (a) Altitude profiles of $A_I$ in the longitudinal WN3 at frequency of 2 cpd (red solid line) and 1 cpd (red dashed line). (b) and (c) are the altitude profiles of the mean log density gradient ($k_N$) and the mean electron density ($N_0$), respectively.

![Figure 8](image_url)

Figure 8. Altitudinal variation in $k_N$ (white lines) in the different $L_S$. 

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changes from positive to negative, the phases of 2-cpd WN3 and 1-cpd WN3 also pass through the point of $\pi$. The inflection point of the phase at approximately 110–120 km corresponds to the valley of $k_N$ at almost the same altitude. The relatively slow phase change in the upper ionosphere (>150 km) shows agreement with the quite stable $k_N$. This modulation is valid for both 1-cpd WN3 and 2-cpd WN3 (Figure 9a); nevertheless, atmospheric DE2 and SE2 tides have different vertical wavelengths, $k_z$.

5. Conclusion

Using the MGS/RO electron density profiles, the ionospheric longitudinal wave structure response to nonmigrating tides on Mars is derived for different altitudes. The altitudinal variation in the ionospheric longitudinal wave structure does not follow that of the electron density profiles. The maximum amplitudes occur 5–15 km lower than the M1 and M2 peaks. The phase of the ionospheric longitudinal wave structure rapidly increases with altitude around the M2 peak and has an inflection point around the M1 peak. Such altitudinal variation is then examined by a tidal coupling equation, which is derived theoretically from the ionospheric governing equation. A reasonable explanation for the altitudinal variation in amplitude and phase is then given. The log density gradient of electrons ($k_N$) modulates the influence of nonmigrating atmospheric tides on the ionosphere and plays an important role in determining the altitudinal variation in the WN3 structure. The main results of this work are summarized as follows:

1. Double peaks at altitudes of $\sim$120–130 km and $\sim$95–105 km are found in the longitudinal WN3 structure in the Mars ionosphere, and the semidiurnal component, the dominant component, could achieve amplitudes greater than $\sim$20%.
2. The seasonal variation shows that the amplitude of the semidiurnal component around the upper peak decreases from early summer ($\sim$70°Ls) to autumn ($\sim$180°Ls) with deep depressions between summer and autumn ($\sim$120°Ls), and the lower peak is similar to the upper peak, except for an enhancement that is present during late autumn in MY27 ($\sim$180°Ls–210°Ls).
3. The theoretical tidal coupling equation implies that the ionospheric response to atmospheric tides is determined by the log density gradient, atmospheric tides in vertical wind, and polarization relationship of neutral wind.
4. MGS/RO observations reveal that the amplitude of the ionospheric WN3 structure achieves a maximum value at coincident altitudes of $k_N$ peaks, and the $L_s$ variation in the $k_N$ peak height is correlated with the ionospheric WN3 structure.
5. The rapid increase in the phase of the ionospheric WN3 structure at $\sim$130–145 km corresponds to a decrease in $k_N$ with a steep slope.

Figure 9. (a) Altitudinal variation in the ionospheric phase in the 2 cpd (red solid line) and 1 cpd (red dashed line) WN3 components. Subplot (b) shows the altitude profiles of the opposite mean log density gradient ($-k_N$).
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Data Availability Statement
The MGS RO data are publicly available and can be downloaded from the website of https://atmos.nmsu.edu/PDS/data/mors_1102/eds/.

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