

Water vapor vertical profiles on Mars in dust storms observed by TGO/NOMAD

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Key Points:

- We present vertical profiles of water vapor in the Martian atmosphere during global and regional dust storms in 2018-2019.
- We show a rapid and significant increase of water vapor in the middle atmosphere (40-100 km) during both global and regional dust storms.
- Water vapor reaches very high altitudes, at least around 100 km, during the global dust storm.

29 **Abstract**

30 It has been suggested that dust storms efficiently transport water vapor from the near-surface to
31 the middle atmosphere on Mars. Knowledge of the water vapor vertical profile during dust
32 storms is important to understand water escape. During Martian Year 34, two dust storms
33 occurred on Mars: a global dust storm (June to mid-September 2018) and a regional storm
34 (January 2019). Here we present water vapor vertical profiles in the periods of the two dust
35 storms ($L_s=162\text{--}260^\circ$ and $L_s=298\text{--}345^\circ$) from the solar occultation measurements by Nadir and
36 Occultation for Mars Discovery (NOMAD) onboard ExoMars Trace Gas Orbiter (TGO). We
37 show a significant increase of water vapor abundance in the middle atmosphere (40–100 km)
38 during the global dust storm. The water enhancement rapidly occurs following the onset of the
39 storm ($L_s\sim 190^\circ$) and has a peak at the most active period ($L_s\sim 200^\circ$). Water vapor reaches very
40 high altitudes (up to 100 km) with a volume mixing ratio of ~ 50 ppm. The water vapor
41 abundance in the middle atmosphere shows high values consistently at $60^\circ\text{S}\text{--}60^\circ\text{N}$ at the growth
42 phase of the dust storm ($L_s=195\text{--}220^\circ$), and peaks at latitudes greater than 60°S at the decay
43 phase ($L_s=220\text{--}260^\circ$). This is explained by the seasonal change of meridional circulation: from
44 equinoctial Hadley circulation (two cells) to the solstitial one (a single pole-to-pole cell). We also
45 find a conspicuous increase of water vapor density in the middle atmosphere at the period of the
46 regional dust storm ($L_s=322\text{--}327^\circ$), in particular at latitudes greater than 60°S .

47

48 **Plain Language Summary**

49 The most striking phenomenon on Mars is a planet-encircling storm, "global dust storm". Once it
50 starts, the floating dust covers the whole atmosphere for more than several weeks. Recent studies
51 suggest that dust storms effectively transport water vapor from the near-surface to the middle
52 atmosphere. In June-September 2018 and January 2019, a strong global dust storm and a regional
53 storm occurred on Mars, respectively. This study investigates altitude profiles of water vapor in
54 the Mars atmosphere measured during the dust storms, by using brand-new measurements by
55 Nadir and Occultation for Mars Discovery (NOMAD) onboard the ExoMars Trace Gas Orbiter
56 (TGO). We confirm that the water vapor expanded into the middle atmosphere and we find that
57 the water vapor reached very high altitudes (up to 100 km) during the dust storms. The dust
58 storms intensify the atmospheric dynamics and heat the atmosphere. As a result, water vapor is
59 lifted to higher altitudes and distributes along the meridional circulation.

60

61 **1. Introduction**

62 Recent observations and studies have revised our understanding of water loss processes
63 on Mars. The variations of the escape rate are not dominated by the solar EUV radiation flux but
64 rather by the variable water vapor abundance in the middle atmosphere (Chaffin et al., 2014,
65 2017; Clarke et al., 2014; Fedorova et al., 2018; Heavens et al., 2018; Clarke 2018). Interestingly,
66 these recent measurements imply that global dust storms may effectively transport water vapor
67 from the near surface to the middle atmosphere and hence increase the escape rate with respect to
68 the atmospheric water loss under no-storm conditions. Heavens et al. (2018) and Fedorova et al.
69 (2018) showed the vertical profile of water vapor before/during/after the global dust storm in
70 2007. They found a significant increase of water vapor abundance in the middle atmosphere and
71 an increase in altitude of the hygropause (where the water content rapidly decreases following

72 saturation and ice cloud formation). Fedorova et al. (2018) found that the water vapor
73 enhancement is asymmetric between the northern and southern hemispheres – the increase of the
74 water vapor abundance due to the global dust storm is remarkable only in the northern
75 hemisphere. This suggests that meridional circulation of the atmosphere is intensified during the
76 dust storms and transports water vapor more efficiently from the southern to the northern
77 hemisphere. Moreover, it is suggested that dust storm related increases of atmospheric
78 temperatures suppress the hygropause, hence reducing ice cloud formation and so allowing water
79 vapor to extend into the middle atmosphere (Heavens et al., 2018; Neary et al., 2019).

80 To date, vertical profiles of water vapor have been investigated by the solar occultation
81 measurements with Spectroscopy for Investigation of Characteristics of the Atmosphere of Mars
82 (SPICAM) onboard Mars Express (MEX) (Maltagliati et al., 2011; Maltagliati et al., 2013;
83 Fedorova et al., 2018) and by the limb measurements with Compact Reconnaissance Imaging
84 Spectral Mapper (CRISM) and Mars Climate Sounder (MCS) onboard the Mars Reconnaissance
85 Orbiter (MRO) onboard Mars Reconnaissance Orbiter (MRO) (Clancy et al., 2017; Heavens et
86 al., 2018). The measurements by SPICAM and MCS revealed unexpected high abundance of
87 water in the middle atmosphere that contributes the atmospheric escape of water. They also
88 found that the water vapor abundances in the middle atmosphere are further increased during the
89 global dust storm that occurred in 2007. However, even though it is proposed that an intensified
90 meridional circulation may transport water vapor efficiently, the complete picture of the water
91 vapor distribution during global dust storms is not yet confirmed. This is because (1) the
92 MEX/SPICAM measurements can be performed only in a limited period because the MEX orbit
93 is not dedicated for solar occultation observations, for instance its measurements during the
94 perihelion season in MY 28 (the year of the previous global dust storm) were performed only
95 during $L_s = 255\text{-}300^\circ$ and no observation is available during $L_s = 200^\circ\text{-}255^\circ$ and $300^\circ\text{-}360^\circ$
96 (Fedorova et al., 2018); (2) the previous MRO/MCS analysis (Heavens et al., 2018) did not
97 directly retrieve water vapor vertical profiles from the water vapor spectral features: water vapor
98 abundances were rather indirectly estimated from the retrieved temperature, pressure, dust, and
99 water ice; (3) the 2009-2016 MRO/CRISM water vapor profiles, derived from O_2 dayglow
100 profiles, did not encounter global dust storm conditions. In order to understand the mechanism of
101 the water vapor transport from the near-surface to the middle atmosphere, it is crucial to
102 investigate the latitudinal, longitudinal, and temporal variation of water vapor vertical profiles
103 before/during/after global dust storms. Solar occultation measurements by two new
104 spectrometers onboard TGO - NOMAD (Vandaele et al., 2018) and Atmospheric Chemistry
105 Suite (ACS) (Korablev et al., 2018) - are now able to monitor the water vapor vertical profiles
106 through the whole a Martian Year and obtain a latitudinal map for every $\sim 20^\circ$ of L_s since the
107 orbit of TGO is optimized for solar occultation measurements, producing 24 occultations per day
108 at a maximum (average 5-7 observations per day).

109 In 2018, for the first time after 2007, a global dust storm occurred on Mars. It lasted for
110 more than two months (from June to August). Moreover, following the global dust storm, a
111 regional dust storm occurred in January 2019. TGO began its science operations on 21 April
112 2018. The NOMAD and ACS observations therefore fully cover the period before/during/after
113 the global and regional dust storms and offer a unique opportunity to study the trace gases
114 distributions during the dust storms. In this paper, we present water vapor vertical profiles from
115 April to September 2018 and from December 2018 to February 2019 retrieved from the
116 NOMAD measurements. We have analyzed those datasets and published two water vapor
117 vertical profiles as early results - one before the global dust storm and the other one during the

118 storm, which present a significantly conspicuous increase of water vapor during the global dust
119 storm (Vandaele et al., 2019). This study presents the results of the extended datasets at the
120 period of the dust storms. The details of the NOMAD observations and the data analysis are
121 described in Sections 2 and 3, respectively. The observational results are discussed in Sections 4
122 and 5 for the global dust storm and regional dust storm, respectively. A full GCM simulation is
123 presented in an accompanying paper (Neary et al., 2019).

124 **2. Observations: NOMAD Solar occultation**

125 **2. 1. Instrument - NOMAD onboard TGO**

126 NOMAD is a spectrometer operating in the spectral ranges between 0.2 and 4.3 μm
127 onboard ExoMars TGO. NOMAD has 3 spectral channels: a solar occultation channel (SO –
128 Solar Occultation; 2.3–4.3 μm), a second infrared channel capable of nadir, solar occultation, and
129 limb sounding (LNO – Limb Nadir and solar Occultation; 2.3–3.8 μm), and an ultraviolet/visible
130 channel (UVIS – Ultraviolet and Visible Spectrometer, 200–650 nm). The infrared channels (SO
131 and LNO) have high spectral resolution ($\lambda/d\lambda \sim 10,000\text{--}20,000$) provided by an echelle grating
132 used in combination with an Acousto Optic Tunable Filter (AOTF) which selects diffraction
133 orders (Neefs et al., 2015). The concept of the infrared channels are derived from the Solar
134 Occultation in the IR (SOIR) instrument (Nevejans et al., 2006) onboard Venus Express (VEx).
135 The sampling rate for the solar occultation measurement is 1 second, which provides better
136 vertical sampling step (~ 1 km) with higher resolution (~ 2 km) from the surface to 200 km.
137 Thanks to the instantaneous change of the observing diffraction orders achieved by the AOTF,
138 the SO channel is able to measure five or six different diffraction orders per second in solar
139 occultation mode. One of the most remarkable capabilities of NOMAD is its high spectral
140 resolution in the near infrared range. It allows us (1) to investigate vertical profiles of the
141 atmospheric constituents (such as carbon dioxide, carbon monoxide, water vapor, and their
142 isotopic ratio), and (2) to perform sensitive search of organic species (such as CH_4 , C_2H_4 , C_2H_6 ,
143 H_2CO) and other trace gases (such as HCl , HCN , HO_2 , H_2S , N_2O , OCS) by solar occultation
144 measurements with the SO channel.

145 **2. 2. Dataset**

146 In this study, we analyze the solar occultation measurements acquired by the NOMAD
147 SO channel during the period from 21 April to 30 September 2018 (corresponding to $L_s = 162\text{--}260^\circ$
148 in Martian Year (MY) 34) and from 1 December 2018 to 23 February 2019 ($L_s = 298\text{--}345^\circ$
149 in MY 34). Measurements of diffraction order 134 ($3011\text{--}3035\text{ cm}^{-1}$) and 168 ($3775\text{--}3805\text{ cm}^{-1}$)
150 are analyzed. Observations of these orders have been regularly conducted since they include
151 strong H_2O bands (both in orders 134 and 168) and CH_4 Q-branch (in order 134). In the period
152 above, a total of 987 occultations that operated with both diffraction orders 134 and 168 were
153 acquired. **Figure 1** shows the latitudinal coverage of these occultations. As shown in this figure,
154 the latitude shifts from orbit to orbit, at the same time, the longitudinal coverage is dispersed
155 over the whole planet. It demonstrates that this dataset allows us to obtain a latitudinal map for
156 every $\sim 20^\circ$ of L_s . The local solar time is generally around 6 AM or 6 PM outside of the polar
157 regions. Note that the gaps in data between $L_s = 171\text{--}179^\circ$, $201\text{--}210^\circ$, and $327\text{--}333^\circ$ are due to
158 the orbital geometry, which prevents solar occultations from being measured when the orbital
159 nadir track is close to the terminator and the Sun is never occulted by Mars.

160 2. 3. Data reduction

161 Solar occultation is a powerful technique for investigating vertical structure of
 162 atmospheres. It observes a strong light source - the Sun - through the atmosphere from very high
 163 altitude (typically from 250 km altitudes for NOMAD) down to the surface. The absolute
 164 calibration of solar occultation spectra is relatively easy since transmittances are obtained by
 165 dividing the spectra measured through the atmosphere by the reference solar spectrum recorded
 166 outside the atmosphere, which basically removes systematic instrumental effects (except small
 167 changes occurring during the occultation). For the calculation of transmittances, we employ the
 168 algorithm developed for the reduction of the SOIR data (Trompet et al., 2016). This algorithm
 169 does not simply average the spectra recorded outside the atmosphere but calculates its linear
 170 regression with altitude for each pixel. The algorithm first calculates the pixel-by-pixel linear
 171 regressions with the solar spectra between 150 and 250 km, it then applies them to the spectra
 172 between 120 and 150 km where no absorption due to Mars atmosphere is expected. If more than
 173 80 % of the transmittances at 120-150 km are equal to one sigma, the algorithm applies the linear
 174 regression to the spectra recorded through the Mars atmosphere (below 120km) and calculates
 175 the transmittances. If not, the algorithm re-calculates the linear regression by excluding the solar
 176 spectrum at the highest altitude. This iterative process continues until accepted. Such a pixel-by-
 177 pixel linear extrapolation allows us to reduce the residual instrumental systematic due to small
 178 changes during an occultation (such as small deviation of the center wavenumber of the AOTF
 179 transfer function, tiny spectral shift due to grating movement/expansion/contraction because of
 180 instrumental oscillations). The instrumental noise is given by

$$181 \quad (I/F)_{err} = \frac{I}{F} \sqrt{\left(\frac{I_{err}}{I}\right)^2 + \left(\frac{F_{err}}{F}\right)^2}$$

182 where I is the spectrum through the atmosphere, F is the reference solar spectrum, and I/F is thus
 183 the transmittance. The noise in the reference solar spectrum F_{err} is given by the standard
 184 deviation of the solar spectra (normalized) that are used to create the reference solar spectrum.
 185 The noise in the spectrum through the atmosphere I_{err} is given by the sum of dark noise (we use
 186 the standard deviation of the signal recorded when the radiation reaching the detector is similar
 187 to the noise level when the planet is in the Line-Of-Sight (LOS)) and estimated noise from the
 188 one in the reference solar spectrum (Vandaele et al., 2013). Typical Signal to Noise Ratio (SNR)
 189 for a single spectrum recorded in diffraction orders 134 and 168 is 1500-2500. **Figure 2** shows
 190 an example of calculated transmittances of the diffraction order 168 during an occultation. When
 191 the LOS to the Sun transects the atmosphere, the slant optical depth along the LOS gradually
 192 increases owing to the presence of aerosols and molecules, until the atmosphere becomes
 193 completely opaque at some tangent altitude. For the particular example in **Figure 2**, the
 194 transmittance drops to zero around 5 km. It usually occurs because of the large amount of dust in
 195 the lowermost part of the atmosphere.

196 2. 4. Calibration

197 Calibration of the NOMAD infrared channels from the first in-flight data was
 198 summarized in Liuzzi et al. (2019) and has been improved since then. Based on the work by
 199 Liuzzi et al. (2019), in this analysis, the AOTF transfer function of the NOMAD SO channel is
 200 characterized as a sinc square function whose side lobes are multiplied by 1.3 and 1.8 for
 201 diffraction order 134 and 168, respectively. The instrumental line shape is assumed to be a
 202 Gaussian function with the full width at half maximum (FWHM) of 0.228 cm^{-1} ($\lambda/\Delta\lambda \sim 13250$) for

203 diffraction order 134, and 0.338 cm^{-1} ($\lambda/\Delta\lambda \sim 11220$) for diffraction order 168. The spectral
204 calibration is first performed based on the results in Liuzzi et al. (2019) and then refined using
205 the solar lines at 3014.960 cm^{-1} (order 134) and 3878.865 cm^{-1} (order 168) for each spectrum in
206 the retrieval process.

207 **3. Data analysis: retrievals of H₂O vertical profiles**

208 **3. 1. Forward model**

209 For the forward calculation of the simulated spectra and inversion of the H₂O abundance,
210 we employ the ASIMUT-ALVL radiative code developed at Royal Belgian Institute for Space
211 Aeronomy (BIRA-IASB) (Vandaele et al., 2006). The code solves the radiative transfer equation
212 for nadir or solar occultation geometries. It was originally developed for the Earth atmosphere
213 and then extended for Venus (e.g., Vandaele et al., 2008) and Mars (e.g., Vandaele et al., 2019)
214 atmospheres. The code has been widely used in the data analysis of the solar occultation
215 measurements by VEx/SOIR.

216 The radiative transfer calculation is performed in the spectral ranges between ± 4
217 diffraction orders from the main order to properly model the contributions from adjacent orders
218 (Vandaele et al., 2008). CO₂ and H₂O molecules absorption are taken into account in the
219 calculation. The absorption coefficients of CO₂ and H₂O molecules are calculated based on the
220 line-by-line method by using the following spectroscopic database: HITRAN 2016 database
221 (Gordon et al. 2016) for CO₂ and the water line list for CO₂-rich atmospheres by Gamache et al.
222 (2016) for H₂O. A Voigt function is adopted for the line shape function.

223 The observing geometry is calculated based on the SPICE kernel of the TGO orbits.
224 Based on the geometry (i.e., latitude, longitude, solar longitude, and the local solar time at the
225 tangential point), we extract vertical profiles of the temperature, pressure, and CO₂ volume
226 mixing ratio predicted by the general circulation model (GCM), Global Environmental
227 Multiscale Mars model (GEM-Mars) (Neary et al., 2019; Daerden et al., 2019; Neary and
228 Daerden, 2018). This is performed for each spectrum corresponding to the probed tangent
229 altitude. The modeled atmosphere starts from the tangent altitude of the measurement and up to
230 150 km altitude with a uniform thickness of 5 km. Note that the tangent altitude is calculated as
231 the shortest distance between the LOS of the center of the field of view and the MGM1025
232 areoid (i.e., the Mars geoid) (Lemoine et al., 2001). In addition, we note that the atmospheric
233 state predicted by GEM-Mars takes into account the effects of the dust storms in MY34 (Neary
234 et al., 2019). Neary et al. (2019) described multiple dust storm simulations. We used the run
235 where dust is scaled to the MY34 climatology and the vertical profile was determined using an
236 adjusted Conrath profile (Conrath parameter=0.0008; “GDS0008” simulation).

237 **3. 2. Retrievals**

238 We performed the retrieval of H₂O abundance for each spectrum at each tangential
239 altitude independently (Vandaele et al., 2019), i.e., using the classical "onion peeling" method.
240 The retrievals start at the top of atmosphere (120 km altitude for this study). At the highest
241 altitude, the initial guess for the H₂O volume mixing ratio is set a small value (typically 1 ppm).
242 Once H₂O absorption features are detected, the retrieved value is used as the initial guess for the
243 next layer below. Since the slant optical depth integrated along the LOS is relatively large for
244 solar occultation measurements comparing to nadir observations, strong H₂O lines (line intensity

245 S larger than $\sim 10^{-20}$ $\text{cm}^{-1}/(\text{molecule cm}^{-2})$ are easily saturated. In our retrieval scheme, if lines
 246 are saturated (i.e., the total optical depth at infinite resolution is greater than 1), their weights are
 247 reduced as follows:

$$248 \quad \sigma'(v) = \sigma(v) \times 10^{\tau(v)}$$

249 where $\sigma(v)$ is the instrumental noise, $\sigma'(v)$ is the de-weighted error used in the retrievals, and $\tau(v)$
 250 is the total slant optical depth integrated along the LOS. De-weighting pixels that contain lines
 251 whose optical depth at line center before considering the instrument exceeds 1 would be a strong
 252 constraint, however, it is useful to reduce possible biases and uncertainties associated with
 253 imperfect knowledge of instrumental functions.

254 The retrievals are performed using the Optimal Estimation Method (OEM) (Rodgers,
 255 2000) implemented in a Gauss-Newton iterative scheme. The spectral ranges for the retrieval are
 256 the full range of diffraction order 134 (3011–3035 cm^{-1}) and a confined spectral range (3783–
 257 3803 cm^{-1}) for diffraction order 168, where more than 50 % of the recorded signal originate from
 258 the main diffraction order. This is because the side-lobes in the AOTF transfer function are
 259 relatively large for diffraction order 168. Since the side-lobes are not yet perfectly characterized,
 260 limiting the retrieval to the spectral region where the majority of signal comes from the main
 261 order reduces the uncertainties associated with imperfect knowledge of the AOTF function. The
 262 free parameters in the retrievals are a factor multiplying the initial guess of H_2O density, and the
 263 parameters of the 5th order polynomial function used to model the continuum of each spectrum.
 264 Note that most of the information comes from the sounded tangent altitude (about 70% of the
 265 slant number density integrated over the LOS is within 4 km from the tangent height), thus the
 266 retrieved local H_2O abundances at the tangential altitudes of the measurements can be considered
 267 as its vertical profiles. The continuum established by the polynomial function removes the effect
 268 of extinctions due to the presence of aerosols (dust and water ice clouds) along the LOS as well
 269 as instrumental features caused by physical changes of the instrument during the occultation
 270 (such as small deviation of the center wavenumber of the AOTF transfer function). **Figure 3**
 271 shows examples of the fitting results. The intensity of the H_2O lines in the diffraction order 168
 272 is about 100 times stronger ($S \sim 10^{-19}$ $\text{cm}^{-1}/(\text{molecule cm}^{-2})$) than those in the diffraction order
 273 134 ($S \sim 10^{-21}$ $\text{cm}^{-1}/(\text{molecule cm}^{-2})$), which allows us to investigate water vapor abundance
 274 from the near-surface to high altitudes. As shown **Figure 3b**, we firmly detect H_2O with a good
 275 SNR up to at least 100 km during the global dust storm. The retrievals are conducted for each
 276 diffraction order (i.e., 134 and 168) independently in order to evaluate the consistency between
 277 orders. Lastly the vertical profiles of H_2O abundances and their errors are calculated from the
 278 weighted averages:

$$279 \quad W_{134} = \left(\frac{1}{H_2O_{134err}} \right)^2, W_{168} = \left(\frac{1}{H_2O_{168err}} \right)^2$$

$$H_2O = \frac{W_{134}H_2O_{134} + W_{168}H_2O_{168}}{W_{134} + W_{168}},$$

$$H_2O_{err} = \sqrt{\frac{1}{W_{134} + W_{168}}},$$

280 where H_2O_{134} , H_2O_{168} , H_2O_{134err} and H_2O_{168err} are retrieved H_2O volume mixing ratio and
 281 their error values from the diffraction order 134 and 168, respectively.

282 3. 3. Uncertainties

283 There are three sources of error in the retrieved H₂O volume mixing ratio: (1)
284 instrumental noise on the measured spectra, (2) uncertainty in the vertical temperature and
285 atmospheric density profile used in the radiative transfer calculation, and (3) uncertainty of
286 imperfect knowledge of the NOMAD instrumental functions.

287 The first error is directly derived from the covariance matrix of the optimal fit parameters
288 in the retrievals. If both diffraction orders 168 and 134 are available, the error values obtained
289 from each diffraction order are weighted averaged, as described above. The median value of this
290 error in the retrievals is about 5 %.

291 The second error can be evaluated by retrievals with temperature profiles shifted by their
292 accuracy. We estimate that the accuracy of the GEM-Mars temperature predictions is about ± 10
293 K. The difference between the water vapor profiles retrieved with the temperature profiles
294 uniformly shifted by ± 10 K and those with retrieved with the original temperature profiles are
295 about 5-8 %, which can then be considered as the error in the retrieved H₂O volume mixing ratio
296 due to uncertainty in the GEM-Mars temperature. Moreover, since the H₂O volume mixing ratio
297 is the ratio between the water vapor number density and the total atmospheric number density
298 that is based on the predictions by GEM-Mars in this study, the uncertainty in the GEM-Mars
299 total atmospheric number density directly affects the retrieved H₂O volume mixing ratio. We
300 estimate that the accuracy of the atmospheric total number density by GEM-Mars is about 10-15
301 %, thus we have another 10-15 % of error in the retrieved H₂O volume mixing ratio.

302 The third error can be roughly estimated by comparing the retrieved H₂O abundances in
303 each diffraction order (168 and 134). **Figure 4** shows examples of the retrieved vertical profiles
304 of the H₂O volume mixing ratio from diffraction orders 134 and 168. The retrieved profiles from
305 each diffraction order are predominantly consistent within 3-sigma (as shown in Fig. 4a),
306 however, the retrieved values are sometimes inconsistent beyond 3-sigma of the standard
307 retrieval error (as shown in Fig. 4b). The median value of the difference between order 168 and
308 134 retrievals is about twice larger than 1-sigma of the standard retrieval error, and inconsistent
309 results above 3-sigma occur in around 35% of occultations. We consider this is mainly due to the
310 fact that instrumental characterization such as the AOTF transfer function is not fully achieved,
311 however this does not invalidate the conclusions of this study.

312 4. The global dust storm in 2018

313 4. 1. Overview of the dust storm

314 The global picture of the 2018 global dust storm was summarized in Guzewich et al.
315 (2019) and references therein. The storm started regionally at the middle/end of May ($L_s \sim 180$ -
316 185°) across Acidalia Planitia (30 - 60°N , 300 - 360°E) and Utopia Planitia (30 - 60°N , 80 - 140°E).
317 Then, it merged with substantial dust lifting occurring independently in the southern hemisphere
318 by the beginning of June ($L_s \sim 190^\circ$), and expanded globally by the middle of June ($L_s \sim 195^\circ$).
319 This most active period of the global dust storm lasted until the beginning of July ($L_s \sim 205^\circ$),
320 and started a long decay phase that ended at the middle of September ($L_s \sim 250^\circ$). As shown in
321 **Figure 1a**, the measurements by the NOMAD-SO channels cover the whole period of the global
322 dust storm except for some gaps due to the geometry of the spacecraft's orbit, including the most
323 active period of the dust storm ($L_s = 202$ - 210°). The colors in **Figure 1a** denote the highest
324 tangent altitude at which the slant-optical depth along the LOS is less than 1, which is basically

325 representative of the top altitude of the most opaque dust region in the atmosphere. The dust top
 326 altitudes reaches ~ 50 km in the seasonal range between $L_s = 197$ - 202° , which corresponds to the
 327 most active period of the global dust storm.

328 **4. 2. Seasonal variation of the water vapor vertical profiles**

329 **Figure 5a-b** shows the seasonal variation of the water vapor vertical profiles at $L_s=162$ –
 330 260° retrieved from the NOMAD measurements taken in the northern hemisphere (Fig. 5a) and
 331 the southern hemisphere (Fig. 5b). The top panels show the latitude and local time of the
 332 measurements as references. It would be interesting to explore the changes in the water vapor
 333 vertical distribution during the 2018 dust storm and for the same period in a non-dust storm year.
 334 However, as the 2018 storm occurred in the first year of operations of NOMAD, we have no
 335 direct self-consistent reference for the non-dust storm conditions. Previous directly and indirectly
 336 retrieved water vapor profiles (see introduction) are sparse and the match in space and time with
 337 the NOMAD profiles is in general poor. Therefore, we prefer to use a GCM as tentative
 338 reference for the non-dust storm water vapor distribution, as it provides a complete coverage for
 339 all times and allows for a full interpolation of the model output to the location and time of all the
 340 NOMAD profiles. The vertical water vapor distribution simulated in the GEM-Mars GCM was
 341 evaluated by both the water vapor vertical profiles and total water columns retrieved from
 342 CRISM in Daerden et al. (2019). **Figure 5c-d** shows the same water maps as Fig. 5a-b but
 343 simulated by GEM-Mars for non-dust storm conditions (Daerden et al., 2019). **Figure 5e-f**
 344 presents the differences between the measured water vapor vertical profiles (Fig. 5a-b) and the
 345 predictions by GEM-Mars for non-dust storm conditions (Fig. 5c-d).

346 The abundance of water vapor in the middle atmosphere suddenly increased around $L_s =$
 347 190° in both hemispheres. This is not seen in the GCM data for non-dust storm conditions (Fig.
 348 5c-d). In contrast, around $L_s = 210^\circ$, the water vapor abundance in the lower atmosphere seems
 349 to have decreased with respect to non-dust storm conditions. However, the retrieval accuracy and
 350 data coverage is poorer at low altitudes. Also, a comparison of the GCM water vapor profiles
 351 with those derived from CRISM (Clancy et al., 2017; Daerden et al., 2019) in non-dust storm
 352 conditions shows that the GCM water vapor is too abundant at low latitudes. Therefore, it
 353 remains hard to estimate the effect of the dust storm on the behavior of water vapor in the lower
 354 atmosphere using the current GCM results. GCM simulations for the MY34 dust storm that
 355 reproduce the observed profiles reasonably well are presented in the accompanying paper Neary
 356 et al. (2019) and allow for a theoretical assessment of the impact of the dust storm.

357 The variation of the water vapor abundances occurs very rapidly: the water abundance in
 358 the middle atmosphere in both hemispheres increases by an order of magnitude in just a few days
 359 (around $L_s = 195^\circ$). Since the timing of this phenomenon corresponds to the onset of the global
 360 dust storm and this is not predicted by the GCM for non-dust storm conditions, we conclude that
 361 the rapid enhancement of the water vapor in the middle atmospheres is due to the effects of the
 362 global dust storm. The water abundances in the middle atmosphere have maximum values
 363 around $L_s = 200^\circ$. At that period, we find large water vapor abundances in the middle
 364 atmosphere, exceeding 200 ppm. Such large water vapor abundances are similar to
 365 MEx/SPICAM observation during the global dust storm in MY 28 (Fedorova et al., 2018).
 366 Moreover, we detect water vapor at very high altitude, reaching at 100 km with a volume mixing
 367 ratio of ~ 50 ppm. After these peaks, the enhanced water vapor in the middle atmosphere
 368 gradually returned to the typical climatological levels. A small local maximum in water vapor

369 also appears around $L_s = 235\text{--}240^\circ$ in the northern hemisphere (Fig. 5a). This is due to the fact
 370 that the measurements are performed at equatorial region where more water vapor is present.
 371 While in contrast, the water vapor abundances in the southern hemisphere have a small local
 372 maximum around $L_s = 230^\circ$, which can be explained by the fact that the water vapor in the
 373 southern hemisphere does not have a maximum at equatorial but at high latitude (see Section 4.3.
 374 in detail). Finally, we note that it is difficult to distinguish local time variation from the seasonal
 375 one since either sunrise or sunset measurements last for $10\text{--}20^\circ$ of L_s as shown in the top panel of
 376 Fig. 5a-b.

377 **4. 3. Latitudinal variation of the water vapor vertical profiles**

378 **Figure 6** shows the seasonal variation of latitudinal maps of the water vapor vertical
 379 profiles observed by NOMAD (the top panel) and predicted by GEM-Mars for non-dust storm
 380 conditions (the bottom panel) during $L_s = 160\text{--}195^\circ$ (before the global dust storm, Fig. 6a), $L_s =$
 381 $195\text{--}202^\circ$ (growth phase of the dust storm, Fig. 6b), $L_s = 210\text{--}220^\circ$ (mature phase of the dust
 382 storm, Fig. 6c), $L_s = 220\text{--}240^\circ$, and $L_s = 240\text{--}260^\circ$ (decay phase of the dust storm, Fig. 6d and
 383 6e). The latitudinal distributions of the water vapor vertical profiles before and after the onset of
 384 the global dust storm are quite different (Fig. 6a, b). After the onset of the global dust storm, the
 385 water vapor abundance in the middle atmosphere is significantly and consistently increased at
 386 $60^\circ\text{S}\text{--}60^\circ\text{N}$. At the mature phase of the dust storm (Fig. 6c), the water vapor abundance is
 387 decreased compared to the growth phase, however the latitudinal distribution of the water vapor
 388 is similar. At the growth/mature phase of the dust storm, the season on Mars is just following the
 389 northern autumn equinox and it is expected that the Hadley circulation (the equatorial cell)
 390 extends to 60° latitude in both hemispheres (e.g., Forget et al., 1999; Takahashi et al., 2003). This
 391 explains that the large water vapor abundance is confined between $60^\circ\text{S}\text{--}60^\circ\text{N}$. Under non-dust
 392 storm conditions, water vapor abundances decrease rapidly by ~ 40 km altitude due to saturation
 393 conditions at 30-50 km. Our results suggest that during the dust storm water vapor was
 394 transported from the lower to the middle atmosphere by the intensified circulation and expanded
 395 above 40 km due to the heating of the middle atmosphere by dust absorption that prevents water
 396 vapor from condensation (Neary et al., 2019). In addition, it has been proposed that strong
 397 convective transport processes at the mesoscale can create high altitude dust layers (Spiga et al.,
 398 2013; Daerden et al., 2015; Wang et al., 2018; Heavens et al., 2018). Such processes will also
 399 transport water vapor and may also contribute to the formation of local high altitude water vapor
 400 maxima. At the decay phase of the dust storm (Fig. 6d, 6e), the water vapor abundance is similar
 401 to that at the mature phase, however the latitudinal distribution has a gradient with a maximum
 402 value at latitudes greater than 60°S . It is well known that the symmetry between two meridional
 403 "equinox" cells is significantly reduced at $L_s = 220^\circ$, as the formation of a single pole-to-pole
 404 "solstice" Hadley circulation develops (e.g., Forget et al., 1999; Takahashi et al., 2003). The local
 405 maximum observed at latitudes greater than 60°S may be explained by a new theoretical study
 406 performed by Shaposhnikov et al. (2019). Although their study focused on the MY 28 global dust
 407 storm which occurred later in the season ($L_s = 250^\circ\text{--}270^\circ$), the authors suggest that transport of
 408 water vapor to the upper atmosphere by the strong upward branch of the meridional circulation at
 409 perihelion occur only at latitudes greater than 60°S , which corresponds to the location of the
 410 peak of water vapor abundances observed by NOMAD.

411 5. The regional dust storm in 2019

412 5. 1. Overview of the dust storm

413 A 2019 regional dust storm started around 7 January ($L_s \sim 320^\circ$) in the southern
 414 hemisphere, peaked around 15 January, and declined into the middle of February ($L_s \sim 340^\circ$)
 415 (Chaffin et al., 2019). This seasonal dust event occurs with significantly variable amplitude in
 416 every Mars year, and can present vertically deep increases in atmospheric temperatures and dust
 417 over low-to-mid latitudes (e.g., Kass et al., 2016), as it did in 2019. As shown in **Figure 1b**, the
 418 measurements by the NOMAD-SO channel cover the whole period of the dust storm except for a
 419 gap due to the geometry of the spacecraft's orbit at the active period of this regional dust storm
 420 ($L_s = 327\text{--}333^\circ$). The dust top altitude reaches ~ 50 km in the seasonal range between $L_s = 322\text{--}$
 421 327° in the southern hemisphere, which corresponds to the period and the region of the dust
 422 storm (reddish colors in Fig. 1b).

423 5. 2. Seasonal variation of the water vapor vertical profiles

424 **Figure 7a-b** shows the seasonal variation of the water vapor vertical profiles at $L_s=298\text{--}$
 425 345° retrieved from the NOMAD measurements taken in the northern hemisphere (Fig. 7a) and
 426 southern hemisphere (Fig. 7b). **Figure 7c-d** shows the same water maps as Fig. 7a-b but
 427 simulated by GEM-Mars for non-dust storm conditions (Daerden et al., 2019). **Figure 7e-f**
 428 presents the differences between the measured water vapor vertical profiles (Fig. 7a-b) and the
 429 predictions by GEM-Mars for non-dust storm conditions (Fig. 7c-d). In the time before the storm
 430 ($L_s \sim 300^\circ$ in the north and $L_s \sim 305\text{--}320^\circ$ in the south) and also at $L_s \sim 240\text{--}260^\circ$ in the southern
 431 hemisphere (Fig. 5b), the GCM water mixing ratios between 10 to 40 km are considerably higher
 432 (by a factor ~ 2) than those measured by NOMAD. However total water columns from the GCM
 433 match very well with observations from TES and CRISM (Neary and Daerden, 2018; Smith et
 434 al., 2018; Daerden et al., 2019). This suggests that water vapor was much more confined to the
 435 lowest scale height than currently predicted by models. The water vapor abundances in the
 436 middle atmosphere suddenly increased around $L_s = 321^\circ$ in both hemispheres and reached
 437 maximum values around $L_s = 325^\circ$. The enhancement lasted at least until $L_s = 327^\circ$, before a
 438 data gap due to orbital geometry. The GCM data also show a distinct increase of water vapor
 439 abundances up to 40 km in this period. However, the increase of water vapor seen in the GCM is
 440 due to the latitude of the measurements (low latitudes) and does not expand above 40 km
 441 altitude. Since the timing of the enhancement of the water vapor in the middle atmosphere seen
 442 by NOMAD corresponds to the period of the regional dust storm and since it expands into the
 443 whole middle atmosphere, it is reasonable to attribute that enhancement to the regional dust
 444 storm. At the period of this storm, water vapor abundances in the middle atmosphere exceed 150
 445 ppm and water vapor is present up to (at least) 90 km. The water vapor abundances in the middle
 446 atmosphere are smaller than those in the 2018 global dust storm and the top altitude is lower.

447 5. 3. Latitudinal variation of the water vapor vertical profiles

448 **Figure 8** shows the latitudinal map of the water vapor vertical profiles measured by
 449 NOMAD and predicted by GEM-Mars for non-dust storm condition at $L_s = 300\text{--}320^\circ$ (Fig. 8a),
 450 $L_s = 320\text{--}330^\circ$ (Fig. 8b, at the time of the regional dust storm), and $L_s = 330\text{--}345^\circ$ (Fig. 8c). It is
 451 found that the water vapor is significantly increased in the middle atmosphere at the time of the
 452 regional dust storm with the maximum values at latitudes greater than 60°S (Fig. 8b). The

453 meridional circulation over the period of this regional dust storm ($L_s = 322\text{--}327^\circ$) is still
 454 expected to be a single pole-to-pole "solstice" cell. Thus, the mechanism may be the same as for
 455 the decay phase of the 2018 global dust storm, i.e., the strong upward branch of the meridional
 456 circulation, as proposed by Shaposhnikov et al. (2019), may be responsible for the local
 457 maximum observed at latitudes greater than 60°S .

458 **6. Conclusions**

459 We have analyzed a selection of the first year solar occultation measurements by
 460 TGO/NOMAD and have presented variations in the vertical profile of water vapor on Mars,
 461 including periods of the global dust storm in 2018 and the following regional storm in 2019. The
 462 main results are:

- 463 1. We find a rapid and significant increase of water vapor in the middle atmosphere
 464 following the onset of the global dust storm (around $L_s = 190^\circ$). The enhancement of
 465 water vapor at the most active period of the dust storm (around $L_s = 200^\circ$) is
 466 remarkable: the water vapor reaches very high altitudes (100 km).
- 467 2. The latitudinal variation of water vertical profiles during the growth and mature
 468 phases of the global dust storm ($L_s = 196\text{--}202^\circ$ and $210\text{--}220^\circ$) shows that the water
 469 vapor abundance in the middle atmosphere is consistently increased at $60^\circ\text{S}\text{--}60^\circ\text{N}$.
 470 The overall increase in middle atmospheric water vapor likely reflects a breakdown
 471 in the 30–50 km hygropause trapping of water vapor due to dust driven atmospheric
 472 temperature increases. A full simulation in a GCM supporting this explanation is
 473 presented in an accompanying paper (Neary et al., 2019).
- 474 3. The latitudinal variation of water vertical profiles during the decay phases of the
 475 global dust storm ($L_s = 220\text{--}240^\circ$ and $240\text{--}260^\circ$) shows that the water vapor
 476 abundance in the middle atmosphere has peaks at latitudes greater than 60°S . At this
 477 time, the global circulation is beginning its transition from an equinoctial to a
 478 (southern summer) solstitial Hadley circulation pattern. Strong upward branch of the
 479 circulation at latitudes greater than 60°S (Shaposhnikov et al., 2019) may be the
 480 responsible for the local maximum.
- 481 4. We also find a conspicuous enhancement of water vapor in the middle atmosphere
 482 during the regional dust storm in 2019 (around $L_s = 325^\circ$). The magnitude of the
 483 water vapor increase is not as large as in the global dust storm but still remarkable:
 484 water vapor increases reach ~ 90 km. Again, this behavior is indicative of dust storm
 485 heating conditions.
- 486 5. Water vertical profiles observed during this regional dust storm ($L_s = 322\text{--}327^\circ$)
 487 exhibit peak upper level water vapor abundance at high southern latitudes (60°S),
 488 which can be explained by the upward branch of the solstitial Hadley circulation.
 489 This annually reoccurring dust storm behavior exhibits strong interannually
 490 variability (Kass et al., 2016), and appears to have been particularly intense in this
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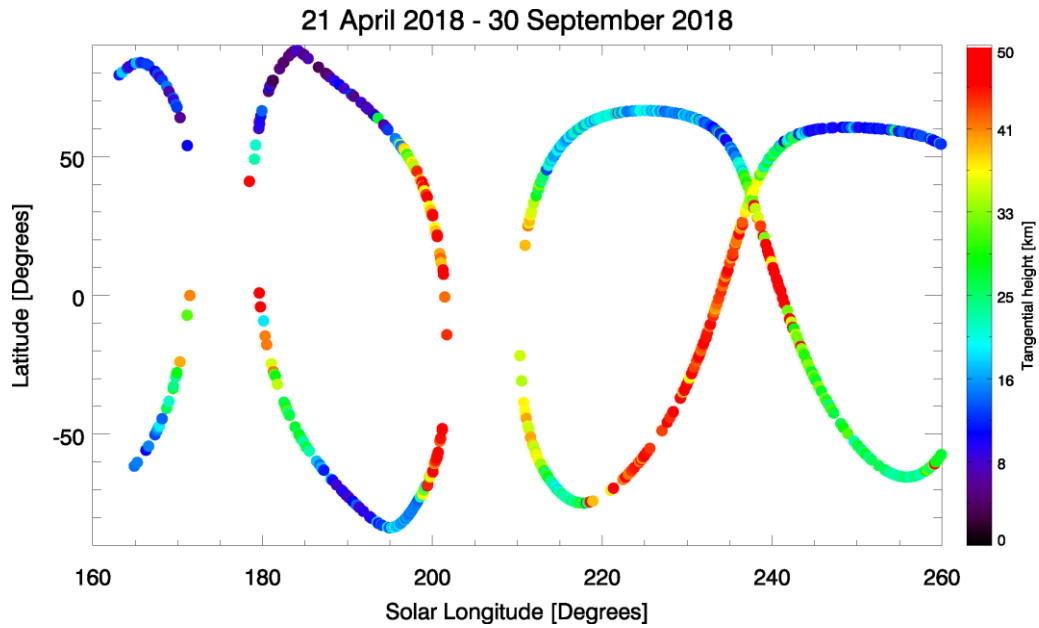
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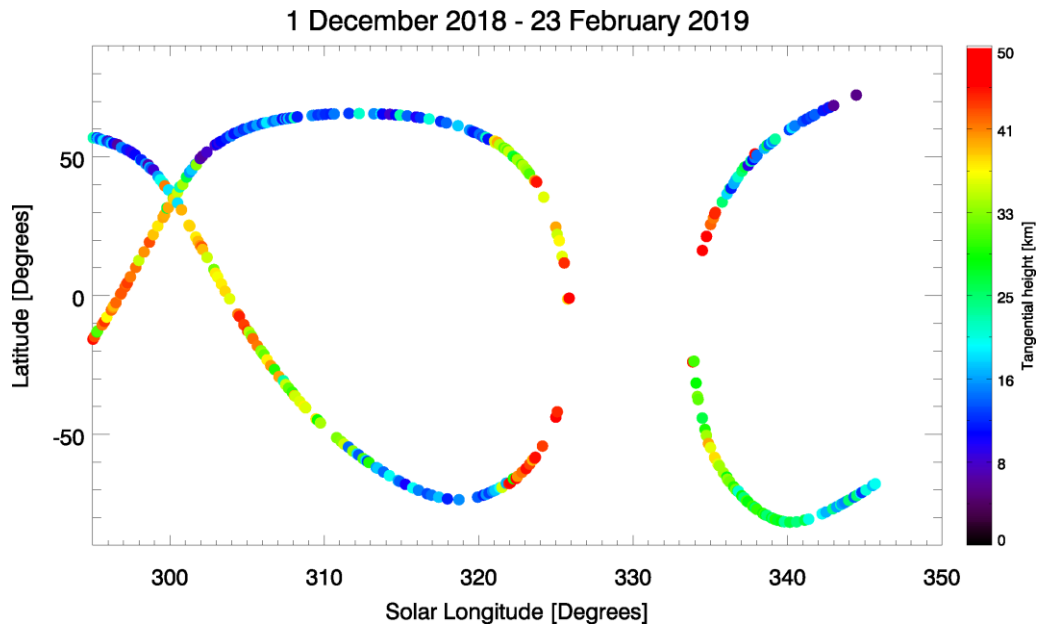
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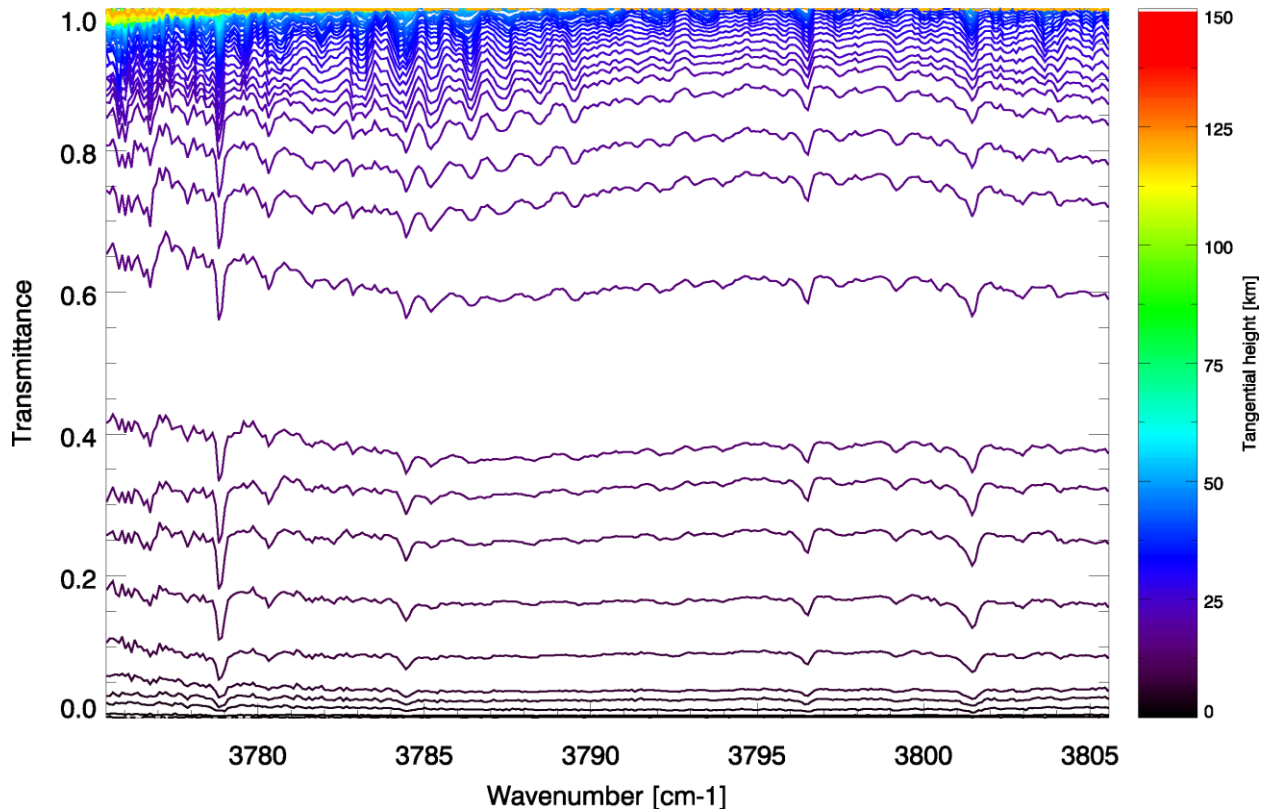
699 **Figures**700 **(a)**

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702 **(b)**

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704 **Figure 1.** Solar longitude (x-axis) and latitude (y-axis) of the solar occultation measurements
 705 taken from (a) 21 April to 30 September 2018 and (b) from 1 December 2018 to 23 February 2019
 706 by TGO/NOMAD used in this study. The color denotes the highest altitude at which the
 707 mean transmittance of spectra at the diffraction order 168 is less than $\exp(-1.0)$ (i.e., the optical
 708 depth is less than 1.0), which basically corresponds to the top altitude (relatively) free from the
 709 dust suspended in the Mars atmosphere.



710

711 **Figure 2.** Examples of spectra obtained during one occultation (21 April 2018; the first date of
712 the solar occultation measurements by TGO/NOMAD) in the spectral range between 3775 and
713 3805 cm^{-1} (diffraction order 168). Each transmittance is obtained by dividing the spectra
714 measured through the atmosphere by the reference spectrum recorded outside the atmosphere.
715 The selection of a spectral interval is achieved through the AOTF. The absorption features
716 presented in the spectra are mainly H_2O lines. Differences in colors represent the tangential
717 altitude of the measurements.

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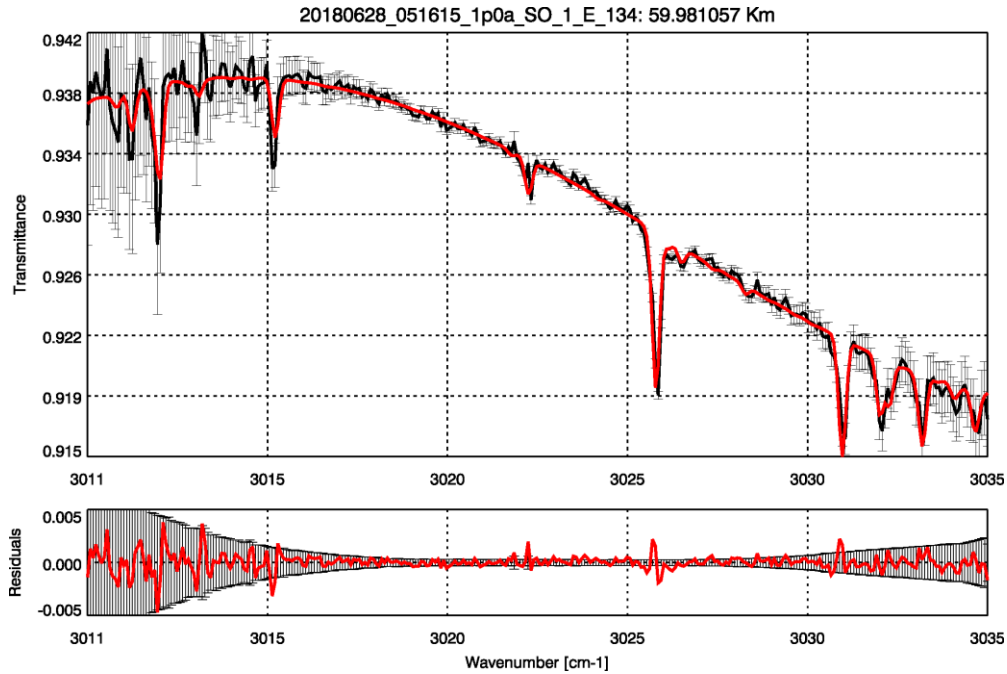
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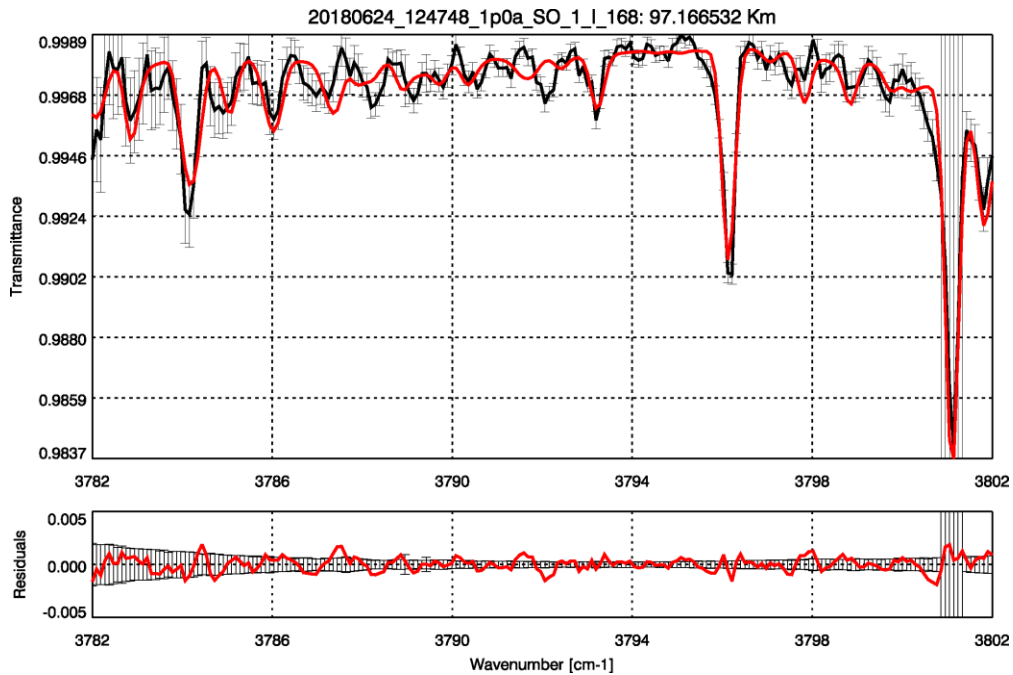
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727 (a)



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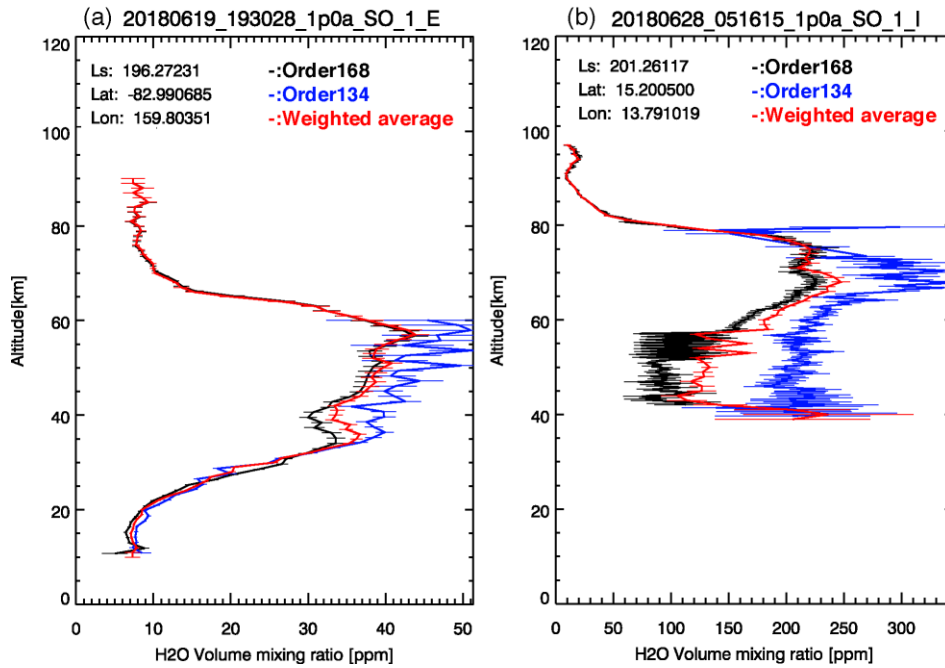
729 (b)



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731 **Figure 3.** Examples of the data analysis of NOMAD spectra to retrieve H₂O abundance. Spectra
 732 of (a) diffraction order 134 taken at 60 km altitude (latitude=51S°, longitude=22E°) on 28 June
 733 2018 ($L_s=201^\circ$) and (b) diffraction order 168 taken at 97 km altitude (latitude=38°N,
 734 longitude=138W°) on 24 June 2018 ($L_s=199^\circ$). The black and red curves show the measured
 735 NOMAD spectrum and the best-fit synthetic spectrum calculated by the radiative transfer model,

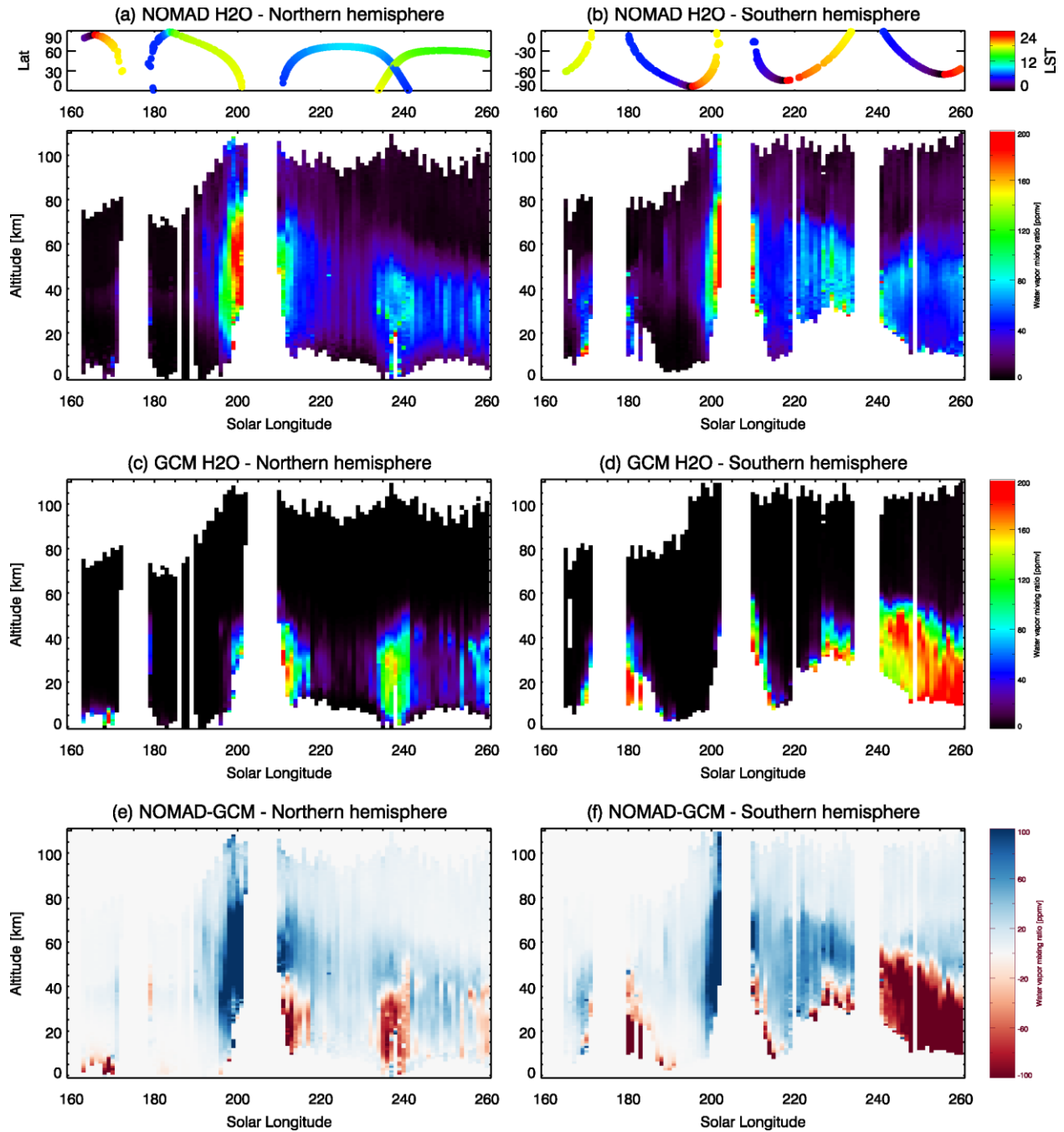
736 respectively. The retrieved H₂O abundance are 201 (± 4) and 233 (± 8) ppm, respectively. The
 737 bottom figures show the residuals of the fits in red and the error bars in black. Note that larger
 738 residuals around the H₂O line at 3026 cm⁻¹ shown in Panel (a) are due to the fact that the spectral
 739 calibration is not achieved with a sub-pixel accuracy. The H₂O line at 3801.4 cm⁻¹ shown in
 740 Panel (b) has larger error bars because the core of the line is saturated.



741

742 **Figure 4.** Examples of the water vapor vertical profiles retrieved from solar occultation
 743 measurements by TGO/NOMAD taken on (a) 19 June 2018 ($L_s=196^\circ$, latitude= 82° S,
 744 longitude= 159° E at the highest altitude), and on (b) 28 June 2018 ($L_s=201^\circ$, latitude= 15° N,
 745 longitude= 14° at the highest altitude). The black and blue curves show the vertical profiles of
 746 water vapor volume mixing ratio retrieved from the spectra in the diffraction orders 168 and 134,
 747 respectively. The red curves are their weighted averages. The horizontal bars represent 1-sigma
 748 errors due to the instrumental noise.

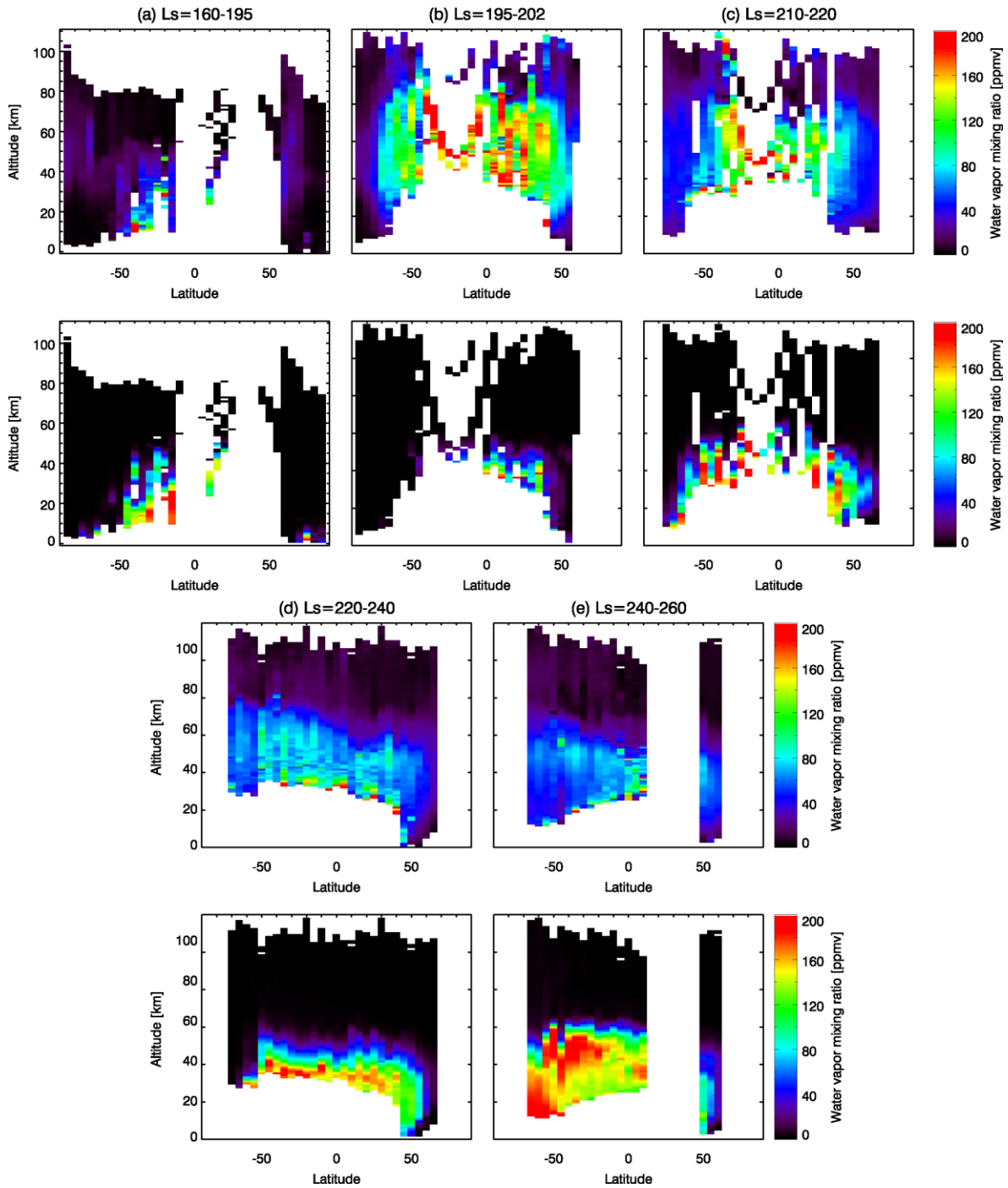
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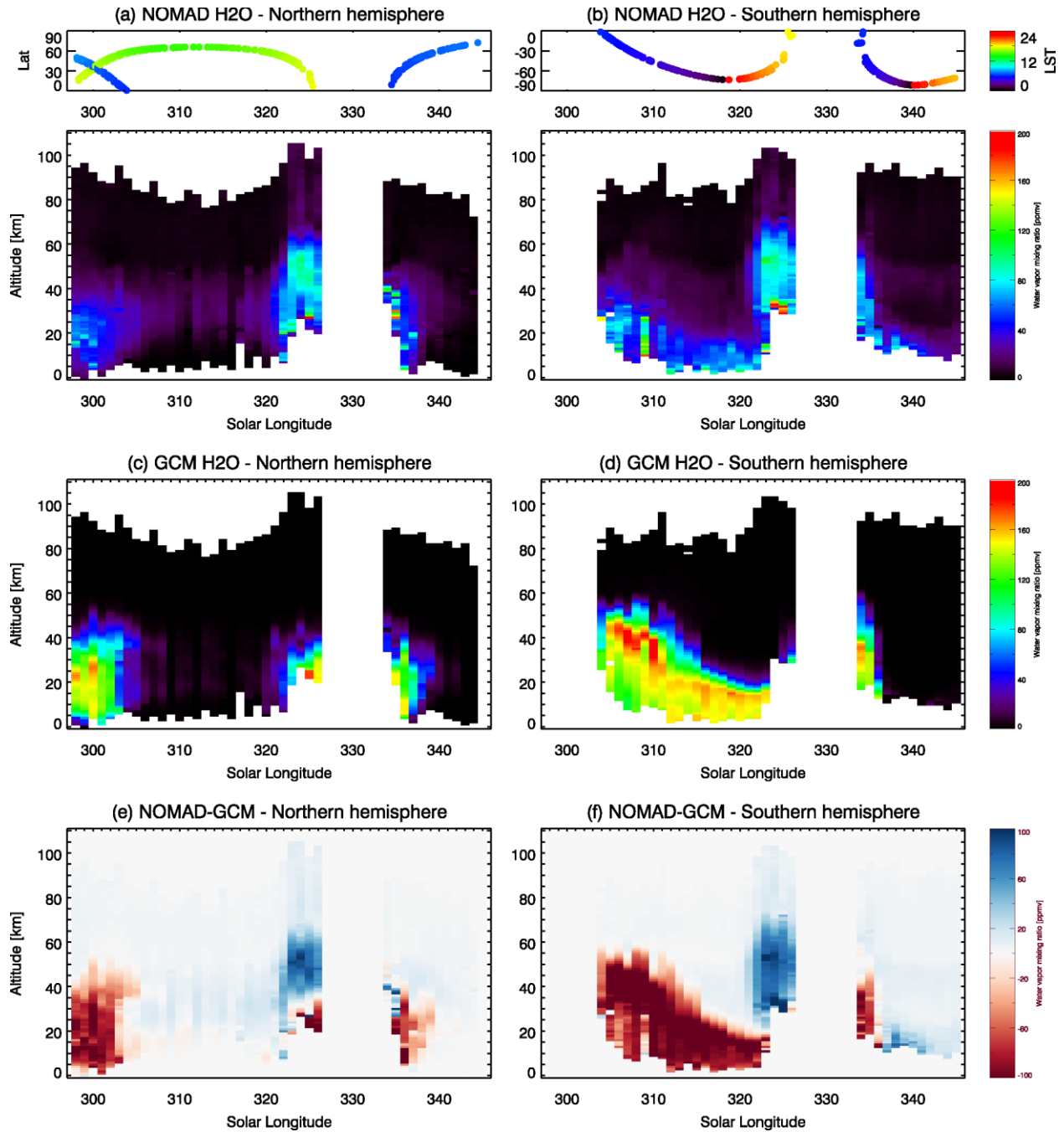
751 **Figure 5.** Seasonal variation of the water vapor vertical profiles at $L_s=162\text{--}260^\circ$ retrieved from
 752 the NOMAD data in (a) the northern hemisphere and (b) the southern hemisphere, (c, d) those
 753 predicted by the GEM-Mars for non-dust storm conditions, and (e, f) the differences between the
 754 NOMAD retrievals and the GCM predictions. The retrievals and GEM predictions are binned in
 755 $1^\circ L_s \times 1 \text{ km}$ altitude grid (averaged in latitude and longitude). The top panels of (a) and (b)
 756 show the latitudes and local solar time of the measurements.

757



758

759 **Figure 6.** Latitudinal variation of the water vapor vertical profiles retrieved from NOMAD data
 760 (the top panels of (a)-(e)), predicted by the GEM-Mars for non-dust storm conditions (the bottom
 761 panels of (a)-(e)) in the seasonal range between $L_s = 180\text{--}195^\circ$ (Fig. (a), before the global dust
 762 storm), $L_s = 195\text{--}202^\circ$ (Fig. (b), during the growth phase of the storm), $L_s = 210\text{--}220^\circ$ (Fig. (c),
 763 during the mature phase of the storm), $L_s = 220\text{--}240^\circ$ (Fig. (d), during the decay phase of the
 764 storm), and $L_s = 240\text{--}260^\circ$ (Fig. (e), during the decay phase of the storm). The retrievals and
 765 GEM predictions are binned in 5° latitude \times 1 km altitude grid (averaged in season and
 766 longitude).

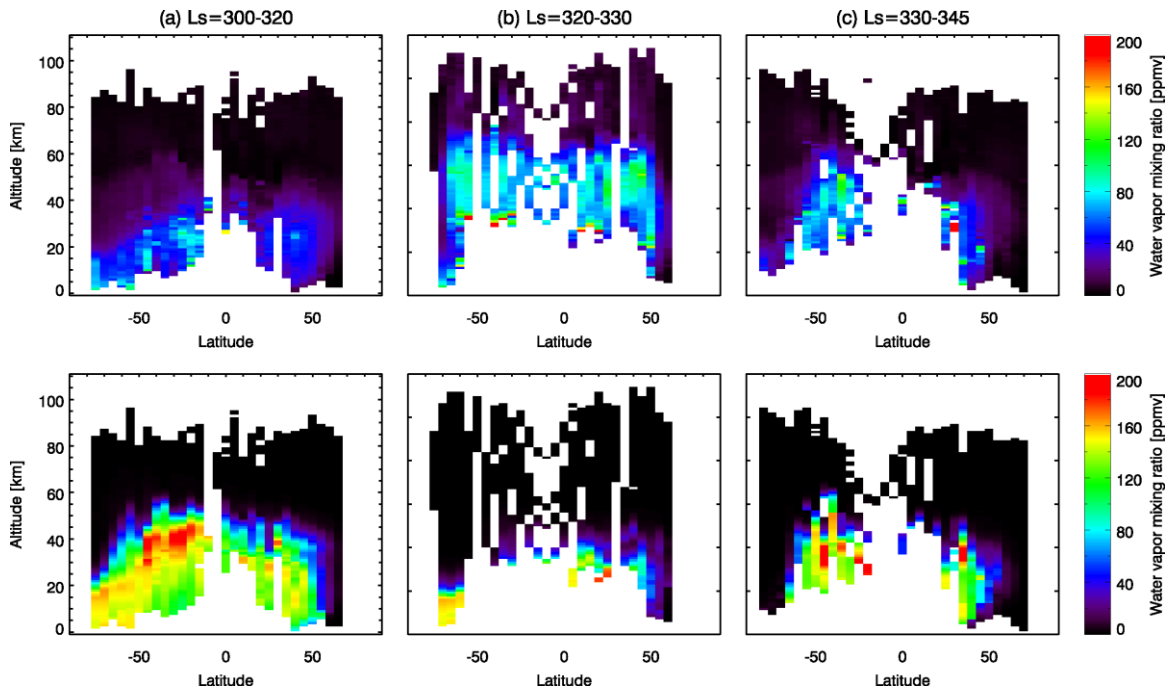


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768 **Figure 7.** Seasonal variation of the water vapor vertical profiles at $L_s = 298\text{--}345^\circ$ retrieved from
 769 the NOMAD data in (a) the northern hemisphere and (b) the southern hemisphere, (c, d) those
 770 predicted by the GEM-Mars for non-dust storm conditions, and (e, f) the differences between the
 771 NOMAD retrievals and the GCM predictions. The retrievals and GEM predictions are binned in
 772 $1^\circ L_s \times 1 \text{ km}$ altitude grid (averaged in latitude and longitude). The top panels of (a) and (b)
 773 show the latitudes and local solar time of the measurements.

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775



776

777 **Figure 8.** Latitudinal variation of the water vapor vertical profiles retrieved from NOMAD data
 778 (the top panels), predicted by the GEM-Mars for non-dust storm conditions (the bottom panels)
 779 in the seasonal range between $L_s = 300\text{--}320^\circ$ (a), $L_s = 320\text{--}330^\circ$ (b, at the time of the regional
 780 dust storm), $L_s = 330\text{--}345^\circ$ (c). The retrievals and GEM predictions are binned in 5° latitude \times 1
 781 km altitude grid (averaged in season and longitude).

782

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807