Spectral Analysis of Gravity Waves in the Martian Thermosphere during Low Solar Activity Based on MAVEN/NGIMS Observations

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Abstract

Gravity waves (GWs) are important for vertical coupling between the lower and upper atmosphere on Mars. Saturated GWs will promote the acceleration of mean flow and turbulence generation, causing diffuse transport of energy and momentum. We analyze the spectral characteristics of GWs in the thermosphere on Mars during the low solar activity of Martian Year (MY) 35, based on CO\(_2\) density measured by the Neutral Gas and Ion Mass Spectrometer (NGIMS) on board Mars Atmosphere and Volatile EvolutioN (MAVEN). The results are compared with those of MY33 at the higher solar activity. The vertical wavenumber spectral density obtained in the Martian thermosphere is similar to the semi-empirically predicted saturation spectrum in the Earth’s atmosphere, with a logarithmic spectral slope around \(-3\). The average spectral density of GWs shows saturation between 160 and 200 km and decreases with increasing altitude. Compared to MY33, GW activity is larger and less attenuated in MY35 as it propagates upward, implying that waves experience more favorable propagation conditions during low solar activity. Also, the dependence of GW activity on local time during this period is found to be different from previous studies, with the strongest GWs in the spring of MY35 occurring within 6–12 hr on the dayside, suggesting that the amplitude of the GWs in the Martian thermosphere during low solar activity is less controlled by the background temperature and is subject to dissipative effects of other factors, such as molecular viscosity and thermal conduction.

Unified Astronomy Thesaurus concepts: Mars (1007); Internal waves (819); Thermosphere (1694); Solar activity (1475)

1. Introduction

Gravity waves (GWs) are small-scale perturbations with the nature of vertical propagation that use buoyancy as a restoring force and are ubiquitous in the stratified atmospheres of planets such as Earth, Mars, Jupiter, and Titan (Young et al. 2005; Fritts et al. 2006; Müller-Wodarg et al. 2006; Medvedev & Yiğit 2019). GWs on Earth are typically generated in the lower atmosphere by a variety of sources such as convection, fronts, jet streams, and topography (Pickersgill & Hunt 1979; Spiga et al. 2013; Imamura et al. 2016). GW amplitudes will grow exponentially as they propagate upward eventually breaking up via convective or shear instability, and will release momentum and energy into the background atmosphere (Fritts & Alexander 2003; Yiğit et al. 2016). These are thought to bring far-reaching thermal and dynamical effects to atmospheric circulation (Lindzen 1981; Palmer et al. 1986). GWs in the lower atmosphere of Mars play a similarly important role in the upper atmosphere as they do on Earth (Yiğit et al. 2021a). The Mars General Circulation Model (MGCM) reveals the direct influence of GWs on the large-scale circulation on Mars, and the various observations provide important insights into the structure of GWs. GWs generated in the lower atmosphere not only affect the spatial and temporal variation of parameters such as density and temperature during their vertical upward propagation process but also modulate the motion of the neutral component (Fritts & Alexander 2003; Parish et al. 2009; Medvedev et al. 2016). The mean flow in the upper thermosphere will be modified by the momentum transport processes of the GWs and the local atmosphere will be heated or cooled, leading to a change in the energy budget of Mars (Medvedev & Yiğit 2012; Kuroda et al. 2015). Temperature perturbations caused by GWs contribute to the formation of carbon dioxide ice clouds in the Martian middle atmosphere (Spiga et al. 2012; Yiğit et al. 2015), which becomes an important issue affecting the aerobraking phase of the Mars missions. There are also model studies showing that the heating effect of GWs on the upper atmosphere contributes to the occurrence of atmospheric escape (Yiğit et al. 2021a). In order to further understand the Martian atmospheric environment and its evolutionary mechanisms, it is particularly important to investigate the GW dynamics characteristics of the Martian atmosphere.

The GWs propagate to a certain height when instability occurs and the wave’s energy is lost. The amplitude no longer increases, limited by the turbulence, at which time the waves are considered to be saturated (Lindzen 1981; Fritts & Alexander 2003). The semiempirical formula for the vertical wavenumber spectra of saturated waves is based on the assumption that these waves are saturated due to convective instability (Smith et al. 1987; Tsuda et al. 1991). The vertical wavenumber spectra of GW disturbances in different regions of the Earth’s atmosphere have been obtained by using radars, rocket smoke trail, and lidar observation, and the spectral characteristics show some information about the sources and dissipation processes of the GWs. These observational studies have shown that spectral shapes of GWs have a considerable degree of universality, and the spectra follow a logarithmic slope of \(-3\), which is consistent with the expectations of theoretical predictions (Smith et al. 1987; Tsuda et al. 1991).
The vertical wavenumber spectra of GWs on Mars have also been partially studied by related scholars in the past. Ando et al. (2012) studied the vertical wavenumber spectra of the temperature profiles in the lower atmosphere of Mars with radio occultation measurements from the Mars Global Surveyor (MGS). They found that the universal spectra of the Earth’s saturated GWs are also applicable to the Martian atmosphere. The spectral density decreases with wavenumber in the vertical wavelength range of 2.5 km to 15 km, and the power-law spectral index is generally around —3, these results are similar to those obtained in the Earth’s stratosphere and mesosphere. They also found similar spectral features within the 65–80 km altitude of the Venusian atmosphere, with the wavenumber spectra tending to follow the semiempirical spectrum of saturated GWs, suggesting that the GWs are dissipated by saturation as well as by radiative damping (Ando et al. 2015). Recently, Nakagawa et al. (2020) analyzed the vertical propagation of wave perturbations in the Martian middle atmosphere using temperature profiles observed by the Imaging Ultraviolet Spectrograph (IUVS) on board the Mars Atmosphere and Volatile EvolutioN (MAVEN). They found that the vertical wavenumber spectrum of waves in the Martian middle atmosphere shows similar characteristics to Earth’s atmosphere. Longer waves grow with height, while CO₂ radiative cooling can effectively dissipate shorter waves. But their results included the combined effects of GWs and thermal tides because of the limitations of the Fourier analysis method.

The spectral characteristics of GWs in the Martian thermosphere have not been studied yet, but the accurate construction of the MGCM requires realistic parametric information on GWs, including dynamical processes such as momentum transport and turbulent diffusion due to GW breaking. In addition to the internal waves from below, solar wind forcing from above is another important factor that may be involved in the atmospheric vertical coupling process (Yiğit et al. 2015, 2016). Solar activity alters the density and temperature of electrons and ions through ultraviolet radiation, thus producing large-scale dynamical and thermodynamic effects on the Martian thermosphere (Yiğit 2018). Therefore, it is more advantageous to study the spectral characteristics of GWs in the Martian thermosphere during periods of low solar activity, when solar variations have minimal effects on the upper atmosphere, and the characteristics of waves propagating upward from below are more direct. A clearer understanding of the dynamical properties such as wave source characteristics and dissipation processes of GWs can be obtained to guide the improvement of GW parameterization schemes.

The present analysis attempts to provide a clear idea of the vertical wavenumber spectrum characteristics of GWs in the Martian thermosphere for the first time, based on Neutral Gas and Ion Mass Spectrometer on board Mars Atmosphere and Volatile EvolutioN (MAVEN/NGIMS) observations. The overall structure of the article is organized as follows. Section 2 presents the derived data of NGIMS, the derivation method of the temperature profile, and the extraction of GW perturbations. The vertical wavenumber spectral features are given in Section 3, the discussion is presented in Section 4, and the conclusion and summary are included in Section 5.

2. Data and Methods

2.1. MAVEN/NGIMS Data and Coverage

The MAVEN spacecraft has been measuring the upper atmosphere of Mars since its launch in 2014 October. It operates in an eccentric orbit with a period of 4.5 hr, with a nominal periapsis altitude of 150 km (Jakosky et al. 2015). The NGIMS instrument on board MAVEN is a quadrupole mass spectrometer designed to measure the abundance of neutrals and ions in the upper atmosphere of Mars below 500 km, with unit mass resolution (Mahaffy et al. 2015). To analyze the spectral characteristics of GWs in the Martian thermosphere at solar minimum, we use CO₂ densities measured in situ by the NGIMS instrument, and only focus on the altitude range of 160–220 km, because the wave perturbations become less significant at higher altitudes. The density data below 220 km have been widely used in previous studies and have been shown to be reliable (Manju & Mridula 2021; Rao et al. 2021).

In addition, only the data of the inbound section of each orbit were used, as mentioned by Stone et al. (2018), the outbound data may be affected by instruments. The present study uses the derived data record of NGIMS products: Level 2, version 08, and revision 01. The vertical variations of observations are dominant in the altitude range considered in this study. As suggested by Manju & Mridula (2021), the vertical variation of density in the 160–220 km altitude is much larger than the horizontal variation within the pass. Siddle et al. (2019) also used temperature and density profiles from NGIMS to extract perturbations and interpret them as vertically propagating gravity waves. Thus, we are essentially studying the vertical propagation properties of GWs.

To properly select the period of low solar activity, we examined the number of sunspots as a function of time, which has often been used previously to indicate the intensity of solar activity (Kiess et al. 2014; Yiğit et al. 2021b). Figure 1(a) shows the time series of sunspot number Sₙ for the last 13 yr since the 24th solar cycle in 2008, containing the daily average (orange), monthly average (blue), and monthly smoothed average (red). The variation of the sunspot number with the corresponding Mars time expressed by the solar longitude is shown in Figure 1(b). The sunspot data are from the SILSO World Data Center of the Royal Observatory of Belgium. We find that MAVEN has been operating in the declining phase of the solar cycle since 2014. To analyze the GW spectrum characteristics during low values of solar activity, we used NGIMS measurements during the most recent solar minimum from 2019 March 24 to 2020 February 25, corresponding to Lₛ = 0–180° for Martian Year (MY) 35, with a very low average sunspot number. Also, the first half of MY33 was chosen for comparative analysis, the first Martian year in which MAVEN acquired a full year of observations since its operation, during which solar activity was relatively more active.

Figure 2 shows the NGIMS data spatiotemporal coverage of MY33 and MY35 used in this study, during which MAVEN has comparable latitude and local time coverage. The orbital coverage of MY33 expressed in altitude–latitude is shown in Figure 2(a), and it can be seen that the spring orbits are mostly scanned in the southern hemisphere, while the summer orbits are concentrated in the northern hemisphere. A similar distribution can be seen in Figure 2(b). In contrast, the orbital coverage of MY35 is more evenly distributed in latitude, with data obtained in the latitudinal range between 75°N and 75°S and longitude coverage is almost global, both in spring and summer. Figures 2(c) and (d) depict the latitude of periapsis as the function of local time and solar longitude. In terms of local time coverage, the sampled orbits are relatively evenly
distributed across latitudes during 0–12 hr for MY33, while the orbital coverage is concentrated at high latitudes in each hemisphere during 12–24 hr. The opposite is true for MY35. However, the orbit varies continuously with local time, and measurements are uniformly covered with latitude for both nighttime and daytime. Also note that there are intermittent gaps in the data coverage, mainly due to the operation of spacecraft with deep dip (DD) campaigns and the unavailability of observations during solar conjunction and spacecraft–Earth communication (Jakosky et al. 2015). Overall, the orbital coverage during MY35 is generally similar to MY33, with slightly better latitude coverage than in MY33. The similar
orbital coverage in MY33 and MY35 allows the comparison of GW activity between these two periods.

2.2. Calculation of Temperature

In this paper, the normalized temperature perturbations of GWs are used to obtain the vertical wavenumber spectrum, so it is necessary to derive the temperature from NGIMS CO2 density profiles. First, according to the hydrostatic equilibrium equation (Snowden et al. 2013; Stone et al. 2018), integrating from the upper boundary to a given altitude can get the local partial pressure of CO2,

$$ P = P_0 + GMm \int_{r_0}^{r} N(r) \frac{dr}{r^2}, $$

where $r$ and $r_0$ are the distance from the center of Mars and Mars’s upper boundary, respectively, and $G$, $M$, and $m$ are the universal gravitational constant, the mass of Mars, and the molecular weight of the CO2, respectively. $N(r)$ is the number density, and $P(r)$ is the pressure at a given altitude. $P_0$ is partial pressure at the upper boundary, determined by fitting the top three points within the height range of the density profile to the hydrostatic distribution model (Manju & Mridula 2021). Specifically, by taking the derivative of the ideal gas law, we get

$$ \frac{1}{P} \frac{dP}{dr} = \frac{1}{T} \frac{dT}{dr} + \frac{1}{N} \frac{dN}{dr}, $$

where $dP/dr$ is substituted using hydrostatic equilibrium and the temperature gradient is equal to constant times the adiabat $\alpha$,

$$ \frac{dP}{dr} = -Ng = -N \frac{GM}{r^2} $$

$$ \frac{dT}{dr} = -\alpha \frac{g}{c_p} = -\frac{GM}{r^2 c_p} $$

by replacing $dP/dr$ and $dT/dr$ in Equation (2), we get

$$ \frac{d\log N}{dr} = \frac{GM}{r^2 T_0} \left( \frac{\alpha}{c_p} - \frac{1}{R} \right). $$

where $R$ is the specific gas constant and $\alpha = 0$ (Snowden et al. 2013; Manju & Mridula 2021). Thus, $T_0$ can be calculated from Equation (5) using the known density profile, and the value of $P_0$ is obtained with the known $T_0$ and density according to the ideal gas law,

$$ P = NKT, $$

where $K$ is Boltzmann’s constant. Then, the pressure $P$ for the entire profile is estimated from Equation (1). Second, combining the pressure $P$ with number density and the ideal gas law, the temperature for the entire profile is calculated as given by

$$ T = \frac{P}{NK}. $$

2.3. Extraction of GW Perturbations

The extraction of GW temperature perturbations can be performed based on the temperature profiles obtained above. The vertical resolution on each profile varies with height, and the average vertical resolution of all profiles used in this study is around 0.5 km. To facilitate processing and simplify the spectral analysis of vertical wavenumber, each temperature profile was interpolated at uniform intervals of 0.5 km. The wave-induced temperature perturbations $T'$ are equal to subtracting the background temperature from the measured instantaneous temperature $T$,

$$ T' = T - \bar{T}. $$

(8)

We use a seventh-order polynomial fit to exclude the effects of large-scale waves and obtain the background temperature $\bar{T}$, which is always applied to extract the GW in the terrestrial (Spiga et al. 2008) and Martian atmosphere (Terada et al. 2017; Jesch et al. 2019; Starichenko et al. 2021), and has been widely proven to give plausible results. Then, the normalized temperature perturbations $(T'/\bar{T})$ of GWs are determined by normalizing the estimated perturbations with the background temperature. The Brunt–Väisälä frequency $N$ can be obtained by,

$$ N = \sqrt{\frac{g}{\bar{T}} \left( \frac{\partial g}{\partial \bar{T}} \right) + \frac{g}{c_p}}, $$

(9)

where $g$ is the acceleration of gravity, and $c_p$ is the specific heat at constant pressure, which is usually taken as 0.844 kJ/(kg K) for the Martian atmosphere.

Figures 3(a)–(d) show the profile of CO2 density, estimated and background temperature, normalized temperature perturbations, and Brunt–Väisälä frequency observed on orbit 2101 in MY33, respectively. Figures 3(e)–(h) are the same, but for orbit 8829 in MY35. Note that the CO2 density profiles used in this study are mostly continuous after interpolating for any small gaps. The estimated temperatures in Figures 3(b) and (f) are in good agreement with the results of previous studies (Stone et al. 2018; Siddle et al. 2019) verifying the accuracy of our calculations. Figures 3(c) and (g) show that the amplitude of wave-induced temperature perturbations is less than 10% of the mean background temperature. In Figure 3(h), the Brunt–Väisälä frequency $N_2$ appears to be less than zero at about 180 km, which indicates the presence of convective instability in the Martian atmosphere at this altitude possibly caused by wave breaking.

2.4. Vertical Wavenumber Spectra of GW

In this study, the vertical wavenumber spectrum of GWs was also calculated from the normalized temperature perturbations. Spectral analysis of the temperature perturbations was performed using a 1024-point fast Fourier transform (FFT) with a Hanning window. To explore the dependence of the spectral density on altitude, two altitude regions 160–200 km and 180–220 km were analyzed separately. Since the vertical resolution is 0.5 km we obtain the minimum vertical wavelength of 1 km. The semiempirical curve of the vertical wavenumber spectrum of saturated GWs in the Earth’s atmosphere is calculated from the theory prediction (Smith et al. 1987; Tsuda et al. 1991). It can be written as

$$ F_{T'/\bar{T}} = \frac{1}{4\pi^2} \frac{N^4}{10g^2 k_z^4}. $$

(10)
as seen in Equation (1) in Ando et al. (2015), where \( N \) is the Brunt–Väisälä frequency and \( g \) is the gravitational acceleration. The vertical wavenumber \( k_z \), in cycles per meter.

3. Results

3.1. Vertical Wavenumber Spectra at Different Latitudes

By dividing the temperature profiles into two altitude ranges, 160–200 km and 180–200 km, we can study the wave propagation with height in the thermosphere; also the latitudinal trend of GWs is studied by dividing it into different latitude intervals. The normalized temperature perturbations and the seventh-order polynomial fit to the data are divided by 30° latitude intervals. The normalized temperature perturbations in Figure 3.

The vertical wavenumber power spectrum during the altitude range of 160–200 km are used to perform FFT containing the Hanning window function to obtain the power spectrum of the GW signal. Figure 4 shows the average vertical wavenumber power spectrum during the first half of MY35 of low solar activity in the two altitude intervals of 160–200 km and 180–220 km. The latitude ranges are divided by 30° intervals and then indicated by lines of different colors, and the theoretical spectrum of saturated GWs is also plotted given by Equation (10). The most obvious feature is that the power spectral density decreases with increasing wavenumber, which is similar to the characteristics of the spectral structure obtained in the terrestrial stratosphere, mesosphere, and thermosphere (Smith et al. 1987).

In the altitude range of 160–200 km, the slope of the logarithmic spectrum is around \(-3\) in most latitudinal intervals at higher wavenumbers \( k > 0.1 \) (vertical wavelengths \(<10\) km), and the spectral density almost coincides with the theoretical prediction developed for the saturated GWs of the Earth’s atmosphere. This indicates that saturation due to convective instability also happened in the Martian thermosphere, which is consistent with the convective instability found in the individual profiles in Figure 3(h). The exceptions are the 45° N–75°N in northern spring and 15°S–15°N in summer in the altitude range of 160–200 km. The spectral slope becomes flatter at low wavenumbers \( k < 0.1 \) (vertical wavelengths \( >10\) km). In contrast, the spectral density is significantly smaller than the saturation spectrum at the altitude of 180–220 km (Figures 4(c)–(d)), indicating that a larger share of unsaturated GWs in the average spectrum. Note that power-law spectral indices close to \(-3\) are also seen in the large wavenumber region for \( k > 0.1 \), but the spectral power is nearly an order of magnitude less than the saturation value.

Figure 3. The altitude profiles of (a) CO$_2$ density, (b) estimated temperature (black) and the seventh-order polynomial fit to the data (red), (c) normalized temperature perturbations, and (d) Brunt–Väisälä frequency observed on orbit 1060 in MY33. Panels (e)–(h) are the same, but for orbit 2101 in MY35.
Figure 4. Mean vertical wavenumber spectra of normalized temperature perturbations divided by $N^4$ at different latitude bins, in the altitude range of 160–200 km (upper panels) and 180–220 km (bottom panels) during northern (a) spring ($Ls = 0°–90°$) and (c) summer ($Ls = 90°–180°$) in MY35. The theoretical spectrum of saturated GWs given by a dashed black line. The different color lines correspond to their respective latitude ranges.

Table 1
The Number of Temperature Profiles in MY35 and MY33 Was Used in the Analysis

| Martian Year | MY35 | MY33 |
|--------------|------|------|
| Altitude     | 160–200 km | 180–220 km | 160–200 km | 180–220 km |
| $Ls$ (Solar longitude) | $0°–90°$ | $90°–180°$ | $0°–90°$ | $90°–180°$ |
| Latitude     | Number | Number | Number | Number | Number | Number | Number | Number |
| 45°–75°N     | 38     | 238   | 38    | 214    | 0     | 546   | 0     | 533    |
| 15°–45°N     | 51     | 124   | 71    | 120    | 0     | 208   | 16    | 225    |
| 15°–15°S     | 147    | 139   | 146   | 136    | 166   | 18    | 188   | 0      |
| 15°–45°S     | 131    | 137   | 87    | 128    | 215   | 0     | 208   | 0      |
| 45°–75°S     | 372    | 68    | 333   | 103    | 332   | 0     | 299   | 0      |

(2017) also shows that wave dissipation and breaking/saturation tend to preferentially suppress shorter wavelengths. It is important to note that the spectral densities in 45°–75°N in Figure 4(a) and 15°S–15°N in Figure 4(b) are not comparable to the theoretical curve of saturated GWs and are decaying with altitude. The small spectral density within 45°–75°N may be due to the lack of data volume. The number of temperature profiles used in the analysis is summarized in Table 1, which shows the sampling bias of the satellite in the spring for the northern hemisphere. However, for the region of 15°S–15°N, the data coverage in this region is complete and uniform in summer, but the spectral values are small and decrease with height, indicating that there are other processes that play a role in dissipating GWs in the thermosphere, such as nonlinear interactions and molecular viscosity.

In terms of the seasonal variations in GW activity, the larger amplitudes in spring occur in the southern hemisphere, followed by the lower latitudes near the equator and then the northern hemisphere. The reasons for this discrepancy are mainly influenced by the uneven latitudinal data coverage. As shown in Table 1, MAVEN samples poorly in the spring for the middle and high latitudes of the northern hemisphere. Specifically, the number of temperature profiles located north of 15°N in the northern hemisphere during spring in MY35 is less than 100, much less than the number for all latitudes in the southern hemisphere. And almost no data are available for the northern hemisphere in the spring of MY33. The data coverage is good during the summer in MY35, the strongest GWs are located in the mid-latitude region of 15°–45°N, followed by the southern hemisphere and the weakest in the equatorial region. Yiğit et al. (2021b) revealed similar results that GW-induced density fluctuations are much larger in the southern hemisphere during the northern spring season, and they argued that this difference should be interpreted with caution due to the poor sampling of MAVEN during spring. Similarly, when interpreting the distribution of GWs at different latitudes in our paper, we should take into account that part of the possible effects is caused by biases in the data coverage.

Figure 5 shows the average vertical wavenumber spectrum for the spring and summer of MY33, and the general characteristics are similar to those of MY35, with the spectrum density decreasing with increasing wavenumber. At 160–200 km altitude, the vertical wavenumber spectrum follows a power-law dependence with a slope of −3, and the spectral density is close to the saturation theoretical value of the Earth’s atmosphere. Compared with MY35, the spectral density...
is smaller in MY33 and varies more as it propagates upward from below. GW attenuates more significantly as it propagates into the upper thermosphere, probably because it encounters stronger damping due to molecular viscosity (Yiğit et al. 2008; Parish et al. 2009; Medvedev et al. 2011a). On the one hand, the higher solar radiation produces higher thermospheric temperature, which results in a stronger dissipative effect of saturation due to convective instability and reduces the GW amplitude. On the other hand, the dissipation due to molecular diffusion is enhanced, which also reduces the GW amplitude. Yiğit & Medvedev (2010) showed that the coefficient of molecular diffusion increases with extreme ultraviolet flux and thermospheric temperature. These spectral features suggest more dependence of the GW perturbation on the background temperature during the high solar activity period of MY33. Meanwhile, the higher vertical wavenumber spectral density and less pronounced attenuation during the upward propagation of GWs in MY35 suggest that the waves experienced more favorable propagation conditions.

### 3.2. Vertical Wavenumber Spectra at Different Longitudes

Figures 6 and 7 depict the spectral features in different longitude intervals. A similar overall spectral signature to that in Figure 4 is presented, with the spectral slope essentially conforming to the saturation spectrum and the spectral density decreasing with altitude. In addition, the spectral densities are essentially similar across the different longitude ranges, and there does not appear to be a correlation between GW activity and the location of the terrain. This suggests that the amplitude of regionally distributed GWs may be controlled more by propagation effects (Fritts & Alexander 2003) than by the sources that excite these waves. Terada et al. (2017) also concluded that the correlation between GW activity in the upper thermosphere and specific topography is weak, and little evidence of GW variation with longitude was found.

### 3.3. Vertical Wavenumber Spectra at Different Local Time

The dependence of GW activity on local time is examined next. Figure 8 shows the seasonally averaged vertical wavenumber spectra in the first half of MY35 according to different local time intervals. In the 160–200 km thermosphere (Figures 8(a), (b)), the spectral densities within each local time interval are basically consistent with the saturated GW theoretical spectra, with a power-law index around −3, indicating the saturation state of GWs at this altitude. Specifically, the maximum spectral density in spring is within 6–12 hr on the dayside, which differs from previous findings that GW activity is more active on the nightside. However, the summer results show that GW amplitudes are significantly stronger on the nightside (0–6 hr, 18–24 hr) than on the dayside (6–12 hr, 12–18 hr). Terada et al. (2017) showed that convective saturation is the main process determining GW amplitudes in the Martian thermosphere. The amplitude is inversely proportional to the background temperature, so the GWs are more active in the nightside. However, these conclusions are based on a windless atmosphere, where the temperature variations can control the behavior of GWs. In general, the GW-induced temperature fluctuations in the thermosphere are created by GW harmonics that survived propagation from the lower atmosphere, depending on the wave sources and filtering by the underlying mean winds. Previous studies have shown how winds influence the GW spectrum as well as the dissipation (Medvedev et al. 2011a; Yiğit 2018). In the present study, the strongest GWs were found on the dayside 6–12 hr during the low solar activity period in the spring of MY35, which is contrary to previous perceptions. Here, we infer that the spring–summer discrepancy in terms of the local time behavior of the maximum power spectrum may be related to the filtering of the mean background wind, but the detailed mechanism of the wind field acting on the GWs is unclear.
The spectral values within 180–220 km are significantly smaller than 160–200 km, suggesting that the GWs attenuate with increasing altitude. Note that the perturbation decays most significantly at 12–18 hr, both in spring and summer. It is suggested that the GW perturbation is controlled by the background temperature from noon to dusk, as the temperature changes more significantly during that period. We believe that it is convective saturation/breaking that exerts the main dissipative effect from noon to dusk. During the local time of 0–6 hr in summer, the spectral density is higher than the saturation spectrum in both altitude intervals and varies little with altitude, suggesting that not only saturation/breaking but also other processes may affect the propagation of GWs in the thermosphere. Consistent with the results obtained in Figure 4(b) above.

Figure 9 shows the spectral characteristics of GWs within different local times of the MY33 year. The spectral density is higher on the nightside than on the dayside during both spring and summer, which is consistent with the previous perception that GW activity is more active on the nightside. For example, Yiğit et al. (2015) showed for the first time with MAVEN data that GW activity is stronger at night. The most recent observational evidence for the strong nightside GW activity is by Yiğit et al. (2021b). The dayside perturbation decays
more significantly during the propagation from below up into the upper thermosphere, probably because the temperature variation is more pronounced under the influence of solar radiation in the daytime. Both of the above points show a close relationship between GW amplitude and atmospheric background temperature.

4. Discussion

4.1. Possible Wave Sources of Perturbations in the Thermosphere

In general, GWs in the thermosphere include harmonics propagating from the lower and middle atmosphere as well as locally excited by the upper solar wind forcing. In terms of GW sources from above, the main source of excitation of perturbations in the thermosphere on Earth comes from the magnetospheric energy deposited in the polar regions in the form of particle deposition. Unlike Earth, there is no planetary-scale magnetic field on Mars, but the deposition of energy from precipitating O\(^+\) ions may excite perturbations in the thermosphere on Mars (Leblanc & Johnson 2002; Chaufray et al. 2007; Fang et al. 2013). Fang et al. (2013) estimated that temperature perturbations in the upper thermosphere of Mars due to O\(^+\) ion deposition can be as high as 3%, with larger amplitudes in the 200–250 km altitude. However, the small-scale perturbations obtained from the MAVEN/NGIMS

Figure 8. Mean vertical wavenumber spectra of normalized temperature perturbations divided by N\(^4\) at different local time bins, in the altitude range of 160–200 km (upper panels) and 180–220 km (bottom panels) in MY35.

Figure 9. As in Figure 8, but in MY33.
measurements in this study are no longer significant in amplitude above 200 km (as shown in Figures 10(c)–(d)). The strongest amplitudes are located below 200 km during both high and low solar activity. This could be that the perturbations from O\(^+\) ions subsidence are not large enough to have any significant effect. Therefore, we believe that the forcing from above has an insignificant excitation effect on the thermospheric perturbation.

GWs in the lower atmosphere on Mars are excited by a variety of mechanisms, including topographic sources, convection, and instability of weather systems (Heavens et al. 2020). Observation and simulation results have shown that GW sources with intermittency vary spatially with the season (Creasey et al. 2006; Kuroda et al. 2019). On average, the main process of GW generation in the lower atmosphere is nontopographic excitation (Yiğit et al. 2021b). Our study also found that the GW spectral density in the thermosphere rarely varies with longitude, and no correlation between GW activity and topography was found. However, there are significant differences in spectral density within different latitudinal intervals. Although our results are partly influenced by the satellite sampling bias, the latitudinal variability of GW activity in the thermosphere is objectively present. Figure 10 shows the altitude versus latitude distributions of the background temperature and normalized temperature perturbations between 160 and 200 km during the first half of MY33 and MY35. In general, the temperature is significantly lower during the first half of MY35 than during MY33, while the small-scale perturbations induced by GWs are stronger. The average amplitude is less than 15%, which is broadly consistent with previous findings (Creasey et al. 2006; Forbes et al. 2006; Terada et al. 2017; Manju & Mridula 2021). The distribution of GW activity varies significantly in latitude. Larger values of perturbations are located around 50°N in the northern hemisphere, as well as in the polar region of the southern hemisphere, while the smallest are near the equator. Simulations of general circulation models (GCMs) on Earth (Yiğit & Medvedev 2009; Miyoshi et al. 2014) and Mars (Medvedev et al. 2011b; Kuroda et al. 2015) show that for GWs from lower wave sources, their energy shifts poleward with increasing altitude, leading to relatively weak GW activity in the lower latitudes of the thermosphere. The latitudinal differences in GW activity in the present study are consistent with these previous findings. Therefore, it is inferred that the disturbances in the thermosphere may be generated by nontopographic processes from the lower atmosphere.

4.2. Possible Dissipative Processes for Perturbations in the Thermosphere

The nature of GWs in the thermosphere is determined by a combination of wave sources, propagation conditions, and damping processes. The amplitude of GWs in the lower atmosphere is usually larger at low latitudes near the equator (Creasey et al. 2006), while strong wave disturbances in the thermosphere occur in the polar regions. Clearly, the distribution of GWs observed in the thermosphere cannot be explained by the wave sources alone. We analyze the possible dissipation processes of GWs based on the vertical wavenumber spectrum of temperature perturbations. The spectral density of GWs shown in Figures 4 and 5 is essentially consistent with the theoretical prediction of saturated GWs, and the greater
statistical weight of saturated waves in the average spectrum indicates that the amplitude of GWs is mainly determined by the saturation due to convective instability. This implies that the quasiempirical saturation spectrum theory developed for the Earth’s atmosphere is also applicable to the Martian thermosphere. Eckermann et al. (2011) estimated the relative importance of various GW damping processes on Mars assuming a wind-free atmosphere by numerical simulations. In the upper Martian thermosphere, wave saturation is considered to be the dominant energy deposition mechanism for waves with vertical wavelength $\lambda_z > 15$ km. Our results suggest that wave saturation processes caused by convective instability do exist in the Martian thermosphere. Saturated GWs will accelerate the mean flow and induce turbulence, thus causing a redistribution of momentum and energy. However, the linear convective instability theory can only explain the characteristics of GW activity in the thermosphere to a limited extent. It is known that winds and the associated wave refraction is a particularly important process in gravity wave propagation and dissipation (Yiğit et al. 2008; Medvedev et al. 2011a). For GWs in the thermosphere, the main mechanism of wave damping is molecular diffusion and thermal conduction, while nonlinear wave damping is secondary, i.e., convective instability plays only a minor role in the thermosphere in limiting the amplitude of GWs. The spectral values of unsaturated GWs decay with increasing altitude in the $15^\circ$S–$15^\circ$N latitude interval during the summer of MY35, suggesting the role of other dissipative processes, such as molecular diffusion (molecular viscosity and heat conduction). Forbes et al. (2006) showed that the response of temperature variations to solar extreme ultraviolet fluxes with a period of 27 days is larger in the upper thermosphere of Mars than on Venus and that CO$_2$ 15 $\mu$m radiative cooling is less important in the Martian thermosphere. The day–night difference in the amplitude of GWs in the upper thermosphere of Venus may be due to the modulation effect of CO$_2$ 15 $\mu$m radiative damping (Terada et al. 2017). In this study, observational data are used to show the possible role of GW dissipation processes beyond saturation in the Martian thermosphere in limiting the amplitude of GW.

To explain the dependence of GW activity on local time, we examined the local time–altitude cross-section of background temperature and GW perturbations during the first half of MY33 and MY35, as shown in Figure 11. Overall, the disturbances are stronger in MY35 than in MY33, with the largest amplitudes occurring within 5–10 hr in the morning, consistent with the results shown in Figure 8. These results differ from the previous understanding that GW activity is more active on the nightside than on the dayside. And the distribution of the strongest perturbations does not correspond to the regions of low temperature, suggesting that the perturbations in the Martian thermosphere are not only controlled by the background temperature but may also be influenced by other factors such as molecular diffusion. Yiğit et al. (2021b) investigated the dependence of GWs on local time using a one-dimensional spectral nonlinear GW model, suggesting that convective instability mechanisms may play a limited role in explaining diurnal differences in GW amplitudes in the upper thermosphere.

5. Summary and Conclusion

In this study, we analyze the spectral characteristics of GW activity in the Martian thermosphere during the low solar activity in MY35, using CO$_2$ density data from 160 to 220 km measured by NGIMS on board MAVEN. Also, the relatively higher solar activity of MY.33 was chosen for comparative analysis. MAVEN has good local time and latitude/longitude coverage during the selected season, and the orbital scans are similar globally. During this period, the spectral density of GW perturbations in the thermosphere shows significant variation with altitude, latitude, season, and local time. Such complex variations are influenced by the combination of solar irradiance and the slow progress of MAVEN’s orbit. The main conclusions obtained are as follows:

1. The slope of the vertical wavenumber spectrum of the normalized temperature perturbation in the Martian thermosphere is around $-3$, and the spectral density almost matches the theoretical prediction curve of saturated GWs. The power spectral density decreases with increasing wavenumber, which is similar to the spectral structure on Earth. The quasiempirical saturation spectrum theory developed for the Earth’s atmosphere is also applicable to the GWs in the Martian thermosphere.

2. The maximum perturbation of the GWs in the Martian thermosphere is between 160 and 200 km, which is consistent with previous observational results (Yiğit et al. 2021b), where GWs are attenuated during upward propagation and the shorter waves are preferentially dissipated.

3. Compared to MY33, the GW activity is more active and less attenuated as it propagates upward in the low solar activity of MY35, indicating that waves experience more favorable propagation conditions.

4. The amplitude of GWs in the thermosphere varies significantly in latitude, with small perturbations near the equator and large values in the polar regions. In addition, the spectral density of GWs shows a significant seasonality, with higher values in the southern hemisphere in spring, while for summer it is higher in the northern hemisphere. This seasonality may be due to a combination of the variability of the GWs themselves and the limited sampling range of the satellite.

5. The strongest spectral density in the spring of MY35 was found in 6–12 hr, suggesting that the amplitude of perturbations in the Martian thermosphere during periods of low solar activity is much less controlled by background temperature, along with other factors such as dissipative effects of molecular viscosity and thermal conduction.

Most of the GWs in the thermosphere obtained in this study have vertical wavelengths in the range of 4–20 km. We discuss the possible sources of GWs in the Martian thermosphere, which are more likely to be upward-propagating harmonics generated by nontopographic processes, while GWs excited locally in the thermosphere by precipitating particles are not evident. Wave saturation due to convective instability is important at lower altitudes in the thermosphere in determining GW amplitudes. However, for the thermosphere, where dissipative processes modulated by background winds dominate on Earth (Yiğit & Medvedev 2009) and Mars (Yiğit et al.
radiative damping is not important in the Martian thermosphere (Medvedev et al. 2011b). Our statistical analysis cannot clearly distinguish the nature of the wave sources and the relative importance of different dissipative processes in the Martian thermosphere. More refined observations of the global distribution of GW activity in specific seasons and comprehensive theoretical modeling of wave excitation, propagation processes, and dissipation are needed.

All CO2 density profiles used in this paper correspond to Level 2, version 08, and revision 01 data that are publicly available at the Planetary Data System https://atmos.nmsu.edu/PDS4/data/PDS4/MAVEN/ngims_bundle/12. The sunspot data were provided by SILSO World Data Center of the Royal Observatory of Belgium https://wwwb.isb.be/silso/datafiles. We thank the MAVEN team and SILSO World Data Center for data support. The study was also supported by the National Natural Science Foundation of China (Grant 41875045), Hunan Outstanding Youth Fund Project (Grant 2021JJ10048), and the “Western Light” Cross-Team Project of the Chinese Academy of Sciences, Key Laboratory Cooperative Research Project.

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Figure 11. Local time vs. altitude distributions of temperature (a)–(b) and normalized temperature perturbations (c)–(d) for the first half of MY33 and MY35. The data are binned in 2 hr × 5 km (local time × altitude) intervals.

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