Mars Climate Sounder Observations of Gravity-wave Activity throughout Mars’s Lower Atmosphere

Nicholas G. Heavens1,2, Alexey Pankine1, J. Michael Battalio3, Corwin Wright4, David M. Kass5, Armin Kleinböhl5, Sylvain Piqueux5, and John T. Schofield5,6

1 Space Science Institute, 4765 Walnut Street, Suite B, Boulder, CO 80301, USA; nheavens@spacescience.org
2 Department of Earth Science and Engineering, Imperial College, London, UK
3 Department of Earth and Planetary Sciences, Yale University, New Haven, CT, USA
4 Centre for Space, Atmospheric and Oceanic Science, University of Bath, Bath, UK
5 NASA Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA, USA

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Abstract

Gravity waves are one way Mars’s lower atmospheric weather can affect the circulation and even composition of Mars’s middle and upper atmosphere. A recent study showed how on-planet observations near the center of the 15 μm CO2 band by the A3 channel (635–665 cm−1) of the Mars Climate Sounder on board the Mars Reconnaissance Orbiter could sense horizontally short, vertically broad gravity waves at ∼25 km above the surface by looking at small-scale radiance variability in temperature-sensitive channels. This approach is extended here to two additional channels closer to the wings of the 15 μm CO2 band, A1 (595–615 cm−1) and A2 (615–645 cm−1), to sense gravity waves throughout the lower atmosphere. Using information from all three channels demonstrates that gravity-wave activity in Mars’s lowermost atmosphere is dominated by orographic sources, particularly over the extremely rough terrain of Valles Marineris. Much of this orographic population is either trapped or filtered in the lowest two scale heights, such that variations in filtering and nonorographic sources shape the gravity-wave population observed at 25 km above the surface. During global dust storms, however, gravity-wave activity in the first scale height decreases by approximately a factor of 2, yet trapping/filtering of what activity remains in the tropics substantially weakens. Exceptionally high radiance variability at night in the tropics during the less dusty part of the year is the result of observing mesospheric clouds rather than gravity waves.

Unified Astronomy Thesaurus concepts: Mars (1007); Planetary atmospheres (1244); Atmospheric variability (2119)

1. Introduction

In a planetary atmosphere, disturbances in a stably stratified fluid can be restored by buoyancy, resulting in the formation of an atmospheric gravity wave (GW; Fritts & Alexander 2003). A GW can propagate vertically through the atmosphere and become unstable, transporting energy and momentum from lower to higher levels of the atmosphere (e.g., Holton et al. 1995; Yamanaka 1995; Yiğit & Medvedev 2015; Medvedev & Yiğit 2019). In the atmospheres of the Earth and Mars, GW activity can have a variety of analogous dynamical consequences, including the closure of middle atmospheric jets (e.g., Holton 1982; Barnes 1990; Medvedev et al. 2011), the formation of middle atmospheric clouds under otherwise unfavorable thermodynamical conditions (e.g., Fritts et al. 1993; Spiga et al. 2012), and the setting of the vertical extent of the homopause altitude (e.g., Offermann et al. 2006; Slipski et al. 2018). In the case of Mars, turbulence generated by GW breaking enhances mixing at middle to upper atmospheric altitudes (70–140 km), where eddy diffusion is otherwise weakening (Slipski et al. 2018).

The potential impact of GW activity on homopause altitude suggests that understanding GW activity is necessary for understanding not only the behavior of the present-day atmosphere but also the evolution of Mars’s atmospheric composition through time. As noted by Slipski et al. (2018), an atmosphere with a higher homopause, i.e., one in which atmospheric gases are well mixed to higher altitude, is one in which escaping species are less isotopically fractionated. In addition, GWs in the upper atmosphere strongly affect exospheric temperatures locally and possibly enhance atmospheric escape (Parish et al. 2009; Walterscheid et al. 2013; England et al. 2017; Williamson et al. 2019; Leelavathi et al. 2020; Yiğit et al. 2020).

In the upper atmosphere, GW activity is diurnally and seasonally variable. It is strongest, on average, during southern spring afternoons, possibly because of strong dust storm activity during that season (Liu et al. 2019). However, the relationship between upper atmospheric GW activity and dust storm activity is complicated by the effects dust storms can have on the density structure of the thermosphere (Liu et al. 2019). Nevertheless, there is independent evidence that global dust storms of 2001 and 2018 significantly raised the altitude below which Mars’s atmospheric composition was dominated by higher molecular mass CO2 as opposed to lower molecular mass O (Xu et al. 2015; Elrod et al. 2020). (The dust storms of 2001 and 2018 will be referred to hereafter as 25P and 34P, because they occurred in Mars Years (MYs) 25 and 34 in the sense of Clancy et al. 2000; Piqueux et al. 2015a.) Mars’s well-mixed atmospheric composition is dominated by CO2, so CO2 dominance at higher altitude implies a higher homopause altitude and thus potential breaking of GWs at
higher altitude, a type of seasonal variability observed by Liu et al. (2019) and reported in 34P by Elrod et al. (2019). In addition, Leelavathi et al. (2020) and Yiğit et al. (2020) recently showed that upper atmospheric GW amplitudes increased by a factor of 2 or more during 34P, despite the high likelihood that GWs would be more strongly filtered by convective instability in the warmer conditions near 100 km that prevailed during 34P (England et al. 2017; Vals et al. 2019; Leelavathi et al. 2020; Yiğit et al. 2020).

One pathway to understanding variability in upper atmospheric GW activity is to study the observational record of GW activity in the lower and middle atmosphere. Because of various dissipative processes and the possible production of secondary GWs during breaking (e.g., Vadas & Fritts 2001; Chun & Kim 2008; Vadas et al. 2018; Heale et al. 2020), the upper atmospheric GW population can be very different from what is observed in the lower and middle atmosphere. Nevertheless, observing the lower and middle atmospheric GW population is necessary to accurately predict the upper atmospheric population from first principles, as well as to simulate the effects of GWs through the whole atmosphere, because this population is ultimately what is being dissipated and forming secondary GWs (e.g., Medvedev et al. 2011; Yiğit et al. 2015; Imamura et al. 2016; Kuroda et al. 2016, 2019; Gilli et al. 2020).

Recently, Heavens et al. (2020) argued that the observational record of lower and middle atmospheric GW activity at Mars could be vastly expanded by using small-scale variability in infrared observations sensitive to temperature as a proxy for GW activity. As reviewed by Heavens et al. (2020), previous studies of GW activity in the lower and middle atmosphere had been based on relatively infrequent occultation-type measurements (e.g., Creasey et al. 2006; Altieri et al. 2012), so leveraging the high frequency of the measurements by mapping instruments like the Thermal Emission Spectrometer on board the Mars Global Surveyor or the Mars Climate Sounder (MCS) on board the Mars Reconnaissance Orbiter (MRO) could have major advantages over radio occultation, even if there were no difficulty in distinguishing tidal from GW oscillations in radio occultation profiles.

Heavens et al. (2020) also provided a proof of concept of this idea by analyzing the variance in calibrated radiance (expressed as brightness temperature) in the MRO-MCS A3 channel (635–665 cm\(^{-1}\); near the center of the 15 \(\mu\m\text{m}\) CO\(_{2}\) band) in nadir geometry and an “off-nadir” geometry, in which the surface was viewed at emission angles of \(\approx 70^\circ\). The MRO-MCS was designed to scan the limb, as well as observe the surface, with a detector array divided into lines of 21 detectors corresponding to an individual channel (McCleese et al. 2007). In the limb, these 21 detectors observe the atmosphere from the side at a vertical resolution of 5 km, where detector 1 is at the top and detector 21 is at the bottom. In nadir geometry, these detectors observe the surface at a horizontal resolution near 1 km, with detector 1 being most forward in the direction of spacecraft motion and detector 21 being farthest aft. In the typical off-nadir geometry, the horizontal resolution stretches to \(\approx 2.9\) km (Hayne et al. 2012). Off-nadir observations from MRO-MCS far outnumber nadir observations because of elevation actuator problems after the first 3 months of science operations, which prevented MRO-MCS from regularly scanning to nadir thereafter (Jau & Kass 2008). Off-nadir views in A3 were simulated to capture GW activity at \(\approx 25\) km above the surface, and nadir views were simulated to capture GW activity \(5\) km lower. Simulations suggested that the nadir views were sensitive to GWs with horizontal wavelengths of \(10–30\) km and vertical wavelengths no less than \(35\) km, while the off-nadir views were sensitive to GWs with horizontal wavelengths of \(10–100\) km and vertical wavelengths around half of the corresponding horizontal wavelength.

However, these simulations were based on a parameterization of the A3 vertical weighting function in off-nadir geometry derived from the calculated nadir weighting function and did not consider possible changes to the vertical weighting function under high dust opacity conditions. Incorrectly estimating the off-nadir weighting function under clear conditions risks incorrect assessment of the vertical wavelength of the observed GWs and an incorrect estimate of their propagation potential into the thermosphere. Imamura et al. (2016) suggested that GWs with vertical wavelengths \(<20\) km would be strongly filtered during the transition between the middle and upper atmosphere, which is somewhat supported by the finding of Siddle et al. (2019) that the dominant vertical wavelength in the thermosphere is \(\approx 20\) km. Neglecting changes in vertical weighting function under high dust opacity conditions risks misinterpreting the altitude to which the GW observations are sensitive, which could result in mistaking changes in the altitude of sensitivity for changes in GW activity where GW dissipation is strong.

From analyzing brightness temperature variances in on-planet observations in the A3 channel, Heavens et al. (2020) concluded that (1) strong GW activity followed the westerly jets in the extratropics and tropical easterly jet over the Tharsis volcanoes, (2) GW activity in areas of climatologically low–moderate GW activity strongly decreased during regional and global dust storms (contradicting the inference from the upper atmospheric GW observations but also raising questions about the effect of dust on the vertical weighting function), and (3) strong but infrequent nighttime GW activity appears to be high in some parts of the tropics during much of the year, which could indicate either a significant GW source there, such as convective water-ice clouds driven by radiative cooling (Spiga et al. 2017), or contamination of the analysis by reemission of radiation by CO\(_{2}\) clouds in the middle atmosphere.

An additional yet unhighlighted point that could be drawn from the analysis of Heavens et al. (2020) was the strong mismatch between the apparent distribution of GW activity and a “topographic hypothesis” based on the estimated wind stress. This aspect of the analysis recapitulated an argument made by Creasey et al. (2006) that if observed GWs were orographically generated, their distribution should follow the global distribution of wind stress, which is strongest over Valles Marineris. Thus, GW activity at 25 km and, by extension, GWs that reach the middle atmosphere (30–50 km altitude) are principally generated by nonorographic sources (presumably because GWs from orographic sources are nonexistent or have been strongly filtered).

The purpose of this study is to reassess the multiannual record of GW activity presented by Heavens et al. (2020), focusing on the three areas of uncertainty outlined above rather than on all possible aspects of GW climatology that could be addressed with the data, i.e., whether (1) GW activity in the lower atmosphere increases or decreases during regional and global dust storm activity, (2) there is a strong source of GW...
activity in the nighttime tropics due to water-ice cloud convection driven by radiative cooling, and (3) GW activity appears to be controlled by nonorographic sources throughout the lower atmosphere. To make this reassessment, we will analyze variance in nadir and off-nadir views by MRO-MCS in a broader range of channels that includes A3. The A1 (595–615 cm\(^{-1}\)) and A2 (615–645 cm\(^{-1}\)) channels in the wings of the 15 \(\mu\)m CO\(_2\) band will be analyzed to sense GW activity/mesoscale variability in the lower atmosphere below 25 km. As noted by Heavens et al. (2020), A1 and A2 are also more transparent to CO\(_2\) ice than A3, so these channels might be able to be used to look for GW activity related to nighttime water-ice cloud convection in the tropics without the ambiguity introduced by the presence of middle atmospheric CO\(_2\) ice clouds.

The outline of this paper is as follows. In Section 2, the sensitivity of A1–A3 to GW activity under low and high dust conditions will be evaluated, and the importance of the surface contribution to variance analysis over high-elevation surfaces will be assessed. In Section 3, the analysis of the data set and any other analyses necessary for its interpretation will be described. In Section 4, the results of the analysis that are relevant to the four areas of focus of the study will be presented. These results will be discussed in Section 5 and summarized in Section 6.

2. Sensitivity of the Observations to GW Activity

2.1. Visibility Analysis

The GWs propagate with different wavelengths and angles within a planetary atmosphere, which can be incompletely resolved by a remote-sensing observation (Wu et al. 2006). This effect can be captured and quantified by simulating a quantity called visibility, which is the proportion of brightness temperature variance due to GWs that is recovered by an observational method as a function of GW horizontal and vertical wavelength (Wu & Eckermann 2008). To calculate the visibility of GW activity in channels A1–A3, the basic technique outlined by Heavens et al. (2020) was followed. Here the relevant details about the MRO-MCS observations will be provided in the proper place, but a more thorough account is provided by Heavens et al. (2020) and will not be fully recapitulated here.

2.1.1. Radiative Transfer Modeling

Radiative transfer modeling was used to calculate the vertical and horizontal weighting functions for nadir and off-nadir observations by MRO-MCS in the relevant channels. It was assumed that off-nadir observations were made at a typical angle of 8°9 below the limb, which corresponds to a surface emission angle of 67°04′ in the detector closest to the surface, 21°, and 73°20′ in the detector farthest from the surface, 1°. The surface contributions were estimated from the difference between the sum of the vertical weighting functions and unity.

These weighting function calculations were made for three input atmospheres, two of which were the Kliore (1978) standard atmospheres for the northern and southern hemispheres (compiled under the auspices of the Committee on Space Research, COSPAR, and thus often referred to as the COSPAR standard atmospheres), which were assumed to be clear of aerosol. The third atmosphere enabled calculation of the weighting functions under extreme aerosol conditions by adopting a profile from the Mars Climate Database (MCD) for \(L_s = 280°\) in the Martian Year 28 scenario at 45°S, 40°E at 15:00 LST (Millour et al. 2015; MCD, cited 2022a). This profile approximates conditions during the mature phase of 28P in the southern hemisphere. The total visible dust column optical depth in the profile is 2.69, which, while on high end for this scenario and MCD scenarios in general, may significantly underestimate typical dust column opacities in global dust storms. The Mars Exploration Rovers, Spirit and Opportunity, measured maximum visible opacities near 5 during 28P, while the Mars Science Laboratory, Curiosity, measured a maximum visible opacity of \(\approx 8\) (Guzewich et al. 2019) in 34P. That said, unlike the rover observations or satellite observations that could potentially have been used to supplement them to build an input atmosphere, the MCD scenario provides a physically consistent temperature profile along with a dust profile over the entire atmospheric column, which is essential to the radiative transfer modeling.

Dust was incorporated into the radiative transfer modeling by assuming a 2 \(\mu\)m dust particle size with optical properties corresponding to those derived by Wolff & Clancy (2003). The resulting column optical depth in the dusty profile at 9.3 \(\mu\)m under these assumptions is 1.5.

The simulated A3 vertical weighting functions agree well with those previously simulated by Kleinböhl et al. (2010), which were used by Heavens et al. (2020) as the basis of their visibility calculations (Figures 1(a) and 2(c) and (f)), particularly in the northern hemisphere. This close agreement enables confident assessment of the altitude to which each channel is most sensitive by looking at the peak of the vertical weighting function.

In the standard atmospheres, the nadir views peak at the surface in A1 (with a strong additional surface contribution of up to 40% at 10 km in A2 and 21 km in A3; Figures 1(a) and (b)). In dusty conditions, the altitude of peak sensitivity increases in all channels by 5–10 km because of the additional opacity from dust, so that A1’s vertical weighting function under dusty conditions is similar to that of A2 under clear conditions, though A2’s vertical weighting function does not rise enough to match A3’s vertical weighting function under clear conditions (Figures 1(a)–(c)).

Off-nadir vertical weighting functions for the lowest and highest detectors in off-nadir geometry slightly diverge at low and high altitudes but peak at nearly the same altitude \(\approx 5\) km above the corresponding vertical weighting function in the nadir. Surface contributions are smaller than the nadir by a factor of 2–3 in A1 and A2 but similar in A3. As in the nadir case, dusty conditions raise the peak altitude of sensitivity in the off-nadir, such that the A2 and A1 vertical weighting functions under dusty conditions approximate the A3 and A2 vertical weighting functions under clear conditions (Figures 2(a)–(i)).

Thus, A1 will be most sensitive to near-surface variability in clearer conditions and 15 km in dusty conditions, A2 will be most sensitive to variability at 15 km in clearer conditions and 25 km in dusty conditions, and A3 will be most sensitive to variability at 25 km in clearer conditions and 30 km in dusty conditions.

The horizontal weighting functions for the off-nadir observations show that off-nadir observations are capturing information over a 100–200 km range from the center of the observation at the surface (Figures 3(a)–(i)). One consequence
of this is that if a diagnosis of GW activity is assigned a position in space based on the intersection of detector 11’s optical path with the surface (the center of the line of 21 detectors), the true position of the measurement projected to the surface is actually 30–90 km closer to the subspacecraft point than assumed.

Unlike weighting functions in nadir geometry, those in off-nadir geometry are slanted, which is easily illustrated by constructing two-dimensional weighting functions for nadir and off-nadir views in an \(x-z\) plane of 1000 × 100 km binned at 1 km resolution (Figures 4(a) and (b)). The surface contribution was added to the lowest level of the vertical weighting.
function. The vertical weighting functions were then distributed across bins to account for the width of the off-nadir optical path (≈3 km) and smearing of each measurement by spacecraft motion during each 2.048 s measurement (≈6 km) and then renormalized to sum to unity. Two-dimensional weighting functions like this were constructed for each channel, detector, and atmosphere.

2.1.2. GW-perturbed Model Atmospheres

To evaluate the sensitivity of the GW diagnosis technique, GW-perturbed model atmospheres were constructed over the same $x-z$ domain as the two-dimensional weighting functions corresponding to the three model atmospheres used to calculate the weighting functions. In other words, the unperturbed atmospheric state was the horizontally uniform model atmosphere used to calculate the weighting function. The perturbed

Figure 3. Comparison of the horizontal weighting functions for the MRO-MCS A1–A3 channels and off-nadir geometry for detectors 1, 11, and 21 for the three input atmospheres, as labeled. The weighting functions have been plotted so that they are relative to the intersection point of the optical path of detector 11 (the central detector) with the surface.

Figure 4. Two-dimensional weighting functions (km$^{-2}$) for the MRO-MCS A3 channel and off-nadir geometry for detectors 21, 11, and 1 for the COSPAR northern hemisphere atmosphere, as labeled. The weighting functions have been plotted so that they are relative to the intersection point of the optical path of detector 11 (the central detector of the line of detectors in a channel) with the surface. Intersection points with the surface for each of the plotted detectors are indicated with white markers at 0 km altitude.
state was constructed by adding a 1 K amplitude \( \tilde{T} \) GW in the \( x-z \) plane whose temperature perturbation, \( T' \), had the functional form

\[
T' = \tilde{T} e^{(ikx + mz)},
\]

where \( k \) and \( m \) are the horizontal and vertical wavenumbers.

This equation is a simplification of Equation (1) of Heavens et al. (2020), which accounts for waves being observed much faster than their intrinsic period so that they can be regarded as being observed at a fixed instant of time equivalent to \( t = 0 \).

By varying the wavenumbers, the visibility was then calculated for each model atmosphere and channel by (1) convolving the weighting functions for each detector with the model atmosphere to calculate the simulated brightness temperature; (2) calculating the brightness temperature variance measured across the simulated detector array according to the protocol outlined by Heavens et al. (2020); (3) normalizing this variance by 0.5 K\(^2\), the variance of a sine or cosine wave with an amplitude of 1 K; and (4) converting to a percentage.

2.1.3. Results of the Visibility Analysis

Although calculated more rigorously, the results of the visibility analyses are consistent with those of Heavens et al. (2020). Nadir views in A2 and A3 are most sensitive to horizontal wavelengths, \( \lambda_h \), of 10–30 km and vertical wavelengths, \( \lambda_v \), of >50 km (Figures 5(b), (c), (e), (f), (h), and (i)). Substantial visibility for A1 under clear conditions extends to the smallest vertical wavelengths sampled but is strongest at \( \lambda_h = 10–30 \) and \( \lambda_v > 50 \) km (Figures 5(a) and (d)). Under dusty conditions, A1 visibility is similar to A2 and A3 visibility under all circumstances (Figures 5(g)).

Following Heavens et al. (2020), visibility plots like this can be interpreted to constrain the intrinsic period of the observed waves by adopting the relation that for \( \lambda_v \) sufficiently smaller than 120 km in Mars’s atmosphere, the intrinsic frequency, \( \Omega_{GW} \) (wind-relative frequency of a GW), is

\[
\Omega_{GW} = \frac{N}{\sqrt{1 + \left( \frac{\lambda_v}{\lambda_h} \right)^2}},
\]

(Fritts & Alexander 2003), where \( N \) is the Brunt–Väisälä frequency (0.008 s\(^{-1}\); Imamura et al. 2007; Ando et al. 2012).

So the A2 and A3 channels in nadir geometry generally observe GWs with periods <20% greater than \( N^{-1} \), that is, a few minutes, while A1 has some sensitivity to periods of up to 1 hr (30\( N^{-1} \)), though generally much shorter.

The calculated off-nadir visibilities also have forms very similar to those presented in Heavens et al. (2020; Figures 6(a) –(j)). There is some sensitivity starting at a \( \lambda_h \) of 10 km and \( \lambda_v \) of 5 km and continuing along a parabolic slope to a \( \lambda_h \) of 80 km and \( \lambda_v \) of 30 km, where there is weak visibility at larger vertical wavelengths, particularly in A1 and A2. The magnitude

![Figure 5. Comparison of the estimated visibility to GW (%) with the given horizontal and vertical wavelengths of the MRO-MCS A1–A3 channels and nadir geometry for the three input atmospheres, as labeled.](image-url)
of visibility generally peaks near a $\lambda_h$ of 30–40 km and $\lambda_z$ of 10–20 km. The periods corresponding to this range are $\approx (10N - 1)$ or 20 minutes.

Under clear atmospheric conditions, A1 can have a peak visibility $>100\%$ and substantial visibility at all wavelengths (Figures 6(a) and (d)). This is a consequence of its weighting function sloping and peaking near the surface, which introduces detector-to-detector oscillations connected to the increase or decrease of temperature with altitude as opposed to the oscillations introduced by the GW perturbations. This effect vanishes if the temperature profile is vertically uniform and can be strengthened or weakened by adjusting the lapse rate near the surface in the visibility simulations. Thus, A1 could be sensitive not just to GW variability but also to the vertical lapse rate in the lowest 15–20 km of the atmosphere. Peak visibility is $\approx 75\%$ in A2 and $\approx 60\%$ in A3 in clear conditions.

3. Methods

3.1. Analysis of the MRO-MCS Level 1B Data Set

The MRO-MCS Level 1B data set (MCD. cited 2022b) was analyzed according to the procedure outlined in Heavens et al. (2020) with a few modifications.

The data analyzed were restricted to all forward, in-track (180° azimuth), on-planet (scene altitude of zero) observations in the A1, A2, A3, and B1 channels between the beginning of the mission and $L_e = 128^\circ 84$ of MY 35 (the end of 2019). Nadir observations were defined as on-planet observations with elevations between 177° and 183°. Off-nadir observations were defined as on-planet observations with elevations between 117° and 123°. In addition, observations were only included if they had “Gqual” and “Moving” flags equal to zero and a top detector (detector 1) radiance greater than zero.

To exclude the calibration-related artifacts described by Heavens et al. (2020), the observations were filtered by including on-planet radiance measurements only if the detector 1 radiances in the A1, A2, A3, and B1 channels in the preceding space view had a magnitude less than 0.4 mW m$^{-2}$ (cm$^{-1}$)$^{-2}$ sr$^{-1}$; the sum of the radiances over all detectors in each of the A1, A2, A3, and B1 channels in the preceding space view had a magnitude less than 1.52 mW m$^{-2}$ (cm$^{-1}$)$^{-2}$ sr$^{-1}$; and the prior and subsequent limb views had detector 1 radiances in the A1, A2, A3, and B1 channels greater than $-0.4$ mW m$^{-2}$ (cm$^{-1}$)$^{-2}$ sr$^{-1}$.

In Heavens et al. (2020), these criteria were only applied to A3. Heavens et al. (2020) noted that the first and third criteria are based on the noise threshold of the instrument. The second is presumably empirical but likewise should scale with instrument noise. These criteria seemed reasonable to apply to all channels because the estimated radiance noise in A1 (the noisiest channel among the four studied) is only 33% greater than that of A3 (Kleinböhl et al. 2009). An additional consequence of this is that the uncertainties in GW variance diagnoses for the channels other than A3 will be within a factor of 2 of those in A3 for equivalent temperatures.

We identified additional calibration-related artifacts associated with breaks in the normal observation pattern of on-planet views that lasted an orbit or more. We therefore excluded any pair of on-planet views that took place more than 300 s after the previous set of on-planet views. This filter reduced the number of diagnoses by $<0.1\%$ from the unfiltered condition.
Then, $\Omega_{GW}$ (the GW variance in an observation), $\epsilon_{GW}$ (the 1σ uncertainty in GW variance in an observation), $\bar{\Omega}_{GW}$ (the GW variance in an observation normalized by the mean of the squared temperature in each measurement within the observation), $\epsilon_{GW}^\text{norm}$ (the 1σ uncertainty in the normalized variance), and relevant time/location/elevation/roughness information were calculated for each individual observation. The variance and error calculations were made for each of the A channels in the same way as in Heavens et al. (2020), just using the appropriate band central wavenumber and radiance error in Kleinböhl et al. (2009) for each individual channel studied. However, the central wavenumber for B1 was varied by detector to account for the more variable intrachannel spectral response of B detectors (Kleinböhl et al. 2011).

The position of each detector in each channel relative to the position of the scene observed by the center of the detector array was computed by reprojecting the MCS detector array in Cartesian space along the orbital track of the spacecraft, taking account of crossing the anti-prime meridian. The mean latitude and longitude of the observation were then estimated to be the position of detector 11, the central detector of each channel. The mean elevation relative to the areoid was calculated by interpolating the positions of each detector on the 16 point deg$^{-1}$ Mars Orbiter Laser Altimeter (MOLA) elevation map available from the NASA Planetary Data System (Smith et al. 2003) and taking the mean of the elevations for all detectors.

The maximum elevation in all detectors was also recorded to enable identification of observations with potentially significant surface contributions. The roughness at the scale of the observation was then calculated from the variance of the elevations corresponding to the detectors. Nightside observations were distinguished from dayside observations by determining from the scene location information on whether the spacecraft was ascending in latitude on the dayside or determining from the scene location information on whether the maximum elevation in all detectors was also recorded to enable identification of observations with potentially significant surface contributions. The roughness at the scale of the observation was then calculated from the variance of the elevations corresponding to the detectors. Nightside observations were distinguished from dayside observations by determining from the scene location information on whether the spacecraft was ascending in latitude on the dayside or descending in longitude on the nightside.

One consequence of simultaneously applying the space view–based filter to A1, A2, A3, and B1 was to substantially reduce the density of coverage. This can be illustrated by comparing the number of total off-nadir diagnoses between our survey and that of Heavens et al. (2020) over the period studied by Heavens et al. (2020; MY 28, $L_s = 111°2823$–MY 34, $L_s = 232°643$). In Heavens et al. (2020), there were 6,661,256 total off-nadir diagnoses on the nightside, compared to 6,314,719 in our survey (5.2% less); on the dayside, there were 6,994,855 total diagnoses by Heavens et al. (2020) but only 4,573,202 diagnoses in our survey (34.6% less). Therefore, the survey was repeated with A1, A2, and A3 alone, to eliminate the need to include B1 in the filter. There were 6,646,544 nightside diagnoses over the period considered by Heavens et al. (2020; 0.2% less), while there were 6,963,000 dayside diagnoses over the same period (0.5% less). Therefore, the A1–A3 channel data set is nearly identically dense in coverage to the A3 data set of Heavens et al. (2020), with typically 50,000–170,000 dayside or nightside observations in a bin of 30° of $L_s$ (Figures 7(a)–(d)). Because the focus of this study is variability in atmospheric temperature due to atmospheric waves rather than surface temperature, we restricted the remainder of our analysis to the A1–A3 channel data set. While a recent analysis by Hinson & Wilson (2021) suggests that B1 may be sensitive to atmospheric variability in the lowermost atmosphere under some circumstances in the extratropics, this may not necessarily be globally applicable.

In map view, the $\Omega_{GW}$ and $\epsilon_{GW}$ data were binned by dayside/nightside and 30° of $L_s$ and averaged at $1° \times 1°$, $2° \times 2°$, and $5° \times 5°$ resolution. Likewise, $\bar{\Omega}_{GW}$ and $\epsilon_{GW}^\text{norm}$ were calculated and averaged in the same way. Averages were made of the available data from individual MYs, as well as all of the data in MYs 29–33 and 35, to better infer differences between years with (that is, MYs 28 and 34) and without global dust storm activity. In order to better study intraseasonal variability, the $\Omega_{GW}$ and $\epsilon_{GW}$ data were binned by dayside/nightside and 9° of $L_s$ and averaged at $5° \times 5°$. Following Heavens et al. (2020), a momentum flux metric was also calculated from the off-nadir observations,

$$\hat{F}_{ph} = \frac{1}{(T^2)^{1/2}} \Omega_{GW},$$

where $(T^2)^{1/2}$ is the square root of the mean squared temperature.

To characterize the intermittency of GW sources, we followed Heavens et al. (2020) in using the intermittency metric of Hertzog et al. (2008) and Wright et al. (2013): the percentage of total momentum flux in the top 10% of momentum flux diagnoses relative to the integrated momentum flux in the distribution. To minimize spurious extrapolations, this intermittency metric was only calculated when at least 10 diagnoses of $\Omega_{GW}$ were available in the given spatial/seasonal/time-of-day bin. Therefore, there is only enough data from non–dust storm years to generate visually acceptable intermittency maps with minimal unfilled bins at a resolution of 30° of $L_s$ and a spatial resolution of $2° \times 2°$.

Zonal averaging was performed on a grid binned by MY, dayside/nightside, 2° in $L_s$, 2° in latitude, and 10° in longitude. The MRO makes approximately 50 orbits per 2° of $L_s$, and MCS makes a pair of on-planet observations approximately every 2° in latitude, so this averaging resolution enables there to be at least one data point per longitude bin when there are 50,000 diagnoses per 30° of $L_s$ at a given time of day. Averaging is done by longitude bin at a given latitude, and then the average of all longitude bins is made. Global averaging was performed by taking the average of the zonal average data weighted by cosine of the central latitude of each latitude bin.

While averages were calculated with and without observations where the maximum elevation was >12 km, inspection of Heavens et al. (2020) and the analysis in the Appendix suggest that only the results with high-elevation surfaces excluded are worth presenting. Likewise, Heavens et al. (2020) demonstrated that the behavior of the nonnormalized quantities (e.g., $\Omega_{GW}$ and $\epsilon_{GW}$) is generally similar to the normalized quantities, only the normalized quantities will be presented. These normalized quantities can be related to GW specific energy density and the vertical flux of horizontal momentum (Ern et al. 2004; Heavens et al. 2020).

3.2. Roughness Information

Creasey et al. (2006) observed that orographic GW parameterizations in global climate models typically related the GW flux to the surface stress, which is a function of atmospheric density ($\rho$), $N$, the magnitude of low-level wind, and the topographic variance (surface roughness). Heavens et al. (2010) followed Creasey et al. (2006) in trying to estimate surface stress based on information from the MCD (Millour

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and the average of the surface roughness estimates in the MRO-MCS radiance observations. This analysis suggested that surface stress varied minimally seasonally, implying that the surface stress is dominantly a function of the surface roughness. Therefore, the surface roughness distribution should be a good hypothesis for the distribution of orographic GW activity.

**Figure 7.** Number of analyzed nadir and off-nadir observations, binned by 30° of $L_s$, and nightside vs. dayside, as labeled. The titles report the total number of observations analyzed for the A1–A3 survey. The red bars indicate the total number of observations for the A1–A3 survey, the blue horizontal lines indicate the number of observations available for the survey with B1 included, and the green horizontal lines indicate the number of observations for the A1–A3 survey where the maximum surface elevation for any detector in A3 is less than 12 km (the numbers in the other channels are similar).

**Figure 8.** The $\log_{10}$ of the estimated surface roughness (m$^2$) labeled with important geographic features mentioned the text. The scale is saturated at 4 and 7 log units to emphasize the roughest features on Mars. Abbreviations OLY (Olympus Mons), FF (Fortuna Fossae), and SCH (Schiaparelli Crater) are used to label smaller-scale features. The rectangular box marks a region of interest plotted in Figures 14–16.
The surface roughness was therefore estimated by averaging the topographic variance in all off-nadir observations in the A2 channel at surface elevations <12 km at a spatial resolution of 1° × 1°, 2° × 2°, and 5° × 5°. The A2 channel is directly between A1 and A3 in the MRO-MCS detector array (McCleese et al. 2007). Topographic maps at the same resolutions were generated by averaging similarly filtered elevation data. A roughness map stretched to emphasize major roughness features has been included as Figure 8 and can be used as a geographic reference throughout the remainder of the paper.

4. Results

4.1. A Changing Global Distribution of GW Activity with Altitude

4.1.1. Nadir Observations

Nadir observations were limited to a brief period early in the MRO-MCS mission. The GW activity observed from the nadir during this time was generally stronger lower in the atmosphere and stronger on the dayside than the nightside (Figures 9(a)–(f)). Heavens et al. (2020) argued that GW activity in the dayside tropics in A3 observations was stronger over high-elevation areas than low-elevation areas. However, the A1 and A2 nadir observations (particularly A2 at night) show substantial GW activity associated with Valles Marineris (15° S–0°N, 100°–45°W), which reaches elevations as low as −4500 m (Figures 9(a), (b), (d), and (e)). Another notable feature in A1 and A2 dayside observations (Figures 9(a) and (b)) is a planet-encircling band of substantial GW activity at 50°S, which is absent in A3.

4.1.2. Off-nadir Observations

As in the nadir, GW activity viewed in the off-nadir is generally stronger lower in the atmosphere. Valles Marineris is a prominent area of GW activity in A1 and A2 (but not A3), and there is a band in GW activity on the dayside at 50°S that perhaps spreads in width with altitude (Figures 10(a)–(f)). The variability in A1 has significant structure near the resolution of averaging, which has disappeared in most tropical areas in A3.

The presence of a narrow band of substantial GW activity in the southern extratropics (near 50°) on the dayside in A1 and A2 is only apparent in two of the periods presented and in A3 during only one (Figures 11(b), (c), (f), (g), and (k)). But GW activity in Valles Marineris and the structure in GW activity near the resolution of averaging are apparent in A1 and A2 throughout the year (Figures 11(a)–(h) and 12(a)–(h)).

This high-resolution structure maps closely onto large craters. The most prominent craters with such structure in all channels are the Argyre (50°S, 60°W) and Hellas (40°S, 80°E) Basins in the southern extratropics. But smaller circular features can be seen in A1. One example is centered near 5° S, 15°E, corresponding to the 458 km (≈8°) diameter crater Schiaparelli (Figures 12(a) and 13). Higher activity is observed near the raised edge of this crater than its center, resulting in a circular feature. In other words, the distribution of GW activity in A1 and A2 appears to be shaped by variability in surface roughness.

4.1.3. Correlation with Surface Roughness

Direct comparison of GW activity in A1 with surface roughness suggests a close relationship. Fine structure near the resolution of the map is clearer in the surface roughness map (Figure 14(b)) than in the \( \rho_{GW} \) map for A1 (Figure 14(a)) and accentuates Schiaparelli and other similar-sized craters. The surface roughness map also shows that Valles Marineris is the roughest area of Mars. In the elevated volcanic province of Tharsis and Valles Marineris, roughness features at a variety of scales are areas of high GW activity. Valles Marineris is prominent, but so are Olympus Mons and the Tharsis Montes (the circular features with white space in the middle, where
observations over high-elevation data have been excluded; Figures 14(c) and (d)). Activity is also high around some roughness features of both positive and negative topography near 20°N, 90°W. These include Ceraunius and Uranus Tholi (near 25°N, 100°W), Uranus Mons (near 25°N, 95°W), Tharsis Tholus (near 15°N, 90°W), and Fesenkov Crater (near
20°N, 85°W). Further east, GW activity seems to weakly illuminate the area around Sacra Mensa (near 25°N, 65°W). There is one feature in the A1 GW activity map that does not appear to have anything to do with roughness, the band of activity near 65°S (Figure 14(a)), but otherwise, roughness appears to be explanatory.

In A2, the fine structure mostly disappears (Figure 15(a)), but Valles Marineris, Olympus Mons, the Tharsis Montes, and some of the small topographic features (generally the positive ones) are marked by GW activity (Figure 15(c)). In parallel, rough regions in the northern extratropics that were less prominent in A1, such as the areas around Ascuris Planum and Mareotis Fossae (near 45°N, 80°W) and the dichotomy boundary northwest of the Isidis Basin (15°N, 90°E), are more prominent in A2 (Figure 15(a)). In A3, GW activity associated with the low-altitude areas of Valles Marineris disappears, but...
Olympus Mons and the Tharsis Montes are still apparent (Figure 16(c)). Argyre and Hellas are visible, as is the roughness associated with the margins of Valles Marineris and the southern margin of Solis Planum (near 40°S, 90°W; Figure 16(a)). The areas near Ascuris Planum and Mareotis Fossae and to the northwest of Isidis are the strongest areas of GW activity in A3, though they appear broader in area than the roughness features and are joined by other features in the northern extratropics that do not have obvious associations with roughness, such as the area centered near 75°N, 120°W.

Figure 14. Spatial distribution of GW activity compared with surface roughness. (a) Mean spatial distribution of GW activity viewed from dayside off-nadir observations in A1 during an $L_s$ of 240°–270° of MYs 29–33 and 35 averaged at 1° resolution expressed as $\log_{10} I_{GW}$ (K$^2$ K$^{-2}$). White space indicates that no data are available in the averaging bin. The lower end of the plotting color scale is saturated at $10^{-6}$ K$^2$ K$^{-2}$, about an order of magnitude larger than the approximate uncertainty in the planet’s coldest areas (the winter polar cap). (b) The $\log_{10}$ of mean roughness (m$^2$) at the baseline of off-nadir observations averaged at 1° resolution. (c) Identical to panel (a) but focused on 30°S–30°N, 160°–40°W. (d) Identical to panel (b) but focused on 30°S–30°N, 160–40°W.

Figure 15. Spatial distribution of GW activity compared with surface roughness. (a) Mean spatial distribution of GW activity viewed from dayside off-nadir observations in A2 during an $L_s$ of 240°–270° of MYs 29–33 and 35 averaged at 1° resolution expressed as $\log_{10} I_{GW}$ (K$^2$ K$^{-2}$). White space indicates that no data are available in the averaging bin. The lower end of the plotting color scale is saturated at $10^{-6}$ K$^2$ K$^{-2}$, about an order of magnitude larger than the approximate uncertainty in the planet’s coldest areas (the winter polar cap). (b) The $\log_{10}$ of mean roughness (m$^2$) at the baseline of off-nadir observations averaged at 1° resolution. (c) Identical to panel (a) but focused on 30°S–30°N, 160°–40°W. (d) Identical to panel (b) but focused on 30°S–30°N, 160–40°W.
available in the averaging bin. The lower end of the plotting color scale is saturated at 10⁻⁶ K² K⁻², about an order of magnitude larger than the approximate uncertainty in the planet’s coldest areas (the winter polar cap). (b) The log₁₀ of mean roughness (m²) at the baseline of off-nadir observations averaged at 1° resolution.

(c) Identical to panel (a) but focused on 30°S–30°N, 160°W–40°W. (d) Identical to panel (b) but focused on 30°S–30°N, 160°–40°W.

Figure 16. Spatial distribution of GW activity compared with surface roughness. (a) Mean spatial distribution of GW activity viewed from dayside off-nadir observations in A3 during an L₆ of 240°–270° of MYs 29–33 and 35 averaged at 1° resolution expressed as log₁₀GW (K² K⁻²). White space indicates that no data are available in the averaging bin. The lower end of the plotting color scale is saturated at 10⁻⁶ K² K⁻², about an order of magnitude larger than the approximate uncertainty in individual diagnoses of GW activity. (b) The log₁₀ of mean roughness (m²) at the baseline of off-nadir observations averaged at 1° resolution. (c) Identical to panel (a) but focused on 30°S–30°N, 160°W–40°W. (d) Identical to panel (b) but focused on 30°S–30°N, 160°–40°W.

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(Figures 16(a) and (b)). The band in the southern extratropics remains apparent in A3 but smaller in magnitude than the GW activity in the northern extratropics.

At 1° resolution, surface roughness moderately correlates with GW activity in A1 (Figures 17(a) and (d)). This correlation is weaker near the southern summer solstice (L₆ = 270º, the boundary between Mars Months 9 and 10) and at night but is generally stronger and less variable with season around the Tharsis Montes and Valles Marineris. The correlation between roughness and GW activity in A2 is weaker than in A1 (Figures 17(a), (b), (d), and (e)); roughness and GW activity do not appear to be correlated in A3 (Figures 17(c) and (f)).

Heavens et al. (2020) demonstrated some correlation between elevation and tropical GW activity in A3 dayside nadir observations and moderate anticorrelation on the nightside. This relationship was less apparent in off-nadir observations. Adding the A1 and A2 channels and roughness to a similar correlation analysis suggests that surface roughness more strongly correlates with GW activity at the low altitudes sensed by A1 and A2 in the nadir observations. The GW activity in A3 correlates or anticorrelates more strongly with elevation than roughness (Table 1). In the off-nadir, elevation weakly correlates with GW activity in dayside A3 observations, but nightside GW activity in A3 is poorly explained by roughness or elevation. In A1 off-nadir observations, surface roughness correlates more strongly with GW activity than elevation. The same is true in A2 on the nightside but not on the dayside (Table 1). Thus, the tropical GW population observed from the nadir appears to evolve with altitude from one correlated with surface roughness to one correlated or anticorrelated with elevation, whereas the GW population observed from the off-nadir appears to evolve with altitude from one correlated with roughness to one independent of surface roughness or elevation.

However, the correlation between the logarithms of surface roughness and GW activity in A1 does not seem as strong as inspection of the spatial distribution plots would suggest. One way to understand this is look at the set of one-dimensional probability distribution functions of GW activity within each discrete interval of surface roughness (again on logarithmic scales) and how they change with surface roughness, as in Figures 18(a) and (b). First, the most probable values of GW activity (the brighter colors) only correlate strongly with roughness at roughness >10⁵ m². Below this value, there may be some weak correlation between surface roughness and GW activity, but surface roughness and GW activity are certainly independent at surface roughness <10⁵ m². Second, the probability distribution functions are quite broad. Below the range of roughness where surface roughness and GW activity strongly correlate, the minima of the distributions are flat near $\Omega_{GW} = 10^{-7} \text{ K}^2 \text{ K}^{-2}$, which is approximately the characteristic uncertainty in individual diagnoses of GW activity (assuming a variance of 0.005 K² at 170 K; Heavens et al. 2020), while the maxima of the distributions increase with a slope with an exponent of <1. In the region where surface roughness and GW activity strongly correlate, the minima of the distributions generally increase along the same slope as the center of the
distribution, while the maxima of the distributions flatten. So, depending on how the widths of the distributions are measured, areas with the same roughness can experience an order of magnitude of variability in GW activity or more.

Where the logarithms of roughness and GW activity in A1 do correlate strongly, the slope of their relationship is approximately 1, indicating that surface roughness and GW activity are linearly proportional over the roughest surfaces (Figures 18(a) and (b)). This is expected behavior for orographic GWs, as outlined by Creasey et al. (2006), who noted that orographic GW schemes used the surface wind stress as a proxy for orographic GW activity by relating surface wind stress to the vertical flux of horizontal momentum, absent dissipation. For A1, GWs are presumably being observed close to their sources under conditions of near-perfect visibility (Figures 6(a) and (d)). In that case, the expected relationship

![Figure 17](image)

**Figure 17.** Correlation product moment (weighted by cosine of latitude of the bin) between log10 of roughness (m²) averaged at 1° and log10ΩGW (K² K⁻²) averaged at 1° resolution for the channel, time of day, region, and Mars Month labeled. Mars Month is defined as Ls = 0°–30° for Month 1, etc. Blue bars indicate the global correlation, while orange bars are for the region of the Tharsis Montes and Valles Marineris 30°S–30°N, 160°–40°W.

**Table 1**

|            | Nadir | A1 | Elevation | Roughness | A2 | Elevation | Roughness | A3 | Elevation | Roughness |
|------------|-------|----|-----------|-----------|----|-----------|-----------|----|-----------|-----------|
|            | Day, r|    |           |           |    |           |           |    |           |           |
|            |        |    | 0.12      | 0.60      |    | 0.13      | 0.38      |    | 0.28      | 0.10      |
|            |        |    | 45,354    |           |    | 45,351    |           |    | 45,351    |           |
|            | Night, r|    | 0.15      | 0.43      |    | 0.14      | 0.34      |    | −0.47     | −0.05     |
|            |        |    | 45,390    |           |    | 45,389    |           |    | 45,379    |           |
|            | Off-nadir | A1 | Elevation | Roughness | A2 | Elevation | Roughness | A3 | Elevation | Roughness |
|            | Day, r |    | 0.32      | 0.40      |    | 0.35      | 0.29      |    | 0.20      | 0.05      |
|            |        |    | 214,588   |           |    | 214,595   |           |    | 214,635   |           |
|            | Night, r|    | 0.13      | 0.49      |    | 0.15      | 0.32      |    | −0.05     | 0.00      |
|            |        |    | 210,887   |           |    | 210,897   |           |    | 210,912   |           |

**Note.** Here n is the number of observations used. Observations where the maximum elevation was greater than 12 km have been excluded.
between surface roughness and $\Omega_{GW}^{*}$ in A1 (assuming 100% visibility) can be derived by combining Equation (7) of Ern et al. (2004) with Equation (2) of Creasey et al. (2006),

$$
\Omega_{GW} = \frac{2KN^2U}{g^2} \frac{\lambda_b}{\lambda_c} \sigma_r^2,
$$

where $\kappa$ is a “tunable parameter” of $10^{-4}$ m$^{-1}$ in Creasey et al. (2006), $U$ is the wind velocity, $g$ is the gravitational acceleration, and $\sigma_r^2$ is the surface roughness.

If we estimate $\frac{\lambda_b}{\lambda_c}$ to be 1.5–3 and $N \approx 8 \times 10^{-3}$ s$^{-1}$, the expected magnitude of the term on the right-hand side of Equation (4) multiplying surface roughness is $1-2 \times 10^{-11}U$. The fit by inspection in Figures 18(a)–(f) suggests a value of $\approx 3\times10^{-11}$ for this term, which implies typical surface winds of 1.5–3 ms$^{-1}$. Note that the $\kappa$ in Creasey et al. (2006) has a physical meaning taken from Palmer et al. (1986) and Lewis et al. (1999), where it is a characteristic horizontal wavenumber, $k_h$. However, Lewis et al. (1999) seem to have omitted a factor of $\frac{\lambda_b}{\lambda_c}$ included by Palmer et al. (1986). So $\kappa$ is at most $7.5 \times 10^{-5}$–$1.5 \times 10^{-4}$ m$^{-1}$ to align with the sensitivity of the A1 off-nadir observations.

Potential variables other than the wind are constrained by a factor of 3 or better. Therefore, the scatter around the fit line in Figures 18(a)–(f) is generated either by direct wind variability or because the local wind and stability conditions lead to topography obstructing rather than disturbing the circulation (Lott & Miller 1997). In the former “blocking” case, amplitudes are reduced by up to an order of magnitude from the expected linear wave solution (Lott & Miller 1997). For nonblocked waves, GW generation by surface winds of 30 ms$^{-1}$ seems plausible in rare cases. The flattening of the maxima of the distributions at high roughness implies that winds $>10$ ms$^{-1}$ over the roughest surfaces are generating waves with unstably high amplitudes. It is also possible that rough, low-elevation surfaces like Valles Marineris have apparently higher levels of GW activity because they are observed somewhat higher in altitude relative to the surface than typical, and the GWs they excite have grown with amplitude over that altitude range.

Yet dissipation seems to outpace amplitude growth with decreasing density over the next two scale heights. In A2, surface roughness and GW activity correlate at higher surface roughness values, but GW activity is somewhat weaker in magnitude at high surface roughness, though similar in magnitude to A1 at lower surface roughness (Figures 18(c) and (d)). In A3, surface roughness and the most probable values of GW activity are uncorrelated at all surface roughnesses. Even the maxima of the probability distributions are independent of surface roughness, though the maxima of the probability distributions at intermediate surface roughness are greater than the maxima at low and high surface roughness (Figures 18(e) and (f)). The weaker amplitudes could be partly explained by the lower visibility of GWs in A2 and A3, but the decorrelation between activity and roughness with altitude is positive evidence for dissipation.

The high degree to which variability in wind and stability likely drives topographic GW activity precludes isolating the
topographic GW component of $\hat{\Omega}_{GW}$ from the nontopographic component in A1 or A2 observations to assess their spatial distribution without information about surface winds. But it is possible to argue from the independence of GW activity from surface roughness over areas with surface roughness $<10^5$ m$^2$ and the unrealistic surface winds that would be implied by the GW activity observed over a smooth area that significantly GW activity over smooth areas is nonorographic. For example, the GW activity of $10^{-5}$ K$^2$ K$^{-2}$ in the areas with surface roughness $<10^4$ m$^2$ would require $>45$–90 ms$^{-1}$ surface winds (and no blocking), so GW activity observed in northern fall and winter ($L_s = 180^\circ$–360$^\circ$) over smooth areas of the northern extratropics (e.g., most of the area in the 45–80$^\circ$N band in Figures 14(a) and (b)) is likely nonorographic.

4.2. Seasonal Variability and Dust Storm Activity

Heavens et al. (2020) found that GW activity in A3 off-nadir views had a strong seasonal cycle, with a maximum in northern summer, a minimum in northern fall, and pronounced minima during regional and global dust storm activity. There is a similarly phased seasonal cycle in A1 and A2 as well, but it has a higher relative amplitude on the dayside in A1 and A2 than in A3 (Figures 19(a)–(f)). The seasonal cycle of activity on the nightside is very similar in all three channels (Figures 20(a)–(f)). The normalization factors suggest that GW activity is globally weaker at higher altitudes and globally stronger on the dayside than the nightside. Given that $\hat{\Omega}_{GW}$ should increase exponentially with altitude normalized by the scale height rather than decrease by a factor of 4, the differences in the normalization factors between the channels alone suggest that there is strong dissipation of GW activity in the lower atmosphere. Accounting for the decrease in GW visibility in the channels, which is no more than a factor of 2 for the waves with wavelengths to which off-nadir observations are most sensitive, still would imply that GW activity should increase by a factor of 2 between A1 and A3. And the strong contrast between the seasonal cycles of A1 and A2 and that of A3 on the dayside (Figures 19(a)–(f)) suggests that GW activity experiences stronger dissipation in northern summer than in any other season.

And, as in A3, GW activity in A1 and A2 seems to weaken during regional and global dust storm activity. This phenomenon can be seen most easily on the dayside during the C-type (in the sense of Kass et al. 2016) regional dust storms toward the end of northern winter ($L_s = 300^\circ$–330$^\circ$) in MYs 29 and 31–34 (Figures 19(a) and (c)–(f)) but is present in other large regional dust storms and 34P, which started just after northern fall equinox ($L_s = 187^\circ$; Figure 19(f)). One explanation for the minima in GW activity during dust storms would be observational. As discussed in Section 2, dust opacity raises the altitude to which on-planet views in A1–A3 are sensitive.
The weighting function of A1 and the visibility of GWs to observations under high dust opacity resemble those of A2 under normal conditions. The weighting function of A2 and the visibility of GWs to observations under high dust opacity resemble those of A3 under normal conditions (Figures 2(a)–(i)). And the altitude of the peak of A3’s vertical weighting function rises by \( \approx 10 \) km under high dust opacity. Given that GW activity is strongly dissipated in the lower atmosphere and peak visibilities seem to decrease with altitude, raising the vertical range of sensitivity of the channel would result in a lower magnitude of GW activity being observed, all else being equal. But if the near-surface GW sources were unaffected by the dust storm, we likewise would expect the magnitude of GW activity observed in A1 and A2 to be similar to that observed in A2 and A3 under normal conditions.

However, in most cases, the global mean GW activity observed during MY 34 in the channel with the lower altitude of peak sensitivity fell to a lower level than the channel with the higher altitude of peak sensitivity during the dust storm–quiet year of MY 30 (Figures 19(g) and (h) and 20(h)). The one exception is nightside A1 during MY 34, when global mean GW activity fell to a level indistinguishable from A2 during MY 30 (Figure 20(g)).

Yet the evolution of the spatial distribution of GW activity at night as 34P developed is inconsistent with A1 during the dust storm sampling the same GW distribution as A2 during normal conditions. In the period before 34P expanded to regional scale, nightside GW activity in A1 was nearly indistinguishable between MYs 34 and 30 (Figures 21(a) and (d)). In the next period, as 34P expanded from regional to global scale, GW activity decreased in the tropics and some smoother midlatitude areas relative to MY 30 (Figures 21(b) and (e)). And as 34P peaked in intensity (see Heavens et al. 2019) during the next period, the spatial distribution of GW activity in A1 somewhat resembled that of A2 during MY 30 (Figures 21(c) and (i)).

But GW activity in A1 was somewhat weaker in smoother areas and Valles Marineris than A2 during MY 30 (e.g., Hellas but somewhat stronger in a region running to the NE of Valles Marineris toward the typical hot spot of GW activity near 30°N, 60°E).

And in the case of A2 and A3 on the dayside, where global mean GW activity during MY 34 in A2 fell below the global mean GW activity in A3 during MY 30, the spatial distribution of GW activity during 34P evolves to contrast more strongly with A3 during MY 30 by \( L_s = 198°–207° \) (Figures 22(c) and (i)). The GW activity is much weaker in A2 than A3 except in a few hot spots in the northern midlatitudes and an area NE of the Tharsis Montes (0°–30°N, 90°W).

This last area (particularly near Fortuna Fossae) was also identified by Heavens et al. (2020) as an area of unusually high GW activity during 34P. Dayside GW activity in this area does not correlate strongly with roughness in this area.
In addition, it would require surface winds of at least 45–90 ms$^{-1}$ to generate $\hat{\Omega}_{GW}$ in A1 of $10^{-5}$ K$^2$ K$^{-2}$, a level of activity typically observed during most of the year (Figure 23(b)). Therefore, significant GW activity observed in this region is almost certainly nonorographic.

Here GW activity in A1 around $L_s = 200^\circ$ was slightly higher than was typical at this season during the other MYs but much below the level sometimes observed during northern spring and summer (Figure 23(b)). The GW activity in A2 around $L_s = 200^\circ$ of MY 34 was similar to that observed in A2 during northern spring and summer (Figure 23(c)), while GW activity around $L_s = 200^\circ$ of MY 34 exceeded GW activity in northern spring and summer by a factor of 3–7. Thus, GW activity observed in this region typically decreases by an order of magnitude in the roughly two scale heights between the peak sensitivity of A1 and A3 (Figure 2). But the GW activity observed around $L_s = 200^\circ$ of MY 34 was unusual in growing substantially over the same vertical range.

Inspection of the individual observations, including the pair with the highest A3 variance in Figure 23(d) and the adjacent pair, suggests that the GW activity here is not an artifact. The first pair of A1 observations cools with increasing distance (and increasing altitude as the observation is advancing across the array from detector 21 to detector 1; Figure 24(a)), which implies a steep lapse rate. This steep lapse rate is also inferred from the first two observations in A4 (centered at 843 cm$^{-1}$), A5 (centered at 463 cm$^{-1}$), and B1 (Figures 24(d)–(f)).

Normally, these channels are used to retrieve water ice, dust, and surface temperature, respectively (Kleinböhl et al. 2009, 2011), but likely gained dust opacity and thus stronger sensitivity to lower atmospheric temperatures during the dust storm. The second pair of observations in A1 departs from a straight line form at the ends of the observations, suggesting that a low-amplitude wave is mixed with the steep decrease in temperature with altitude (Figure 24(a)). This pair of observations looks similar in A5, only 4 K warmer (Figure 24(c)). The first two observations in A2 likewise bend at the lower end, while the second two observations bend at both ends (Figure 24(b)). But it is in A3 that the waves become most pronounced (Figure 24(c)). The observations in the different channels are not observing the exact same place but do seem to observe growing wave activity from the lower atmospheric opacity channels of B1, A4, and A5 (in that rough order) through A1, A2, and A3. None of the activity observed has unusually low brightness temperatures in the first few detectors, which is the most typical artifact in MRO-MCS on-planet radiance observations.

The increase in GW activity between A1 and A3 is a factor of 10–20, greater than the factor of $e^2$ that would be expected for conservative growth of a GW with height, especially with a
scale height of ≈11 km implied by the observed brightness temperatures. (Recall that GW amplitude increases in proportion to \( \exp \left( \frac{z}{H} \right) \), so \( \Omega_{GW} \) increases in proportion to \( \exp \left( \frac{z}{H} \right) \).) One explanation is that the relationship between GW energy and \( \Omega_{GW} \) is such that increasing \( N^2 \) for constant energy will increase \( \Omega_{GW} \) linearly. This seems unlikely. Stability seems to have decreased with altitude from the lower to the middle atmosphere in the tropics during 34P (Heavens et al. 2019). Another possibility is that additional GWs are being generated between the levels of A1 and A3. A third explanation is that meridional wind speed significantly changes between A1 and A3, changing the intrinsic phase speed of the wave (and thus its vertical wavelength) so that GWs are more visible. The principal meridional overturning circulation has long been simulated to intensify substantially during dust storm activity (e.g., Wilson 1997), and there is evidence for extremely rapid cross-equatorial advection of dust in the early stages of 34P (Shirley et al. 2020) and a few hundred kilometers to the west of Fortuna Fossae within a few sols of the GW observations in Figures 24(a)–(f).

Distinguishing between intrinsic phase speed change and additional GW generation is not straightforward. Assuming the second pair of observations for A1–A3 and A5 captures the peak and trough of a single wave implies an apparent horizontal wavelength of 80–100 km (Figures 24(a)–(c) and (e)). The horizontal alignment of the negative phase of this wave near 270 km shifts by ≈18 km between A1 and A3, implying a vertical wavelength of 50–65 km. A wave like this would be just on the edge of visibility (Figure 6) and would not gain significant visibility by decreasing in vertical wavelength.

Thus, we are restricted to concluding that GWs near Fortuna Fossae around \( L_s = 200^\circ \) were nonorographic and grew with height, atypical for GWs observed in this area outside 34P, as well as for GWs in general. Given the nonlinear growth of the apparent disturbance with decreasing density, critical level filtering of GWs and other GW dissipation with altitude (hereafter filtering) here must have been minimal.

By taking the ratio of zonal mean GW activity between A3 and A1, it is possible to investigate how GW activity grew or did not grow with height and thus the extent of filtering on a more global scale (Figures 25(a)–(h)). Filtering of GWs seems to have been minimal throughout the tropics during 34P. Under normal conditions, GW activity is least filtered in the winter extratropics, moderately filtered in the summer extratropics, strongly filtered in the northern and southern tropics during southern spring and some of southern summer, and almost completely filtered in polar day and the southern tropics during southern fall and winter (Figures 25(a)–(f)). During 34P, filtering in the tropics weakened to levels similar to that of their respective extratropics. If GW dissipation were weaker below
≈15 km, as well as between ≈15 and 35 km, the reduction in GW activity in A1 during 34P by an order of magnitude on the dayside (Figure 19(g)) could not be explained by A1 observing higher in the atmosphere and with lower visibility to GWs, which would explain a factor of 2 decrease at most.

4.3. Tropical Nightside Activity: A Carbon Dioxide Ice Cloud Artifact?

The GW activity in the tropics observed in A3 off-nadir observations on the nightside during the less dusty half of the year by Heavens et al. (2020) was distinctive because of the contrast in the magnitude of GW activity between day and night and the high intermittency of GW activity in some areas of the tropics. That is, there were relatively few observations with extremely high variances. Heavens et al. (2020) considered whether this result could arise from reemission by absorbing mesospheric CO₂ clouds with cloud particle grain sizes ≤1 μm. Having found that high variances in this location and season were always observed in association with “loop features” indicative of clouds, Heavens et al. (2020) suggested that clouds with nadir optical depths in A3 of >0.02 could be responsible.

If these clouds did consist of submicron particles, their signal should be limited to A3. Modeling by Hayne et al. (2012; see Figure S3 of Heavens et al. 2019) showed that the extinction coefficient, Q_{ext}, of a CO₂ ice particle ≤1 μm was an order of magnitude higher in A3 than A2. And because the variance due to reemission scales with the optical depth and thus Q_{ext} for largely absorbing particles for optical depth ≪1 (Heavens et al. 2020), the variance artifact introduced by such a cloud would be 2 orders of magnitude less in A2 than in A3 (and even smaller in A1 by a similar chain of reasoning). However, Figures 26(a)–(f) show normalized variance and intermittency in A1, A2, and A3 on the nightside during Lₚ = 0°–30°, the most active period for this apparent nightside GW activity. Areas of the tropics can be identified with similar levels of normalized variance in both A2 and A3 and high intermittency. In A2, these areas of high normalized variance tend to blend into the high activity associated with Valles Marineris and other topographic features, but the spatial distribution of intermittency in the tropics is quite similar between A2 and A3 (Figures 26(e) and (f)). Moreover, just to the east of Valles Marineris (near 45°W), the magnitude of GW activity is similar between A2 and A3 and easy to compare (Figures 26(b) and (d)). These areas of activity are absent in A1 (Figures 26(a) and (d)).

In some cases, it is possible to show that substantial A1–A3 variance in the off-nadir is directly attributable to a CO₂ ice cloud. Figure 27 shows a thorough breakdown of A1–A3 observations of a nightside loop feature near the equator from the northern summer of MY 34. Loop features have been interpreted to be discrete cloud layers whose true altitude is at the peak of the loop but are observed on either side of their true...
location as they rise and set with respect to the limb observation (Sefton-Nash et al. 2013). In this case, the curve of the loop seems to be interrupted at 5°S. Behind this feature, radiance in the lower detectors is anomalously low compared to the observations to the north and south because the cold, high-altitude cloud is well forward of the tangent point. Thus, its emission is observed in preference to warmer emission from the surface (Figures 27(a)--(f)). The profiles at the center of the loop suggest the presence of a cloud at 60 km altitude (Figures 27(d)--(f)). (Observations like this are regularly used to detect polar stratospheric and polar mesospheric clouds on Earth; e.g., Massie et al. 2007; DeLand & Gorkavyi 2021.)

Off-nadir observations near this loop feature enable its extent along the direction MCS is observing to be quantified more precisely than is possible in limb observations, where the loop feature extends \( \approx 18^\circ \) (\( \approx 1000 \) km). One pair of high-variance off-nadir observations intersects the surface near 1°S and shows 8–12 K depressions in brightness temperature near the center of the observation and decreasing temperature across the observation (that is, from detector 1 to detector 21; Figures 27(g)--(i)). This decrease in temperature contrasts with the slight increase in temperature across the detectors in the neighboring observations and is strongest in A3. The importance of this pair of observations is clear when their observational paths are projected toward the spacecraft. These observations would intersect the peak of the loop feature at 60 km altitude and 1°S (Figures 27(a)--(c)), whereas the neighboring observations would be 1°–1°S (60–90 km) to the north and south. That the depression in brightness temperature does not cover the full off-nadir observation that intersects the cloud also implies that the cloud is \(< 60 \) km in width in the direction MCS is observing.

With its strong signal in A1–A3 (and moderate–strong signals in all of the other thermal infrared channels, which are not shown here), the loop feature in the limb can be interpreted as a strongly scattering CO\(_2\) cloud at 60 km with a particle size of \( > 5 \) \( \mu \)m, based on modeled extinction coefficients reported in the supplementary material of Heavens et al. (2019). The strong signals in off-nadir observations in A1 and A2, as well as A3, are consistent with this idea (Figures 28(a)--(c)). This signal is absent in A4 and B1 off-nadir observations, which resemble one another (Figures 28(d) and (e)). The CO\(_2\) ice is relatively transparent in B1, but A4 has a similar extinction coefficient to A1 at a CO\(_2\) particle size \( > 5 \) \( \mu \)m. The single scattering albedo in A4 at 5–6 \( \mu \)m is \( \approx 0.999 \) and 0.982 for A1 (Hayne et al. 2012), which suggests that the A1 and A4 observations of CO\(_2\) ice particles of \( \approx 5 \) \( \mu \)m should be broadly similar. The simplest explanation of the discrepancy would be that the cloud was narrow enough cross-track to be missed by A4, which is the rightmost A channel (the most western on the nightside;
Thus, off-nadir views in A4 likely observed the same surface emission observed by B1, but the cloud was thick enough to allow some emission contribution in A1. Clancy et al. (2019) inferred CO$_2$ particle sizes $>5$ $\mu$m from nightside loop features. The nightside loop feature in Figure 24 of Clancy et al. (2019) is impressive, but it is only evident in the off-nadir in a pair of off-nadir observations with A3 brightness temperature depressions of $\approx$2 K; this signal is minimal or absent in all other channels. Indeed, A4 agrees with B1 within $\approx$1 K (not shown). But it should be noted that the horizontal path length through the cloud (and thus the opacity) is at least an order of magnitude higher in the limb than the off-nadir, so two optically thick clouds in the limb could have much different off-nadir opacities. And if the cloud in Figure 24 of Clancy et al. (2019) is thin enough, its small effect on A3 could translate to effects indistinguishable from noise in other channels. Thus, the imperfect correspondence between high off-nadir A3 variance and loop features in the limb reported by Heavens et al. (2020) for a sample of tropical nightside data can probably be explained by similar variability in cloud optical thickness.

5. Discussion

In Section 1, we identified three key questions to be addressed by a more thorough study of GW activity in on-planet observations by MRO-MCS. Here we will focus on these questions, briefly summarizing the key results, evaluating related questions that arose from the analysis, and considering the broader implications for understanding GW activity in Mars’s atmosphere.

5.1. GW Activity Changes during Dust Storm Activity

As outlined in Section 4.2, GW activity throughout the lower atmosphere largely decreases during regional and global dust storm activity. Even compensating for the potential change in the altitude to which particular channels are sensitive to GWs, GW activity (in terms of a quantity proportional to GW potential energy) decreased by approximately a factor of 2 during 34P (Figures 19 and 20). Heavens et al. (2020) noted...
that this decrease primarily occurred in areas of low–moderate activity in A3, but it is more precise to say that the decrease during 34P was largest in the northern tropics, moderate in the southern tropics and midlatitudes, and largely absent in the northern extratropics (both midlatitudes and near the poles). Moreover, the decrease seems to have been larger on the dayside than the nightside (Figures 21 and 22).

Even under mature dust storm conditions, GW activity in the lowest two scale heights during 34P remained high in Tharsis and just to its east (Figure 22(c)). Unusually, GW activity in a smoother part of this region was far less attenuated than normal by the time it reached the 30–40 km altitude range sampled by A3 under dust storm conditions (Figure 23(d)). However, attenuation of GW activity throughout the tropical lower atmosphere was weaker during 34P than any other time in the record (Figures 25(g) and (h)), so local enhancement in GW activity here could originate as easily from a near-surface source unaffected by 34P as one specific to the event, such as one connected to the persistent, anomalously dusty mesoscale circulation(s) to the east of Tharsis inferred by Heavens et al. (2019).

High-resolution global climate modeling by Kuroda et al. (2020) reproduced the global reduction in GW activity and a disproportionate reduction in the tropics and the southern extratropics and improved GW vertical propagation conditions during 34P. But this study also simulated that GW activity near Tharsis changed minimally. Kuroda et al. (2020) ran $1.5^\circ \times 1.5^\circ$ (67 km) resolution GCM simulations with a dust scenario based on 34P column dust opacity measurements (Montabone et al. 2020) and compared it to a simulation with a climatological “low-dust” scenario. Indeed, Kuroda et al. (2020) noted that the zonal average reduction in GW activity in their simulations is typically about a factor of 2, which is smaller than that found by Heavens et al. (2020). This difference is attributed to sampling of a longer horizontal wavelength portion of the spectrum by the GCM simulations. However, this analysis suggests that this difference plausibly arises from high dust opacity raising the sensitivity altitude range of the A3 channel, such that the change in GW activity is more accurately assessed by comparing GW activity sampled by A3 under dust-free conditions with GW activity in A2 sampled under dusty conditions, which implies a lower reduction in GW activity than found by Heavens et al. (2020). If Kuroda et al.’s (2020) simulations accurately diagnose the dynamics behind the GW activity reduction, the reduction arises from a weaker diurnal cycle of surface heating below the dust haze suppressing generation of GWs from planetary waves and geostrophic adjustment generally. The minimal or small positive impact of 34P on northern extratropical GW activity is attributed to orographic GW sources, which are apparently less affected by the effects of high dust opacity, though it seems possible that the persistence of GW activity in the northern extratropics during 34P is the result of the robustness of the winter westerly jet, embedded baroclinic activity, and the polar vortex. The observational test for this would be to look for changes in GW activity from a regional dust storm or PEDE that significantly disturbed the polar vortex, such as 26C (a regional dust storm in MY 26) or 28P (Wang 2007; Guzewich et al. 2016; Kass et al. 2016; Battalio & Wang 2020), rather than 34P, which minimally impacted 50 Pa temperatures north of 60°N (Kass et al. 2020).

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**Figure 26.** Nightside GW variability and intermittency at $L_s = 0^\circ–30^\circ$. (a)–(c) $WF_{\text{GW}}$ (K2 K$^{-2}$) averaged at 1° resolution for all off-nadir observations at surface elevations <12 km and MYs 29–33 and 35 for the labeled channel. (d)–(f) Intermittency (%) at 2° resolution for the labeled channel. The difference in resolution is to ensure sufficient sampling to reliably calculate intermittency.
Figure 27. Limb and off-nadir observations of a loop feature/presumptive cloud associated with high $\Omega_{\text{GW}}$ in A3. (a)–(c) Cross sections of the brightness temperature in the labeled channel near 43°W, MY 34, $L_s = 150.27$ (2018 March 28, 07:45–07:59 UTC). The color scale has been saturated at 120 and 180 K to emphasize the loop feature. The latitudes of the limb observations (at the tangent point of each detector) in panels (d)–(f) are indicated with horizontal lines, while the latitudes of off-nadir observations in panels (g)–(i) are depicted with slanted lines to show the projected observational path of off-nadir observations relative to the limb observations. The latitude axis has been reversed to show that the direction of spacecraft motion is toward the southern pole. (d)–(f) Comparison of two limb observations in the labeled channel. The dashed line plots a profile near the peak altitude of the loop. The squares plot a profile in the shadow region on the far side of the loop feature from the spacecraft. (g)–(i) Off-nadir observations in the labeled channel, contrasting a high-$\Omega_{\text{GW}}$ observation (one of a pair) with observations from the pair to the south and north. The $\Omega_{\text{GW}}$ for the observation is provided in the legend.
Kuroda et al. (2020) also noted that the longer horizontal wavelengths resolved by the simulations are much longer than the expected scale of GWs generated by dry convection, so, while it is expected that higher dust opacity will stabilize the atmosphere, suppress convection, and suppress convectively generated GWs, the simulations are unable to simulate this effect. The similarity in magnitude and spatial distribution in the reduction in GW activity between observations at the shorter horizontal wavelengths sampled by MRO-MCS off-nadir views and those resolved by Kuroda et al. (2020) seem to support the suppression of dry convection by 34P. The subtlety identified by the observations analyzed here is that there is significant orographic GW activity near the surface at the observationally resolved scales within the tropics (e.g., Valles Marineris), but activity in these locations, too, is significantly weaker during 34P, even at night (Figures 21(c), (f), and (i)). This reduction likely stems from the suppression of dry convection, which normally generates intense mesoscale winds along steep slopes in Valles Marineris and similarly rough areas and potentially substantial orographic GW activity (Toigo & Richardson 2002; Rafkin & Michaels 2003; Michaels et al. 2006; Clancy et al. 2021).

The other prediction by Kuroda et al. (2020) that seems confirmed by this analysis is that dissipation of GWs with altitude would have decreased in the tropics, at least in the sampled vertical range of the lowest three to four scale heights. The mechanism invoked is stronger jets and reduced critical level filtering. This change is analogous to the reduced critical level filtering that takes place near the westerly jets in the mid-high latitudes as they strengthen in fall and winter. Recent modeling by Rajendran et al. (2021) has proposed that the westerly jet in the northern hemisphere expanded toward the equator during 34P, which would be consistent with the reduced critical level filtering we have inferred.

5.2. Carbon Dioxide Ice Clouds and Other Phenomena that Might Be Mistaken for GWs

As outlined in Section 4.3, highly intermittent GW activity in the nightside tropics in the clear season of the year ($L_s = 330°−150°$) is the artifact of reemission by CO$_2$ ice clouds with larger particle sizes than assumed by Heavens et al. (2020). These anomalies are detectable as depressions in off-nadir brightness temperature that are significant in A1–A3 and potentially removable by statistical techniques for outlier exclusion. Because tropical mesospheric CO$_2$ ice clouds are rarer, thinner, and have larger particle sizes on the dayside than the nightside (Clancy et al. 2019), these artifacts are probably of minimal importance outside the nightside tropics in the clear season.

While MCS observations cannot distinguish GW activity from CO$_2$ ice clouds in the nightside tropics, the formation of these clouds is thought to be impossible without GW-driven thermal fluctuations (Listowski et al. 2014; Yiğit et al. 2018). Thus, the relative spatial distribution of GW activity at 60 km is probably captured by our analysis in the nightside tropics, making even this artifactual information potentially useful for model validation.
Along with potential impacts from non-GW mesoscale variability above the surface, on-planet observations in A1 and A2 have a large enough potential surface contribution to enable surface temperature/composition variability to resemble GW activity (Figures 1–2), which suggests that caution should be exercised when interpreting A1 and A2 brightness temperature variability in areas where surface temperature could be changing rapidly because of composition, such as the southern summer polar cap (Titus et al. 2003; Piqueux et al. 2008).

Analogously, the correlation between A1 and A2 brightness temperature variability and roughness at higher values of roughness poses an additional question: whether this relationship is due to a balance between the momentum flux of GW activity and the surface stress, that is, it is an indicator of significant orographic GW sources, or whether it arises from roughly uniform surface temperature lapse rates being translated into the atmosphere by radiative–convective processes. For example, if surface temperature uniformly increased or decreased with elevation, it is possible that

\[ \Omega_{GW} = \sigma^2 \Gamma^2, \]

where \( \Gamma \) is the lapse rate. In that case, \( \Gamma^2 \) is typically equal to \( 9 \times 10^{-2} \text{ K}^2 \text{ m}^{-2} \) (assuming our fit by inspection of the data and a typical temperature of 170 K) or \( 1 \times 10^{-2} \text{ K m}^{-1} \) (0.96 K km\(^{-1}\)). On one hand, this is a low but still plausible atmospheric lapse rate near the surface of Mars, where the dry convective lapse rate is approximately 4.5 K km\(^{-1}\). On the other hand, low-elevation surfaces do not seem to be monotonically colder or warmer on Mars. Figures A1(d) and A2(d) show a rough decreasing trend in B1 brightness temperature with elevation in nadir and off-nadir views, but this relationship is not monotonic. Elevations close to 20 km can be as warm as elevations near 0 km or lower. Moreover, the trend in brightness temperature with elevation in A1 is positive in the nadir and insignificant in the off-nadir (Figures A1(a) and A2(a)) with the same caveats about scatter.

Even on Earth, the free atmospheric lapse rate can be a poor approximation for the decrease in surface temperature and the near-surface air temperature with elevation because of preferential solar heating of slopes, drainage of cold air into valleys, and variations in surface cover by snow or vegetation (Minder et al. 2010; Pepin et al. 2016). On Mars, physical considerations suggest that departure of the decrease in surface temperature and the near-surface air temperature with elevation from the free atmospheric lapse rate should be more common. The surface energy balance is primarily controlled by radiative fluxes. Stronger radiative cooling at night at higher elevations likely drives katabatic flows down slopes all over the planet at night (Spiga 2011), while solar heating on mountain summits during the day and the resulting high infrared surface fluxes drive anabatic flow up slopes during the day (Rafkin et al. 2002). Thus, mesoscale atmospheric circulations generally tend to equilibrate temperature gradients along slopes. It is thus more likely that a correlation between surface roughness and temperature variability at the surface or in the atmosphere arises from a correlation between mesoscale variability and rough surfaces than observation of variable topography in a region with a uniform lapse rate.

Another alternative explanation for the strong relationship between brightness temperature variance in A1 and surface roughness is inhomogeneous heating of rough surfaces. A smooth surface of uniform composition will heat uniformly, but one with slopes will heat preferentially according to the orientation of the slopes relative to the Sun. Thus, greater surface temperature variability would be expected over rougher terrain during the day. At night, a similar effect might be caused by cliff and crater walls having higher thermal inertia than crater/cliff floors (Edwards et al. 2011).

To test the idea that the observed variance in the lowermost atmosphere or at the surface might arise from inhomogeneous heating/cooling of rough surfaces, the annual mean surface temperature field output by the KRC numerical thermal model from Kieffer (2013) was analyzed. The KRC model simulates the surface and subsurface energy balance for Mars, accounting for slope (at model resolution), thermal inertia, albedo, insolation, and infrared downwelling radiation (Kieffer 2013). Analyzing the annual mean field allows constant factors of surface composition and morphology that might affect kilometer-scale variability in surface temperature from variable factors like insolation and infrared downwelling radiation. Normalized variance on the approximate scale of the MRO-MCS off-nadir observations was estimated by (1) dividing the 0.05 KRC model output into 1° × 1° bins, consisting of 400 data points each; (2) taking each of the 20 sets of 20 points running north–south in each bin at constant longitude, linearly detrending the surface temperature data for these points, calculating the variance, and normalizing by the square of the surface temperature to find the normalized variance for each set; and (3) averaging the normalized variances in each bin.

The resulting normalized variance field is in Figure 29. Surface temperature variability because of surface inhomogeneities is highest along the southern rim of Argyre and the dichotomy boundary to the northeast of Arabia Terra. Several additional features with high normalized variance in Figure 29 have high normalized variance in the A1 nightside, e.g., Figure 13. But many features do not match. Moderate variance is predicted by KRC in the lower thermal inertia regions approximately spanning the tropics from 180°W–60°W and 30°W–60°E, but these regions are not clear features in Figure 13. Normalized variance in A1 is also concentrated more strongly around the rims of the Tharsis and Elysium volcanoes (Figure 13) than in KRC (Figure 29). Moreover, KRC-predicted normalized variance is moderate over Valles Marineris, while Valles Marineris is typically the global maximum of A1 normalized variance. This analysis suggests that an atmospheric phenomenon dependent on roughness is a better explanation for the normalized variance distribution in A1 than inhomogeneous heating/cooling of the surface being translated into heating/cooling of the lowermost atmosphere.

5.3. Implications for GW Sources

Having considered potential non-GW contributions to small-scale brightness temperature variability, it is possible to discuss the implications of this study for GW sources. Heavens et al. (2020) concluded that nadir observations by MRO-MCS generally sampled GW and mesoscale variability associated with boundary layer convection, while off-nadir observations sampled boundary layer convection and three other sources: (1) near the winter poles, (2) one that migrated with the winter westerly jets, and (3) one near the equator at night during the L\(_s\) = 330°–150° period. Information from lower in the atmosphere suggests that high brightness temperature variability in the tropical night is mainly
dominated by the effects of mesospheric CO$_2$ ice clouds rather than GW activity. Analysis of A1 and A2 observations also somewhat weakens the reasoning behind nadir observations sampling boundary layer convection alone. Heavens et al. (2020) partly argued for boundary layer convection being sampled by nadir views on the basis of high-elevation areas having a higher $\Omega_{GW}$ in A3 than low elevation and the absence of a band of high $\Omega_{GW}$ at 50$^\circ$S in dayside nadir A3 observations that was present in dayside off-nadir A3 observations. It now appears that low-elevation areas like Valles Marineris have $\Omega_{GW}$ in A1 and A2 observations,
weakening the correlation between elevation and dayside $\Omega_{GW}$ in the nadir lower in the atmosphere (Figures 9(a) and (b), Table 1). Moreover, there is a band of high $\Omega_{GW}$ at 50°S in A1 and A2 dayside nadir observations that is absent in A3 (Figures 9(d)–(f)). One argument that is robust throughout the lower atmosphere is that $\Omega_{GW}$ is higher during the day than at night, suggesting that GW activity is driven by the diurnal cycle of solar heating. This line of argument, as well as the small horizontal scale of the waves sampled in the nadir, argues for some contribution by boundary layer convection. Off-nadir observations suggest that the high-$\Omega_{GW}$ band at 50°S tracks the seasonal cap edge (Piqueux et al. 2015b). The disappearance of this band in A3 nadir observations but not in A3 off-nadir observations suggests that short horizontal, large vertical wavelength GWs from this source are less likely to propagate vertically than horizontally longer, vertically narrower waves. Filtering like this would imply that total internal reflection of GWs is common in the southern winter extratropics (Fritts & Alexander 2003).

In all channels during winter, GW activity near the winter poles is present and largely structureless at the north pole (Figures 11(d), (h), and (j)) but has some embedded structures at the south pole in the winter in A1 that disappear higher in the atmosphere (Figures 11(a), (e), and (j)). The attribution of this activity to CO₂ ice clouds made by Heavens et al. (2020) remains plausible, but caution must be exercised in distinguishing variability in channels sensitive to the surface resulting from optically thick CO₂ clouds and snowfall from GW.

The GW activity that migrates with the winter westerly jets probably results from a combination of weakened filtering of GW activity in the winter westerly jets (particularly on the dayside) and weaker orographic and nonorographic GW activity associated with the jets (Figures 25(a)–(h)). These twin effects are easiest to see in northern hemisphere summer, where midlatitude GW activity (even in rough areas) is unusually weak in A1 (Figure 11(b)) and the A3/A1 activity ratio is at its minimum (Figures 25(a)–(h)).

Yet the major new insight provided by probing GW activity throughout the lower atmosphere is that GW activity near the surface mostly results from orographic GWs over rough surfaces. This was the initial expectation in past GCM design (Forget et al. 1999) and observational studies (Creasey et al. 2006), but its validation with global remote-sensing data is somewhat unexpected and has no parallel in GW studies at the Earth. This population, however, is largely trapped near the surface. Orographic GWs that propagate out of the lower atmosphere seem to do so in the westerly extratropical jets or easterly equatorial jet. The limited GW activity that reaches the middle atmosphere elsewhere may be nonorographic.

6. Summary

In this study, we investigated small horizontal-scale radiance variability in on-planet views by three MRO-MCS channels that sample GW activity from horizontal wavelengths of <100 km throughout most of the lower atmosphere. Our results confirm that orographic GW activity is widespread over rough...
areas of the planet near the surface but largely does not reach 25 km altitude outside of regions with significant jet streams. The GW activity throughout the lower atmosphere seems strongly driven by the diurnal cycle of solar heating, with dayside GW activity being generally stronger than nightside GW activity in most areas, as previously found for GW activity at 25 km by Heavens et al. (2020).

We also demonstrate that GW activity (as measured by a quantity proportional to potential energy) decreased by a factor of 2 globally in the lower atmosphere during the planet-encircling dust event of MY 34 (2018: 34P). The larger magnitude of this effect reported by Heavens et al. (2020) was biased by changing the altitude sensitivity of the relevant channels under high dust opacities. The reduction of GW activity in dust storms is largely driven by suppression of near-surface GW sources by reduced solar heating at the surface and the resulting processes, particularly in the tropics and southern extratropics. The northern extratropics were largely unaffected. But there is some evidence of reduced dissipation of GWs in the tropics during this event, which may indicate stronger winds in the lower few scale heights and reduced critical level filtering.

Finally, we demonstrate that caution must be exercised when attributing all small-scale radiance variability to GWs. Mesospheric clouds and significant variability in surface composition between ice and bare ground can all create small-scale radiance variability in the sampled channels and viewing geometries. It is often difficult to distinguish the signature of one of these phenomena from a poorly resolved GW. Artifacts in this analysis resulting from mesospheric CO₂ ice cloud activity, however, may be a good proxy for GW activity at the altitude of cloud formation.

Beyond identifying areas where there is GW activity and the local roughness is too low for orographic GWs to occur, we have not attempted a substantive investigation of nonorographic GW sources. In the case of convective sources above the boundary layer, it is difficult to confidently disentangle the GW signal from the associated clouds. Isolating potential frontal sources would require extensive image analysis and comparison with time series over smoother areas, which is well beyond the already large scope of this study. Nevertheless, the data set derived in this study and the artifacts identified in it should be extremely useful for further investigating nonorographic sources of GWs at Mars in the lower atmosphere, as well as better understanding orographic GWs and the conditions under which they propagate into the middle and upper atmosphere.

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The input to and results of the visibility calculations, diagnoses of variance and associated quantities, and averaged spatial fields of variance and associated quantities have been archived on Mendeley (Heavens 2022a, 2022b). The KRC mean temperature field is available in the supporting information of Kieffer (2013).

Appendix

The Surface Contribution over High-elevation Surfaces

A GW variance might be erroneously diagnosed in circumstances when surface elevation or thermal inertia changes rapidly within the same observation (Heavens et al. 2020). This phenomenon arises because observations over high-elevation surfaces have higher surface contributions than those over low-elevation surfaces (Wilson & Richardson 2000), so brightness temperatures over high-elevation surfaces are sensitive to warm surface temperatures rather than the atmosphere. Mixing low- and high-elevation surfaces in an observation can thus produce apparently high variance in brightness temperature.

To investigate whether this effect could occur in the channels considered in this study, the relationship between brightness temperature and surface elevation in individual measurements (that is, at a detector-by-detector level) was studied. Data were selected within a narrow seasonal window and a small area of Mars with a broad range of elevations to determine if there was a critical elevation at which brightness temperatures were much higher than the typical brightness temperatures observed at low elevation, as in Figure A.1 of Heavens et al. (2020). The main methodological change from Heavens et al. (2020) required to do this is to extend the filtering procedure based on A3 observations used in Heavens et al. (2020) to other channels. The details and consequences of this filtering procedure are discussed in Section 3.1. To get a more direct estimate of the surface contribution, we also investigated the elevation–brightness temperature relationship for the B1 channel (290–340 cm⁻¹). Here B1 is used to retrieve surface temperatures in MRO-MCS retrievals (Kleinböhl et al. 2011), so its brightness temperature is a reasonable approximation to the surface temperature (particularly in nadir geometry), except under high aerosol conditions.

The best example of the high-elevation surface contribution issue and its resolution is for the A3 nadir views (Figure A1(c)). Brightness temperatures slowly increase with surface elevation, but there is considerable scatter in a range between 182 and 192 K. However, beyond 12 km surface elevation, brightness temperatures increase nearly monotonically to 220 K at 20 km surface elevation. Excluding measurements from surface elevations greater than 12 km mostly eliminates this problem.

Interpreting the additional channels is more difficult. The A2 nadir views display a relation to elevation similar to A3, but brightness temperature increases faster with elevation at ∼5 to 5 km surface elevation (Figure A1(d)). The A1 brightness temperatures also increase with surface elevation, but the range of brightness temperatures at elevations lower than 12 km overlaps with brightness temperatures at elevations greater than 12 km (Figure A1(a)). In the B1 nadir views, brightness temperature actually decreases with surface elevation, and its trend is mostly flat near ∼250 K at surface elevations between 15 and 20 km, such that brightness temperature elevations greater than 12 km fall roughly in the middle of the range of brightness temperatures at elevations lower than 12 km (Figure A1(d)). Brightness temperatures at elevations greater than 12 km in A1 and A2 trend toward 250 K, the typical brightness temperature in B1. But brightness temperatures in A1 and A2 have the opposite trend with surface elevation at elevations lower than 12 km than B1, so A1 and A2 likely observe the atmospheric temperature rather than the surface temperature at elevations.
lower than 12 km. The positive trend of temperature with elevation may indicate that these measurements sense the convective boundary layer, whose height increases significantly with surface elevation (Hinson et al. 2008; Tellmann et al. 2013).

This analysis is little changed in the off-nadir, except that the magnitudes of increases and decreases in brightness temperature with surface elevation are generally smaller and not as monotonic at high elevation (Figures A2(a)–(d)), suggesting that the surface contributions over high-elevation surfaces are less of a problem than in the nadir. The one feature of interest is the rapid increase of brightness temperature with elevation near 12 km in A2, A3, and possibly A1 (Figures A2(a)–(c)). This increase corresponds to measurements within two observations made near 20°N, 132°W, that is, around 80 km to the north of the summit of Olympus Mons. On the dayside, MCS would make these observations to the south of the surface intersection point.

The horizontal weighting function of A3 in a case like this would peak around 80 km closer to the spacecraft than the intersection of the detector with the surface (see Off-Nadir_11 in Figure 3(c)). Therefore, some measurements are strongly sensitive to the summit of Olympus Mons and its surface contribution at up to 21 km, rather than the atmosphere/surface at 12–13 km surface elevation.

Therefore, because of the strong sensitivity of A2 and A3 nadir measurements to surface temperatures at surface elevations greater than 12 km, it seems justifiable to exclude observations at surface elevations greater than 12 km to reduce false-positive diagnoses of GW activity over high topography. The A1, A2, and A3 off-nadir measurements seem to be less sensitive to surface temperatures at surface elevations greater than 12 km or may transition in sensitivity at significantly greater surface elevation. However, 12 km still seems to be a reasonable criterion for exclusion because of the difficulty of registering any measurement at an exact surface elevation.

**References**

Altieri, F., Spiga, A., Zasova, L., Bellucci, G., & Bibring, J.-P. 2012, *JGRE*, 114, E10004

Ando, H., Inamura, T., & Tsuda, T. 2012, *JAtS*, 69, 2906

Barnes, J. R. 1990, *IGARSS*, 95, 1401

Battalio, R. M., Bougher, S. W., Roeten, K., Sharrar, R., & Murphy, J. 2019, *LPICo*, 2089, 6338

Elrod, M. K., Bougher, S. W., Roeten, K., Sharrar, R., & Murphy, J. 2020, *GeoRL*, 47, e84378

Elrod, M. K., Bougher, S. W., Roeten, K., Sharrar, R., & Murphy, J. 2020, *GeoRL*, 47, 2906

England, S. L., Liu, G., Yiğit, E., et al. 2017, *JGRA*, 122, 2310

Erni, M., Preusse, P., Alexander, M. J., & Warner, C. D. 2004, *JGRE*, 109, D20103

Forget, F., Hourdin, F., Fournier, R., et al. 1999, *JGR*, 104, 24155

Fritts, D. C., & Alexander, M. J. 2003, *BV**, 41, 1003

Fritts, D. C., Isler, J. R., Thomas, G. E., & Andreassen, O. 1993, *GeoRL*, 20, 2039

Gilli, G., Forget, F., Spiga, A., et al. 2020, *JGRE*, 125, e05873

Guzewich, S. D., Lemmon, M., Smith, C. L., et al. 2019, *GeoRL*, 46, 71

Guzewich, S. D., Toigo, A., & Waugh, D. 2016, *Icar*, 278, 100

Hayne, P. O., Puige, A. D., Schofield, J. T., et al. 2012, *JGRE*, 117, E08014

Heale, C. J., Bossert, K., Vadas, L. S., et al. 2020, *JGRE*, 125, e13662

Heaven's, N. 2022a, Brightness Temperature Variences from On-Planet Views in the A1–A3 Channels by the Mars Climate Sounder, version 2, Mendeley Data, doi:10.17632/5knbynd921.4

Heaven's, N. 2022b, Brightness Temperature Variences from On-Planet Views in the A1–A3 and B1 Channels by the Mars Climate Sounder, version 1, Mendeley Data, doi:10.17632/y75kzct93d.1

Hinson, N. G., Benson, J. L., Kass, D. M., et al. 2010, *GeoRL*, 37, L18202

Heaven's, N. G., Kass, D. M., Kleinböhl, A., & Schofield, J. T. 2020, *Icar*, 341, 113638

Kass, D. M., Kleinböhl, A., McCleese, D. J., Schofield, J. T., & Smith, M. D. 2016, *GeoRL*, 43, 6111

Kass, D. M., Kleinböhl, A., McCleese, D. J., Schofield, J. T., & Smith, M. D. 2016, *GeoRL*, 43, 6111

Kass, D. M., Kleinböhl, A., & McIver, Jr., E. 2019, *JGRE*, 124, 1618

Kiefier, H. H. 2013, *JGRE*, 118, 451

Kleinböhl, A., Schofield, J. K., et al. 2010, 38th COSPAR Scientific Assembly, 38, 7

Kleinböhl, A., Schofield, J. K., Abdou, W. A., Irwin, P. G. J., & de Kok, R. J. 2011, *JQSRT*, 112, 1568

Kleiner, A. 1978, The Mars Reference Atmosphere. Proc. of the Twenty-first Plenary Meeting (Pasadena, CA: JPL, Caltech)

Kuroda, T., Medvedev, A. S., & Yiğit, E. 2020, *JGRE*, 125, e06556

Kuroda, T., Medvedev, A. S., Yiğit, E., & Hartogh, P. 2016, *JAS**, 73, 4895

Kuroda, T., Yiğit, E., & Medvedev, A. S. 2019, *JGRE*, 124, 1618

Leelavathi, V., Venkateswararao Rao, N., & Rao, S. V. B. 2020, *JGRE*, 125, e06649

Lewis, S. R., Collins, M., Read, P. L., et al. 1999, *JGR*, 104, 24177

Lis Autos, C., Miattačinov, A., Montmessin, F., Spiga, A., & Lefèvre, F. 2014, *Icar*, 237, 239

Liu, J., Jin, S., & Li, Y. 2019, *JIRA*, 124, 9315

Lott, F., & Miller, M. J. 1997, *QJRMS*, 123, 101

MCD. cited 2022a, *Mars Climate Database v5.3: The Web Interface*, LMD du CEA, Paris

MCD. cited 2022b, *MRO MCS Reduced Data Records* (RDR), NASA

Parish, H. F., Schubert, G., Hickey, M. P., & Walterscheid, R. L. 2009, *Icar*, 203, 28

Pepe, N. C., Maida, E. E., & Williams, R. 2016, *JGRE*, 121, 9998

Offermann, D., Jarisch, M., Oberheide, J., et al. 2006, *JASTP*, 68, 1709

Montabone, L., Spiga, A., Kass, D. M., et al. 2020, *JGRE*, 125, e06111

Minder, J. R., Mote, P. W., & Lundquist, J. D. 2010, *JGRD*, 115, D14122

Millour, E., Forget, F., & Spiga, A. 2015, *EPSC*, 10, 438

MCD. cited 2022a, *Mars Climate Database v5.3: The Web Interface, LMD du CNRS, Paris, France, http://www.mars.lmd.jussieu.fr/.*

MCD. cited 2022b, *MRO MCS Reduced Data Records* (RDR), NASA

Parish, H. F., Schubert, G., Hickey, M. P., & Walterscheid, R. L. 2009, *Icar*, 203, 28

Pepe, N. C., Maida, E. E., & Williams, R. 2016, *JGRE*, 121, 9998
Piqueux, S., Byrne, S., Kieffer, H. H., Titus, T., & Hansen, C. J. 2015a, Icar, 251, 332
Piqueux, S., Edwards, C. S., & Christensen, P. R. 2008, JGRE, 113, E08014
Piqueux, S., Kleinböhl, A., Hayne, P. O., et al. 2015b, Icar, 251, 164
Rafkin, S. C. R., & Michaels, T. I. 2003, JGRE, 108, 8091
Rafkin, S. C. R., Sta. Maria, M. R. V., & Michaels, T. I. 2002, Natur, 419, 697
Rajendran, K., Lewis, S. R., Holmes, J. A., et al. 2021, GeoRL, 48, e94634
Selton-Nash, E., Teanby, N., Montabone, L., et al. 2013, Icar, 222, 342
Shirley, J. H., Kleinböhl, A., Kass, D. M., et al. 2020, GeoRL, 47, e84317
Siddle, A., Mueller-Wodarg, I., Stone, S., & Yelle, R. 2019, Icar, 333, 12
Slipski, M., Jakosky, B. M., Benna, M., et al. 2018, JGRE, 123, 2939
Smith, D., Neumann, G., Arvidson, R. E., Guinness, E. A., & Slavney, S. 2003, Mars Global Surveyor Laser Altimeter Mission Experiment Gridded Data Record: MGS-M-MOLA-5-MEGDR-L3-V1.0, NASA Planetary Data System Geosciences Node, https://pds.nasa.gov/ds-view/pds/viewProfile.jsp?dsid=MGS-M-MOLA-5-MEGDR-L3-V1.0
Spiga, A. 2011, P&SS, 59, 915
Spiga, A., González-Galindo, F., López-Valverde, M. Á., & Forget, F. 2012, GeoRL, 39, L02201
Spiga, A., Hinson, D. P., Madeleine, J.-B., et al. 2017, NatGe, 10, 652
Tellmann, S., Pätzold, M., Häusler, B., Hinson, D. P., & Tyler, G. L. 2013, JGR, 118, 306
Titus, T. N., Kieffer, H. H., & Christensen, P. R. 2003, Sci, 299, 1048
Toigo, A. D., & Richardson, M. I. 2002, JGRE, 107, 5049
Vadas, S. L., & Fritts, D. C. 2001, JAtS, 58, 2249
Vadas, S. L., Zhao, J., Chu, X., & Becker, E. 2018, JGRD, 123, 9296
Vals, M., Spiga, A., Forget, F., et al. 2019, P&SS, 178, 104708
Walterscheid, R. L., Hickey, M. P., & Schubert, G. 2013, JGRE, 118, 2413
Wang, H. 2007, Icar, 189, 325
Williamson, H. N., Johnson, R. E., Leclercq, L., & Elrod, M. K. 2019, Icar, 331, 110
Wilson, R. J. 1997, GeoRL, 24, 123
Wilson, R. J., & Richardson, M. I. 2000, Icar, 145, 555
Wolff, M. J., & Clancy, R. T. 2003, JGRE, 108, 5097
Wright, C. J., Osprey, S. M., & Gille, J. C. 2013, JGRD, 118, 10,980
Wu, D. L., & Eckermann, S. D. 2008, JAtS, 65, S695
Wu, D. L., Preusse, P., Eckermann, S. D., et al. 2006, AdSpR, 37, 2269
Xu, S., Liemohn, M., Bougher, S., & Mitchell, D. 2015, GeoRL, 42, 9702
Yamanaka, M. 1995, AdSpR, 15, 47
Yiğit, E., England, S. L., Liu, G., et al. 2015, GeoRL, 42, 8993
Yiğit, E., & Medvedev, A. S. 2015, AdSpR, 55, 983
Yiğit, E., Medvedev, A. S., Benna, M., & Jakosky, B. M. 2020, GeoRL, 48, e92095
Yiğit, E., Medvedev, A. S., & Hartogh, P. 2018, AnGeo, 36, 1631