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1 **Gravity wave activity in the Martian atmosphere at**
 2 **altitudes 20-160 km from ACS/TGO occultation**
 3 **measurements**

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13 **Key Points:**

- 14 • Observations of gravity waves from the Atmospheric Chemistry Suite instrument
 15 on board ExoMars Trace Gas Orbiter are presented
- 16 • Global distributions of the observed wave activity, potential energy, momentum
 17 fluxes and wave drag agree well with model predictions
- 18 • We found no correlation between wave amplitudes and buoyancy frequency, an ex-
 19 tension of previously observed anticorrelation with temperature

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20 **Abstract**

21 The paper presents observations of gravity wave-induced temperature disturbances in
 22 the Martian atmosphere obtained with the mid-infrared (MIR) spectrometer, a chan-
 23 nel of the Atmospheric Chemistry Suite instrument on board the Trace Gas Orbiter (ACS/TGO).
 24 Solar occultation measurements of a CO₂ absorption band at 2.7 μm were used for re-
 25 trieval density and temperature profiles between heights of 20 and 160 km with ver-
 26 tical resolution sufficient for deriving small-scale structures associated with gravity waves.
 27 Several techniques for distinguishing disturbances from the background temperature have
 28 been explored and compared. Instantaneous temperature profiles, amplitudes of wave
 29 packets and potential energy have been determined. Horizontal momentum fluxes and
 30 associated wave drag have been estimated. The analyzed data set of 144 profiles encom-
 31 passes the measurements made over the second half of Martian Year 34, from the So-
 32 lar longitude 165° through 355°. We observe enhanced gravity wave dissipation/breaking
 33 in the mesopause region of 100-130 km. Our analysis shows no direct correlation between
 34 the wave amplitude and Brunt-Väisälä frequency. It may indicate that convective insta-
 35 bility may not be the main mechanism limiting gravity wave growth in the middle at-
 36 mosphere of Mars.

37 **Plain Language Summary**

38 Gravity waves (GWs) of lower atmospheric origin continuously disturb the Mar-
 39 tian atmosphere. While propagating upward, their amplitudes grow and eventually GWs
 40 break up or dissipate. The deposited momentum and energy are the major mechanisms
 41 driving the circulation in the thermosphere above 100 km. Since spatial scales of GWs
 42 are relatively small, they are difficult to measure. Atmospheric Chemistry Suite (ACS)
 43 instrument on board the ExoMars Trace Gas Orbiter allows for extracting altitude pro-
 44 files of density and temperature from the troposphere to the thermosphere (20-160 km)
 45 with high vertical resolution, around 2 km. The instrument measures the solar spectrum
 46 occulted by the atmosphere with the carbon dioxide absorption in the middle infrared
 47 wavelength range. The observations provide latitudinal and seasonal coverage of the GW
 48 activity and its characterization on Mars. Our results allow for the first observational
 49 validation of model predictions, quantifying dynamical effects of GWs and constraining
 50 Martian general circulation models.

51 **1 Introduction**

52 The structure and circulation of planetary atmospheres are strongly affected by grav-
 53 ity waves (GWs), which are ubiquitous in any convectively stable atmosphere. They are
 54 primarily responsible for energy and momentum transfer from the lower to the upper at-
 55 mosphere. Historically, GW-induced coupling was extensively studied in Earth’s atmo-
 56 sphere (e.g., see reviews by Fritts & Alexander, 2003; Yiğit & Medvedev, 2015). With
 57 the progress in space exploration, the atmosphere of Mars has become the second best-
 58 studied example. Numerous space missions accompanied by numerical modeling have
 59 delivered ample evidence for the importance of GWs on Mars. Some of the Martian GW
 60 effects, their commonality and specifics with those on Earth, have been summarized in
 61 the recent review by Medvedev and Yiğit (2019). Observational knowledge of GW ac-
 62 tivity on Mars is crucial but still insufficient for quantifying their effects and constrain-
 63 ing Martian general circulation models (MGCs). Our paper addresses this problem by
 64 utilizing high-resolution occultation data obtained from the Atmospheric Chemistry Suite
 65 (ACS) instrument on board the Trace Gas Orbiter (TGO).

66 Observations of the Martian GWs have been conducted from orbiters by different
 67 remote sensing techniques and in situ methods. In situ measurements of GW-induced
 68 density fluctuations in the thermosphere were performed with accelerometers during aer-
 69 obraking operations by several spacecraft including Mars Global Surveyor (MGS), Mars

70 Odyssey (ODY), Mars Reconnaissance Orbiter (MRO), Mars Atmosphere and Volatile
 71 EvolutioN (MAVEN) and Trace Gas Orbiter (TGO) (Keating et al., 1998; Creasey et
 72 al., 2006a; Fritts et al., 2006; R. H. Tolson et al., 2005; R. Tolson et al., 2008; Withers,
 73 2006; Jesch et al., 2019; Vals et al., 2019; Siddle et al., 2020). GWs in the upper ther-
 74 mosphere were also measured by Neutral Gas and Ion Mass Spectrometer (NGIMS) on
 75 board MAVEN (Yiğit et al., 2015; England et al., 2017; Terada et al., 2017). Temper-
 76 ature and density disturbances associated with GWs have been remotely retrieved from
 77 stellar, solar and radio occultation data as well as from limb observations (Hinson et al.,
 78 1999; Creasey et al., 2006b; Ando et al., 2012; Wright, 2012; Heavens et al., 2020; Nak-
 79 agawa et al., 2020).

80 The shortcoming of many previous GW observation techniques is their limited alti-
 81 tude coverage. For example, in situ measurements were confined to a relatively nar-
 82 row vertical range in the thermosphere, while radio occultation and infrared limb sound-
 83 ing allowed for studying the lowermost (0–40 km) part of the atmosphere. Remote sens-
 84 ing in UV permitted the extension of the altitude coverage. The Spectroscopy for the
 85 Investigation of the Characteristics of the Atmosphere of Mars (SPICAM) instrument
 86 on board Mars Express (MEX) measured temperature and density profiles between 60
 87 and 130 km (Forget et al., 2009). The Imaging Ultraviolet Spectrograph (IUVS) on board
 88 MAVEN explored the thermospheric layers (100–150 km) (Medvedev et al., 2016; Gröller
 89 et al., 2018). Recently, Nakagawa et al. (2020) obtained temperature profiles from IUVS
 90 data spanning the atmosphere from 20 to 140 km with a vertical sampling better than
 91 6 km. ACS/TGO is the first IR instrument that allows for measuring in the solar oc-
 92 cultation mode temperature and density distributions within an even broader range of
 93 altitudes (20–160 km). Since April 2018, the Middle InfraRed (MIR) spectrometer, one
 94 of the three spectrometers of ACS, delivered several hundreds of profiles, some of which
 95 having vertical resolution allowing for the study of GWs.

96 Smaller-scale GW-induced temperature perturbations have to be first separated
 97 from the background temperature field associated with the large-scale variations. Although
 98 this procedure has been frequently performed in the terrestrial and Martian context, there
 99 is no universal technique that can be applied to vertical profiles obtained from different
 100 instruments (Ehard et al., 2015). In this work, we explore the sensitivity of several meth-
 101 ods for extracting GWs from the ACS MIR data and describe their applications for der-
 102 iving various characteristics of the GW field (amplitudes, wave potential energy, mo-
 103 mentum fluxes and wave drag) from the first available set of ACS/TGO data.

104 The paper is structured as follows. The ACS/TGO experiment and the instrument
 105 itself are outlined in Section 2. Section 3 describes the methods used in this study. In
 106 particular, retrievals of temperature profiles from measured spectra are presented in sec-
 107 tion 3.1, the techniques for extracting wave disturbances are given in section 3.2. Sub-
 108 section 3.3 describes the derivation of wave activity (amplitude of wave packets) and po-
 109 tential energy, and 3.4 outlines the calculation of the absolute vertical flux of horizon-
 110 tal momentum and momentum forcing of the mean flow. The results are presented in
 111 section 4. They include a case study (4.1), the spatial distribution of wave characteris-
 112 tics (4.2), and the relationship between wave amplitudes and the Brunt-Väisälä frequency
 113 (4.3). Conclusions are given in section 5.

114 2 Atmospheric Chemistry Suite Instrument on Board Trace Gas Or- 115 biter

116 ACS is a set of three infrared spectrometers for ExoMars 2016 TGO mission. It
 117 has been operating in the Martian orbit since April 2018. ACS consists of the near-(NIR),
 118 middle-(MIR) and thermal-infrared (TIRVIM) channels, that altogether cover the broad
 119 spectral range of 0.7–17 μ m. (Korablev et al., 2018). In this paper, we use the data re-
 120 trieved from the cross-dispersion echelle MIR spectrometer working in the solar oc-
 cul-

121 tation mode in the 2.3-4.2 μm range. This spectral coverage is achieved with a secondary
 122 dispersion grating, which can be rotated to one of 12 positions. During an occultation,
 123 the instrument is pointed to the Sun. Each measurement consists of an image at the $640 \times$
 124 512 pixels focal plane array (FPA), which accommodates up to 20 diffraction orders dis-
 125 persed over FPA by the secondary grating. One occultation covers 0.15-0.3 μm range.
 126 The instrument's resolving power is $\lambda/\Delta\lambda \sim 25\,000$ and the signal-to-noise ratio varies
 127 between 1000 and 10000. The vertical resolution of MIR depends on the integration time
 128 (~ 2 s per image) and ranges from 0.5 to 2.5 km. The transmission is obtained by divi-
 129 sion of the solar spectrum passed through the atmosphere to the reference one, which
 130 is measured above the altitude of 200 km, where the absorption by the atmosphere is
 131 negligible.

132 In this study, we use the 2.66-2.68 μm portion of the spectrum from the grating
 133 position #4, the echelle diffraction order #223, which includes a wing of the 2.7 μm CO_2
 absorption band (Figure 1a). Strong absorption lines of CO_2 allow for retrieving tem-

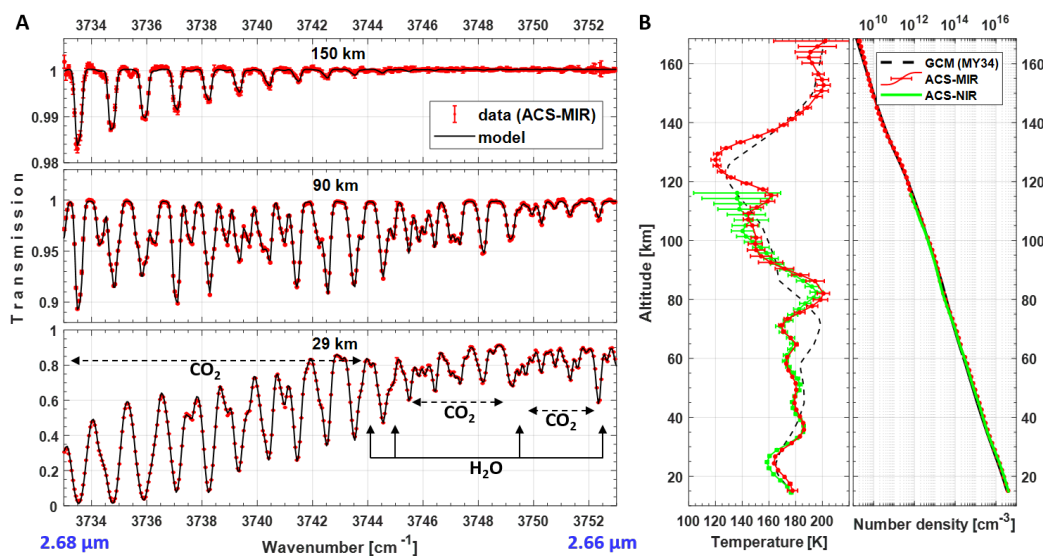


Figure 1. Spectroscopy of CO_2 and H_2O absorption in the diffraction order #223 of ACS-MIR (panel A) and an example of retrieved atmospheric temperature and density vertical profiles (panel B). a) Transmission spectra measured at tangent altitudes of 150, 90 and 29 km (red dots) on a background of the best-fitted models (black solid lines); b) Vertical profiles of temperature (left) and atmospheric number density (right) derived from the MCD (black dashed line), from ACS-MIR (red dots), and from ACS-NIR (Fedorova et al., 2020) (green dots). Error bars for the temperature values express 1- σ uncertainties of the retrievals.

134 perature and density in the Martian atmosphere with good sensitivity.
 135

136 3 Methods

137 3.1 Retrieval of Temperature Profiles

138 The retrieval scheme consists of several iterations. On the first step, we retrieve
 139 temperature and pressure from the rotational structure of CO_2 absorption bands in spec-
 140 tral intervals without H_2O lines (see Figure 1a). A priori altitude profiles of $T(z)$ and
 141 $p(z)$ as well as one of the CO_2 VMR, are taken from the Mars Climate Database (MCD)

142 for a specified occultation in MY34 (Millour et al., 2018). On the second step, we simul-
 143 taneously retrieve temperature and CO₂ concentration, while the pressure profile is kept
 144 constant assuming the hydrostatic equilibrium $p_{hyd}(z) = p_0(z_0) \exp[-\int_{z_0}^z \frac{g(z')M(z')}{RT(z')} dz']$,
 145 where g is the acceleration of gravity, M is the atmospheric molar mass and R is the gas
 146 constant. The reference pressure p_0 is chosen at an altitude z_0 , usually around 30-50 km,
 147 where uncertainties of the fitting are smallest. We repeat the second step 5-7 times until
 148 the profiles reach convergence. In each iteration, we apply the Tikhonov regulariza-
 149 tion (Tikhonov & Arsenin, 1977) for the temperature and concentration altitude pro-
 150 files with a smoothing coefficient less than 5 km. It defines the shortest wavelength to
 151 5-6 km when analyzing vertical wavy structures. The third step focuses only on CO₂ and
 152 H₂O concentration retrievals over the entire wavenumber range in order #223 (Figure 1a)
 153 using the $p(z)$ and $T(z)$ profiles already found. This step is not a subject of the present
 154 paper.

155 A similar fitting procedure, including the hydrostatic approximation, has been used
 156 in the work by Fedorova et al. (2020) (proprietary code) and Alday et al. (2019) (the NEME-
 157 SIS code, (Irwin et al., 2008)) in their retrievals of temperature and pressure from the
 158 ACS data. We validated our atmospheric temperature and number density profiles with
 159 simultaneous and collocated occultation measurements by ACS-NIR (Fedorova et al., 2020).
 160 An example comparison is presented in Figure 1b. A weaker CO₂ absorption band at
 161 1.58 μm measured by NIR allows for detection up to 110-120 km, or the density of \sim
 162 10^{12} cm^{-3} , while the band at 2.7 μm observed by MIR is measurable up to 160-170 km,
 163 or $\sim 10^9 \text{ cm}^{-3}$. The lowermost altitude of the temperature profile retrieval is conditioned
 164 by the aerosol opacity and by the saturation of the CO₂ absorption lines.

165 Each temperature value in a vertical profile was retrieved by fitting a modeled trans-
 166 mission spectrum J_{mod} to the measured one J_{mes} at a specified altitude. We model the
 167 spectra by the Beer-Lambert law

$$J_{mod}(\nu, z) = \exp \left[- \int (\sigma_{CO_2}(T, p)n_{CO_2}(z') + \sigma_{H_2O}(T, p)n_{H_2O}(z') + \tau_a) dz' \right], \quad (1)$$

168 where $n(z)$ are gaseous concentrations, $\sigma(T, p)$ are absorption cross-sections of CO₂ and
 169 H₂O correspondingly for specific temperature $T(z)$ and pressure $p(z)$ at an altitude z ,
 170 and τ_a is aerosol slant opacity. A transfer between the linear [cm^{-2}] and the volume [cm^{-3}]
 171 concentrations is performed using the well-known “onion-peeling” method with the nu-
 172 meric integration over all altitude layers z_i above the i -th one. Molecular cross-sections
 173 are calculated line-by-line on a basis of the HITRAN2016 database (Gordon et al., 2017)
 174 considering pressure-broadening coefficients of the H₂O lines suitable for a CO₂-rich at-
 175 mosphere (Gamache et al., 2016) and self-broadening in the case of CO₂. Then we con-
 176 volve the modeled spectrum by the previously determined instrument line shape (ILS)
 177 using wavenumber calibrations (see details in Alday et al., 2019). The fitting procedure
 178 is conducted by minimizing the “chi-square” function
 179 $\chi^2 = \sum_i A^2(\nu_i)$, $A(\nu_i) = [J_{mod}(\nu_i) - J_{mes}(\nu_i)]/\delta J$, where δJ are transmittance uncer-
 180 tainties, and the sum is taken over all considered spectral points (pixels). Our optimiza-
 181 tion algorithm to search for the χ^2 minimum is based on partial derivatives of the Ja-
 182 cobian matrix $\partial A/\partial X$ (Marquardt, 1963), where X is a vector of free parameters, i.e.,
 183 temperature, CO₂ concentration, H₂O mixing ratio, and aerosol slant opacity. Here, a
 184 significant contribution to the Jacobian comes from the rotational absorption lines, which
 185 are strongly sensitive to the temperature variability in the spectral range of interest.

186 3.2 Derivation of Wave Disturbances

187 Gravity wave-induced perturbations of temperature T' are sought by separating
 188 the mean, or background profile $\bar{T}(z)$ from the measured one $T(z)$:

$$T' = T - \bar{T}, \quad (2)$$

189 where the bar denotes an appropriate averaging. Generally, it implies averaging over wave
 190 phases, or spatial and temporal scales that are larger than the periods and wavelengths
 191 of contributing GW harmonics. In the case of almost instantaneous (with respect to the
 192 periods of GWs) occultation profiles, only separation in vertical scales is possible.

193 John and Kumar (2013) and Ehard et al. (2015) reviewed several common meth-
 194 ods of the partition of measured temperature and/or density profiles into the “mean”
 195 and wave components. They work well if a clear separation in vertical wavelengths does
 196 exist between GWs and large-scale motions belonging to the background. This is not al-
 197 ways the case in the Martian atmosphere, because vertical scales of disturbances asso-
 198 ciated with tides, planetary waves, and other motions may overlap with those due to GWs.
 199 It is desirable to retain the former in the background, but one still has to set a vertical
 200 scale Λ_z that separates GWs from the larger-scale features. In the following, we assumed
 201 $\Lambda_z = 30$ km. This value may lead to an overestimation of the retrieved wave activity
 202 by including non-GW perturbations, but at least no large-scale GW components are missed.
 203 Concerning the short-wavelength part of the spectrum, the limited vertical resolution
 204 favors detection of larger-scale waves, leaving out harmonics with smaller scales unob-
 205 served. Alexander (1998) has quantified this “observational filter” and pointed out that
 206 some large-scale harmonics refracted by the mean wind beyond the lowest resolution may
 207 be missing in observations as well.

208 We explore three methods: spectral filtering, sliding least-square polynomial fit and
 209 high-order polynomial fit. The former two have been discussed in relation to lidar and
 210 space-based measurements in the atmosphere of Earth (John & Kumar, 2013; Ehard et
 211 al., 2015, and the references therein), while the latter was applied to profiles obtained
 212 in the terrestrial (e.g., Spiga et al., 2008) and Martian atmosphere (Yiğit et al., 2015;
 213 Terada et al., 2017; Jesch et al., 2019). Since the ACS data are distributed irregularly
 214 over the altitude, they were first interpolated (oversampled) to an evenly spaced 500-m
 215 grid. We used only the temperature data with errors ≤ 10 K. Spectral filtering was per-
 216 formed using Fourier decomposition within sliding 60-km intervals (± 30 km around each
 217 point), and zero-order Fourier coefficients were used to calculate the background tem-
 218 perature. The examples are shown in Figure 2 for two characteristic profiles $T(z)$. They
 219 visibly differ: the profile in Figure 2a (orbit 2892n1) contains large-scale disturbances,
 220 while the one in Figure 2c (orbit 3251n1) comprises mostly smaller-scale fluctuations.
 221 This method yields smooth mean temperature profiles and, as a result, large deviations
 222 from the mean (Figure 2b and d). This is in particular obvious below 60 km and in the
 223 upper part of the domain (panels b and d).

224 For the sliding polynomial fit, we used a procedure described in the work of Whiteway
 225 and Carswell (1995). The background profiles are obtained by fitting cubic polynomi-
 226 als within the 60-km sliding intervals. Observational errors were used as weights, that
 227 assign a significance to the measurements at each altitude. At first, the intervals were
 228 shifted up from the bottom to top by a certain distance (shown in Figure 2a and c for
 229 2 and 11 km), and then the procedure was repeated for the downward shifts starting from
 230 the top. The overlapping values of fits from each range were then averaged. Thus ob-
 231 tained profiles were then smoothed using a moving average. At the bottom of the pro-
 232 files, we had to decrease the width of the sliding windows due to large spurious varia-
 233 tions in fitted polynomials and in order to make most of the observational data. The up-
 234 per and lower 4 km of thus obtained profiles have to be excluded anyway, because of the
 235 poor behavior of fitting polynomials, which cannot be averaged with counterparts from
 236 other sliding windows. This method occasionally produces disturbances oscillating not
 237 around zero. To correct for these numerical biases, we perform detrending by applying
 238 the Theil-Sen estimator (Theil, 1950; Sen, 1968) and fitting a linear function to the per-
 239 turbation profile. The Theil-Sen estimator is a robust method, which is used for deter-
 240 mining the linear regression taking the median of the slopes of all lines that can be drawn

241 through the given dataset. The linear function is then subtracted from the profile to obtain
 242 the corrected temperature.

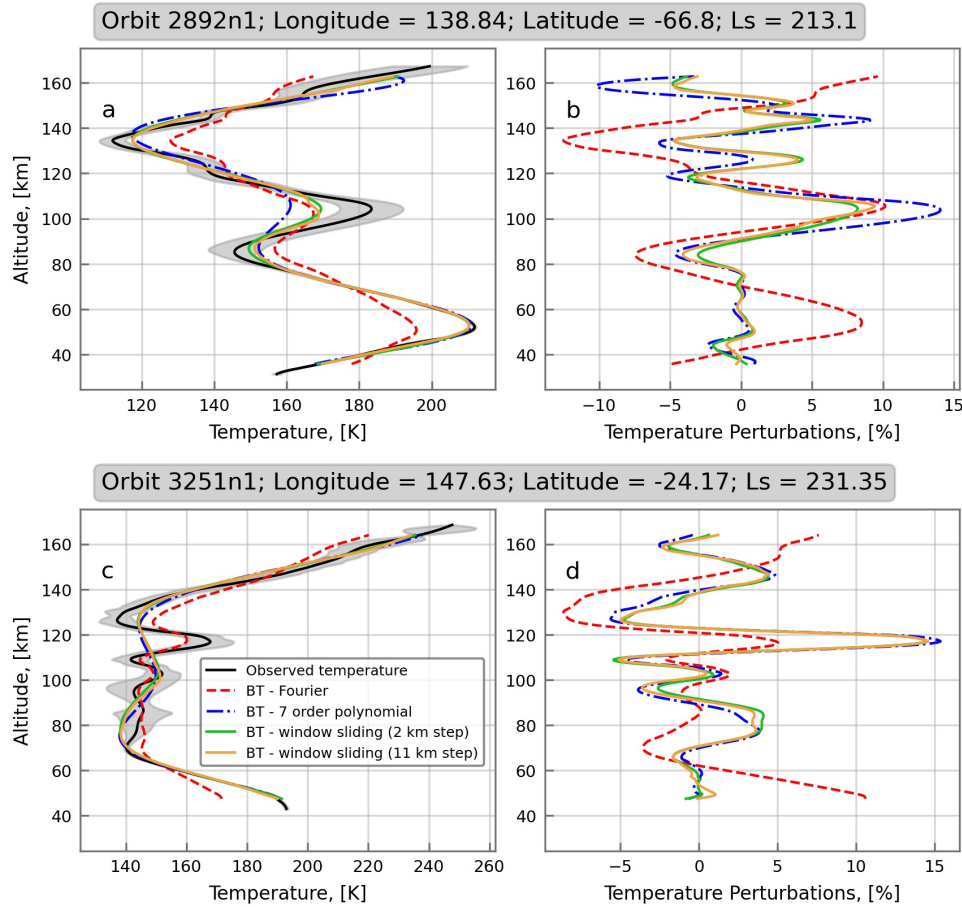


Figure 2. Separation of the observed temperature into the mean and wave components for two characteristic profiles: dominated by large vertical-scale (orbit 2892n1, upper row) and small-scale disturbances (orbit 3251n1, lower row). Left column is for the mean temperature $\bar{T}(z)$, the right one is for the relative perturbations $T'(z)/\bar{T}(z)$ (in percent). The legend describes the applied methods. Red dashed lines correspond to the Fourier decomposition, green and yellow lines are for the sliding polynomial fit with 2-km and 11-km shift steps, correspondingly, and the blue lines are for the 7-th order polynomial fit. The observed temperature profiles are given with the solid black lines. Shaded area denotes the uncertainty of the measurements.

243 The results for the sliding polynomial fit are plotted in Figure 2 for the 2 and 11
 244 km shift steps with green and yellow lines, correspondingly. It is seen that they are very
 245 close and, thus, the background and disturbances depend on the sliding step to a minor
 246 degree. The method shows some useful features in comparison with spectral filtering.
 247 The fitted mean curves in the regions of large-scale disturbances (Case 1) follow the ob-
 248 served temperature profiles closer (Figure 2a) and are smoother where small-scale struc-
 249 ture dominates (Case 2) (Figure 2c, between 70 and 130 km). This produces smaller wave
 250 amplitudes in Case 1, and reveals more wavy structures in Case 2. Especially plausible
 251 results are in the bottom of the profiles, where GWs are expected to have smaller am-
 252 plitudes (due to larger density).

253 We next explored the technique of fitting higher-order polynomials in the entire
 254 interval of heights. In particular, the seventh-order polynomial fit, which was previously
 255 used for extracting GWs on Mars (Yiğit et al., 2015; Jesch et al., 2019), produces most
 256 plausible results. They are presented in Figure 2 with dashed and dotted blue lines. It
 257 is immediately seen that thus obtained wave disturbances are in a very good agreement
 258 with those derived by the sliding polynomial fit method, especially for profiles contain-
 259 ing small-scale features (Figure 2d). For profiles dominated by large-scale perturbations,
 260 the agreement is also good in terms of the determined vertical structure of the wave, al-
 261 though the magnitudes are often exaggerated (Figure 2b). The weak point of the method
 262 is that it occasionally produces spurious disturbances near the edges of the vertical do-
 263 main with vertical gradients of the mean temperature directed opposite to the measured
 264 profiles. After careful consideration of the three methods applied to the available mea-
 265 surements, we selected the sliding third-order polynomial fit as the most appropriate and
 266 robust.

267 3.3 Wave Activity and Potential Energy

268 The GW field is often characterized by the magnitude of fluctuations $|T'| = (\overline{T'^2})^{1/2}$
 269 and wave potential energy (per unit mass)

$$E_p = \frac{1}{2} \left(\frac{g}{N} \right)^2 \overline{\left(\frac{T'}{T} \right)^2}, \quad (3)$$

270 where N is the Brunt-Väisälä frequency

$$N = \sqrt{\frac{g}{T} \left(\frac{dT}{dz} + \frac{g}{c_p} \right)}, \quad (4)$$

271 g is the acceleration of gravity and c_p is the specific heat capacity at constant pressure.
 272 The amplitude of the wave packet at a given height $|T'(z)|$ (hereafter called “wave ac-
 273 tivity”) represents an envelope of the measured profile $T'(z)$. We calculated it by per-
 274 forming Fourier decomposition in each sliding 60-km vertical interval and, based on Parce-
 275 val’s identity, summing up contributions of all harmonics. Examples of thus obtained en-
 276 velopes and potential energy for the same selected profiles as in section 3.2 are presented
 277 in Figure 3. Blue and red dashed lines denote the quantities calculated from the entire
 278 spectrum and by accounting for contributions of only two largest harmonics. It is seen
 279 that the neglect of shorter-scale harmonics, as was occasionally done in analyses of satel-
 280 lite observations (e.g., Ern et al., 2004), introduces little error to the estimated GW ac-
 281 tivity. However, the neglect of short-scale harmonics may lead to a noticeable underes-
 282 timation of wave potential energy, (cf. Figures 3b and d).

283 3.4 Momentum Flux and Momentum Deposition

284 Another useful characteristic of the GW field is the vertical flux of horizontal mo-
 285 mentum, or “momentum flux” for brevity, $\mathbf{F} = (F_x, F_y, 0) = \rho_0(\overline{u'w'}, \overline{v'w'}, 0)$, where
 286 ρ_0 is the mean density and (u', v', w') are the components of wave-induced perturbations
 287 of wind velocity \mathbf{u}' along with the two horizontal and the vertical axis, correspondingly.
 288 Momentum flux is constant for conservatively propagating waves. Breaking/dissipating
 289 GWs deposit their momentum to the mean flow, thus inducing an acceleration or decel-
 290 eration (depending on the sign) of the horizontal flow

$$(a_x, a_y) = -\frac{1}{\rho_0} \frac{d\mathbf{F}}{dz}. \quad (5)$$

291 The direction of the flux cannot be determined from the occultation measurements, how-
 292 ever total (or absolute) momentum fluxes for a harmonic $F_{k,m} = \sqrt{F_{x,k,m}^2 + F_{y,k,m}^2}$ can

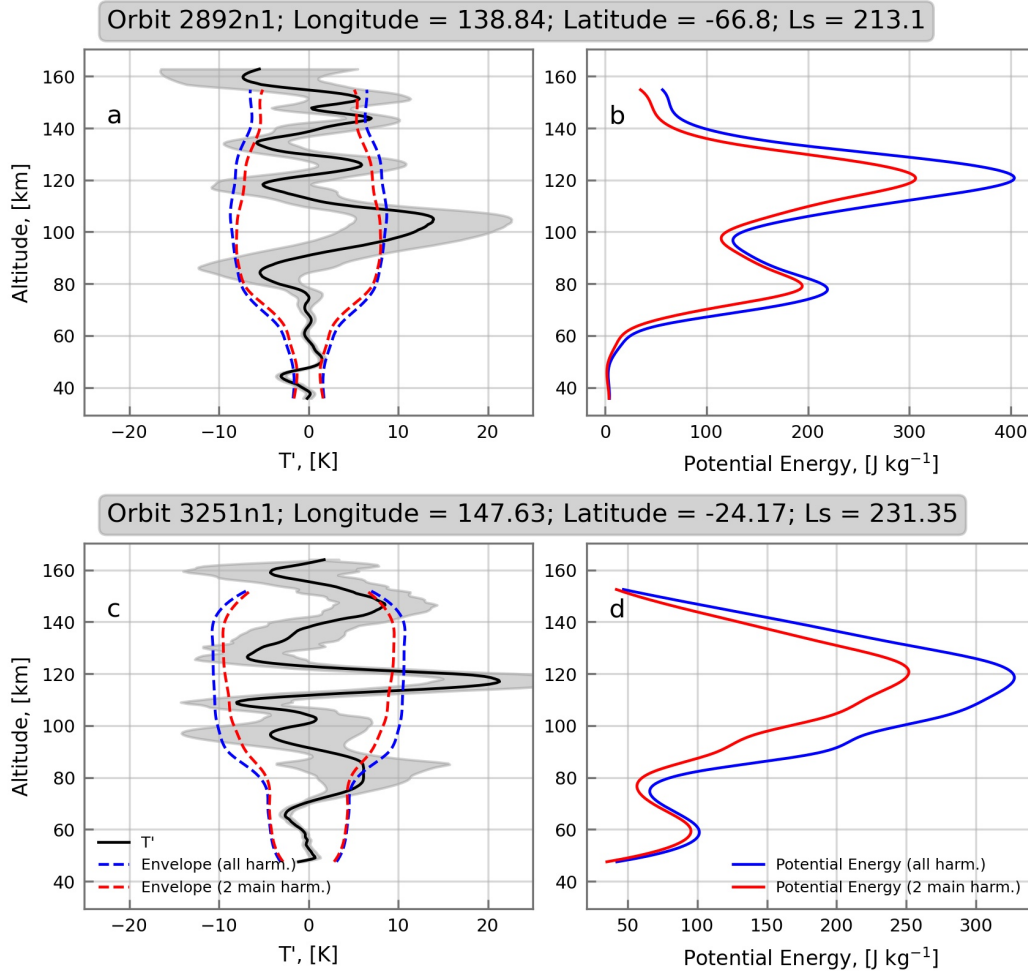


Figure 3. Wave activity $|T'|$ (left column) and potential energy (per unit mass, right column) for the same as in Figure 2 representative profiles. Dashed blue lines indicate quantities calculated for the entire spectrum, dashed red lines are for accounting two longest harmonics. Shaded areas denote observational errors.

293 be estimated (e.g., Ern et al., 2004, sect. 4):

$$F_{k,m} = \frac{1}{2} \rho_0 \frac{k_h}{m} \left(\frac{g}{N} \right)^2 \left(\frac{|T'_{k,m}|}{\bar{T}} \right)^2, \quad (6)$$

294 where k_h and m are the horizontal and vertical wavenumbers, correspondingly, and $|T'_{k,m}|$
 295 is the amplitude. The latter two are found from the Fourier decomposition, whereas k_h
 296 cannot be derived from our measurements.

297 The total flux F is the sum of contributions of individual harmonics $F = \sum_m F_{k,m}$.
 298 Since the horizontal wavenumber k_h cannot be obtained from the measurements, it, there-
 299 fore, serves as a scaling factor for the derived profiles of F and momentum forcing (5).
 300 The densest atmospheric footprint at a target point in occultation geometry is 400-500
 301 km horizontally, depending on the height. This constrains the upper limit for unresolved
 302 wavelengths. In our calculations, we assumed a representative horizontal wavelength $\lambda_h =$
 303 $2\pi/k_h = 300$ km, the value typically used in numerical general circulation models (Yiğit

304

et al., 2018), and allowing for more direct comparison with simulations. The results for two representative profiles, same as in Figures 2 and 3, are given in Figure 4. To demon-

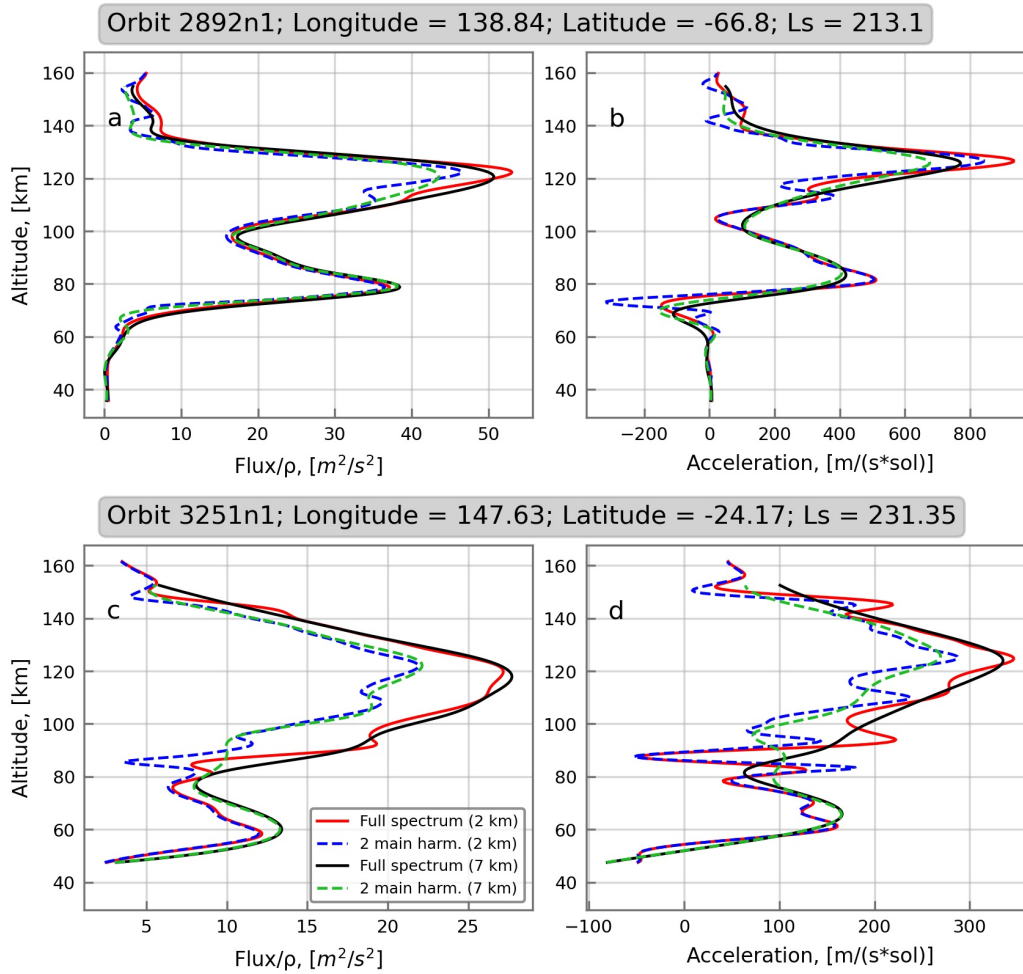


Figure 4. Absolute momentum flux (per unit mass) and the momentum forcing for two representative profiles (orbits 2892n1 and 3251n1, upper and lower rows, correspondingly). The legend describes the profiles calculated using the full spectrum and only two major harmonics along with sliding interval steps 2 and 7 km.

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strate the sensitivity of the calculations to the used parameters of the technique, we plotted with different colors the profiles of momentum fluxes (per unit mass) F/ρ_0 and GW momentum deposition, i.e., wave drag a obtained from the full spectrum and taking account of only two major harmonics. In addition, the results are shown for the interval shifts 2 and 7 km. It is immediately seen that these details play little role, and the calculations of fluxes and wave drag are very robust when the measured temperature profile is dominated by large-scale features (Figure 4, the upper row). It is different for profiles containing smaller vertical-scale disturbances (Figure 4, the lower row): their neglect leads to an underestimation of the fluxes and wave drag, and the smaller vertical shifts reveal finer structure associated with dissipation of individual spectral harmonics.

317 **4 Results and Discussions**

318 **4.1 Case Study**

319 Spectral analysis of the obtained set of profiles (described in the next subsection)
 320 has demonstrated greater contribution of larger-scale disturbances in all cases. However,
 321 each individual profile was unique. Two examples with and without small vertical-scale
 322 components have been presented above. We next consider a case with a relatively broad
 323 spectrum of wave-like perturbations with large amplitudes (about twice as large as those
 324 in orbit 3251n1). The retrieved temperature for the orbit 4926n1 along with the fitted
 325 background profile are plotted in Figure 5a. The envelope in Figure 5b clearly shows that
 326 the amplitude gradually ceases its exponential growth with height and becomes nearly
 constant above ~ 110 km. The reason for this so-called wave “saturation” can be seen

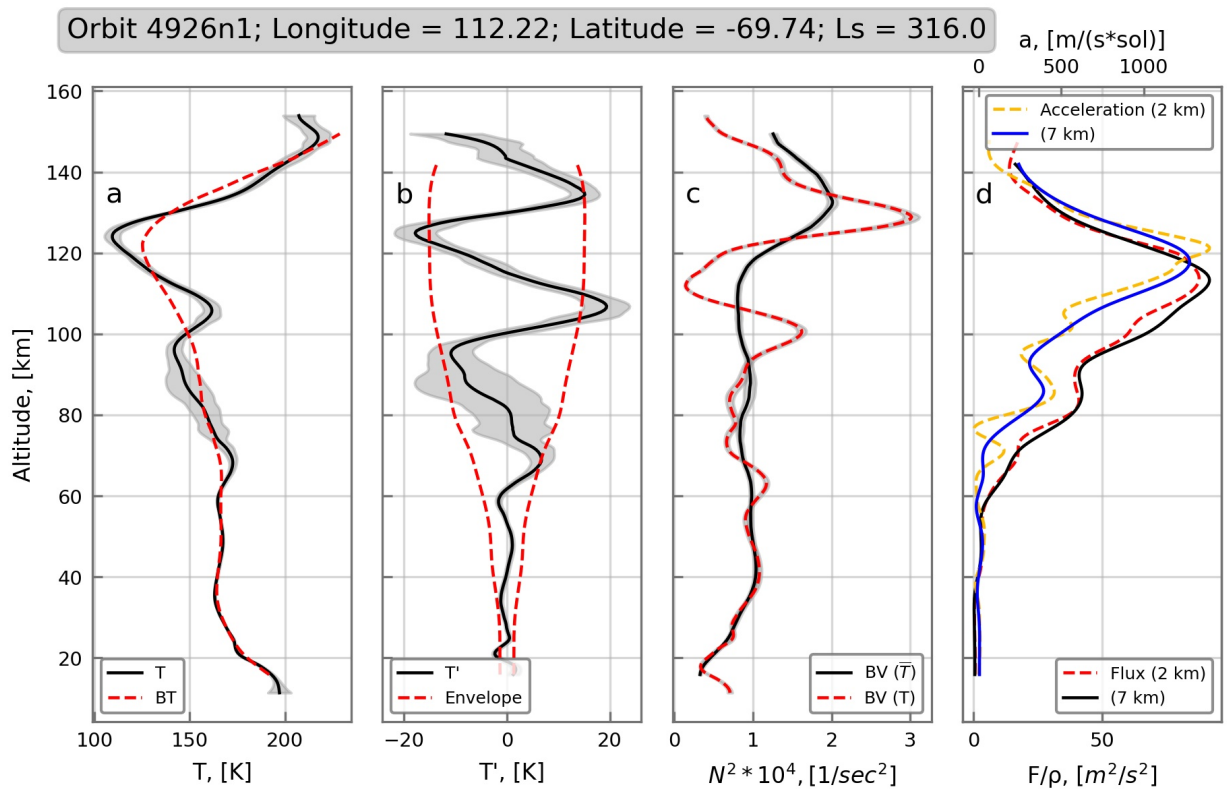


Figure 5. Vertical profiles for the orbit 4926n1. a) The measured (solid black) and fitted mean temperature (red dashed); b) wave temperature disturbance (solid black) and envelope (red dashed); c) Brunt-Väisälä frequency calculated for the mean (black) and net temperature (red dashed); d) momentum flux calculated using 2- and 7-km sliding window shifts (bottom axis, red dashed and solid black lines, correspondingly, and mean flow acceleration (“wave drag”, upper axis, yellow dashed and solid blue for the 2- and 7-km steps, respectively). Shading denote observational uncertainties.

327 from the behavior of the squared Brunt-Väisälä frequency $N^2(z)$ (Figure 5c, black). N^2
 328 calculated from the background profiles (Figure 5c) remains relatively constant with height
 329 (up to about 120 km) suggesting convective stability of the mean state. N^2 from the orig-
 330 inal profiles (see Figure 5c, red-dashed) shows large swings associated with temperature
 331

332 disturbances. Near 110 km, N^2 drops almost to zero as the result of the temperature
 333 gradient (associated with a large amplitude of the disturbances) approaching the adi-
 334 abatic lapse rate. Enhanced wave dissipation due to a combination of physical processes
 335 (Yiğit et al., 2018) in the vicinity of the convective instability severely limits the GW
 336 amplitude, leading to the decrease of the momentum flux above this altitude and peak-
 337 ing of the mean flow acceleration (Figure 5d) at almost $2000 \text{ m s}^{-1} \text{ sol}^{-1}$. In the ana-
 338 lyzed data set, such large numbers are not common and occur only occasionally. Appli-
 339 cation of a smaller vertical shift of sliding intervals shows finer structure of the GW mo-
 340 mentum flux and drag, but do not significantly modify the magnitudes.

341 4.2 Spatial Distribution of Gravity Wave Activity

342 In this section, we use the data obtained by the ACS instrument in MY34, at so-
 343 lar longitudes from $L_s = 164^\circ$ to 354° . The data set contains altogether 144 occulta-
 344 tion profiles: 84 in the northern hemisphere and 60 in the southern one. The latitude-
 345 solar longitude coverage is shown in Figure 6 with red and blue dots representing morn-
 346 ning and evening occultation measurements, correspondingly. The longitudinal orbit cov-
 erage was fairly uniform, and is not discussed here.

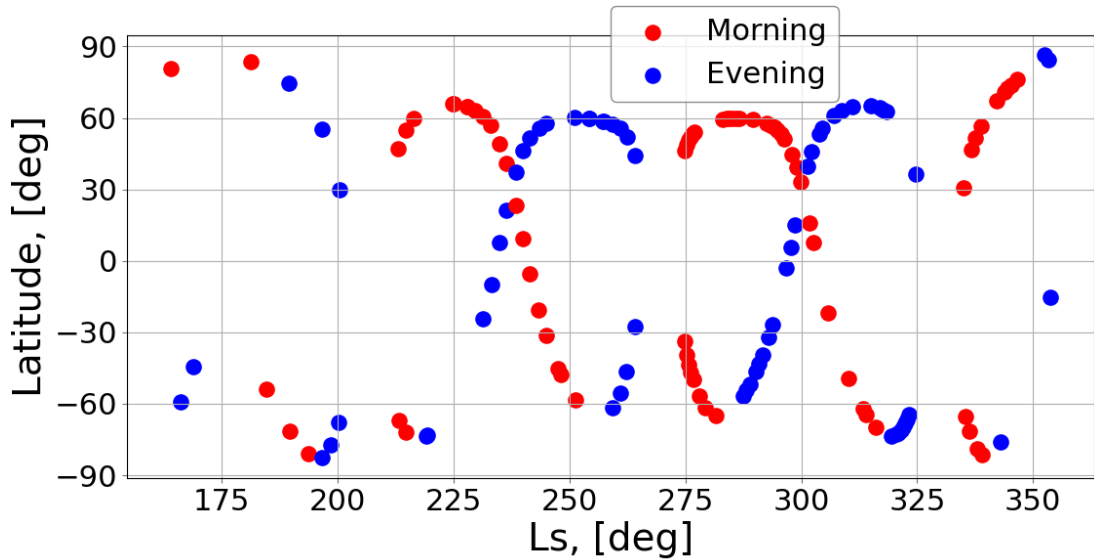


Figure 6. Latitude-solar longitude (L_s) distribution of the ACS MIR occultation profiles used in this study. Morning and evening measurements are shown in red and blue, correspondingly.

347

348 A significant portion of observations were made during the global dust storm of MY34,
 349 which started between $L_s = 185^\circ$ and 190° , attained its maximum around $L_s = 220^\circ$,
 350 and gradually decreased until $L_s \approx 290^\circ$. A regional storm occurred at the end of MY34
 351 between approximately $L_s = 325^\circ$ and 345° . Figure 7 presents latitude-altitude dis-
 352 tribution of the derived GW parameters averaged over the entire period of observations
 353 depicted in Figure 6. It shows that the mean amplitude of GW-induced temperature fluc-
 354 tuations ($|T'|$, Figure 7a) grows with height reaching up to $\sim 10 \text{ K}$ near the top of the
 355 domain. At higher altitudes (170-220 km), the in situ measurements with Neutral Gas
 356 and Ion Mass Spectrometer (NGIMS) on board MAVEN revealed even larger GW mag-
 357 nitudes over the same time (Leelavathi et al., 2020; Yiğit et al., 2021). The latitudinal

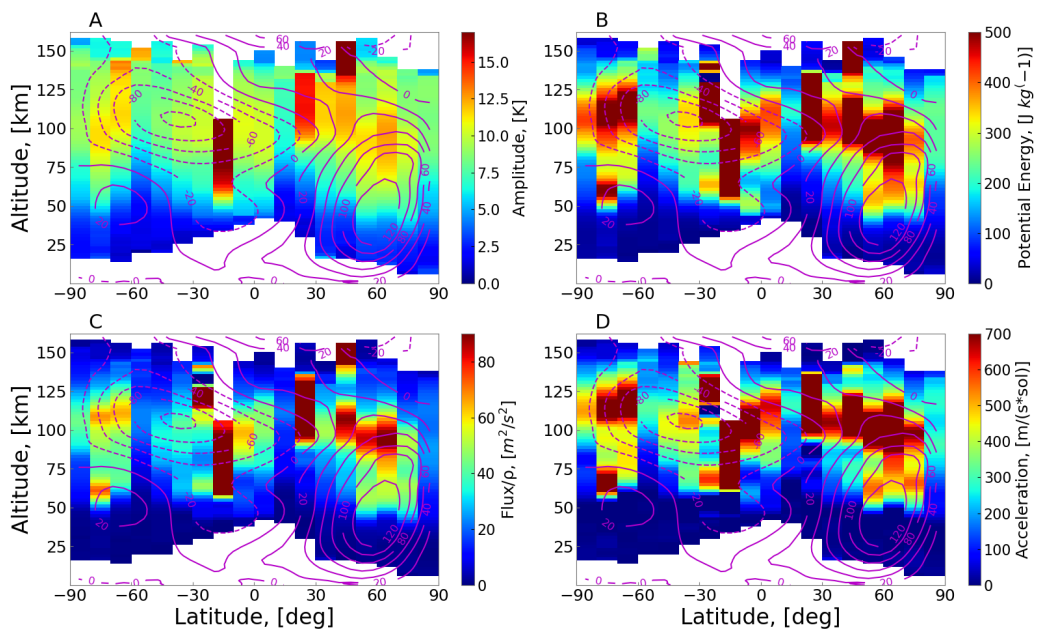


Figure 7. Latitude-altitude cross-sections of the retrieved GW a) amplitudes (in K), b) potential energy (per unit mass), c) vertical fluxes of absolute horizontal momentum (per unit mass) and d) associated momentum forcing (GW drag). The size of the employed latitudinal bins is 10° . Contour lines present the zonal wind (in m s^{-1}) simulated with the MAOAM MGCM for MY34 (<https://mars.mipt.ru/data.php>) and averaged over the same as in Figure 7 period of observations.

358 structure of the GW activity in the mesosphere and lower thermosphere is not uniform.
 359 For comparison, we overplotted the zonal wind simulated with the Max Planck Institute
 360 (MAOAM) MGCM <https://mars.mipt.ru/> for MY34 and averaged over the same in-
 361 terval of L_s as in the observations. The wind distribution varied during this time from
 362 the equinoctial to solstitial and back to the equinoctial types. The result reflects the largest
 363 contribution of the prograde and retrograde jets during the perihelion solstice. It is seen
 364 that the regions with large wave amplitudes encircle the upper edges of two midlatitude
 365 jets. This is the result of intensive filtering of individual harmonics by strong background
 366 winds. For the wave potential energy, which is a quadratic function of wave amplitudes,
 367 this pattern is even more obvious (Figure 7b).

368 Figure 7c shows that GW momentum fluxes reach local maxima near the mesopause
 369 (100-125 km) giving evidence of very intensive wave breakdown/dissipation in this re-
 370 gion. The peaks of the associated momentum deposition approximately coincide (Figure
 371 7d). They too wrap around the edges of the jets in the middle atmosphere. It is note-
 372 worthy that such distribution of the GW drag is very similar to that predicted by a Mar-
 373 tian GCM (Medvedev et al., 2011, Figures 3 and 7) for the solstice and equinox, respec-
 374 tively, and represents the first (to the best of our knowledge) observational validation
 375 of the model predictions. The magnitudes of the GW drag, although defined up to the
 376 constant k_h , agree with the simulations (using a similar k_h) as well.

4.3 Amplitude Dependence on Mean Temperature and Brunt-Väisälä Frequency

In situ measurements with NGIMS on board MAVEN showed a clear anti-correlation between relative density fluctuations in the upper thermosphere and the ambient temperature (Yigit et al., 2015; England et al., 2017; Terada et al., 2017; Vals et al., 2019). It was linked to convective instability as a dominant mechanism that limits growth of GW amplitudes with height (wave saturation). The arguments were based on the relation for a single harmonic (e.g., Fritts et al., 1988, Eq. 6)

$$\frac{|T'|}{\bar{T}} = \frac{|u'|}{|c - \bar{u}|} \frac{N^2}{mg}, \quad (7)$$

where $|u'|$ is the amplitude of fluctuations of horizontal velocity in the wave, c is its horizontal phase velocity and \bar{u} is the background wind. When $|u'|$ approaches $|c - \bar{u}|$, increasing dissipation limits $|u'|$ thus that the ratio $|u'|/|c - \bar{u}|$ becomes constant. The linear convective instability threshold demands a unit ratio, however observations suggested a ratio of 0.7 (Fritts et al., 1988, Eq. 2), and the theoretical consideration of the non-linear diffusion mechanism yielded $1/\sqrt{2} \approx 0.707$ (Medvedev & Klaassen, 2000, Sect. 7). Regardless of the precise number, (7) establishes proportionality between the amplitude of relative temperature/density perturbations and squared mean Brunt-Väisälä frequency under the saturation condition. Near the exobase, where the majority of NGIMS/MAVEN observations were taken, the vertical gradient $d\bar{T}/dz$ is small and can be neglected in (4), thus giving the inverse proportionality of relative perturbation amplitudes and \bar{T} .

ACS/TGO occultation data cover altitudes below the exobase, where $d\bar{T}/dz$ can no longer be neglected. Therefore, we plotted in Figure 8a the amplitudes of relative temperature perturbations for all orbits as functions of N^2 . It is seen that red and blue dots corresponding to morning and evening measurements show no clear dependence on N^2 at all altitudes. To explore this further, we over-plotted the linear regression of the form $|T'|/\bar{T} = \alpha N^2 + \beta$ and put the values of α and β in the legend. The coefficients α are far less than those expected from (7), i.e., several tens or hundreds, depending on the characteristic vertical wavenumber m . The distinction between morning and evening amplitudes is also insignificant, except above 100 km, where morning values are slightly larger.

Figure 8b presents the dependencies of amplitudes of relative temperature disturbances as functions of the mean temperature. They are nearly uniform. Although regression coefficients show a weak negative trends at all altitudes, their magnitudes are much smaller than to those observed previously (of the order of 0.5 to 1) near the exobase (Yigit et al., 2015; England et al., 2017; Terada et al., 2017; Vals et al., 2019). A similar lack of correlation between GW amplitudes and atmospheric temperature was found from TGO aerobraking measurements at altitudes between 100 and 130 km (Jesch et al., 2019, Figure 12). The atmospheric drag data were collected between $L_s = 332^\circ$ of MY33 and $L_s = 132^\circ$ of MY34. The ACS observations after the aerobraking cover the dusty second half of MY34. Thus, the absence of correlation between GW amplitudes and the background temperature in the lower thermosphere appear to be independent of the season and dust conditions. In the upper thermosphere, (Leelavathi et al., 2020, Figure 10d) found a positive correlation during the same second half of MY34, instead of a clear negative correlation over the first (“non-dusty”) half of the year. Our results in the adjacent region (around 140 km) show no visible change, neither strong negative trend previously found in the MAVEN/NGIMS observations, nor indication of a positive trend. This means that convective instability may not be the main mechanism responsible for damping GWs in the thermosphere, at least during dust storms.

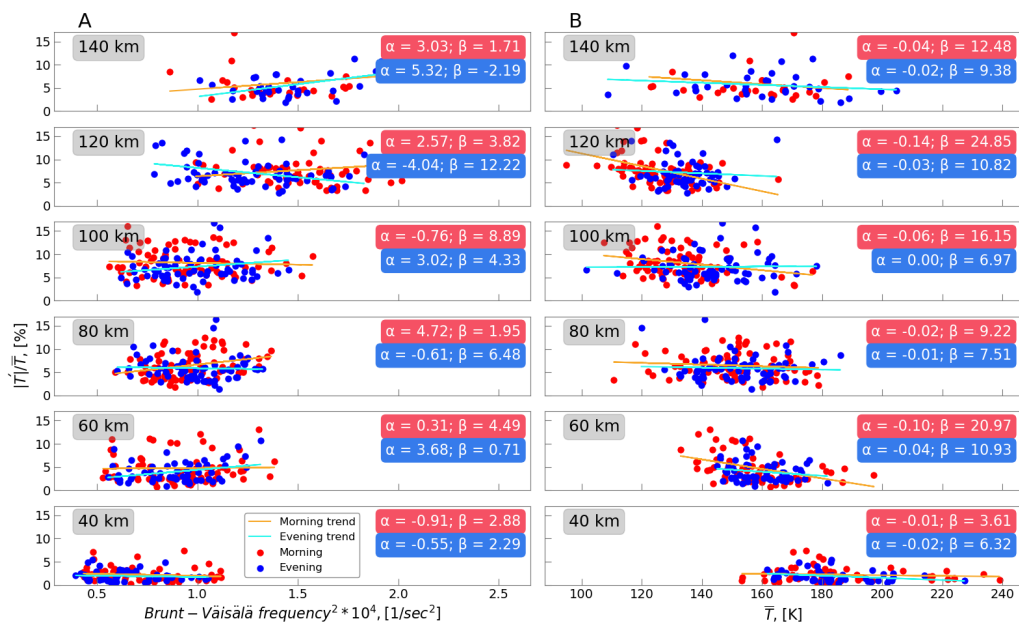


Figure 8. Amplitudes of relative temperature disturbances as functions of the squared Brunt-Väisälä frequency (a) and mean temperature (b) at different heights. Red and blue dots are for the morning and evening measurements, correspondingly. Linear regressions of the form a) $|T'|/\bar{T} = \alpha N^2 + \beta$ and b) $|T'|/\bar{T} = \alpha \bar{T} + \beta$ are shown with thin solid lines, and the values of the respective coefficients are given in the legends.

423 **5 Summary and Conclusions**

424 We have presented the results of gravity wave (GW) retrievals obtained from the
 425 Atmospheric Chemistry Suite instrument on board the ExoMars Trace Gas Orbiter (ACS/TGO),
 426 which observed solar occultation spectra. GW disturbances are derived from the verti-
 427 cal temperature profiles retrieved from one of the three instrument channels - the mid-
 428 infrared ACS/MIR. The uniqueness of the data is that they continuously cover a broad
 429 range of altitudes from the Martian troposphere to the thermosphere (20-160 km) and
 430 have a relatively high (0.5 to 2.5 km) vertical resolution.

431 Several techniques of separating GW components from the background tempera-
 432 ture have been studied. The sliding-window least square polynomial fitting method have
 433 demonstrated to be the most robust and effective. The procedure was applied to 144 mea-
 434 surements collected over the second half of MY34 to derive vertical profiles of GW dis-
 435 turbances as well as further wave characteristics: amplitude, wave potential energy, ab-
 436 solute vertical flux of horizontal momentum and absolute momentum forcing produced
 437 by breaking/dissipating GWs (“GW drag”). The main results are listed below.

- 438 1. Amplitudes of GW-induced temperature fluctuations, generally, grow with height,
 439 while breaking/saturation processes often limit the wave amplitude growth at higher
 440 altitudes. Based on a half-year average, wave amplitudes are around 8–14 K near
 441 the mesopause, and often exceed these values in individual profiles.

- 442 2. The mesopause (100-120 km) is the region of the strongest GW breaking/dissipation,
 443 which is evidenced by a local maximum of momentum fluxes and their vertical di-
 444 vergence, i.e., GW drag. Similarly, a large GW drag of hundreds of $\text{m s}^{-1} \text{sol}^{-1}$
 445 in the mesopause region has been demonstrated by MGCMs (e.g., Yiğit et al., 2018).
 446 3. The spatial (altitude-latitude) distribution of the wave drag also agrees well with
 447 modeling results (e.g., Medvedev et al., 2011). This is the first direct observational
 448 validation of model predictions.
 449 4. We did not find positive correlation between amplitudes of relative temperature
 450 perturbations and the Brunt-Väisälä frequency at all heights. This correlation is
 451 a more general formulation of the anti-correlation found near the exobase (Yiğit
 452 et al., 2015; England et al., 2017; Terada et al., 2017; Vals et al., 2019) that ac-
 453 counts for vertically varying mean temperature.

454 The presented GW activity retrievals extending from the middle troposphere to
 455 the thermosphere, as derived from the ExoMars data, highlight the role of atmospheric
 456 gravity waves as a whole atmosphere phenomenon on Mars. Mars' thin and windy at-
 457 mosphere favors strong gravity wave generation, thus an accurate characterization of grav-
 458 ity waves is absolutely essential for a better understanding of the Martian climate (Yiğit
 459 & Medvedev, 2019).

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 465 able from ESA's Planetary Science Archive at <https://archives.esac.esa.int/psa/>
 466 #!Table%20View/ACS=instrument. The data for latitude-altitude cross-sections of GW
 467 characteristics such as amplitudes, potential energy, fluxes, accelerations and zonal wind
 468 are available at the Mendeley database (Starichenko, 2021).

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