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Key Points:

- IUVS limb scan observations show significant variations of N_2/CO_2 at 140 km, with both seasonal sinusoidal and latitudinal trends.
- Seasonal change of inferred dayside homopause is suggested to be mainly controlled by inflation and contraction of the lower atmosphere.
- Latitudinal dependences of homopause altitudes are, for the first time, suggested by observed latitudinal dependences of N_2/CO_2.

Abstract

The dayside N_2/CO_2 at 140 km altitude in the Martian thermosphere has been investigated by the Imaging UltraViolet Spectrograph (IUVS) aboard the Mars Atmosphere and Volatile EvolutioN (MAVEN) spacecraft during the period from October 2014 to May 2018. We find that N_2/CO_2 at 140 km altitude varies significantly in the range of 0.02 to 0.20 and shows a seasonal sinusoidal trend. The higher value appears during aphelion and the lower value during perihelion. Variations of observed N_2/CO_2 ratio at 140 km are mainly associated with CO_2 number density. Thus, while we found that N_2/CO_2 varies at a given altitude, we could not identify variations at a given pressure level. In order to reveal the drivers of N_2/CO_2 variations at 140 km, we examine the effects of surface N_2/CO_2, thermospheric temperature, and homopause altitude. The variations of homopause altitude could be the dominant driver of changes in N_2/CO_2. Inferred dayside homopause altitudes derived from IUVS observations show an anti-correlation with the trend of N_2/CO_2 at 140 km. Distributions of CO_2 density at the inferred homopause altitudes suggest that dayside homopause altitude is mainly controlled by inflation and contraction of the lower atmosphere. Additionally, N_2/CO_2 shows a clear latitudinal dependence for L_s = 80° - 100°. Higher N_2/CO_2 values appear in the southern winter hemisphere, corresponding to lower homopause altitude by ~30 km. Meanwhile, the latitudinal dependence of both N_2/CO_2 and homopause altitude disappears for L_s = 340° - 360°.

Plain Language Summary

Changes of atmospheric composition in the upper atmosphere could be a source of variation of gas species escaping to space. We report the seasonal and latitudinal variations of dayside N_2/CO_2 ratios at 140 km altitude by a remote-sensing instrument aboard the Mars Atmosphere and Volatile EvolutioN spacecraft. Our observations showed that changes in N_2/CO_2 could be caused by the variations of homopause altitude. Distributions of CO_2 density at the inferred homopause altitudes suggest that dayside homopause altitude is mainly controlled by inflation and contraction of the lower atmosphere.

1 Introduction

Planetary atmospheres are generally characterized by a compositional boundary, called the homopause, below which gases are well-mixed by eddy diffusion (homosphere) and above which gases are diffusively separated according to their own scale heights by molecular diffusion (heterosphere). In the heterosphere, the mixing ratio of lighter species is expected to increase with altitude above the homopause. The location of the homopause altitude influences the thermospheric composition and thereby the escape of species to space. In addition, the fractionation between the homopause and the exobase determines the relative abundance of species that escape to space and the total atmospheric loss from the isotope record (Jakosky et al., 1994, 2017; Lammer & Bauer, 2003; Chassefière & Leblanc, 2004; Slipski & Jakosky, 2016).

Early measurements of atmospheric composition in the Martian thermosphere were performed by the neutral mass spectrometers mounted on the aeroshell of Viking 1 and 2 landers (Nier & McElroy, 1977). Viking observed vertical profiles of CO_2, N_2, Ar, CO, and O_2 in the range between 120 and 200 km. The Viking observations showed that the relative abundance of lighter species increases with altitude due to diffusive separation. After the Viking observations, the Mars Atmosphere and Volatile EvolutioN (MAVEN) spacecraft measures atmospheric composition in detail in the thermosphere using in-situ and remote-sensing instruments; e.g., the Neutral Gas and Ion Spectrometer (NGIMS) (Mahaffy, Benna, King et al., 2015) and the Imaging UltraViolet Spectrograph (IUVS) (McClintock et al., 2014). In the beginning of MAVEN mission, vertical profiles of several species observed by both the NGIMS and IUVS instruments confirmed diffusive separation in the thermosphere (Mahaffy, Benna, Elrod et al., 2015; Withers et al., 2015; Evans et al., 2015). Vertical profiles between the Viking and MAVEN observations are different in both scale heights and background temperature. The scale heights of CO_2 are larger in the MAVEN NGIMS measurements (15 km) than from Viking (10 km), with corresponding temperatures of 270 K and 180 K, respectively (Withers et al., 2015). Withers et al. (2015) mentioned that the warmer temperature observed by MAVEN is due to the Sun-Mars distance and the solar cycle. Vertical profiles of N_2/CO_2 observed by the Viking probes and MAVEN/IUVS are compared in Evans et al. (2015). Diffusive separation of the two species is more effective in the Viking observations and the difference with IUVS observations is obvious above 140 km altitude. It is noted that the comparison of atmospheric composition by Evans et al. (2015) was conducted in the beginning of the MAVEN mission. The vertical profile measurements by IUVS aboard MAVEN have now been taken for over three Martian Years, allowing an investigation into compositional distribution and variation of N_2 and CO_2 densities in the thermosphere of Mars.

The Martian homopause altitude was considered to be located around 120 and 130 km from observations by the Viking landers (Nier & McElroy, 1977). However, numerical simulations after this suggested that the Martian homopause altitude varies due to the atmospheric global circulation (González-Galindo et al., 2009), and the saturation of small-scale gravity waves, planetary-scale waves, and thermal tides (Imamura et al., 2016; Leovy, 1982). Recently, Jakosky et al. (2017) and Slipski et al. (2018) reported substantial variations of the homopause altitude by using vertical profiles of $N_2/^{40}Ar$ observed by NGIMS. The inferred homopause altitudes are located between 60 and 140 km. Slipski et al. (2018) demonstrated the possibility of a seasonal change in homopause altitude at noon by ~30 km, with the largest values observed between L_s = 240° and 360° and the smallest values between ~50° and 180°. Slipski et al. (2018) also showed the latitudinal dependence of homopause altitudes, but a clear latitudinal dependence predicted in González-Galindo et al. (2009) was not obtained because latitudinal and local time effects could not be distinguished due to the NGIMS observing geometry.

In addition to homopause altitude variations, it is noted that Mars' thermospheric temperature is variable because solar extreme ultraviolet (EUV) irradiance changes with time and it is absorbed above ~100 km altitude. Gröller et al. (2018) showed diurnal temperature variations for a pressure level of 3×10^{-5} Pa (around 130 km) and for latitudes below 50° over one Martian year in their Figure 19. The average temperature between 12:00 and 16:00 local time (LT) is ~170 K, but observed temperatures showed broad variability between ~150 and ~200 K. Gröller et al. (2018) mentioned that the relatively large variability could be due to variations with latitude and season. Bougher et al. (2017) implied that long-term variations of temperature in the Martian dayside upper thermosphere correlate with solar EUV irradiance at Mars. The observed temperature range corresponding to scale heights over 150 - 180 km in Bougher et al. (2017) is 170 to 240 K. Thiemann et al. (2018) also showed the correlation between thermospheric temperature and solar EUV irradiance using solar occultations of EUV irradiance at Mars. The thermospheric temperature has an impact on vertical profiles of atmospheric composition because each gas has its own scale height above the homopause altitude, $H=kT/mg$, where k is the Boltzmann constant, T is temperature, m is molecular mass according to species, and g is gravity. Therefore, the atmospheric composition above the homopause can be affected by the thermospheric temperature.

In addition, the atmospheric composition at the homopause would vary associated with variations of mixing ratio near the surface. Due to the tilt of the rotation axis, Mars poles become cold enough for CO_2 to condense out during local fall and winter, with CO_2 subliming from the poles during local spring and summer. The change in atmospheric mass results in variations of surface pressure with two peaks at L_s ~ 60° and L_s ~ 270° each year, corresponding to times when the greatest amount of CO_2 is in the atmosphere (Hess et al., 1980). Both of two Viking landers observations showed a similar variation in pressure at the surface, although an absolute value of pressure is different due to their locations. Similar seasonal peaks of surface pressure were obtained by Mars Pathfinder lander, Phoenix lander, and Mars Science Laboratory (MSL) (Martínez et al., 2017; Ordonez-Etxeberria et al., 2017). When the main atmospheric constituent condenses out, trace gases that do not condense out with the CO_2, such as Ar and N_2, are relatively enhanced (i.e., their mixing ratio increases). Mixing ratio of N_2 at Gale crater was measured by MSL (Trainer et al., 2019). A maximum mixing ratio of N_2 appears when the surface pressure is minimum. The seasonal variation of mixing ratio of N_2 at the surface is about ±10% (Trainer et al., 2019). Note that the mixing ratio of trace gases varies with location and time, due to transport by advection and mixing. A strong depletion of mixing ratio of trace gases, such as CO and Ar, occurs in the summertime polar regions, associated with sublimation of CO_2 polar caps (e.g., Smith et al., 2018; Sprague et al., 2007). Around the winter polar region, an enhancement of mixing ratio of trace gases is expected. Such variations might affect the atmospheric composition above the homopause.

In this study, we aim to investigate variations of atmospheric composition in the thermosphere and to reveal drivers of the variations. We use remote-sensing data for this study, which enables us to distinguish geophysical and geometrical effects owing to a wider coverage than in-situ observations, for better understanding the drivers of the variations in the thermosphere.

The outline of this paper is as follows. In Section 2, we describe the dataset used in this study. The seasonal, local time, and latitudinal variations of N_2/CO_2 are shown in Section 3. We discuss the drivers of N_2/CO_2 variation, inferred homopause altitudes from IUVS observations, and comparisons with model predictions in Section 4. Finally, our conclusions are summarized in Section 5.

2 MAVEN IUVS

We use limb observations from IUVS aboard MAVEN. The MAVEN orbit is elliptical with apoapsis near 6000 km and periapsis near 160 km altitude, with a 4.5-hour period. IUVS limb measurements are taken near periapsis below 500 km altitude for dayglow observations in the far and mid-UV regime (115 - 340 nm) (Jain et al., 2015). Dayglow emissions of CO_2^+ Ultraviolet Doublet (UVD) at 288 - 289 nm (Jain et al., 2015; Evans et al., 2015) and N_2 Vegard-Kaplan bands (VK) at 258.0 - 287.5 nm (Stevens et al., 2015) are used to retrieve CO_2 and N_2 densities, respectively from 100 to 200 km with an altitude resolution of 5 km (Evans et al., 2015; Stevens et al., 2015). To understand variations of atmospheric composition in the thermosphere, it is better to use non-reactive species that do not vary spatially and temporally. N_2 is weak in photochemical reactivity and non-condensable. The CO_2 density can vary due to photochemistry and meteorology of the condensation as CO_2 ice clouds, but these effects should be negligible because CO_2 is the dominant species in Mars' atmosphere. Since the effect of sublimation and condensation from the polar ice caps is well understood as the variation of surface pressure, CO_2 variations can be corrected using a knowledge of surface pressure. Thus, we analyze the number densities of CO_2 and N_2 to investigate the mixing ratio in the thermosphere. It is noted that Evans et al. (2015) adopted an upper limit of 60° for Solar Zenith Angle (SZA) over concerns that the retrievals may be biased at larger SZA values due to anisotropy effects (i.e. variation in solar illumination along the slant path). The upper limit has since been raised to 75° because no evidence of a bias has been seen in retrievals for SZAs up to this revised limit. This is not surprising, since CO_2^+ UVD and N_2 VK are optically thin such that most of the airglow signal originates from within a few scale heights of the tangent point, or the airglow peak, whichever is larger. In general, contributions along the slant path within two scale heights of a tangent point of 130 km will have an SZA within ~ 6° of the SZA at the tangent point. We use the dataset of level 2 version 13 revision 1 provided in the Planetary Data System (PDS).

In the level 2 dataset, CO_2 densities are provided at fixed altitude bins centered at 130, 140, 150, and 170 km, whereas N_2 densities are provided at 125, 137, 153, and 173 km. The uncertainties of CO_2 densities are small (less than 10%) whereas those of N_2 densities are relatively large, typically up to 40% below 140 km and above 160 km. Below 140 km altitude, there is a possibility of contamination by solar stray light in the N_2 VK emission bands. On the other hand, above 160 km, the intensity of N_2 VK emission becomes dimmer, resulting in a reduced signal-to-noise ratio in the N_2 VK emission and corresponding increased uncertainties of retrieved N_2 densities. In this study, we confine our analysis to ~140 km where the densities have the smallest uncertainties. We interpolate the N_2 density between 137 and 153 km onto the CO_2 altitude grid and evaluate the mixing ratio at 140 km. Data we applied in this study have uncertainties in N_2/CO_2 less than 50%. Under the given constraint, uncertainties of N_2/CO_2 at 140 km used for this study are in the range of 15% to 50% with a mean value of 37%.

We analyzed CO_2 and N_2 densities derived from IUVS limb scan observations from October 2014 to May 2018, which correspond to 9258 profiles. The tangential footprints of IUVS limb observation included in the analysis are shown in Figure 1. The dayside measurements by IUVS have been performed in several groups of orbits during different seasons throughout the MAVEN mission. Local time is limited from 7:00 to 19:00 except for observations around the southern polar region. SZA is in the range from 0° to 75°. IUVS limb observations over the time period considered here provide good coverage in longitude, latitude, local time, and season.

3 Results

Time variations of IUVS observations vs. L_s are shown in Figure 2. Figure 2(a) provides the Sun-Mars distance (blue line) and the orbit averaged solar EUV irradiance at 17 - 22 nm observed by the Extreme UltraViolet Monitor (EUVM) (Eparvier et al., 2015) aboard MAVEN (black line). Figures 2(b)-2(d) show the observed CO_2 density, N_2 density, and N_2/CO_2 at 140 km, respectively. Colors correspond to the observed local time. Uncertainties are shown by gray bars. The dayside N_2/CO_2 varies from 0.02 to 0.20, but variations of CO_2 density, N_2 density, and N_2/CO_2 shown in Figure 2 include effects of season, local time, and latitude. A short-term variation of N_2/CO_2 during L_s = 30° - 70° in Martian Year (MY) 33, L_s = 120° - 160° in MY 33, and L_s = 340° - 40° through MY 33 to MY 34, which correspond to orbits around 1800 - 2200, 2900 - 3500, and 4800 - 5500, respectively, is seen in Figure 2(d). These variations are explained below.

3.1 Seasonal variation

Figure 3 demonstrates a seasonal trend of the CO_2, N_2 number densities, and N_2/CO_2 ratio at 140 km. Each color represents a specific local time bin, including 8:00 - 10:00, 12:00 - 14:00, and 16:00 - 18:00, and symbols represent observed MYs. Figures 3(a) and 3(b) demonstrate that CO_2 and N_2 densities, distinguished by local time, vary with L_s. Higher number densities of CO_2 and N_2 appear around perihelion (L_s ~ 250°). Lower number densities appear around aphelion (L_s ~ 70°). This seasonal variation of CO_2 density obtained in this study agrees well with that found by Mars Express (MEx) Spectroscopy for Investigation of Characteristics of the Atmosphere Mars (SPICAM) observations of CO_2 density in the lower thermosphere (Forget et al., 2009). Although each profile's density depends on the combinations of local time, latitude, and local season, the variations in CO_2 and N_2 densities at 140 km observed by IUVS limb show predominately a seasonal sinusoidal trend. Those variations correspond to the inflation and contraction of the lower atmosphere. It is noteworthy that variations in CO_2 densities have a larger amplitude than variations in N_2 densities and this affects the variation of N_2/CO_2. The seasonal N_2/CO_2 variations, which in the local time bin of 12:00 - 14:00 become ~0.063 on average during aphelion (between 50° and 90° in L_s) and ~0.031 on average during perihelion (between 230° and 270° in L_s), show an anti-correlation with the variations of CO_2 and N_2 densities. The other local time bins also demonstrate a clear seasonal variation.

3.2 Local time variation

Figure 4 shows the short-term variations of IUVS observations vs. local time during three seasons. We selected three groups L_s = 30° - 70° in MY 33, L_s = 120° - 160° in MY 33, and L_s = 340° - 40° in MY 33 to 34 because those well cover the range from morning terminator to evening. In all three groups, local time of IUVS observation varies from evening to morning for each period as shown in Figure 1. We have to note that data taken during L_s = 240° - 300° in MY 33 have wide local time coverage, however, we didn't select this group since observation during this period included data at high southern latitudes (> ~65°S) which are sunlit throughout the sol (i.e. there is no night). The results presented in Figure 4 represent the combination of local time and seasonal variations, thus caution must be taken. Data taken during L_s = 120°- 160° in MY 33, the second column in Figure 4, correspond to an increasing of number density with season. Meanwhile, data taken during L_s = 30°- 70° in MY 33 and L_s = 340°- 40° in MY 33 to 34, the first and the third columns in Figure 4, correspond to a decreasing of number density with season.

In Figures 4(d) and 4(e), observed CO_2 and N_2 densities have their maximum around noon. That would be a result of local time variation. A similar local time variation of Ar density is found in the upper thermosphere (Gupta et al., 2019). On the other hand, CO_2 and N_2 densities in Figures 4(a), 4(b), 4(g), and 4(h) don't have any clear peaks around noon. It could be that the effects of season and local time on number density cannot be distinguished from these datasets. Nevertheless, Figures 4(c), 4(f), and 4(i) show that N_2/CO_2 become minimum around noon and increase toward the dawn side. N_2/CO_2 around dawn is larger than that at noon by a factor of 1.6 on average. This suggests that N_2/CO_2 at 140 km has a local time variation. In order to verify the local time effect on CO_2, and N_2 densities and N_2/CO_2 ratio, additional data are needed.

3.3 Latitudinal variation

Additionally, we have investigated latitudinal dependences of CO_2, N_2, and N_2/CO_2. To constrain the local time effects found in Figure 4, local time is limited from 10:00 to 14:00. We selected the datasets that cover more than 60 degrees in latitude. Latitudinal dependences are investigated for observation in L_s = 80° - 100° and L_s = 340° - 360°. In Figure 5, plots of densities vs. latitude and N_2/CO_2 vs. latitude in the northern summer (L_s = 80° - 100°) are shown in Figures 5(a) through 5(c), and in the northern spring (L_s = 340° - 360°) are shown in Figures 5(d) through 5(f), respectively. It is noted that the latitudinal variation observed in the northern spring contains observations during both MY 32 and MY 33. MYs in the northern spring are distinguished by color in Figure 5. Uncertainties are shown by gray bars. In the northern summer, CO_2 densities become larger toward the northern polar region, which are 3.4×10^9 cm^{-3} around 45°S on average and 8.6×10^9 cm^{-3} around 25°N on average. Although N_2 densities have relatively large uncertainties, N_2 densities slightly increase toward the northern polar region, which are 3.5×10^8 cm^{-3} around 45°S on average and 4.3×10^8 cm^{-3} around 25°N on average. A larger amplitude of CO_2 density variations than that of N_2 density drives the latitudinal dependence of N_2/CO_2. The dayside N_2/CO_2 ratio in the northern summer is larger in the southern hemisphere (~0.11 around 45°S on average) and smaller in the northern hemisphere (~0.05 around 25°N on average). On the other hand, in the northern spring, these latitudinal dependences disappear. Because both CO_2 and N_2 densities have a similar latitudinal variation, N_2/CO_2 is almost constant with latitude. On the other hand, it is noteworthy that both CO_2 and N_2 densities in the southern hemisphere observed in MY 32 (red dots) are slightly higher than those in MY 33 (black dots). Higher densities in MY 32 might be associated with inflation at the upper atmosphere due to solar EUV heating in MY 32, for which EUV irradiance is about two times larger as seen in Figure 2(a). Our observation suggests that the latitudinal variation of N_2/CO_2 in the thermosphere has a strong seasonal dependence.

4 Discussion

4.1 Drivers of seasonal variation

In order to reveal drivers of the seasonal variations of N_2/CO_2 at 140 km, we have examined seasonal variations of surface N_2/CO_2, thermospheric temperature, and homopause altitude. To infer the effects of these parameters on N_2/CO_2 at 140 km, we use the same method described in Mahaffy, Benna, Elrod et al. (2015), Jakosky et al. (2017), and Slipski et al. (2018). Under assumptions of isothermal atmosphere and hydrostatic equilibrium, we can estimate vertical profiles of CO_2 and N_2 densities using observed densities at 140 km and thermospheric temperature. A vertical profile of mixing ratio is obtained from the CO_2 and N_2 profiles, which is expected to match the mixing ratio in the homosphere at the homopause altitude. The mixing ratio in the homosphere is assumed to be the mixing ratio at the surface. Therefore, densities of CO_2 and N_2 at 140 km observed from IUVS are functions of the surface N_2/CO_2, thermospheric temperature, and homopause altitude. For the sake of simplicity, a plausible range of these parameters to explain the observed N_2/CO_2 at 140 km is evaluated under the assumption that the other two parameters are fixed. We constrain the geometry from 10:00 to 14:00 in LT and from 45°S to 45°N in latitude to derive these parameters. Under the given constraints, mean values of N_2/CO_2 at 140 km derived from IUVS are 0.033 near perihelion (between L_s = 230° and 270°) and 0.056 near aphelion (between L_s = 50° and 90°), respectively. The results of our parameter study are summarized in Table 1.

We find that the surface N_2/CO_2 is required to vary by ~70% between 0.016 and 0.027 with season in order to explain the observed variation of N_2/CO_2 at 140 km between 0.033 and 0.056 if we assume an isothermal temperature of 180 K and a constant homopause altitude of 120 km. The seasonal variation of inferred surface N_2/CO_2 shows a similar trend with observed variation of N_2/CO_2 at 140 km, with a minimum value near perihelion and a maximum value near aphelion. In contrast, the typical surface value of mean N_2/CO_2 ratio is 0.0272 and surface N_2 mixing ratio has a seasonal variation in ±10%, which are measured by MSL (Trainer et al., 2019). Since we restricted the data between 45°S to 45°N here, an enhancement and a depletion of mixing ratio of trace gases around the polar region are out of scope. Trainer et al. (2019) showed that N_2 mixing ratio has its maximum at L_s ~ 160° and a second maxima-at L_s ~ 0°, responding to the condensation of the CO_2 polar cap (Hess et al., 1980; Ordonez-Etxeberria et al., 2017; Martínez et al., 2017). However, two peaks of N_2 mixing ratio shown in Trainer et al. (2019) do not agree with the timing and magnitude of seasonal variation of inferred surface N_2/CO_2 from this study. For these reasons, the seasonal variation of surface N_2/CO_2 cannot explain the observed N_2/CO_2 variation at 140 km.

It is also found that a plausible range of the variation of thermospheric temperature cannot explain the observed N_2/CO_2 variation when we assume a constant surface mixing ratio to be 0.020 (Mahaffy et al., 2013) and two cases of homopause altitudes (110 and 120 km). In the case of an assumed homopause altitude at 110 km (120 km), the thermospheric temperature is required to vary between 193 and 403 K (129 and 269 K), a difference of 210 K (140 K), due to the variation of N_2/CO_2 at 140 km. However, previous observations suggested that the dayside upper thermospheric temperature varies by ~70 K in the range from 170 to 240 K due to the effects of changing heliocentric distance and solar cycle (Bougher et al., 2017). In addition, Bougher et al. (2000; 2015) determined the seasonal variation of the exobase temperature to be ~40 K under the solar moderate condition ($F_{10.7}$ = 130) using numerical modeling. In addition to the upper thermosphere, temperature in the lower thermosphere around ~130 km observed in Gröller et al. (2018) varies between ~150 and ~200 K. The inferred range of thermospheric temperatures (210 K and 140 K) are clearly out of the plausible range reported by these previous studies and not sufficient to explain the observed seasonal variation of N_2/CO_2 at 140 km.

Thus, we propose that variations in homopause altitudes are most likely needed to reproduce the observed seasonal variation of N_2/CO_2 at 140 km altitude. Under the assumption of a constant surface mixing ratio of 0.020 and two cases of isothermal temperatures (150 and 200 K), inferred homopause altitudes are estimated between ~110 and ~130 km. Analysis of observations by MAVEN/NGIMS suggested that homopause altitudes on the dayside are located between ~100 and 140 km (Slipski et al., 2018). Their inferred homopause altitudes have seasonal differences of ~30 km. Inferred homopause altitudes in this study show a reasonable agreement with the results reported by Slipski et al. (2018). Consequently, we conclude that variations of the homopause altitude could be the dominant driver of changes in N_2/CO_2 in the thermosphere.

4.2 Estimation of homopause altitude

Hereafter, we determine homopause altitudes considering the variations of surface N_2/CO_2 and thermospheric temperature. Here, we use the N_2/CO_2 ratio at 50 km altitude, in the homosphere, predicted by Mars Climate Database (MCD) version 5.3 (Millour et al., 2018) instead of the surface N_2/CO_2. As the input for the MCD, IUVS observation geometries shown in Figure 1 are applied, and an average condition of solar EUV flux and a typical dust climatology are used. The mean mixing ratio of N_2 in the MCD is designed to match the observational result by MSL/SAM (Mahaffy et al., 2013). N_2/CO_2 reported in Mahaffy et al. (2013) is ~0.020 at L_s = 187.5°. Figures 6(a) and 6(b) show N_2/CO_2 in the homosphere derived from the MCD and temperature at 175 km observed from IUVS, respectively, which are used to infer homopause altitudes. The MCD predicts N_2/CO_2 values within the range between 0.012 and 0.022. The MCD N_2/CO_2 obtained around L_s = 270° in MY 33 (orbits near 4000) are remarkably low because N_2/CO_2 has a gradient toward the higher latitude in the MCD. Roughly speaking, the MCD predictions of N_2/CO_2 in the homosphere except for the data in the polar region are well matched with the observation of surface N_2/CO_2 from MSL in Mahaffy et al. (2013). Temperatures corresponding to an altitude of 175 km are determined from Chapman fits to CO_2^+ UVD brightness profiles observed by IUVS (Lo et al., 2015; Bougher et al., 2017). We have applied the 8607 profiles whose temperatures are valid. We assume an isothermal atmosphere and thus the temperature at 175 km is applied at all altitudes. The standard deviation of derived temperatures is ~50 K, which reflects a combination of atmospheric phenomena, such as nonmigrating tides (e.g., Lo et al., 2015) and thermal tides (e.g., Stevens et al., 2017). A temperature deviation of ~50 K corresponds to a deviation of ~10 km in the inferred homopause altitudes. Values of observed N_2/CO_2 at 140 km would not be expected to be lower than N_2/CO_2 in the homosphere based on an assumption of a well-mixed homosphere, thus we exclude these values from the ensemble data when inferring homopause altitudes. Although the number of excluded profiles is 17, that is negligible compared with the total number of profiles (9258). Inferred homopause altitudes, shown in Figure 6(c), are located between ~60 and 140 km. The uncertainty of inferred homopause altitudes is ± ~8 km based on the uncertainty of N_2/CO_2 at 140 km by IUVS, which is ~37% on average. The inferred homopause altitudes clearly have an anti-correlation with the variations of N_2/CO_2 at 140 km.

To evaluate the effect of an isothermal atmospheric assumption on the inferred homopause altitudes, we assume a constant gradient of temperature below 140 km but still an isothermal atmosphere above that altitude. The constant gradient of temperature is determined such that the temperature at 100 km should be 130 K (cf. Gröller et al., 2018; Forget et al., 2009). In this case, the inferred homopause altitudes are located between 105 and 140 km. It should be noted that the impact~~s~~ of assuming a non-isothermal atmosphere on the inferred homopause altitude is strong below 105 km, where the inferred altitude increases by up to ~45 km and by ~19 km on average. However, even when the non-isothermal atmosphere is assumed, a seasonal variation similar to Figure 6(c) is obtained.

Recently, Franz et al. (2017) revised the volume mixing ratio of both N_2 and CO_2 measured by MSL, which suggests that the surface N_2/CO_2 value is higher by a factor of ~1.5 than that measured by Mahaffy et al. (2013) and used by the MCD. We find that a factor of 1.5 increase in the surface mixing ratio uniformly increases the inferred homopause altitude by ~10 km.

Figure 7 shows time variations of pressure at 140 km as in Figure 2. As the pressure, CO_2 partial pressure derived from CO_2 density and temperature is applied. We assume an isothermal atmosphere; i.e., the temperature at 175 km derived from IUVS observations are used. Uncertainties of pressure are calculated from both that of CO_2 density and temperature, shown as gray bars. Pressure changes between ~4×10^{-6} and ~3×10^{-4} Pa and shows a seasonal sinusoidal trend; higher pressure appears around perihelion and lower pressure around aphelion. A similarity of seasonal trend between N_2/CO_2 and pressure would suggest that N_2/CO_2 variation at 140 km is mainly caused by variation of background pressure due to the inflation and contraction of the lower atmosphere. While we found that seasonal N_2/CO_2 ratio varies at a given altitude, we could not identify the seasonal variation at a given pressure level. It is noted that the N_2/CO_2 variation from MY 33 L_s = 340° to MY 34 L_s = ~30° is somewhat different to the pressure variation over the same period. This might imply that N_2/CO_2 could also change slightly via other mechanisms.

The CO_2 density at the inferred homopause altitude, extrapolated from the CO_2 density at 140 km, is shown in Figure 8, which are extrapolated from CO_2 density at 140 km altitude. The average value is 2.6×10^{11} cm^{-3} but it deviates between 10^{10} and ~10^{13} cm^{-3}. Distributions of CO_2 density at the inferred homopause altitude suggest that dayside homopause altitude changes would be generated by two factors: (1) the inflation and contraction of the lower atmosphere due to changing of solar irradiance that changes the homopause altitude at a fixed pressure level, (2) eddy diffusion via turbulence related to atmospheric wave activity as reported by Slipski et al. (2018) that changes the pressure level at the homopause. Although our study doesn't include dust storm events, the atmospheric heating due to the dust loading in the lower atmosphere, which absorbs the solar irradiation, could cause the change in homopause altitude and N_2/CO_2 variation in the upper atmosphere.

We have also inferred the latitudinal dependence of homopause altitudes. In Figure 9, plots of inferred homopause altitude vs. latitude are shown for the same seasons as in Figure 5. In Figure 9(a), the inferred homopause altitudes in the northern summer decrease toward the southern winter polar region. The difference in altitudes between northern and southern hemispheres is about ~30 km. In Figure 9(b), the latitudinal dependence of inferred homopause altitudes in the northern spring disappears. This is the first observational evidence of a latitudinal dependence of dayside homopause altitudes, which matches well with model predictions by González-Galindo et al. (2009). In their model, homopause altitude is defined where mixing ratio of CO_2 equals to 0.9. They mentioned that lower homopause altitudes in southern winter indicate a relative depletion of CO_2 in the upper atmosphere. Certainly, our observations showed that relative depletion of CO_2 in the southern hemisphere. González-Galindo et al. (2009) indicated that the depletion of CO_2 could result from the condensation of CO_2 due to colder temperatures or the meridional transportation of CO_2-poor air from the upper thermosphere. Latitudinal asymmetrical heating in the lower atmosphere due to its season would also be a potential driver of the latitudinal dependence of homopause altitude. The driver of the latitudinal dependence of homopause altitudes, however, cannot be fully constrained by our study.

4.3 Comparison of densities and composition with Mars Climate Database predictions

Figure 10 shows a comparison of densities of CO_2 and N_2, and N_2/CO_2 derived from IUVS observations with those predicted by the MCD using an average solar EUV input and a typical dust climatology for the corresponding geometries of the IUVS observations. IUVS observations are shown as black dots. The gray bars represent uncertainties of IUVS observations. The MCD predictions are shown as red dots. The MCD well reproduces the seasonal variation of N_2/CO_2 derived from IUVS observations throughout the observation period. However, the MCD overestimates the N_2/CO_2 value at 140 km by a factor of two. The possible reason for this discrepancy is that the homopause altitude is too low in the MCD. As can be seen in Figure 10, the MCD shows large discrepancies of N_2 densities between L_s = 30° and L_s = 70° in MY 33 with an opposite trend in the N_2 densities between L_s = 340° and L_s = ~40° in MY 33 to 34. The opposite trend found in this study suggests that the treatment of minor species in the model is not perfect, and further investigation is required to explain the contradiction with the observations reported here (private communication with F. González-Galindo, E. Millour, & F. Forget). Further studies of discrepancies between the model and observations offer opportunities to improve our understanding of atmospheric composition in the upper atmosphere.

5 Conclusions

In this paper, we have investigated dayside N_2/CO_2 at 140 km using MAVEN/IUVS limb scan observations from October 2014 to May 2018. We find that N_2/CO_2 at 140 km significantly varies in the range of 0.02 to 0.20 and clearly exhibits seasonal and latitudinal dependences. Seasonal dependence is observed throughout the MAVEN mission, which shows a sinusoidal trend that N_2/CO_2 is largest near aphelion (~0.063 on average between 50° and 90° in L_s) and smallest near perihelion (~0.031 on average between 230° and 270° in L_s). A latitudinal dependence of N_2/CO_2 is found in the northern summer. N_2/CO_2 increases toward the southern winter hemisphere by a factor of two (between latitude 45°S and 25°N). Meanwhile, the latitudinal dependence of N_2/CO_2 disappears in the northern spring.

In order to reveal the drivers of seasonal variations of N_2/CO_2 at 140 km, we have examined the effects of surface N_2/CO_2, thermospheric temperature, and homopause altitude. Homopause altitudes inferred from IUVS under the assumption of an isothermal atmosphere and a fixed surface N_2/CO_2 are estimated to vary between 100 and 130 km. The altitudes of inferred homopause and seasonal trends agree reasonably well with Slipski et al. (2018). The variations of the homopause altitude could be the dominant driver of changes in N_2/CO_2 ratios in the thermosphere. In addition, we inferred homopause altitudes by considering the variations of both surface N_2/CO_2 and thermospheric temperature by combining the MCD predictions with IUVS observations. Our inferred homopause altitudes vary between ~60 and 140 km altitude throughout the MAVEN mission and exhibit an anti-correlation with a seasonal sinusoidal trend of N_2/CO_2 at 140 km. Distributions of CO_2 density at the inferred homopause altitude suggest that dayside homopause is mainly controlled by inflation and contraction of the lower atmosphere. Variations of observed N_2/CO_2 ratio at 140 km are associated with CO_2 number density, namely pressure, at 140 km. Thus, we suggest that inflation of the lower atmosphere near perihelion causes an increase in the dayside homopause altitude and a decrease in the mixing ratio of N_2 in the upper atmosphere. Our results show, for the first time, the clear evidence for a latitudinal dependence of homopause altitudes in the northern summer. The inferred homopause altitudes in the northern summer decrease toward the southern winter polar region. The difference in altitudes between northern and southern hemispheres is about ~30 km. On the other hand, the latitudinal dependence of inferred homopause altitudes in the northern spring disappears.

We have compared IUVS observational results with model predictions of N_2/CO_2 from MCD v5.3. The MCD well reproduces the seasonal variation of N_2/CO_2 derived from IUVS observations throughout the MAVEN mission. However, the MCD overestimates IUVS N_2/CO_2 values by a factor of two. A possible reason for this discrepancy is that the homopause altitude is systematically low in the MCD. The observation of the N_2 mixing ratio at 140 km by MAVEN/IUVS offers a source for improvements to atmospheric composition in global circulation models.

IUVS observations imply significant impacts to the variations of the homopause altitude on atmospheric composition in Mars' upper thermosphere. Our results support the idea that atmospheric composition in the exosphere varies with the homopause altitude. Understanding the influence of the homopause altitude on the escape of minor species should be clarified in future work.

Data Availability Statement

MAVEN/IUVS limb level 2 version 13 revision 1 data used in this study are available through the NASA Planetary Data System (PDS) at https://atmos.nmsu.edu/data_and_services/atmospheres_data/MAVEN/limb.html. (Schneider, N. et al., MAVEN IUVS Derived-level Data product Bundle, urn:nasa:pds:maven.iuvs.processed 2015). MAVEN/EUVM level 2 version 11 revision 4 data are available through the NASA PDS at https://pds-ppi.igpp.ucla.edu. (Eparvier, F. G., MAVEN EUV calibrated irradiances Data Collection, urn:nasa:pds:maven.euv.calibrated:data.bands 2020) The full version of MCD v5.3 was provided from E. Millour. The results retrieved from the IUVS limb observation and extracted from the MCD used in this article are available at https://scholar.colorado.edu/concern/datasets/q237ht027, (Yoshida et al., 2020).

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Figure 1. The tangential footprint of IUVS limb observation analyzed in this paper. (a) Solar longitude (L_s; deg), (b) local time (hours), (c) SZA (deg), and (d) latitude (deg). The horizontal axis shows orbit number (bottom) and corresponding Earth date (top).

Figure 2. Time variations of (a) solar EUV irradiance (mW m^{-2}; black), the Sun-Mars distance (AU; blue), number densities (cm^{-3}) of (b) CO_2, (c) N_2, and (d) N_2/CO_2 at 140 km altitude. Uncertainties are shown by gray bars in (b, c, d). Colors in (b, c, d) represent the local time (hours) of observation. The lower horizontal axis shows L_s and Mars Year (MY) and that of top shows corresponding Earth date.

Figure 3. The variations of (a) CO_2 and (b) N_2 density, and (c) N_2/CO_2 at 140 km altitude as a function of L_s. IUVS observations are constrained at specific local time bins, including 8:00 - 10:00 (purple), 12:00 - 14:00 (light blue), and 16:00 - 18:00 (yellow). Data obtained in MY 32, 33, and 34 are indicated with triangles, diamonds, and squares, respectively.

Figure 4. Local time variations of CO_2, N_2, and N_2/CO_2 at 140 km altitude. Uncertainties are shown by gray bars. Data shown are constrained to three seasons. (a, b, c) Observations from L_s 30° to 70° in MY 33, which correspond with orbits around 1800 - 2200. (d, e, f) Observations from L_s 120° to 160° in MY 33, which correspond with orbits around 2900 - 3500. (g, h, i) Observations from L_s 340° to 40° in MY 33 through 34, which correspond with orbits around 4800 - 5500.

Figure 5. Latitudinal variations of CO_2, N_2, and N_2/CO_2 at 140 km altitude. Uncertainties are shown by gray bars. Local time is limited from 10:00 to 14:00. (a, c, e) L_s is limited between 80° and 100°, which corresponds to the northern summer. (b, d, f) L_s is between 340° and 360°, which corresponds to the northern spring. MYs observed between 340° and 360° are distinguished by colors; red is MY 32 and black is MY 33.

Figure 6. Time variations of (a) N_2/CO_2 at 50 km, in the homosphere, predicted by the MCD, (b) temperature (K) at 175 km observed by IUVS, and (c) inferred homopause altitude (km) considering the effect of variations of N_2/CO_2 in homosphere, isothermal thermospheric temperature, and N_2/CO_2 at 140 km.

Figure 7. The same as Figure 2 (b) but CO_2 partial pressure at 140 km altitude is shown. Corresponding CO_2 partial pressure is calculated to be based on IUVS observations both CO_2 density at 140 km and temperature at 175 km under the assumption for isothermal atmosphere.

Figure 8. Time variations of CO_2 density at inferred homopause altitude shown in Figure 6. CO_2 density is extrapolated from the IUVS observation.

Figure 9. (a, b) Latitudinal variations of homopause altitude inferred from IUVS observation. (a) L_s is limited between 80° and 100°. (b) L_s is between 340° and 360°. The limitation in local time is the same as in Figure 3.

Figure 10. Comparisons between IUVS observations and the MCD prediction. (a) CO_2 density (cm^{-3}) at 140 km, (b) N_2 density (cm^{-3}) at 140 km, and (c) N_2/CO_2 at 140 km. IUVS is shown as black dots and the MCD as red dots. Uncertainties of IUVS observations are represented by gray bars. The MCD prediction is for average solar EUV and climatology dust conditions. The geometries of the MCD input are provided by the footprints of IUVS observations shown in Fig. 1.

Table 1. The results of the parameter study. Typical values required to explain N_2/CO_2 at 140 km observed by MAVEN/IUVS are shown. Among three parameters considered (surface N_2/CO_2, thermospheric temperature, and homopause altitude), a plausible value for the parameter of interest in each case is evaluated while the other two parameters are fixed.

Case	Object parameter	Season	Fixed parameters		Result
1	Surface N_2/CO_2 ratio	Perihelion	Temperature; 180[K]	Homopause altitude; 120 [km]	0.016
2	Surface N_2/CO_2 ratio	Aphelion	Temperature; 180[K]	Homopause altitude; 120 [km]	0.027
3	Temperature	Perihelion	Surface N_2/CO_2; 0.02	Homopause altitude; 110 [km]	403 [K]
4	Temperature	Perihelion	Surface N_2/CO_2; 0.02	Homopause altitude; 120 [km]	269 [K]
5	Temperature	Aphelion	Surface N_2/CO_2; 0.02	Homopause altitude; 110 [km]	193 [K]
6	Temperature	Aphelion	Surface N_2/CO_2; 0.02	Homopause altitude; 120 [km]	129 [K]
7	Homopause altitude	Perihelion	Surface N_2/CO_2; 0.02	Temperature; 150 [K]	129 [km]
8	Homopause altitude	Perihelion	Surface N_2/CO_2; 0.02	Temperature; 200 [K]	126 [km]

9	Homopause altitude	Aphelion	Surface N_2/CO_2; 0.02	Temperature; 150 [K]	117 [km]
10	Homopause altitude	Aphelion	Surface N_2/CO_2; 0.02	Temperature; 200 [K]	109 [km]