

# Climatology of Gravity Wave Activity From Two Martian Years of ACS/TGO Observations

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## Key Points:

- GWs' parameters for two Martian years from temperature profiles of ACS spectrometers have been obtained
- The seasonally varying distributions of the GWs' drag agree well with predictions of general circulation models
- Observations reveal high-latitude enhancement of GW activity in both hemispheres during the global dust storm

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**Abstract**

We report the gravity wave (GW) statistics accumulated over two Martian years from the second half of Martian Year 34 (MY34) to the middle of MY36 (May 2018 - February 2022). The observations were performed by the middle- and near-infrared (MIR and NIR, respectively) spectrometers of Atmospheric Chemistry Suite (ACS) on board ExoMars Trace Gas Orbiter (TGO). Temperature profiles obtained independently of both channels during simultaneous measurements show a good agreement, thus providing verification and additional confidence in the data. GW parameters such as temperature fluctuations, potential energy per unit mass, and wave drag are retrieved at altitudes up to 160 km from the MIR channel and up to 100 km from the NIR channel. We present seasonal, intraday and latitude distributions of the wave potential energy and drag, serving to represent the wave activity and impact on the dynamics. A comparison of data obtained during the global dust storm (GDS) of MY34 with the corresponding period of MY35 without a storm reveals a reduction of GW activity in mid-latitudes in agreement with previous observations, and enhancement in the polar regions of both hemispheres, which was predicted by theoretical studies using simulations with a high-resolution circulation model. Seasonal variations of the derived GW activity can be linked to changes in the solar tide.

**Plain Language Summary**

Gravity waves with horizontal scales of tens and hundreds kilometers are present in atmospheres of all planets. They play an important role in the dynamics and thermodynamics of the middle and upper atmosphere of Mars through transporting the energy and momentum from the lower to the upper layers of atmosphere. The knowledge of their spatio-temporal variability is required for characterization of the atmospheric state and circulation. Two independent channels of the Atmospheric Chemistry Suite instrument on board the ExoMars Trace Gas Orbiter provided thousands of temperature profiles over two Martian years, which were used for retrieving wave characteristics. We present the climatology of gravity wave activity in the form of seasonal, latitudinal and local time distributions. They reveal a strong response of the gravity wave field to the global dust storm occurred in 2018. The derived distributions of the deceleration of the large-scale flow imposed by gravity waves can constrain global circulation models and improve their predictive capabilities.

**1 Introduction**

Gravity (or buoyancy) waves (GWs) originating from the balance of gravity and buoyancy forces are ubiquitous in all convectively stable atmospheres. They have been extensively studied in the terrestrial atmosphere, where their important role in the dynamics and vertical coupling of atmospheric layers has been recognized (e.g., see reviews by Fritts & Alexander, 2003; Yiğit & Medvedev, 2015). First GW-like signature in the atmosphere of Mars was detected in entry measurements of Viking 2 (Seiff & Kirk, 1976). Since then, GWs have been observed on Venus (R. E. Young et al., 1987), Jupiter (L. A. Young et al., 1997), Titan (Hinson & Tyler, 1983), Saturn (Brown et al., 2022) and other planets (see a recent review on GWs in planetary atmospheres by Medvedev & Yiğit, 2019). Recently, various GW-related phenomena in the Martian atmosphere and their impact on the whole atmosphere of Mars system have been reviewed (Yiğit, 2023).

In situ accelerometer measurements in the thermosphere of Mars performed during spacecraft aerobreaking demonstrated large amplitudes of GW-induced density disturbances and the associated wave drag (Keating et al., 1998; Creasey et al., 2006a; Fritts et al., 2006). The omnipresence of GWs in the lower Martian atmosphere and first characterization of the GW field were revealed using remote sensing techniques (Hinson et al., 1999; Creasey et al., 2006b; Wright, 2012; Nakagawa et al., 2020; Heavens et al., 2020).

They showed a varying spatio-temporal structure of the GW field and evidence for a limitation of amplitude growth with height, which is an indication of wave momentum transfer to the larger-scale flow (Ando et al., 2012).

GWs exist at all atmospheric heights. In situ measurements using the Neutral Gas and Ion Mass Spectrometer (NGIMS) on board the Mars Atmosphere and Evolution Mission (MAVEN) orbiter delivered a large body of GW statistics in the upper thermosphere (Yigit et al., 2015; England et al., 2017; Terada et al., 2017; Leelavathi et al., 2020; Rao et al., 2021). In particular, the observations found an enhancement of GW activity during the global dust storm (GDS) that occurred in June 2018 (Martian year 34, MY34) (Leelavathi et al., 2020; Yigit, Medvedev, Benna, & Jakosky, 2021). Conversely, observations with the Mars Climate Sounder instrument on board Mars Reconnaissance Orbiter have shown a reduction of such activity in the lower atmosphere (Heavens et al., 2020). Another existing controversy concerns the inverse relation between the amplitudes of GWs and the background temperature in the upper thermosphere. A number of studies attributed it to convective instability that limits wave amplitudes causing the so-called “saturation” (England et al., 2017; Terada et al., 2017; Vals et al., 2019), while Yigit, Medvedev, and Hartogh (2021) argued that the inverse relation occurs because a colder background air reduces the scale height  $H$ , thus facilitating the exponential growth of amplitude, which is proportional to  $1/2H$ . Molecular diffusion and thermal conduction also exponentially grow with height in response to density decrease. They eventually exceed all other damping mechanisms in the thermosphere, and thereby significantly limits the wave growth.

GWs are generated in the lower atmosphere by a variety of mechanisms that vertically displace air parcels, e.g., flow over topography, convection, weather instabilities, etc. While propagating upward, they are partially filtered out by the background mean wind. Amplitudes of the surviving harmonics grow with height. Ultimately, the harmonics reach altitudes, where they are dissipated owing to a combination of nonlinear interactions, molecular diffusion, and thermal conduction, and deposit their momentum and energy to the ambient flow (Yigit et al., 2008). This gravity wave-mean flow interaction produces acceleration/deceleration of the large-scale circulation, which is often called “GW drag”. Its dynamical importance in the middle and upper atmosphere of Mars has been demonstrated with general circulation models (GCMs) where the effects of small-scale GWs are either parameterized (Medvedev et al., 2011a, 2011b; Gilli et al., 2020; Yigit et al., 2018; Roeten et al., 2022), or explicitly resolved (Kuroda et al., 2015, 2016, 2019). These modeling studies have to be validated with observations, and many employed parameters constrained. Therefore, an observational characterization of the GW field and its spatio-temporal variation at all heights is of great importance.

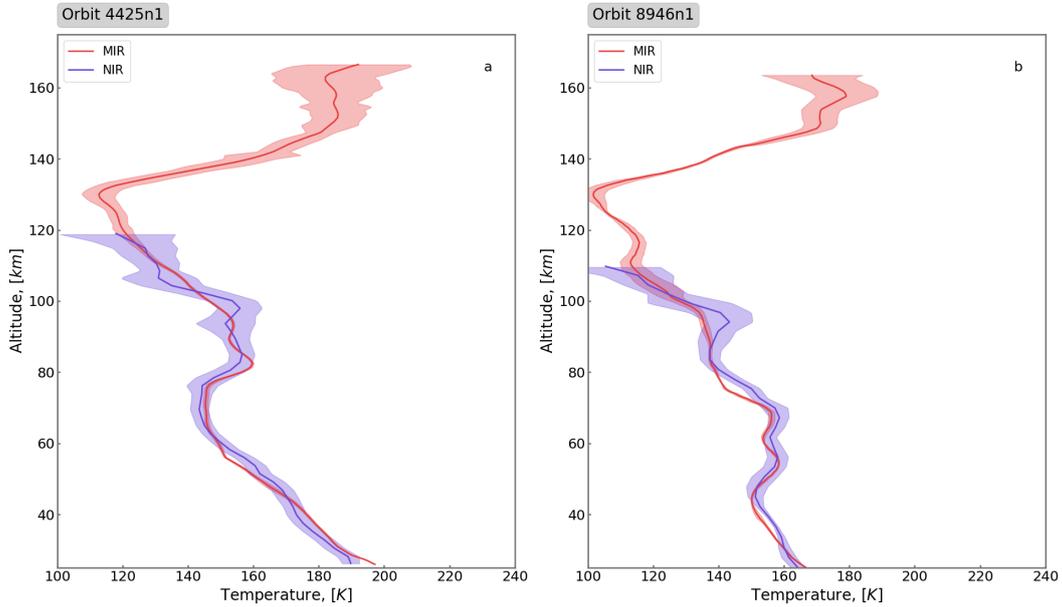
The MIR spectrometer of Atmospheric Chemistry Suite (ACS) experiment on board the Trace Gas Orbiter (TGO) (Korablev et al., 2018) allows filling this gap in the knowledge of atmospheric variability by measuring vertical profiles of density and temperature between 20 and 160–180 km in the Martian atmosphere. The algorithm of retrieving GW profiles and their characteristics along with the first results of its application have been presented in detail in the work by Starichenko et al. (2021). The database of observations has been significantly extended since then. In this work, we present the results on GW activity obtained over the second half of the Martian year 34, the whole MY35 and the first half of the MY36. In addition, we analyzed the profiles measured by another ACS channel - near-IR (NIR). Although they cover altitudes only up to  $\sim 100$  km, their number (several thousand) adds significantly to the overall statistics.

The structure of this paper is the following. In Section 2, observations and temperature retrieval procedures from ACS are outlined. The methods of deriving the GW characteristics are described in Section 3. Section 4 presents the data coverage. Latitude-altitude distributions for four Martian seasons are given in Section 5, the impact of the

119 global dust storm on the GW activity is discussed in Section 6, and local time variations  
 120 are shown in Section 7. Conclusions are presented in Section 8.

## 121 2 Observations and Temperature Retrievals

122 The Atmospheric Chemistry Suite (ACS) is a part of the Trace Gas Orbiter (TGO),  
 123 which represents the ESA-Roscosmos ExoMars 2016 collaborative mission. The instru-  
 124 ment consists of three infrared channels (Korablev et al., 2018): near-IR (NIR, 0.73-1.6  
 125  $\mu\text{m}$ ), middle-IR (MIR, 2.3-4.2  $\mu\text{m}$ ), and thermal-IR (TIRVIM, 1.7-17  $\mu\text{m}$ ). In this work,  
 126 we use the data obtained from the MIR and NIR spectrometric channels operating in  
 127 the solar occultation mode since April 2018. ACS-MIR is a cross-dispersion echelle spec-  
 128 trometer that allows for retrieving temperature and density vertical profiles in the strong  
 129 2.7  $\mu\text{m}$   $\text{CO}_2$  absorption band covering the broad altitude range of 20–180 km (Belyaev  
 130 et al., 2021, 2022). ACS-NIR, an echelle spectrometer combined with an acousto-optic  
 131 tunable filter, measures the atmospheric structure in 1.43  $\mu\text{m}$  and 1.57  $\mu\text{m}$   $\text{CO}_2$  bands  
 132 at altitudes from 10 to 100 km (Fedorova et al., 2020, 2023). Both ACS MIR and NIR  
 133 channels possess a high resolving power exceeding  $\sim 25000$ , the signal-to-noise ratio of  
 134 more than 1000, and sound the atmosphere with the vertical resolution of  $\sim 1$  km. Dur-  
 135 ing simultaneous occultations, the lines of sight (LOS) of both instruments target iden-  
 136 tical tangent points. This provides a confidential cross-validation between the retrieved  
 137 atmospheric profiles. The altitude of the tangent points is determined as the closest dis-  
 138 tance between the instrument’s LOS and the areoid surface of the planet. The atmospheric  
 139 transmission spectrum at each tangential altitude is obtained as a ratio of the solar spec-  
 140 trum transmitted through the atmosphere to the reference one measured at an altitude  
 141 where the absorption at the given  $\text{CO}_2$  band is negligible, that is 200 km for the MIR  
 142 case and 130 km for NIR.



**Figure 1.** Vertical profiles of temperature derived from simultaneous ACS-MIR (red line) and ACS-NIR (blue line) occultations. Shaded area denotes the uncertainty of the measurements. Examples are from orbits a) 4425n1 (20 Nov 2018 MY34,  $L_s=291.7^\circ$ ,  $\text{Lat}=39.2^\circ\text{S}$ ) and b) 8946n1 (25 Nov 2019, MY35,  $L_s=111.7^\circ$ ,  $\text{Lat}=60.9^\circ\text{N}$ ).

143 The procedures of retrieving density and temperature vertical profiles from the trans-  
 144 mission spectra of CO<sub>2</sub> are extensively described in the papers of Belyaev et al. (2021,  
 145 2022) for the MIR channel and of Fedorova et al. (2020, 2023) for the NIR channel. Pro-  
 146 files with estimated 1- $\sigma$  errors exceeding 20 K have been removed from consideration based  
 147 on expected wave amplitudes. Validation between simultaneously measured MIR and  
 148 NIR profiles demonstrates a good coincidence below 100 km with dispersion of 5-10 K  
 149 in more than 90% of occultations (Belyaev et al., 2022). When retrieving the GW pa-  
 150 rameters (see Section 3), the data were checked for reasonably smooth background tem-  
 151 perature profiles (suitable for extracting waves) and for adequate values of the poten-  
 152 tial energy ( $<1000 \text{ J kg}^{-1}$ ). Anomalous values could be found either at inflection points  
 153 of temperature profiles, such as around the mesopause, or near the top of the domain  
 154 (80-100 km) for the NIR data. Such cases account for about 10-15% of all occultations,  
 155 and they were excluded from our consideration.

156 In most of simultaneous observations, both the MIR and NIR individual profiles  
 157 (Figures 1, 2a, 2d) along with the retrieved GW parameters (Figures 2b, 2c) closely match  
 158 each other. Nevertheless, in some cases, the evaluated GW amplitudes and potential en-  
 159 ergy somewhat differ between two channels (see Figures 2d, 2e). The reason for that is  
 160 in different altitude domains used for retrieving the NIR and MIR background temper-  
 161 ature. Consequently, the maximum discrepancy occurs near the upper end (80-100 km)  
 162 of the NIR profiles (Figures 2d, 2e). Overall, statistics of occultations at the MIR 2.7  
 163  $\mu\text{m}$  CO<sub>2</sub> band are about 10 times less frequent than those at the NIR CO<sub>2</sub> spectra. Thus,  
 164 in the analyses to be presented, we complement the MIR profiles with those from NIR,  
 165 whenever measurements with MIR are not available.

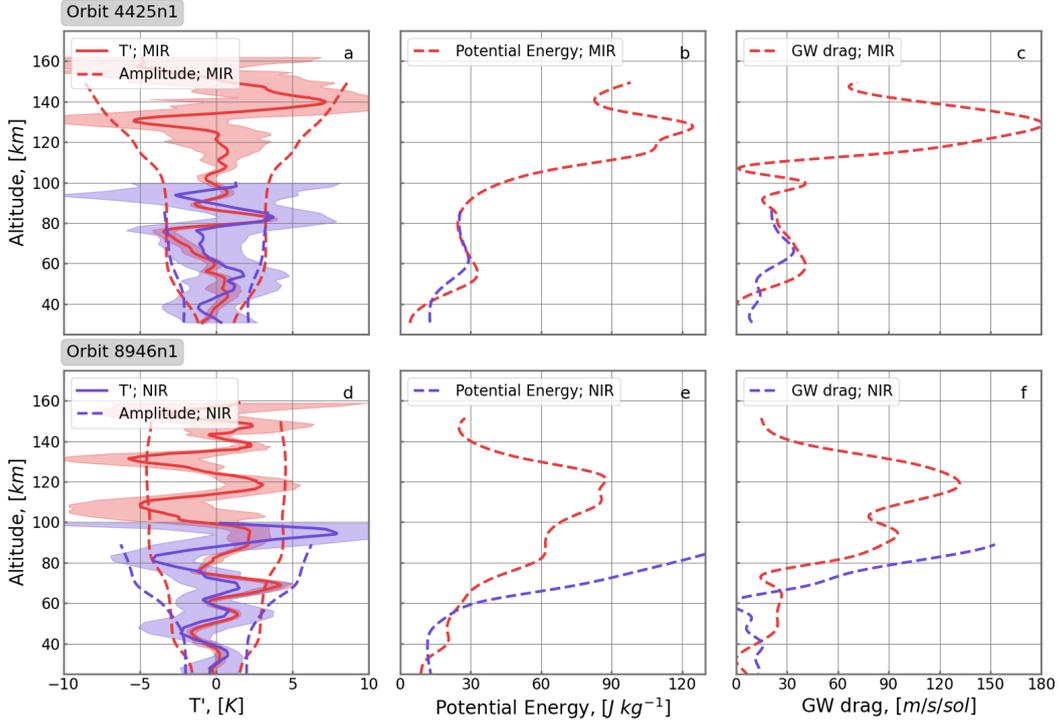
### 166 3 Gravity Wave Characteristics

167 Separation of the observed temperature profile  $T(z)$  into the background  $\bar{T}$  and the  
 168 GW-induced disturbance  $T' = T - \bar{T}$  is an ambiguous procedure, because no unique  
 169 partition exists. The wave component grossly depends on the definition of the mean tem-  
 170 perature  $\bar{T}$ . The routine used in this work was described in detail and extensively tested  
 171 in the paper by Starichenko et al. (2021). It was recently applied to retrievals of GWs  
 172 in the thermosphere of Saturn (Brown et al., 2022). The mean vertical profile is deter-  
 173 mined by fitting cubic polynomials within sliding windows of 60 km width, effectively  
 174 limiting the consideration to relatively short-scale GW harmonics with vertical wavelengths  
 175 smaller than 30 km. The windows are shifted first from the bottom up and then down-  
 176 ward with 7-km steps. Then, all the overlapping polynomials are averaged, and the fi-  
 177 nal profile is smoothed over by a moving average procedure. The uppermost and low-  
 178 est 4 km of each profile have to be dropped due to a spurious behavior of the fitted poly-  
 179 nomials at the edges, which cannot otherwise be averaged. After the mean and wave com-  
 180 ponents for each profile are derived, the Brunt-Väisälä frequency, wave amplitude, wave  
 181 potential energy, vertical flux of horizontal momentum and GW drag can be determined.  
 182 The Brunt-Väisälä frequency characterizes the convective stability of the atmosphere:

$$N^2 = \frac{g}{\bar{T}} \left( \frac{d\bar{T}}{dz} + \frac{g}{c_p} \right), \quad (1)$$

183 where  $g$  is the acceleration of gravity and  $c_p$  is the specific heat capacity at constant pres-  
 184 sure. If  $N^2$  approaches zero (or the temperature gradient approaches the dry adiabatic  
 185 lapse rate), the stability decreases. When  $N^2$  drops below zero, the atmosphere becomes  
 186 convectively unstable and no longer supports GW propagation. Thus, GW harmonics  
 187 experience strong dissipation and/or breaking in the regions of small or negative  $N^2$ .

188 Since GW harmonics usually propagate in wave packets, the observed instantaneous  
 189 peaks and troughs do not fully characterize the wave amplitude. The latter (or “wave  
 190 activity”) is better represented by the envelope for temperature disturbances  $|T'| = \sqrt{T'^2}$ .  
 191 It is calculated by performing the Fourier decomposition in each 60-km sliding window



**Figure 2.** Wave characteristics retrieved for two representative measurements shown in Figure 1: orbits 4425n2 (upper row) and 8946n1 (lower row). The red and blue colors distinguish the MIR and NIR data correspondingly. In the left column (a, d), the solid lines represent wave-induced distributions of temperature  $T'$ , dashed lines are for the wave packet envelopes  $|T'|$ , and shades indicate the errors. The middle column (b, e) compares the potential energy profiles, while the right column (c, f) presents the calculated GW drag.

192 and summation of contributions from all harmonics. The other useful measure of the wave  
 193 field is the potential energy (per unit mass)  $E_p$ :

$$E_p = \frac{1}{2} \left( \frac{g}{N} \right)^2 \left( \frac{|T'|}{T} \right)^2. \quad (2)$$

194 While  $|T'|$  and  $E_p$  describe spatio-temporal distributions of the wave field itself,  
 195 the vertical flux of horizontal momentum (per unit mass) quantifies the momentum trans-  
 196 port by propagating harmonics. It is defined as  $\mathbf{F} = (F_x, F_y, 0) = (\overline{u'w'}, \overline{v'w'}, 0)$ , where  
 197  $u'$ ,  $v'$  and  $w'$  are the components of wave-induced disturbances of velocity in the zonal,  
 198 meridional and vertical directions, correspondingly. The directional part of the flux can-  
 199 not be inferred from a single vertical profile, however the absolute momentum flux  $F =$   
 200  $\sqrt{F_x^2 + F_y^2}$  (e.g., Ern et al., 2004, sect. 4) can be estimated:

$$F = \sum_{k_h, m} \frac{1}{2} \frac{k_h}{m} \left( \frac{g}{N} \right)^2 \left( \frac{|T'_{k,m}|}{T} \right)^2. \quad (3)$$

201 The variables  $k_h$  and  $m$  in (3) are the horizontal and vertical wavenumbers,  $|T'_{k,m}|$  is the  
 202 amplitude of the corresponding harmonic, and the summation over all  $k_h$  and  $m$  is done.  
 203 While amplitudes and vertical wavenumbers of particular harmonics are determined by

204 the Fourier analysis, the horizontal wavenumber cannot be derived from a single verti-  
 205 cal profile. Instead, it serves as a scaling factor, the value of which has to be assigned.  
 206 In our analysis, the characteristic  $k_h^*$  was chosen considering that the densest atmospheric  
 207 footprint at a target point in occultation experiments is  $\sim 400\text{-}500$  km horizontally, and  
 208 harmonics with longer wavelengths remain unresolved. In our calculations, we assumed  
 209 the horizontal wavelength  $\lambda_h^* = 2\pi/k_h^* = 300$  km, which also agrees with that com-  
 210 monly used in GW parameterizations implemented into numerical GCMs (e.g., Yiğit et  
 211 al., 2018). The momentum lost by a given breaking/dissipating harmonic is transferred  
 212 to the background flow, thus producing its acceleration or deceleration, or imposing the  
 213 so-called GW “drag”

$$a_h = \frac{1}{\bar{\rho}} \frac{d\bar{\rho}F}{dz}. \quad (4)$$

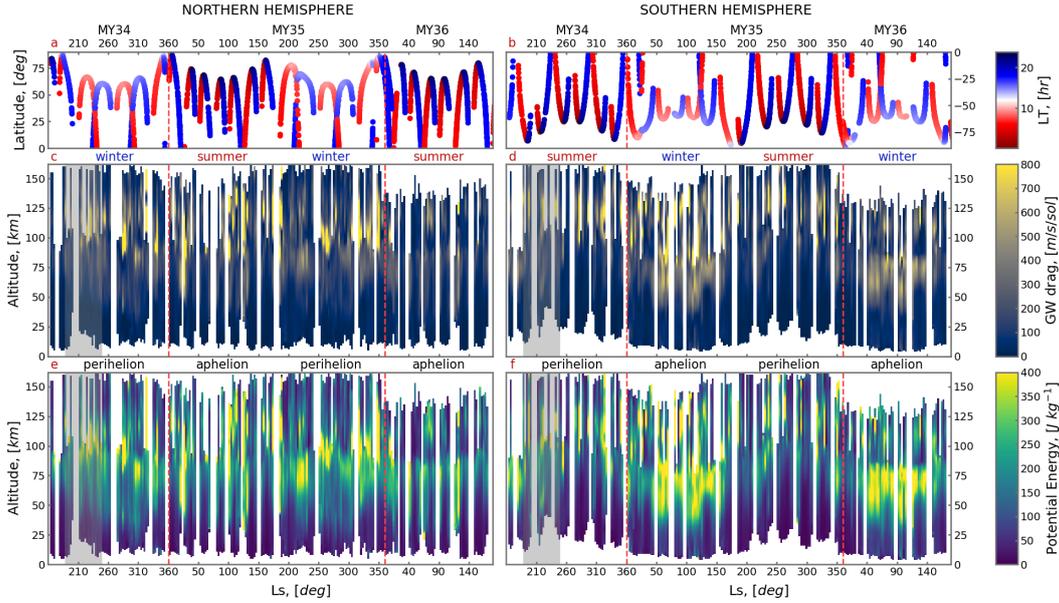
214 In (4),  $\bar{\rho}$  denotes the mean density; the subscript  $h$  indicates that the acceleration oc-  
 215 curs in the horizontal direction. Since the precise direction of  $F$  is not known, only ab-  
 216 solute values of  $a_h$  can be determined from the observations. In the lower parts of the  
 217 profiles, the amplitudes and the momentum flux  $F$  are small. This can lead to big dif-  
 218 ferences of small values in the calculations of  $a_h$ , which can result in negative values. In  
 219 order to avoid this non-physical behavior, we applied to  $F$  the iterative procedure de-  
 220 scribed in the paper of Brown et al. (2022, Supporting Information S1).

221 Results of the derived wave characteristics described above are presented in Fig-  
 222 ure 2 for two representative occultations from Figure 1. For the first example (orbit 4425n1),  
 223 not only the temperature profiles from MIR and NIR coincide, but the retrieved wave  
 224 amplitudes and phases as well (Figure 2a). As a result, the envelopes of the wave packet  
 225 determined from the MIR and NIR data agree well up to  $\sim 90$  km (Fig. 2b, 2c). How-  
 226 ever, for the second profile (orbit 8946n1), the amplitude and estimated potential energy  
 227 disagree between MIR and NIR above 50 km (Fig. 2d, 2e), although short vertical-scale  
 228 features in the temperature profile are resolved well. The altitude distribution of the wave  
 229 drag agrees well up to  $\sim 80$  km for both orbits, peaking with  $60 \text{ m s}^{-1} \text{ sol}^{-1}$  around 80  
 230 km (Figures 2c,f). Above this height, the temperature uncertainties significantly increase  
 231 for the NIR data.

## 232 4 Description of the Data Coverage

233 The analyzed NIR and MIR measurements were taken over two Martian years (MY),  
 234 between the second half of MY34 and the first half of MY36 (May 2018 - February 2022).  
 235 For that period, the MIR  $2.7 \mu\text{m}$   $\text{CO}_2$  band statistics encompasses  $\sim 350$  occultations  
 236 in each hemisphere, while the NIR observations are performed  $\sim 10$  times more frequently.  
 237 The seasonal-latitudinal coverage of individual orbits is presented in Figure 3 (upper row)  
 238 as a function of the solar longitude  $L_s$  for the northern (left column) and southern (right  
 239 column) hemispheres separately. Due to the solar occultation mode, the observations were  
 240 performed either during sunrises or sunsets over morning or evening twilight. However,  
 241 since the local time (LT) of the solar terminator varies with orbit and latitude, it may  
 242 reach midday or midnight closer to the polar regions (Figures 3a,b).

243 To analyse the seasonal variability of the GW parameters, we grouped the individ-  
 244 ual vertical profiles into bins of  $3^\circ$  of  $L_s$  and 1 km of altitude. Thus, each bin represents  
 245 an average of one to seven measured values. Since errors grow with height and profiles  
 246 extend to different altitudes, the contribution of individual profiles in their top 20 km  
 247 was weighted by the coefficient ranging from one to zero. The distributions for the GW  
 248 drag and wave potential energy are plotted in the middle and lower panels of Figure 3,  
 249 correspondingly. The upper altitude spread depends on season and latitude varying from  
 250 10-20 km to 140-150 km at aphelion and from 20-30 km to 160-170 km at perihelion, as  
 251 clearly seen in the southern hemisphere (Figures 3d,f). An increased wave activity of up  
 252 to  $300\text{-}400 \text{ J kg}^{-1}$  is observed in the winter hemispheres at the mesospheric and ther-



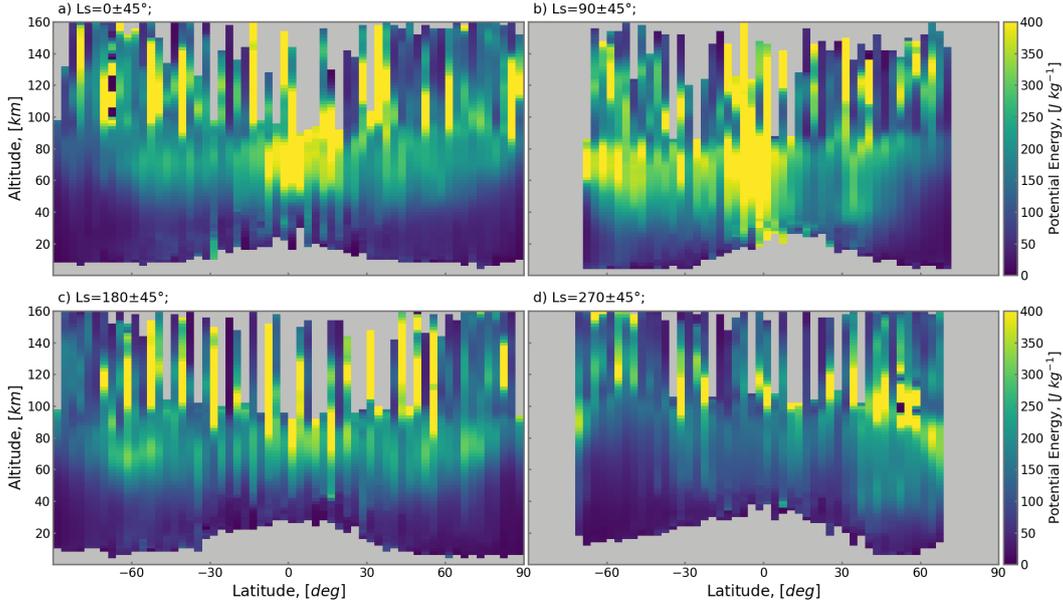
**Figure 3.** Upper row: coverage of the ACS measurements as a function of the solar longitude  $L_s$  and latitude; color indicates the local time of observations. Middle and lower rows: seasonal-altitude distribution for the GW drag and potential energy (per unit mass), correspondingly. The left and right columns present the data for the Northern and Southern hemispheres, respectively. Grey area denotes the period of global dust storm (GDS). Red dashed lines separate Martian years MY34, MY35 and MY36.

253 atmospheric altitudes. In the summer hemispheres, the peaks of wave activity lie higher,  
 254 with GW drag reaching maxima near or above the mesopause.

## 255 5 Latitudinal Distribution

256 We next turn to a more detailed examination and consider the altitude-latitude dis-  
 257 tributions of the GW characteristics. For that, we gathered data into  $3^\circ$  latitude bins  
 258 and organized the results into four seasons centered around  $L_s = 0^\circ, 90^\circ, 180^\circ$  and  $270^\circ$ .  
 259 They represent two equinoctial seasons ( $L_s = 0^\circ$  and  $180^\circ$ ) and two solstitial ones: around  
 260 the aphelion ( $L_s = 90^\circ$ ) and perihelion ( $L_s = 270^\circ$ ). In order to eliminate the influ-  
 261 ence of the major dust storm of MY34 that occurred between  $L_s = 188^\circ$  and  $250^\circ$ , we  
 262 excluded those measurements. The differences in the GW activity introduced by the GDS  
 263 are explicitly considered in the next section. The cross-sections of the GW potential en-  
 264 ergy are plotted in Figure 4. It is immediately seen that the wave activity is stronger  
 265 in the first half of the Martian year. The maxima are located in low latitudes in the up-  
 266 per mesosphere and lower thermosphere during the equinoctial  $L_s = 0^\circ$  season (panel  
 267 a) and shifted to the southern (winter) hemisphere over the solstitial  $L_s = 90^\circ$  season  
 268 (panel b). A similar pattern occurs in the second half of the year, although with a clearly  
 269 smaller magnitude, especially during the northern winter solstice. Note the symmetry  
 270 of the  $E_p$  distribution with respect to the equator during the equinox.

271 Figure 5 provides further insight into the climatology of GWs. It presents the latitude-  
 272 altitude distributions of the zonal GW drag (shaded) along with the mean zonal wind  
 273 (red contours) simulated with the MAOAM Martian general circulation model (MGCM)  
 274 (Hartogh et al., 2005; Medvedev & Hartogh, 2007) for MY34 and 35 and accordingly av-

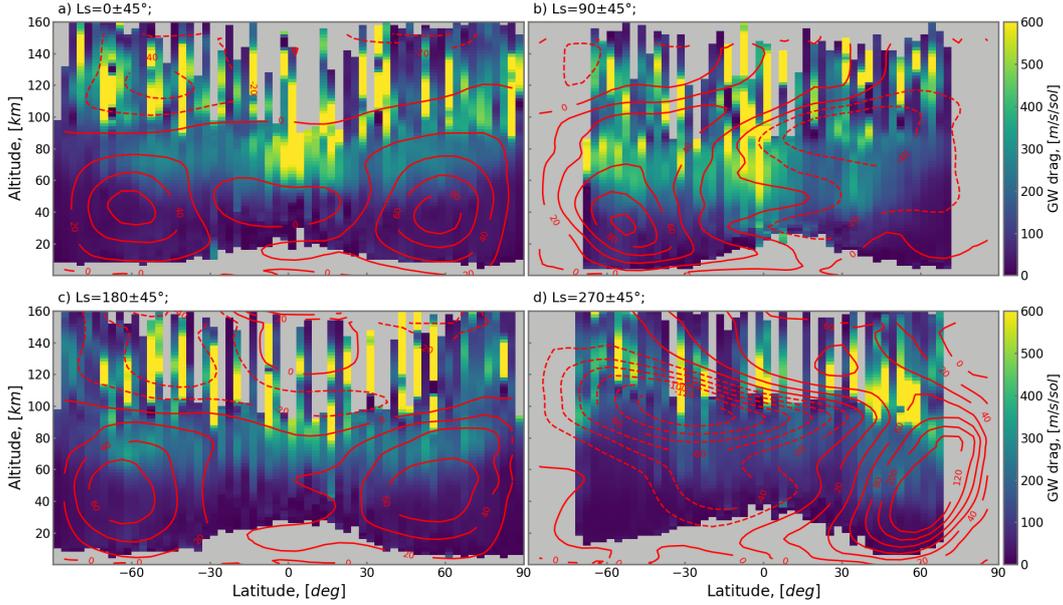


**Figure 4.** Latitude-altitude distributions of the retrieved wave potential energy (per unit mass)  $E_p$  for four representative seasons centered around a)  $L_s = 0^\circ$ , b)  $90^\circ$ , c)  $180^\circ$  and d)  $270^\circ$ . The period of the MY34 GDS is excluded.

275 eraged. It is seen that the regions of large GW drag in general align with the areas of  
 276 relatively weak zonal wind, which agrees with the theoretically expected propagation and  
 277 dissipation characteristics of gravity waves. Harmonics, especially the ones with relatively  
 278 slow (ground-based) horizontal phase speeds  $c$ , are substantially damped, when their phase  
 279 speeds approach the mean wind  $\bar{u}$ . This decrease in the intrinsic horizontal phase speed  
 280  $|c - \bar{u}|$  causes absorption of a significant portion of GWs propagating along the mean  
 281 wind  $\bar{u}$ . Harmonics having  $c > \bar{u}$  or traveling in the opposite to the wind direction can  
 282 avoid wave filtering and propagate higher, grow in amplitude and ultimately break down  
 283 when wave-induced wind fluctuations  $|u'|$  approach the intrinsic phase speed  $|c - \bar{u}|$ . The  
 284 smaller  $\bar{u}$ , the smaller amplitude  $|u'|$  is required for breaking/saturation, which is illus-  
 285 trated by the enhanced momentum deposition in the regions of the weak mean wind shown  
 286 in Figure 5.

287 The equinoctial circulation consists of two prograde (eastward) jets centered in middle-  
 288 to-high latitudes of each hemisphere and the region of weak winds at low latitudes. The  
 289 inferred distribution of the GW drag reflects this pattern of inter-hemispheric symme-  
 290 try. Weaker mean winds in low latitudes allow for GW breaking at lower altitudes. The  
 291 mean wind changes direction to retrograde (westward) above the mesopause. This causes  
 292 harmonics traveling westward ( $c < 0$ ) to break and/or dissipate and deposit their mo-  
 293 mentum there. In fact, the wind reversal itself is the result of the GW drag (Medvedev  
 294 et al., 2011a, 2011b, 2013).

295 The solstitial circulation features eastward/westward jets in the winter/summer hemi-  
 296 spheres. The jets are stronger during the perihelion solstices, as seen from the compar-  
 297 ison of Figures 5b and d, due to greater insolation and larger meridional temperature  
 298 and pressure gradients in this season. Assuming that GW harmonics excited in the lower  
 299 atmosphere have a broad range of phase speeds  $c$  and travel in all horizontal directions,  
 300 stronger background winds  $\bar{u}$  filter out more waves propagating in the same direction.  
 301 The remaining harmonics must acquire larger amplitudes in order to break down and,

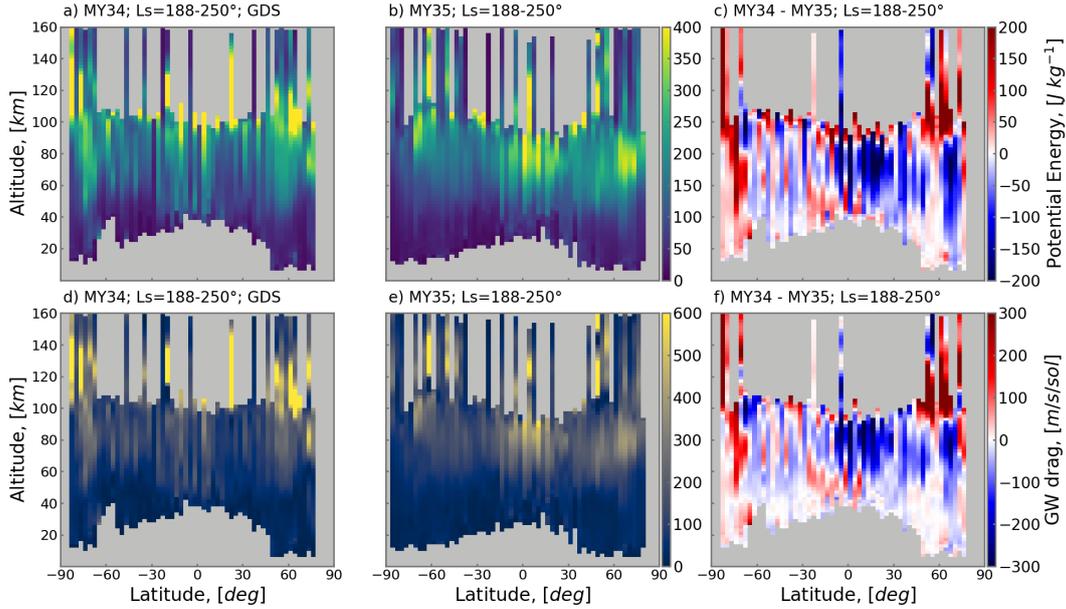


**Figure 5.** The same as in Figure 4, but for the momentum forcing (or GW drag) in  $\text{m s}^{-1} \text{sol}^{-1}$  (shaded). The simulations with the MAOAM MGCM of the mean zonal wind corresponding to the same intervals of  $L_s$  is shown with contour lines. Solid and dashed lines represent eastward and westward winds, respectively. The data for the period of the MY34 GDS are excluded.

302 therefore, have to propagate higher. This mechanism explains why the GW drag is  
 303 localized in a relatively narrow altitude range during the perihelion solstice (Figure 5d)  
 304 compared to that during the aphelion (Figure 5b).

## 305 6 Impact of the Dust Storm

306 As was mentioned above, a global dust storm (GDS) rapidly developed in MY34  
 307 around  $L_s = 188^\circ$  and decayed by  $L_s = 250^\circ$ . The dust load over the same period of  
 308 MY35 was close to normal with a minor enhancement between  $L_s = 230^\circ$  and  $250^\circ$ .  
 309 Figure 6 presents the latitude-altitude distributions of the retrieved GW potential en-  
 310 ergy and drag averaged over the corresponding periods along with their differences (right  
 311 column). The latter show several systematic features introduced by the storm. First, the  
 312 changes are mostly symmetric with respect to the equator, at least in the overlapping  
 313 bins. They may reflect the predominantly symmetric global equinoctial circulation at  
 314 the beginning of the storm, which affects generation and vertical propagation of GW har-  
 315 monics. Second, a distinct reduction of wave activity occurs in middle latitudes between  
 316  $\sim 15^\circ$  and  $70^\circ$  in both hemispheres. A similar behavior during the MY34 GDS was ob-  
 317 served in the Mars Climate Sounder data by Heavens et al. (2020) throughout the lower  
 318 atmosphere below  $\sim 30$  km. Simulations with a high-resolution global circulation model  
 319 also reproduced the approximately factor 2 decrease of the GW potential energy in the  
 320 lower and middle atmosphere (Kuroda et al., 2020). It was related to the reduction of  
 321 wave generation caused by convective and baroclinic stabilization of the atmosphere in-  
 322 duced by the storm. The same simulations predicted a gradual increase of GW activ-  
 323 ity with height, such that  $E_p$  exceeds the “low dust” values near the top of the model  
 324 domain at around 80 km. Further observational evidence for the enhancement of GW  
 325 activity in the upper atmosphere during dust storms was provided in the work by Yiğit,



**Figure 6.** Altitude-latitude cross-sections of wave potential energy (upper row) and drag (lower row) retrieved during the global dust storm of MY34 ( $L_s=188^\circ\text{--}250^\circ$ , left column) and the corresponding low-dust period of MY35 (middle column). The differences between the observations of MY34 and MY35 are shown in the right column.

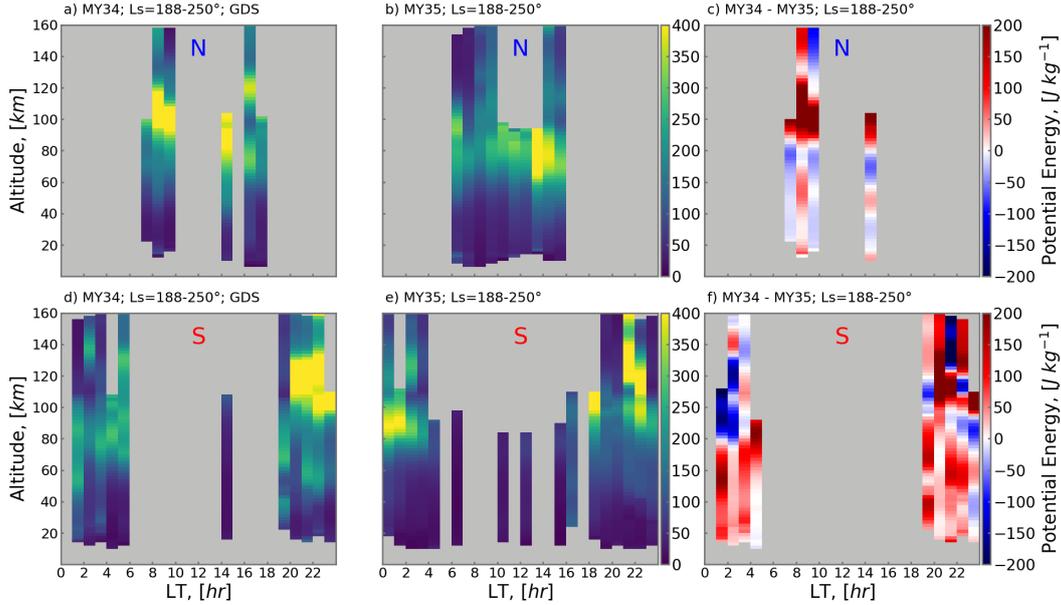
326 Medvedev, Benna, and Jakosky (2021) based on NGIMS measurements on board the MAVEN  
 327 orbiter. Those measurements covered the thermosphere between 160 and 230 km, that  
 328 is above the upper limit of the MIR data presented here. A several latitudinal bins in  
 329 Figures 6a,d with data extending above  $\sim 100$  km and available for comparison point out  
 330 to a dust-induced enhancement of GW activity and drag above the mesopause.

331 The results in Figure 6 show two additional features, which were neither observed,  
 332 nor modeled/predicted before. One of them is the enhancement of the GW activity at  
 333 all heights in the polar regions, which is clearly seen in the southern hemisphere and is  
 334 somewhat less apparent in the northern one. The second is the strong increase of GW  
 335 activity and drag below 60 km during the GDS and a steep reduction above in low lat-  
 336 itudes (approximately within  $\pm 15^\circ$  from the equator). While the polar enhancement is  
 337 a sufficiently robust feature, the equatorial pattern can be an artifact of a much sparser  
 338 coverage (see Figure 3, upper panels). Timing of observations can also be a source of bi-  
 339 ases due to variations of the GW field with local time. They are discussed in the next  
 340 section.

## 341 7 Local Time Variations

### 342 7.1 Changes due to the Dust Storm

343 The relatively small number of observations in low to middle latitudes provides no  
 344 opportunity to further evaluate the dust storm-induced pattern in the equatorial zone  
 345 shown in Figure 6, or the local time behavior there. However, the high-latitude enhance-  
 346 ment can be considered in more detail. For that, we plotted in Figure 7 the local time  
 347 variations of the GW potential energy averaged over latitudes higher than  $60^\circ$  in both  
 348 hemispheres. It is seen that, in both Martian years, most of the observations in the south-  
 349 ern high latitudes were taken during nighttime, and during daytime in the northern hemi-

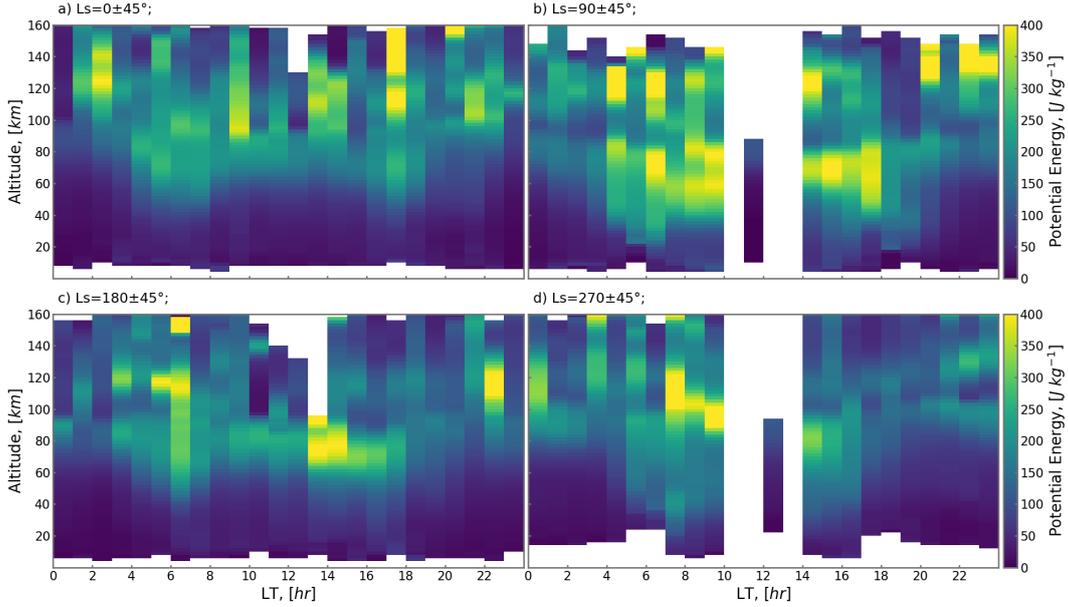


**Figure 7.** Local time variations of the globally averaged wave potential energy observed during the GDS of MY34 ( $L_s=188^\circ-250^\circ$ , left column) and the corresponding “low dust” period of MY35 (middle column). The differences between the observations of MY34 and MY35 are shown in the right column. The observations in high northern (Lat  $> 60^\circ\text{N}$ ) and high southern (Lat  $> 60^\circ\text{S}$ ) latitudes are shown in the upper and lower rows, respectively.

350 sphere. Given that the location (latitudes) and timing (seasons and local times) of the  
 351 observations are the same in MY34 and 35, the differences in the GW activity can be  
 352 attributed to the dust conditions. A clear enhancement of  $E_p$  during the dust storm oc-  
 353 curs in the southern high latitudes of the middle and upper atmosphere. There are fewer  
 354 overlapping observations in the northern hemisphere, but those available demonstrate  
 355 the increased GW activity in the thermosphere as well. Note that high-latitude dust storm-  
 356 induced enhancements in the middle atmosphere of the similar magnitude were predicted  
 357 in simulations with a wave-resolving MGCN up to  $\sim 80$  km (Kuroda et al., 2020, Fig-  
 358 ure 4b). The results presented here provide the first observational validation for this mod-  
 359 eling prediction and, also demonstrate that the enhancement extends higher into the ther-  
 360 mosphere up to  $\sim 160$  km.

## 361 7.2 Seasonal Behavior

362 After exploring the impact of the major dust storm on local time variations of the  
 363 GW field, we next consider how they evolve seasonally. For that, we grouped the data  
 364 into the same four seasons discussed in Section 5 and plotted them as functions of lo-  
 365 cal time in Figure 8. The figure reveals more intraday features of the GW activity. They  
 366 include, in particular, a downward phase progression, which is more clearly seen during  
 367 equinoctial seasons (Figures 8a,c). Such local time variations can reflect a modulation  
 368 of GWs in the middle and upper atmosphere by the diurnal and semi-diurnal thermal  
 369 tides, which have a distinct latitudinal structure and vary with seasons (Yiğit & Medvedev,  
 370 2017; Kumar et al., 2022). Tides are more symmetric with respect to the equator dur-  
 371 ing equinoxes due to the position of the Sun. Therefore, averaging over all latitudes does  
 372 not mask the tidal signal in Figures 8a,c that much as it does during the solstitial sea-  
 373 sons. A mixture of modulation by the semidiurnal and diurnal tides is seen during the



**Figure 8.** Local time variations of the GW potential energy at four representative seasons centered around  $L_s = 0^\circ$ ,  $90^\circ$ ,  $180^\circ$  and  $270^\circ$ . The data for the MY34 GDS are excluded.

374  $L_s = 0^\circ$  equinox, while the diurnal (24-hour period) variation dominates the  $L_s = 180^\circ$   
 375 season. Figures 8b,d show stronger GW activity in the first half of the day with signs  
 376 of the downward phase progression. The latitudinal distribution of tides is more com-  
 377 plex due to the shifted center of solar heating and differences in the background winds.  
 378 Thus, the peculiarities in the intraday variations of  $E_p$  in Figures 8b,d may reflect this  
 379 and the biases due to the measurements sampling. Note that variations can be caused  
 380 by intraday changes in GW sources, which, however, were not detected in the consid-  
 381 ered data set. Besides, the behavior of the phase provides unambiguous evidence for the  
 382 tidal nature of the GW activity variations.

## 383 8 Conclusions

384 We presented the results of the analysis of gravity waves (GWs) retrieved from solar  
 385 occultation measurements by NIR and MIR channels of the ACS instrument on board  
 386 the Trace Gas Orbiter (TGO) taken over two Martian years (mid-MY34 to mid-MY36).  
 387 The retrieved temperature profiles spanning altitudes up to 100 km (NIR) and 160 km  
 388 (MIR) were separated into mean component and disturbances, which were used for char-  
 389 acterizing the wave field. In particular, the wave activity represented by wave potential  
 390 energy  $E_p$  and the dynamical impact on the mean flow in terms of the GW momentum  
 391 deposition (“drag”) are considered here. The main inferences of this study are as follows:

- 392 1. GWs are present at all times and places in the Martian atmosphere. Within the  
 393 considered dataset, we did not find any time period or location (except for a few  
 394 profiles of questionable quality) when and where GW disturbances were absent.
- 395 2. Wave activity is distributed symmetrically with respect to the equator during the  
 396 equinoctial seasons, while the maximum is shifted to the winter hemisphere dur-  
 397 ing solstices.

- 398 3. Maxima of GW drag align with the areas of weak zonal wind along the edges of  
 399 seasonally varying zonal jets. This feature agrees well with physics of GW-mean  
 400 flow interactions.
- 401 4. During the MY34 GDS, observed GWs were depleted in middle latitudes of both  
 402 hemispheres. In contrast, GW activity increased at high latitudes (poleward of  
 403  $\sim 60^\circ$ ).
- 404 5. During both equinoctial seasons we observed diurnal and semidiurnal modulation  
 405 of the GW activity and drag with the downward phase progression.

406 The climatology of the GW activity and drag in the middle and upper atmosphere  
 407 based on 2 Martian years of ACS observations confirms theoretical/modeling predictions,  
 408 on one hand. On the other hand, it reveals new features (like enhancements of wave ac-  
 409 tivity in low latitudes), which were not anticipated and, given the dynamical importance  
 410 of GWs, have to be accounted for in numerical models. ACS continues observations, and  
 411 new data will help to further elucidate the spatio-temporal behavior of the GW field.

## 412 9 Data Availability Statement

413 The ACS data are available from ESA Planetary Science Archive (PSA) ([https://](https://archives.esac.esa.int/psa/%23!Table%20View/ACS=instrument#!Home%20View)  
 414 [archives.esac.esa.int/psa/%23!Table%20View/ACS=instrument#!Home%20View](https://archives.esac.esa.int/psa/%23!Table%20View/ACS=instrument#!Home%20View)).  
 415 The temperature vertical profiles retrieved from ACS-NIR and ACS-MIR measurements  
 416 are described in (Fedorova et al., 2023; Belyaev et al., 2022) and available at (Fedorova,  
 417 2022; Belyaev, 2022), respectively. The most recent MAOAM model output can be ac-  
 418 cessed at <https://mars.mipt.ru>. The vertical profiles of background temperature, wave  
 419 temperature disturbance, amplitude and GW drag are available at [https://data.mendeley](https://data.mendeley.com/datasets/7d9b2kjfy/1)  
 420 [.com/datasets/7d9b2kjfy/1](https://data.mendeley.com/datasets/7d9b2kjfy/1) (Starichenko, 2023).

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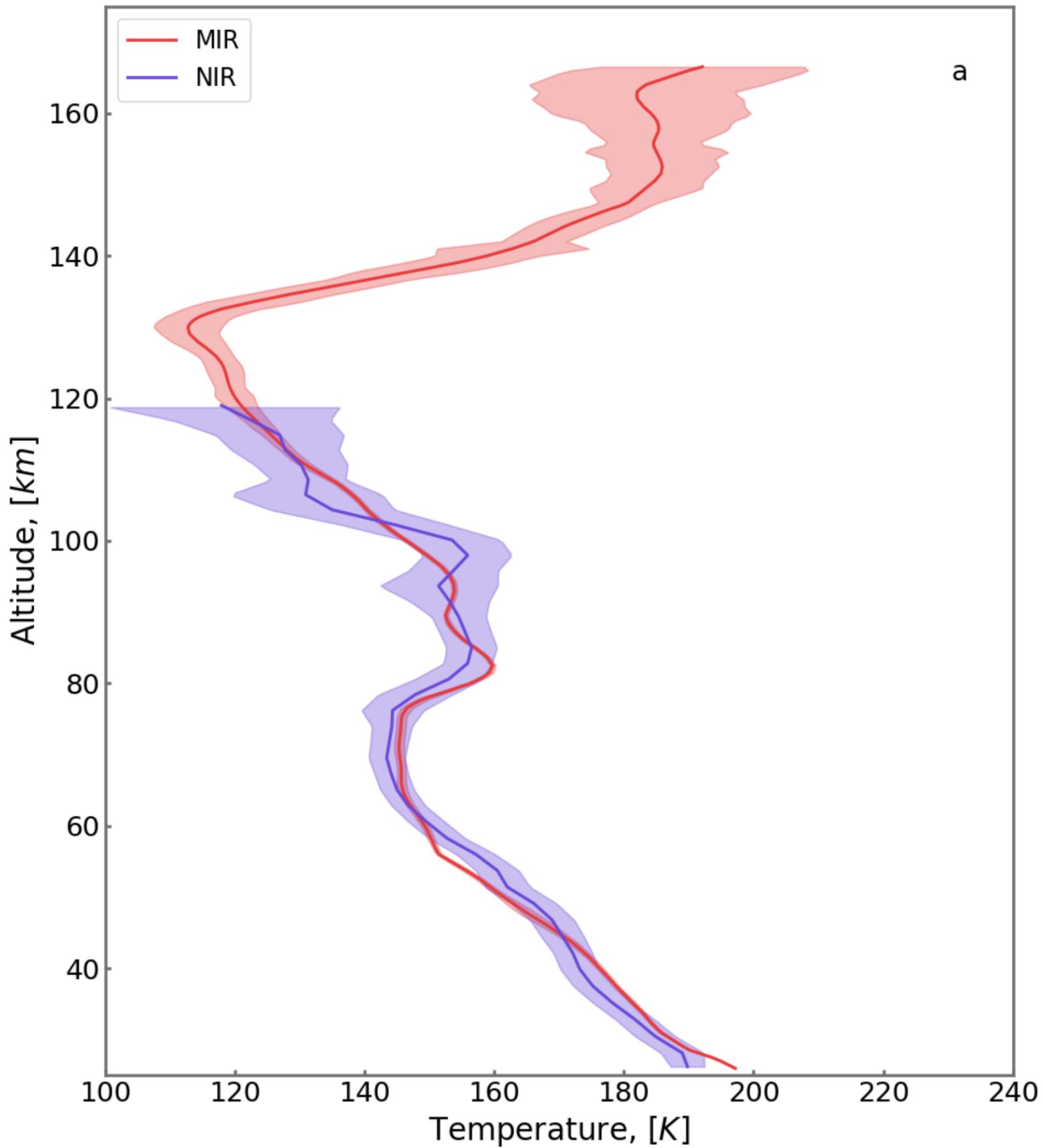
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622 COGWGB>2.0.CO;2

Figure 1.

Orbit 4425n1



Orbit 8946n1

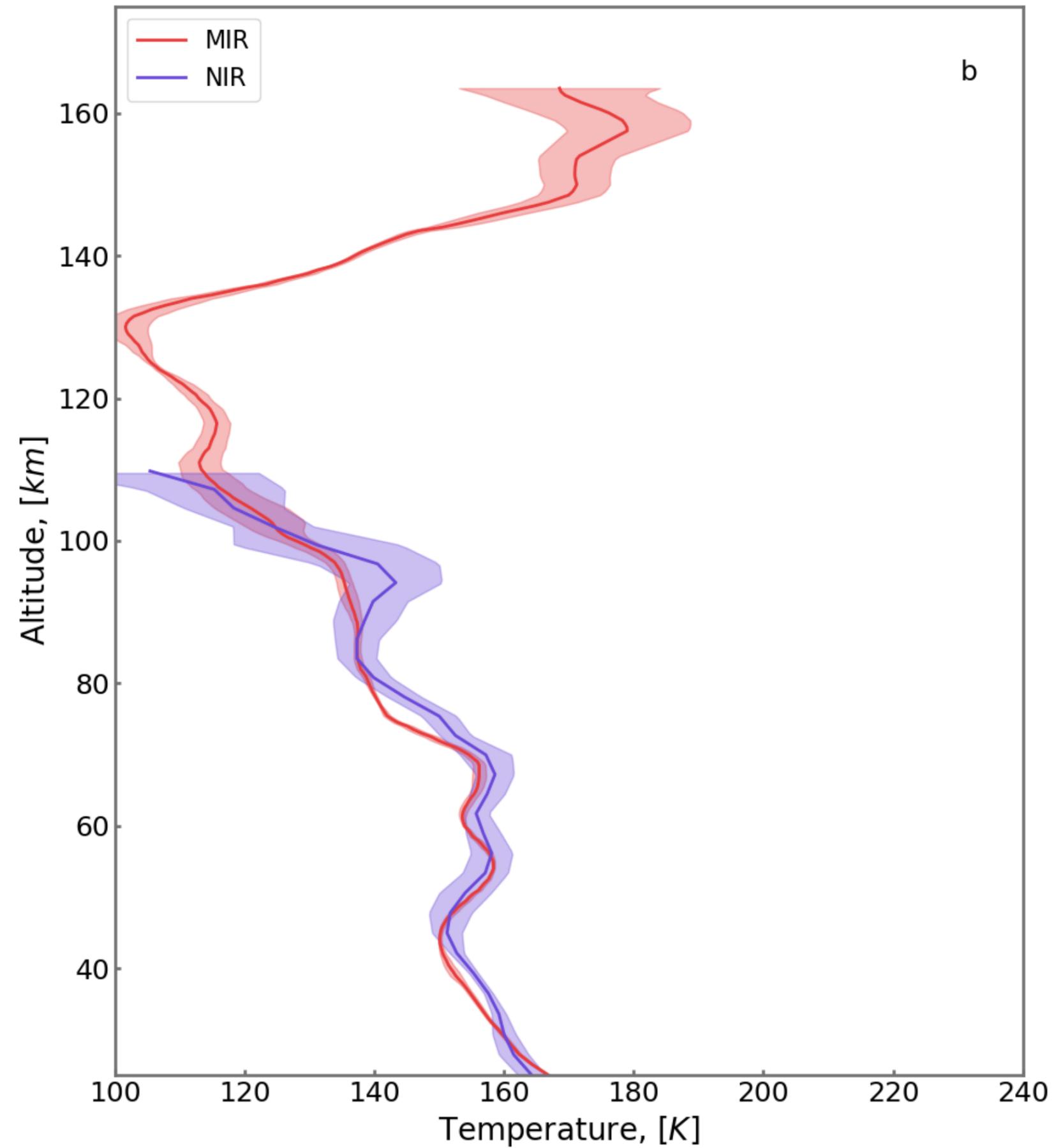
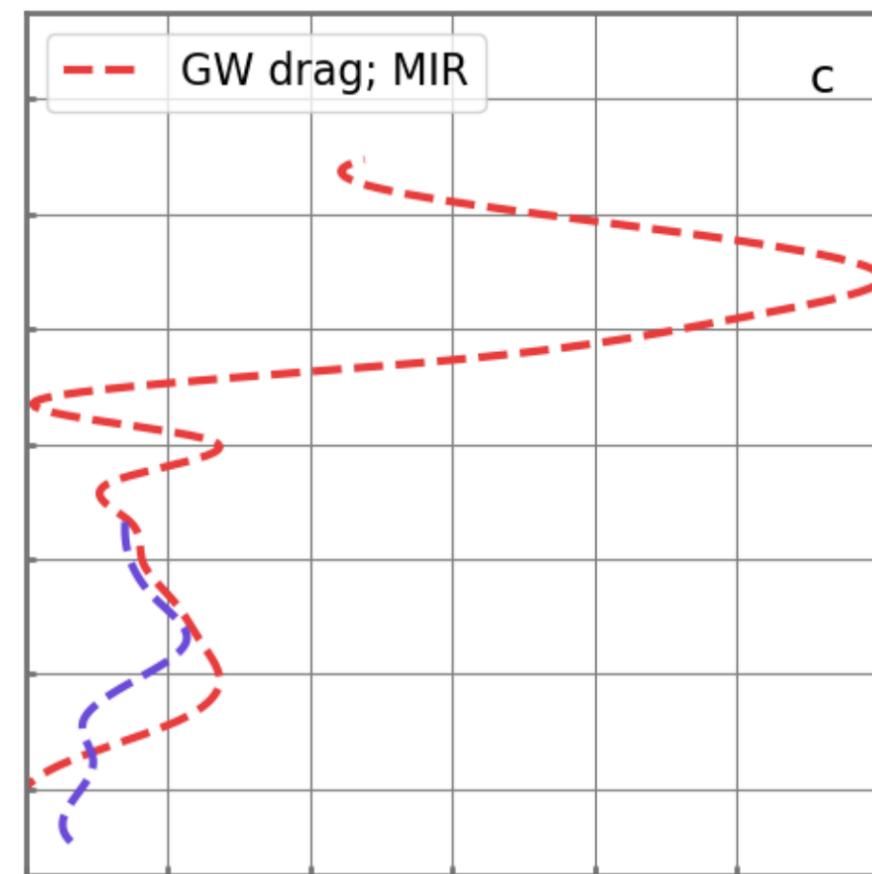
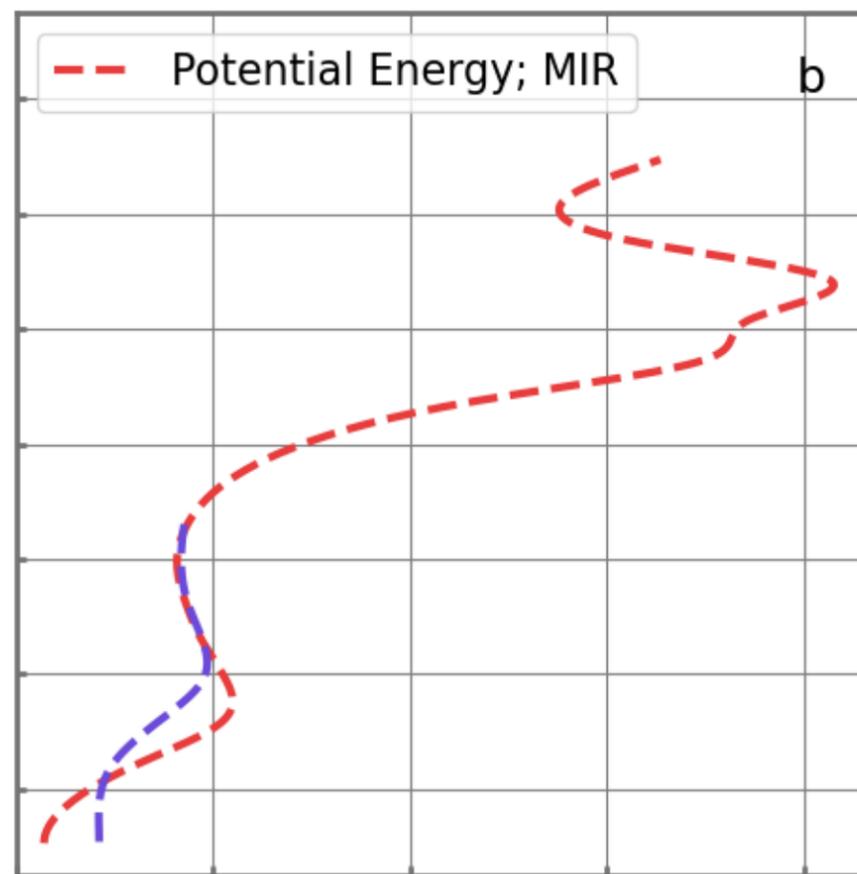
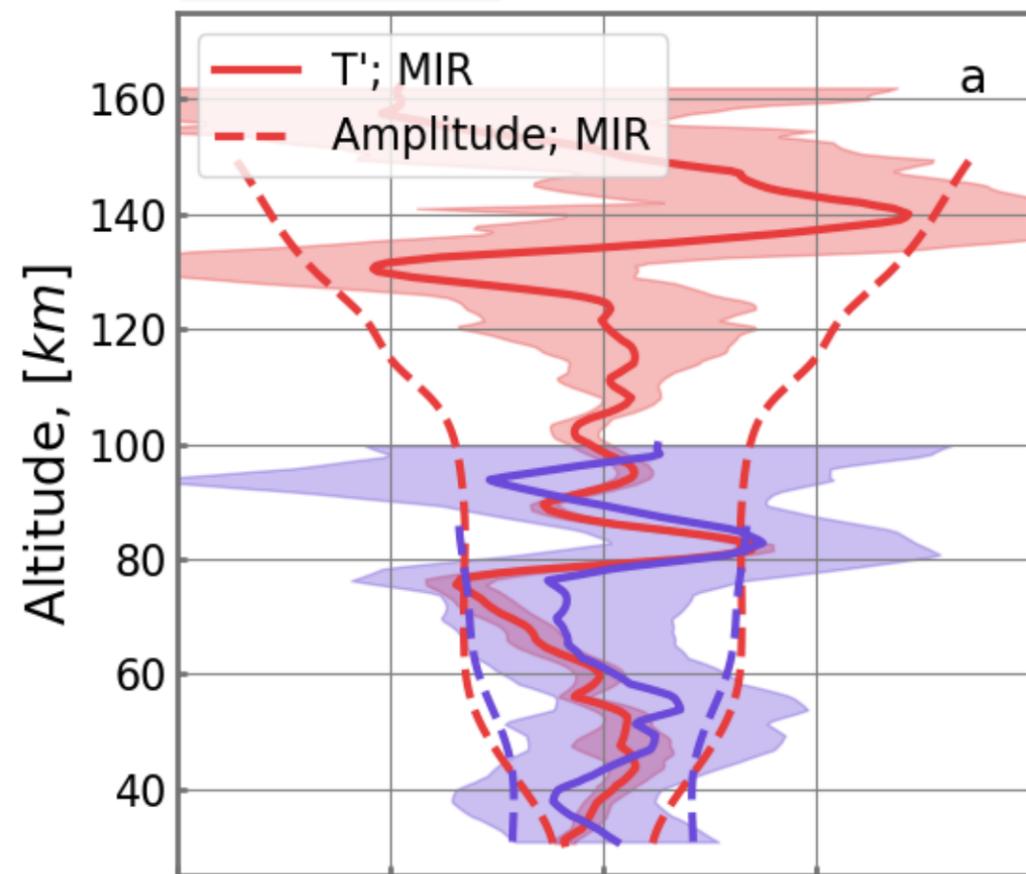


Figure 2.

Orbit 4425n1



Orbit 8946n1

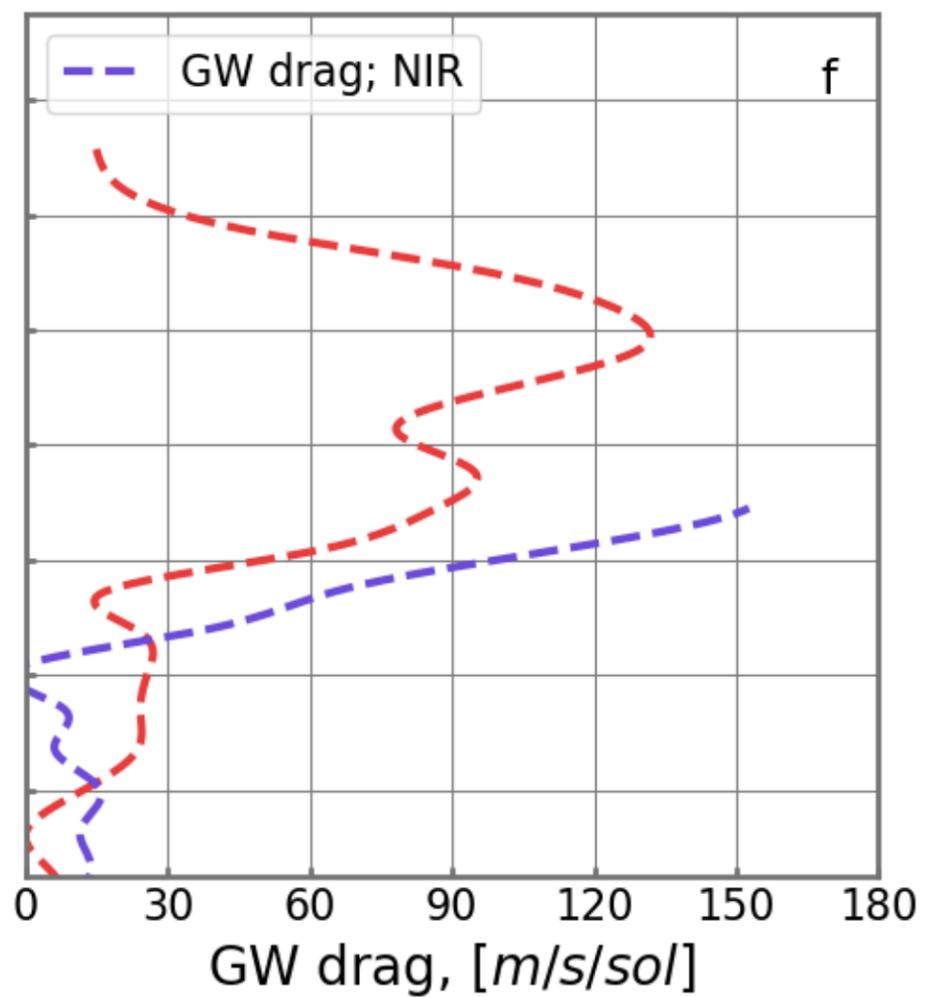
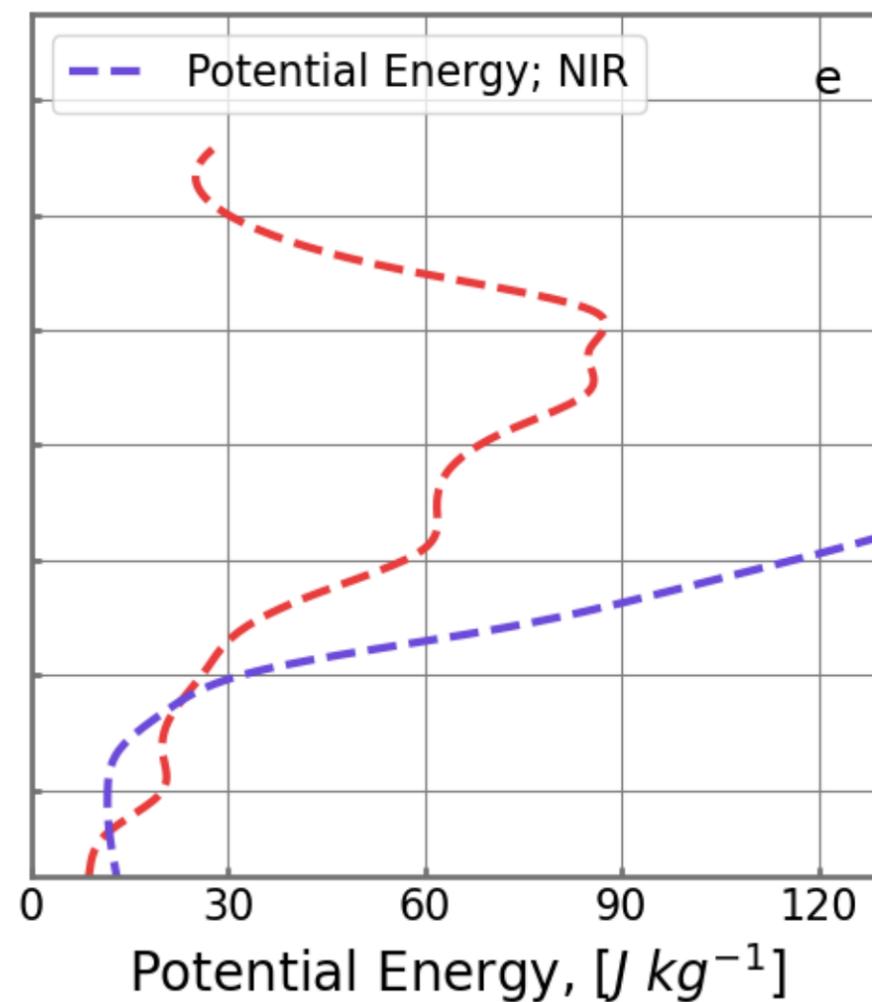
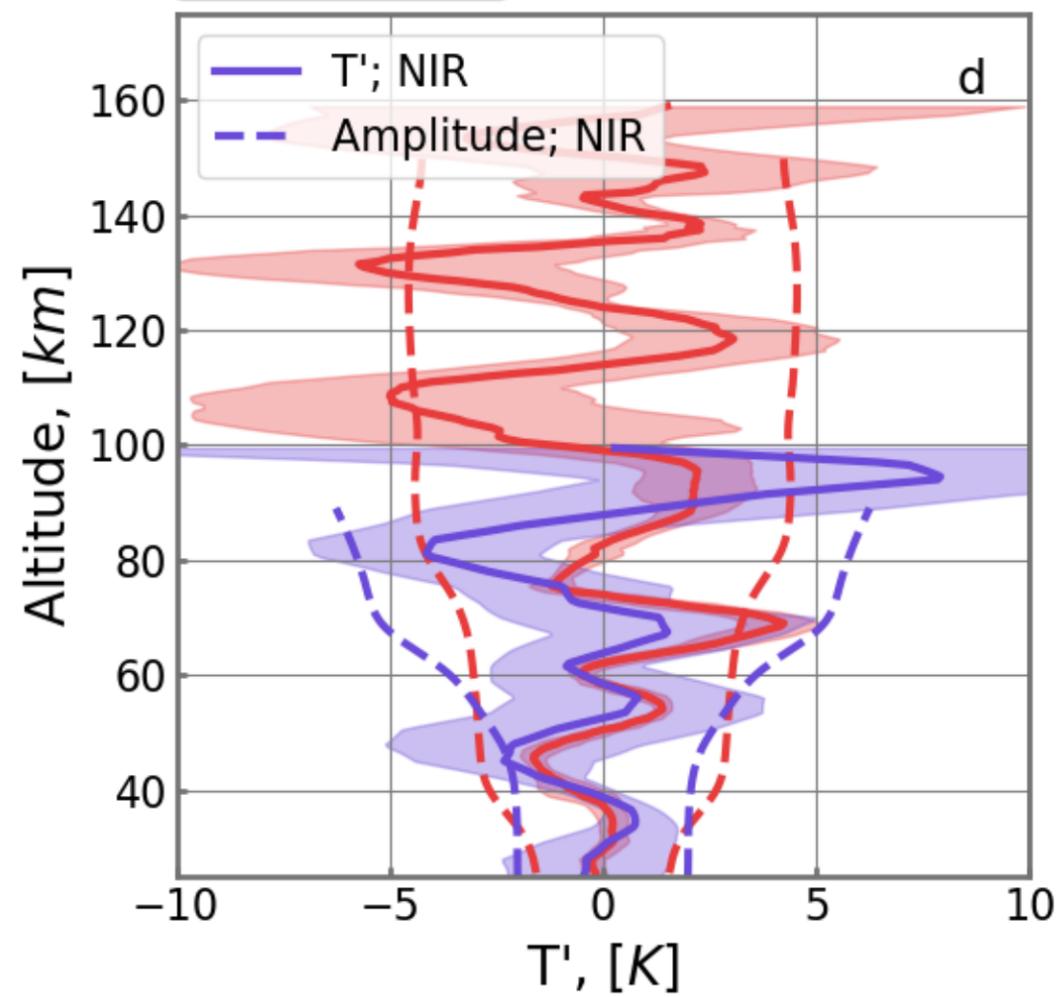


Figure 3.

# NORTHERN HEMISPHERE

# SOUTHERN HEMISPHERE

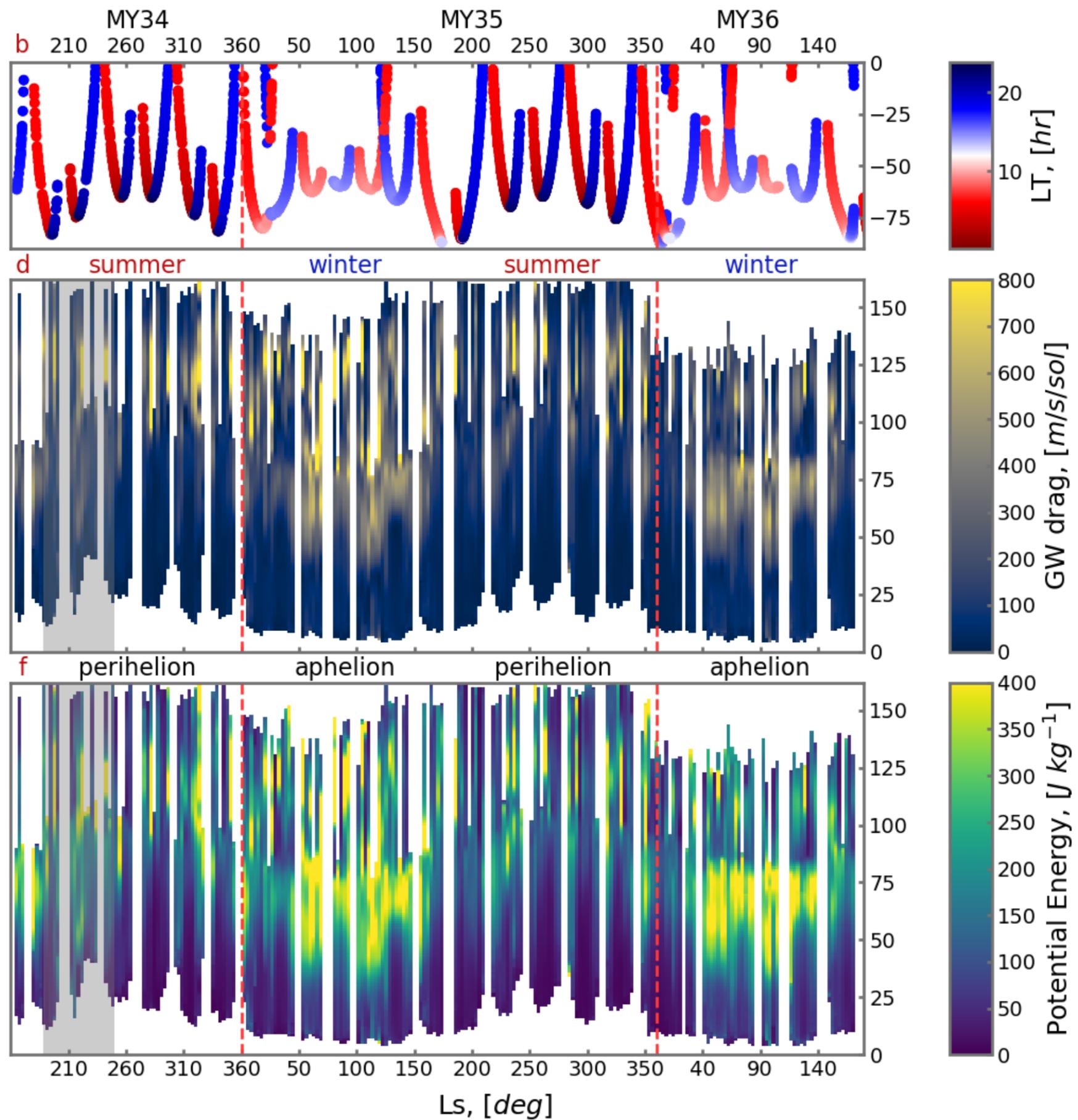
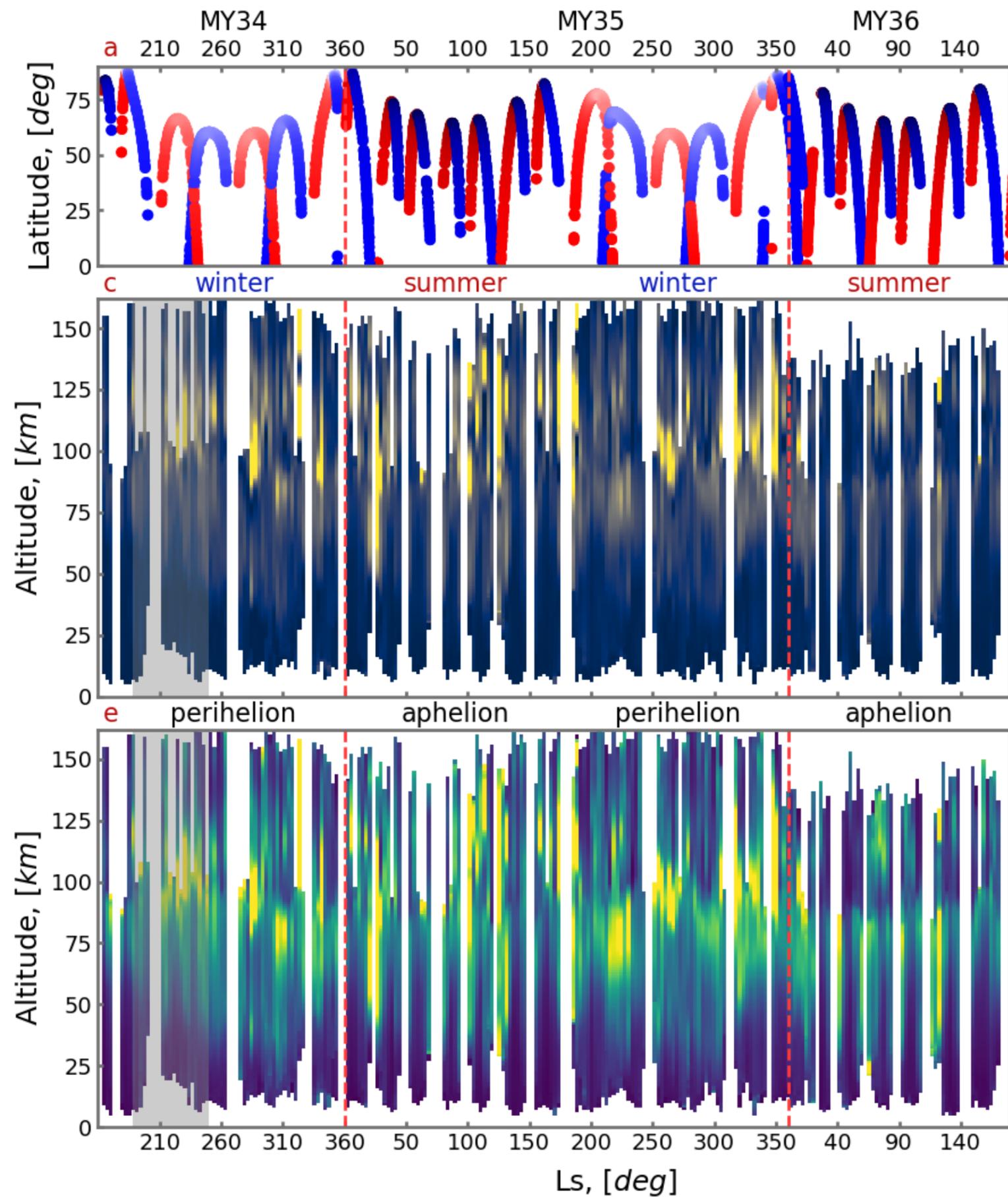


Figure 4.

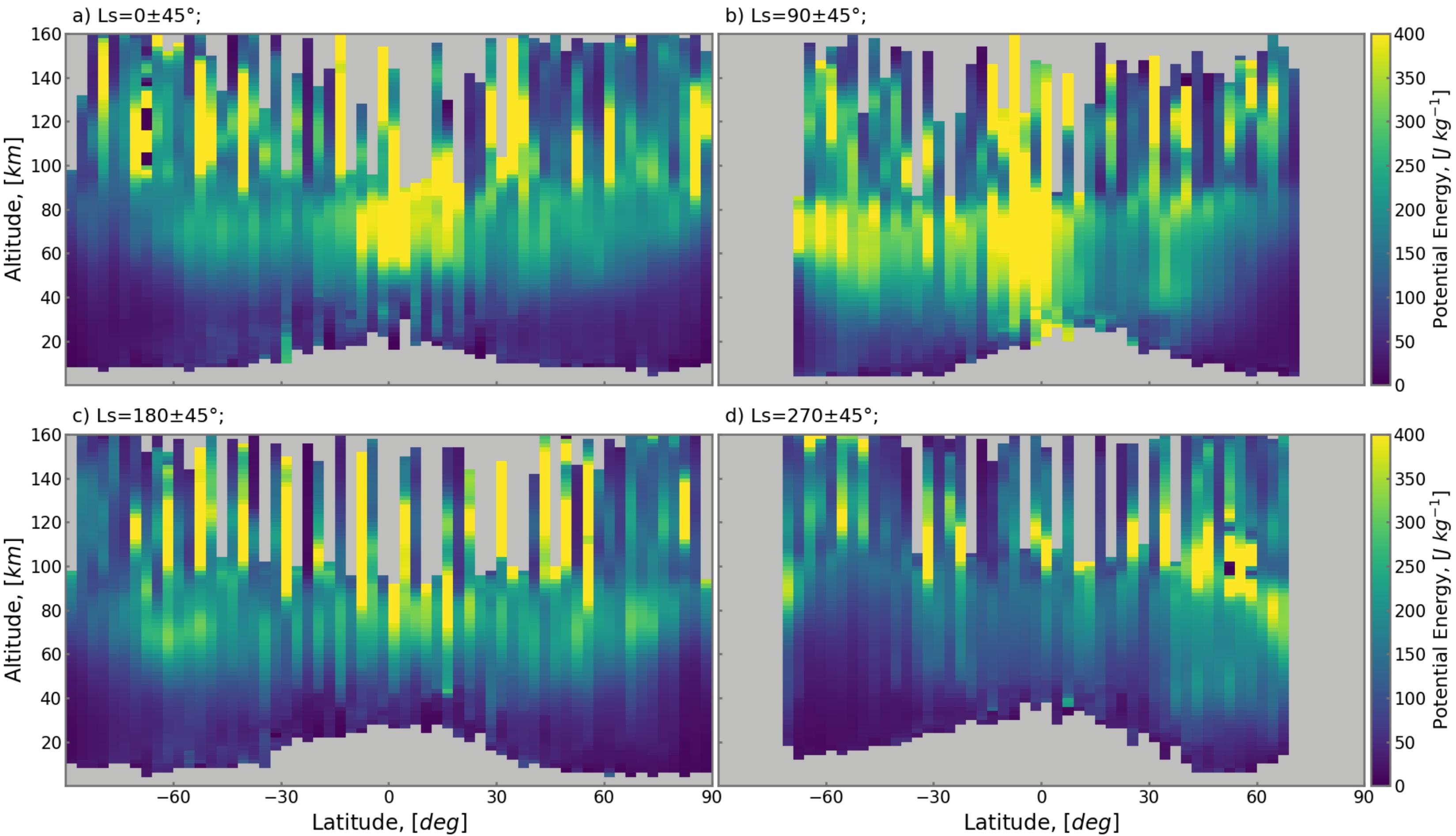


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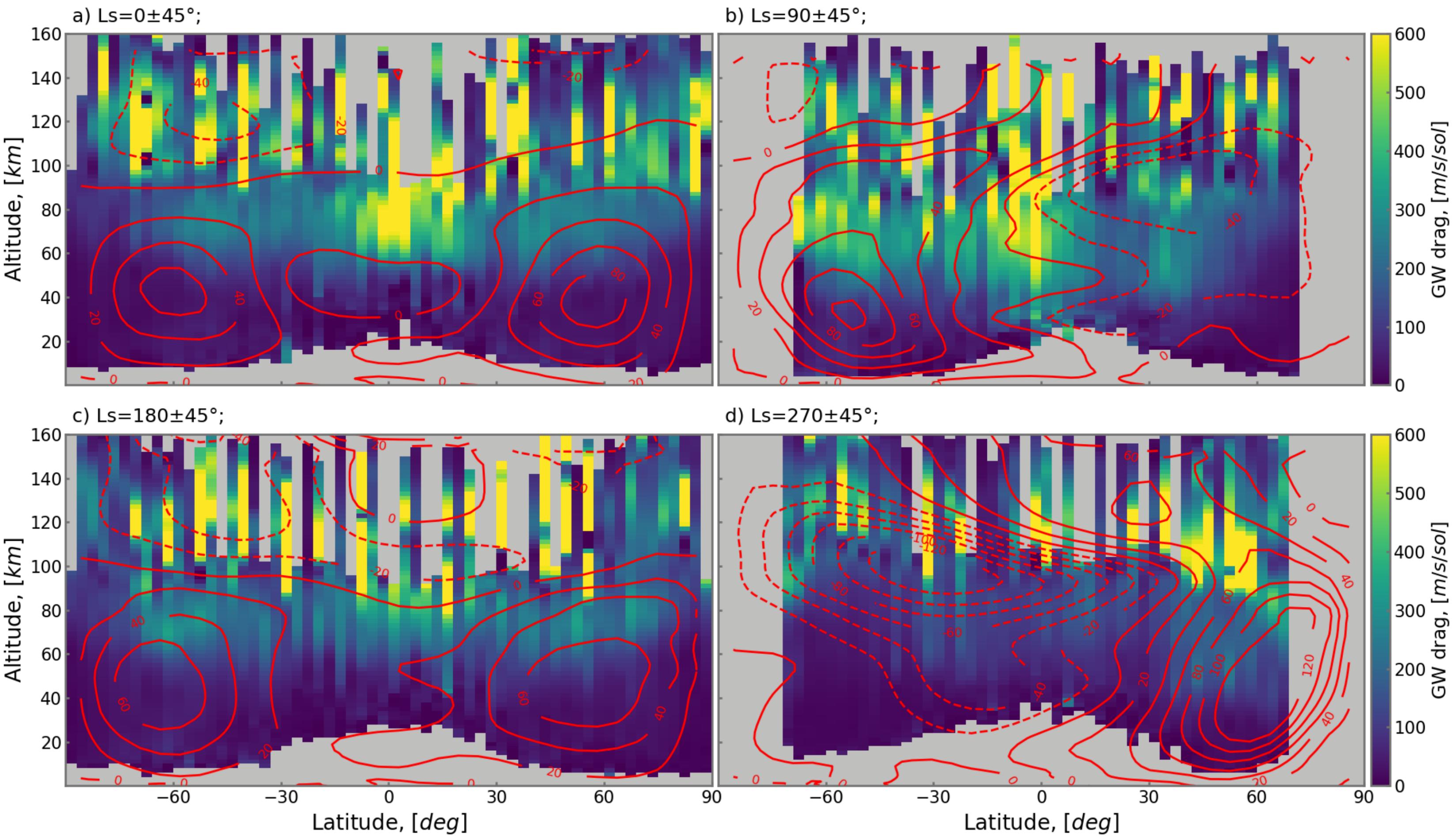
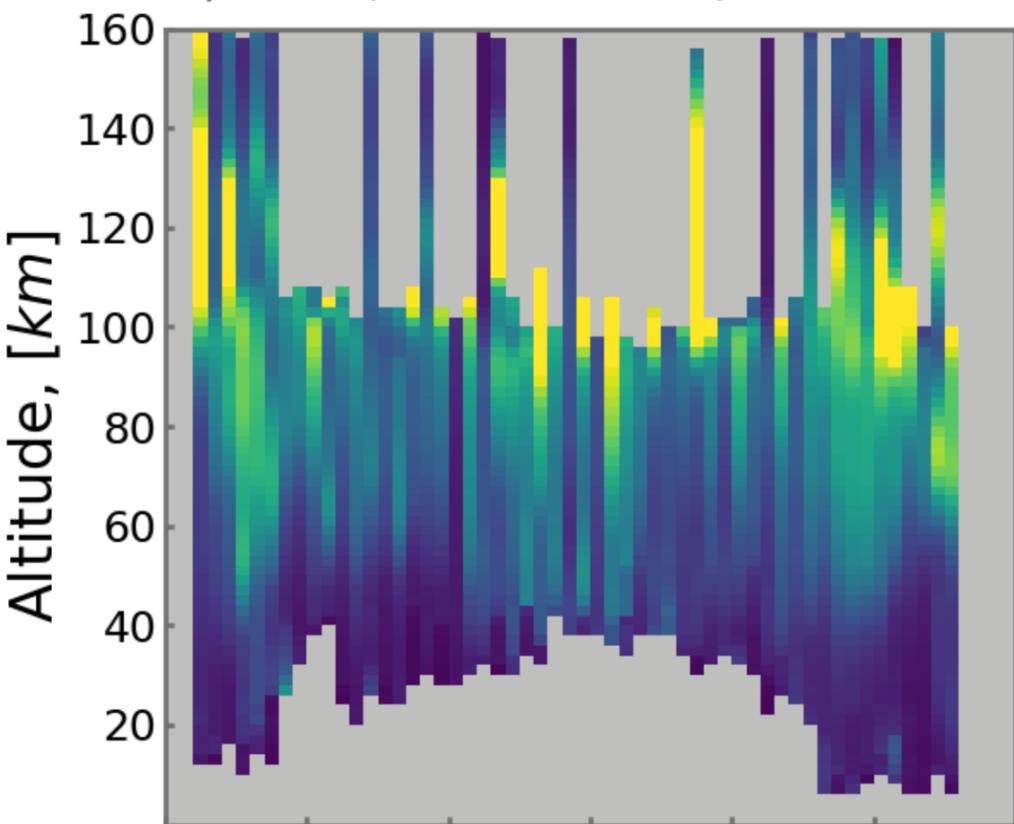
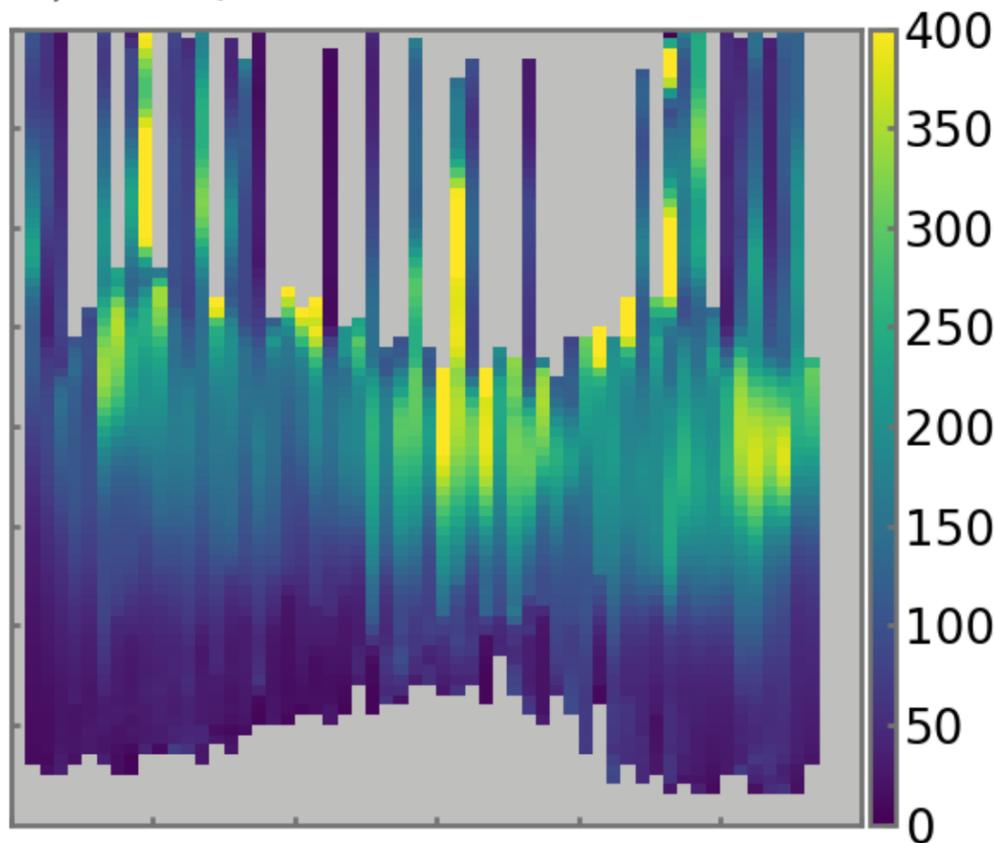


Figure 6.

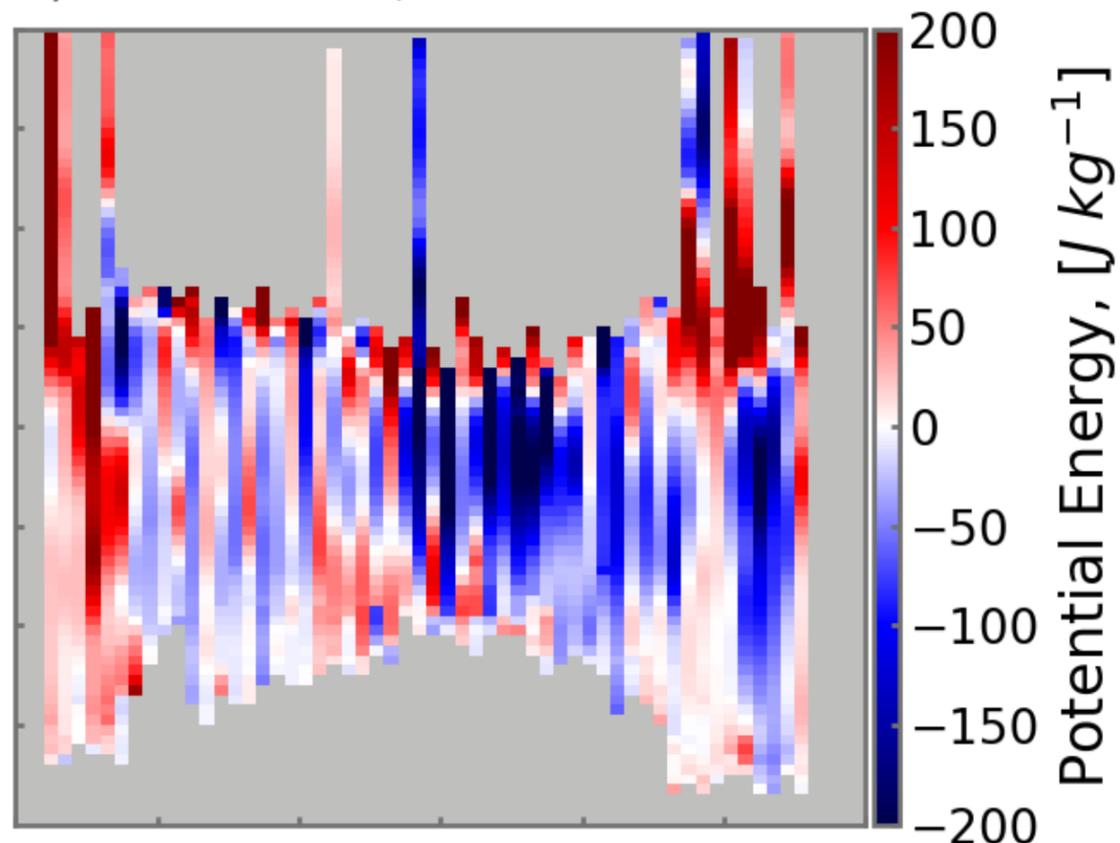
a) MY34; Ls=188-250°; GDS



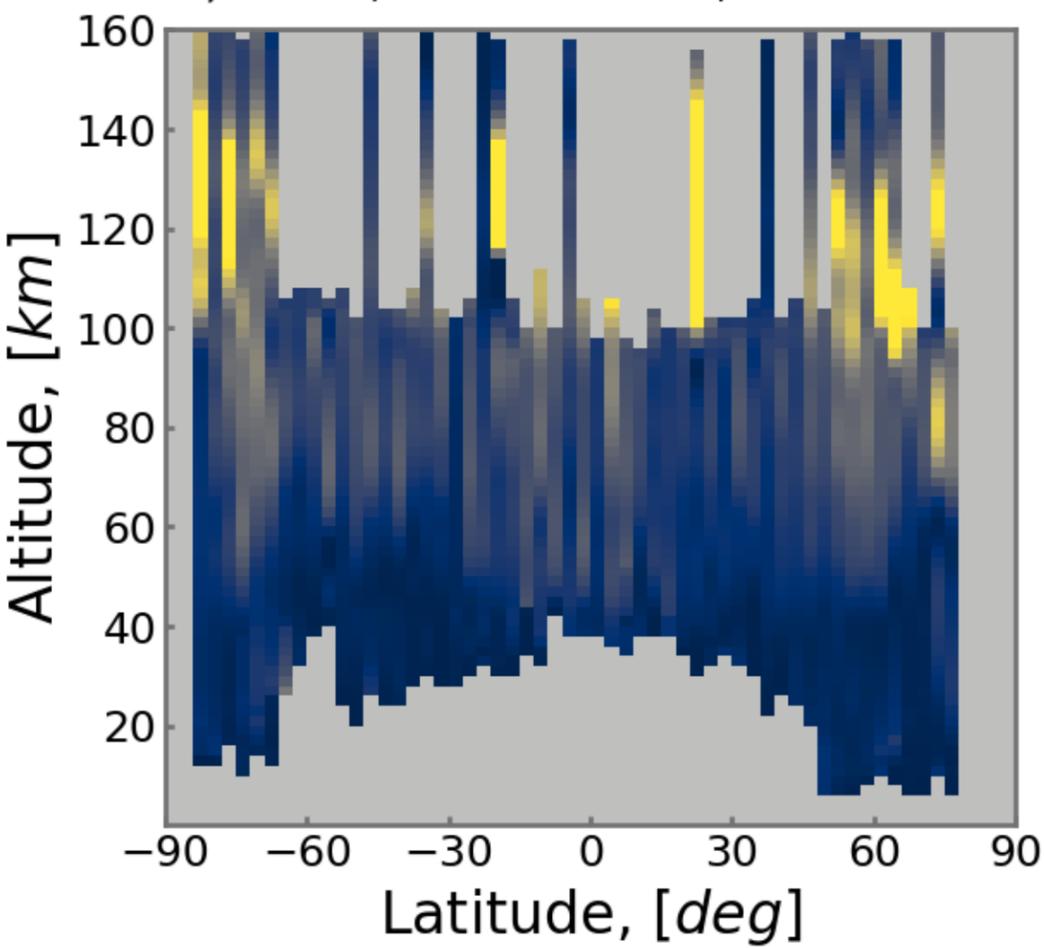
b) MY35; Ls=188-250°



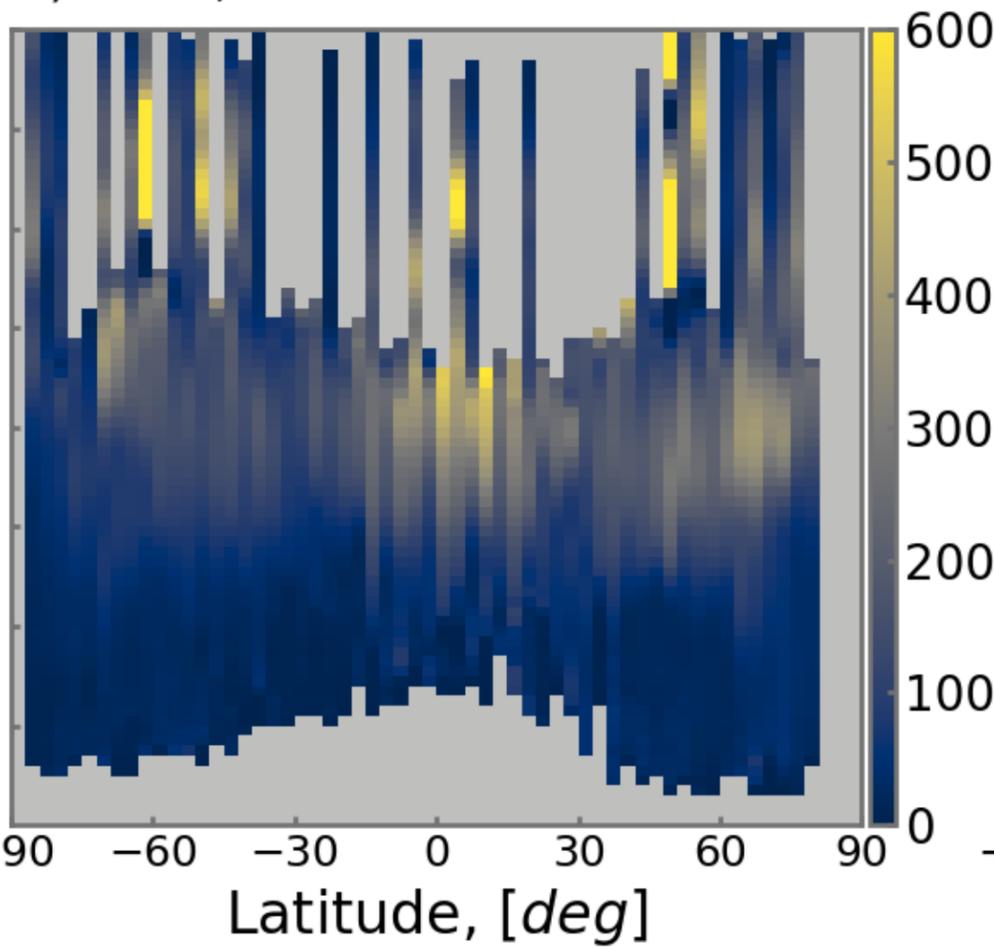
c) MY34 - MY35; Ls=188-250°



d) MY34; Ls=188-250°; GDS



e) MY35; Ls=188-250°



f) MY34 - MY35; Ls=188-250°

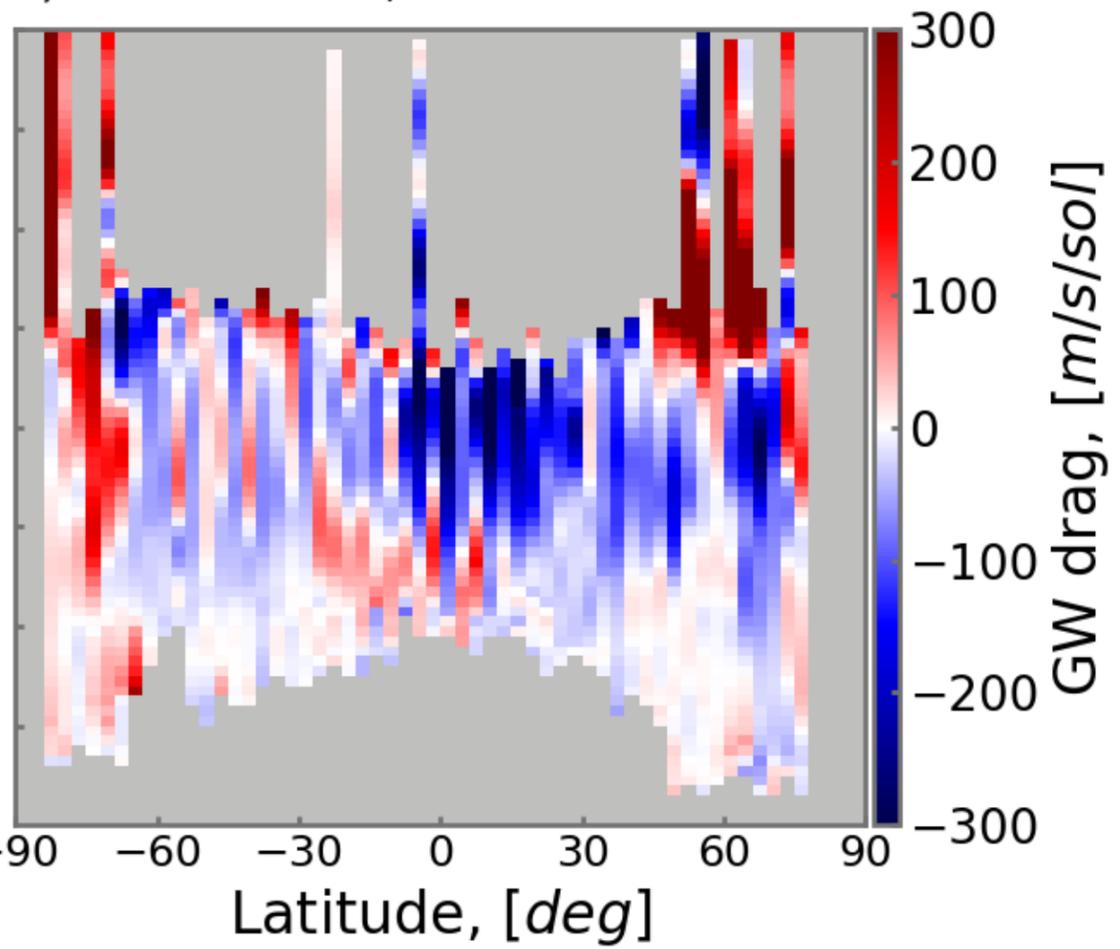


Figure 7.

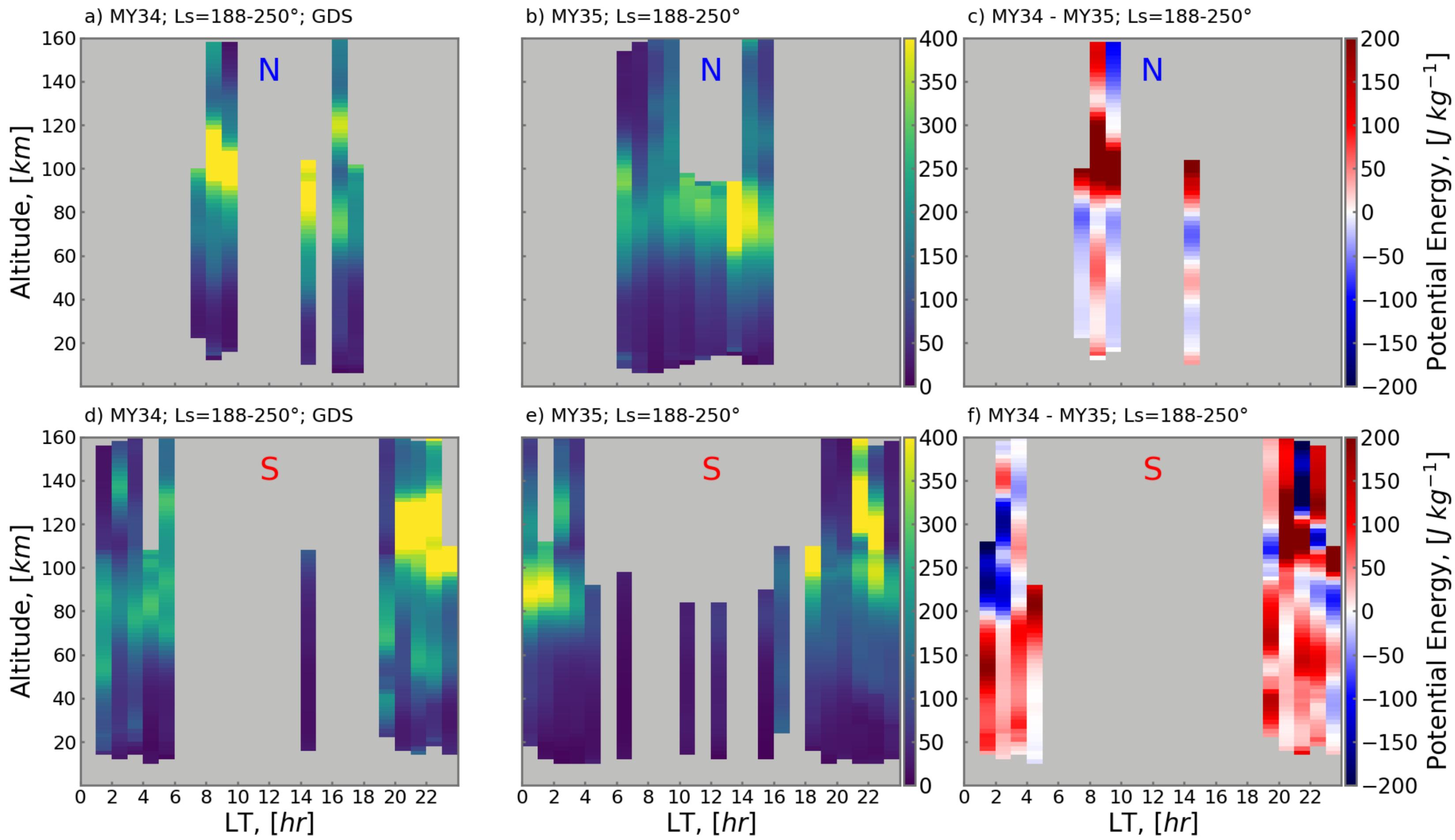
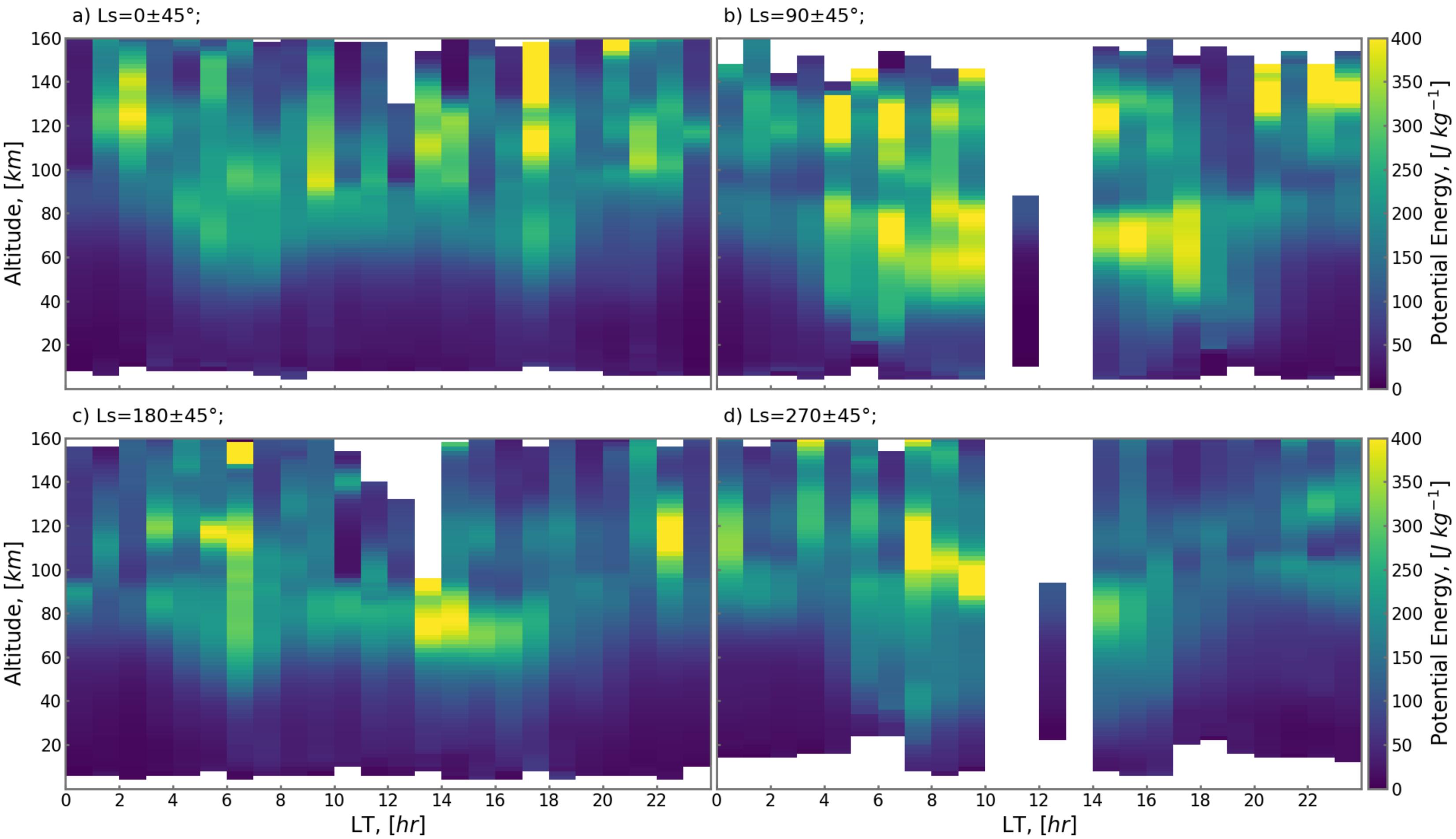


Figure 8.



# Climatology of Gravity Wave Activity From Two Martian Years of ACS/TGO Observations

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## Key Points:

- GWs' parameters for two Martian years from temperature profiles of ACS spectrometers have been obtained
- The seasonally varying distributions of the GWs' drag agree well with predictions of general circulation models
- Observations reveal high-latitude enhancement of GW activity in both hemispheres during the global dust storm

**Abstract**

We report the gravity wave (GW) statistics accumulated over two Martian years from the second half of Martian Year 34 (MY34) to the middle of MY36 (May 2018 - February 2022). The observations were performed by the middle- and near-infrared (MIR and NIR, respectively) spectrometers of Atmospheric Chemistry Suite (ACS) on board ExoMars Trace Gas Orbiter (TGO). Temperature profiles obtained independently of both channels during simultaneous measurements show a good agreement, thus providing verification and additional confidence in the data. GW parameters such as temperature fluctuations, potential energy per unit mass, and wave drag are retrieved at altitudes up to 160 km from the MIR channel and up to 100 km from the NIR channel. We present seasonal, intraday and latitude distributions of the wave potential energy and drag, serving to represent the wave activity and impact on the dynamics. A comparison of data obtained during the global dust storm (GDS) of MY34 with the corresponding period of MY35 without a storm reveals a reduction of GW activity in mid-latitudes in agreement with previous observations, and enhancement in the polar regions of both hemispheres, which was predicted by theoretical studies using simulations with a high-resolution circulation model. Seasonal variations of the derived GW activity can be linked to changes in the solar tide.

**Plain Language Summary**

Gravity waves with horizontal scales of tens and hundreds kilometers are present in atmospheres of all planets. They play an important role in the dynamics and thermodynamics of the middle and upper atmosphere of Mars through transporting the energy and momentum from the lower to the upper layers of atmosphere. The knowledge of their spatio-temporal variability is required for characterization of the atmospheric state and circulation. Two independent channels of the Atmospheric Chemistry Suite instrument on board the ExoMars Trace Gas Orbiter provided thousands of temperature profiles over two Martian years, which were used for retrieving wave characteristics. We present the climatology of gravity wave activity in the form of seasonal, latitudinal and local time distributions. They reveal a strong response of the gravity wave field to the global dust storm occurred in 2018. The derived distributions of the deceleration of the large-scale flow imposed by gravity waves can constrain global circulation models and improve their predictive capabilities.

**1 Introduction**

Gravity (or buoyancy) waves (GWs) originating from the balance of gravity and buoyancy forces are ubiquitous in all convectively stable atmospheres. They have been extensively studied in the terrestrial atmosphere, where their important role in the dynamics and vertical coupling of atmospheric layers has been recognized (e.g., see reviews by Fritts & Alexander, 2003; Yiğit & Medvedev, 2015). First GW-like signature in the atmosphere of Mars was detected in entry measurements of Viking 2 (Seiff & Kirk, 1976). Since then, GWs have been observed on Venus (R. E. Young et al., 1987), Jupiter (L. A. Young et al., 1997), Titan (Hinson & Tyler, 1983), Saturn (Brown et al., 2022) and other planets (see a recent review on GWs in planetary atmospheres by Medvedev & Yiğit, 2019). Recently, various GW-related phenomena in the Martian atmosphere and their impact on the whole atmosphere of Mars system have been reviewed (Yiğit, 2023).

In situ accelerometer measurements in the thermosphere of Mars performed during spacecraft aerobreaking demonstrated large amplitudes of GW-induced density disturbances and the associated wave drag (Keating et al., 1998; Creasey et al., 2006a; Fritts et al., 2006). The omnipresence of GWs in the lower Martian atmosphere and first characterization of the GW field were revealed using remote sensing techniques (Hinson et al., 1999; Creasey et al., 2006b; Wright, 2012; Nakagawa et al., 2020; Heavens et al., 2020).

67 They showed a varying spatio-temporal structure of the GW field and evidence for a lim-  
 68 itation of amplitude growth with height, which is an indication of wave momentum trans-  
 69 fer to the larger-scale flow (Ando et al., 2012).

70 GWs exist at all atmospheric heights. In situ measurements using the Neutral Gas  
 71 and Ion Mass Spectrometer (NGIMS) on board the Mars Atmosphere and Evolution Mis-  
 72 sion (MAVEN) orbiter delivered a large body of GW statistics in the upper thermosphere  
 73 (Yigit et al., 2015; England et al., 2017; Terada et al., 2017; Leelavathi et al., 2020; Rao  
 74 et al., 2021). In particular, the observations found an enhancement of GW activity dur-  
 75 ing the global dust storm (GDS) that occurred in June 2018 (Martian year 34, MY34)  
 76 (Leelavathi et al., 2020; Yigit, Medvedev, Benna, & Jakosky, 2021). Conversely, obser-  
 77 vations with the Mars Climate Sounder instrument on board Mars Reconnaissance Or-  
 78 biter have shown a reduction of such activity in the lower atmosphere (Heavens et al.,  
 79 2020). Another existing controversy concerns the inverse relation between the amplitudes  
 80 of GWs and the background temperature in the upper thermosphere. A number of stud-  
 81 ies attributed it to convective instability that limits wave amplitudes causing the so-called  
 82 “saturation” (England et al., 2017; Terada et al., 2017; Vals et al., 2019), while Yigit,  
 83 Medvedev, and Hartogh (2021) argued that the inverse relation occurs because a colder  
 84 background air reduces the scale height  $H$ , thus facilitating the exponential growth of  
 85 amplitude, which is proportional to  $1/2H$ . Molecular diffusion and thermal conduction  
 86 also exponentially grow with height in response to density decrease. They eventually ex-  
 87 ceed all other damping mechanisms in the thermosphere, and thereby significantly lim-  
 88 its the wave growth.

89 GWs are generated in the lower atmosphere by a variety of mechanisms that ver-  
 90 tically displace air parcels, e.g., flow over topography, convection, weather instabilities,  
 91 etc. While propagating upward, they are partially filtered out by the background mean  
 92 wind. Amplitudes of the surviving harmonics grow with height. Ultimately, the harmon-  
 93 ics reach altitudes, where they are dissipated owing to a combination of nonlinear inter-  
 94 actions, molecular diffusion, and thermal conduction, and deposit their momentum and  
 95 energy to the ambient flow (Yigit et al., 2008). This gravity wave-mean flow interaction  
 96 produces acceleration/deceleration of the large-scale circulation, which is often called “GW  
 97 drag”. Its dynamical importance in the middle and upper atmosphere of Mars has been  
 98 demonstrated with general circulation models (GCMs) where the effects of small-scale  
 99 GWs are either parameterized (Medvedev et al., 2011a, 2011b; Gilli et al., 2020; Yigit  
 100 et al., 2018; Roeten et al., 2022), or explicitly resolved (Kuroda et al., 2015, 2016, 2019).  
 101 These modeling studies have to be validated with observations, and many employed pa-  
 102 rameters constrained. Therefore, an observational characterization of the GW field and  
 103 its spatio-temporal variation at all heights is of great importance.

104 The MIR spectrometer of Atmospheric Chemistry Suite (ACS) experiment on board  
 105 the Trace Gas Orbiter (TGO) (Korablev et al., 2018) allows filling this gap in the knowl-  
 106 edge of atmospheric variability by measuring vertical profiles of density and tempera-  
 107 ture between 20 and 160–180 km in the Martian atmosphere. The algorithm of retriev-  
 108 ing GW profiles and their characteristics along with the first results of its application  
 109 have been presented in detail in the work by Starichenko et al. (2021). The database of  
 110 observations has been significantly extended since then. In this work, we present the re-  
 111 sults on GW activity obtained over the second half of the Martian year 34, the whole  
 112 MY35 and the first half of the MY36. In addition, we analyzed the profiles measured by  
 113 another ACS channel - near-IR (NIR). Although they cover altitudes only up to  $\sim 100$   
 114 km, their number (several thousand) adds significantly to the overall statistics.

115 The structure of this paper is the following. In Section 2, observations and tem-  
 116 perature retrieval procedures from ACS are outlined. The methods of deriving the GW  
 117 characteristics are described in Section 3. Section 4 presents the data coverage. Latitude-  
 118 altitude distributions for four Martian seasons are given in Section 5, the impact of the

119 global dust storm on the GW activity is discussed in Section 6, and local time variations  
 120 are shown in Section 7. Conclusions are presented in Section 8.

## 121 2 Observations and Temperature Retrievals

122 The Atmospheric Chemistry Suite (ACS) is a part of the Trace Gas Orbiter (TGO),  
 123 which represents the ESA-Roscosmos ExoMars 2016 collaborative mission. The instru-  
 124 ment consists of three infrared channels (Korablev et al., 2018): near-IR (NIR, 0.73-1.6  
 125  $\mu\text{m}$ ), middle-IR (MIR, 2.3-4.2  $\mu\text{m}$ ), and thermal-IR (TIRVIM, 1.7-17  $\mu\text{m}$ ). In this work,  
 126 we use the data obtained from the MIR and NIR spectrometric channels operating in  
 127 the solar occultation mode since April 2018. ACS-MIR is a cross-dispersion echelle spec-  
 128 trometer that allows for retrieving temperature and density vertical profiles in the strong  
 129 2.7  $\mu\text{m}$  CO<sub>2</sub> absorption band covering the broad altitude range of 20–180 km (Belyaev  
 130 et al., 2021, 2022). ACS-NIR, an echelle spectrometer combined with an acousto-optic  
 131 tunable filter, measures the atmospheric structure in 1.43  $\mu\text{m}$  and 1.57  $\mu\text{m}$  CO<sub>2</sub> bands  
 132 at altitudes from 10 to 100 km (Fedorova et al., 2020, 2023). Both ACS MIR and NIR  
 133 channels possess a high resolving power exceeding  $\sim 25000$ , the signal-to-noise ratio of  
 134 more than 1000, and sound the atmosphere with the vertical resolution of  $\sim 1$  km. Dur-  
 135 ing simultaneous occultations, the lines of sight (LOS) of both instruments target iden-  
 136 tical tangent points. This provides a confidential cross-validation between the retrieved  
 137 atmospheric profiles. The altitude of the tangent points is determined as the closest dis-  
 138 tance between the instrument’s LOS and the areoid surface of the planet. The atmospheric  
 139 transmission spectrum at each tangential altitude is obtained as a ratio of the solar spec-  
 140 trum transmitted through the atmosphere to the reference one measured at an altitude  
 141 where the absorption at the given CO<sub>2</sub> band is negligible, that is 200 km for the MIR  
 142 case and 130 km for NIR.

143 The procedures of retrieving density and temperature vertical profiles from the trans-  
 144 mission spectra of CO<sub>2</sub> are extensively described in the papers of Belyaev et al. (2021,  
 145 2022) for the MIR channel and of Fedorova et al. (2020, 2023) for the NIR channel. Pro-  
 146 files with estimated  $1\text{-}\sigma$  errors exceeding 20 K have been removed from consideration based  
 147 on expected wave amplitudes. Validation between simultaneously measured MIR and  
 148 NIR profiles demonstrates a good coincidence below 100 km with dispersion of 5-10 K  
 149 in more than 90% of occultations (Belyaev et al., 2022). When retrieving the GW pa-  
 150 rameters (see Section 3), the data were checked for reasonably smooth background tem-  
 151 perature profiles (suitable for extracting waves) and for adequate values of the poten-  
 152 tial energy ( $< 1000 \text{ J kg}^{-1}$ ). Anomalous values could be found either at inflection points  
 153 of temperature profiles, such as around the mesopause, or near the top of the domain  
 154 (80-100 km) for the NIR data. Such cases account for about 10-15% of all occultations,  
 155 and they were excluded from our consideration.

156 In most of simultaneous observations, both the MIR and NIR individual profiles  
 157 (Figures 1, 2a, 2d) along with the retrieved GW parameters (Figures 2b, 2c) closely match  
 158 each other. Nevertheless, in some cases, the evaluated GW amplitudes and potential en-  
 159 ergy somewhat differ between two channels (see Figures 2d, 2e). The reason for that is  
 160 in different altitude domains used for retrieving the NIR and MIR background temper-  
 161 ature. Consequently, the maximum discrepancy occurs near the upper end (80-100 km)  
 162 of the NIR profiles (Figures 2d, 2e). Overall, statistics of occultations at the MIR 2.7  
 163  $\mu\text{m}$  CO<sub>2</sub> band are about 10 times less frequent than those at the NIR CO<sub>2</sub> spectra. Thus,  
 164 in the analyses to be presented, we complement the MIR profiles with those from NIR,  
 165 whenever measurements with MIR are not available.

## 166 3 Gravity Wave Characteristics

167 Separation of the observed temperature profile  $T(z)$  into the background  $\bar{T}$  and the  
 168 GW-induced disturbance  $T' = T - \bar{T}$  is an ambiguous procedure, because no unique