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► **To cite this version:**

Aurélien Stcherbinine, Franck Montmessin, Mathieu Vincendon, Michael J. Wolff, Margaux Vals, et al.. A Two Martian Years Survey of Water Ice Clouds on Mars with ACS onboard TGO. *Journal of Geophysical Research. Planets*, 2022, 127 (12), pp.e2022JE007502. 10.1029/2022je007502 . insu-03908848v2

HAL Id: insu-03908848

<https://insu.hal.science/insu-03908848v2>

Submitted on 20 Dec 2022 (v2), last revised 3 Jan 2023 (v3)

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A Two Martian Years Survey of Water Ice Clouds on Mars with ACS onboard TGO

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Key Points:

- Cloud altitude varies by 20 to 40 km between winter and summer, and between polar and midlatitudes.
- The GDS in MY 34 increases the altitude of the clouds by 10 to 20 km compared to the same season in MY 35.
- Clouds are predicted to be lifted at lower altitudes in the Mars PCM compared to ACS detections.

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This article has been accepted for publication and undergone full peer review but has not been through the copyediting, typesetting, pagination and proofreading process, which may lead to differences between this version and the [Version of Record](#). Please cite this article as doi: [10.1029/2022JE007502](https://doi.org/10.1029/2022JE007502).

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Abstract

The middle infrared (MIR) channel of the Atmospheric Chemistry Suite (ACS) instrument onboard the ExoMars Trace Gas Orbiter (TGO) ESA-Roscosmos mission has performed Solar occultation measurements of the Martian atmosphere in the 2.3–4.2 μm spectral range since March 2018, which now covers two Martian Years (MY). We use the methodology previously developed for the study of the MY 34 Global Dust Storm (GDS) (Stcherbinine et al., 2020) to monitor the properties (effective radii, extinction, altitude) of the Martian water ice clouds over the first two Martian years covered by ACS-MIR. The observations encompass the period $L_s = 163^\circ$ in MY 34 to $L_s = 181^\circ$ in MY 36. We determine that the typical altitude of the clouds varies by 20 to 40 km between the summer and winter, with a maximum extension up to 80 km during summer in the midlatitudes. Similarly, we also note that for a limited temporal range, the altitude of the clouds also varies by 20 to 40 km between the polar regions and the midlatitudes. We also compare observations acquired during the MY 34 GDS to observations from the same period in MY 35, using that latter as a reference to characterize the effects of this GDS on the clouds' properties. In addition, we compare our retrievals with the predictions of the Mars Planetary Climate Model (PCM), which shows a reasonable agreement overall for the altitude of the clouds, although the model usually predicts lower altitudes for the top of the clouds.

Plain Language Summary

We use data from the Middle InfraRed (MIR) channel of the Atmospheric Chemistry Suite (ACS) instrument onboard the ExoMars Trace Gas Orbiter mission, which has been probing the Martian atmosphere for two Martian years (MY), to study the properties of the water ice clouds in the Martian atmosphere. We observe that the clouds' altitude increases by 20 to 40 km during Summer and that they are also observed 20 to 40 km higher around the equator compared to the polar regions. This highlight the spatial and temporal diversity of the Martian water ice clouds, along with the large amplitude of the annual variations of the atmospheric structure. In addition, we contrast observations acquired during the Global Dust Storm (GDS) event of Summer 2018 (MY 34) with data from the following MY that we use as a reference to understand the effects of the GDS on the clouds, which reveals that the presence of the GDS increases the altitude of the clouds by 10 to 20 km. Finally, we compare our results to the predictions of the Mars Planetary Climate Model which shows a reasonable agreement overall, although the model tends to predict lower altitudes for the top of the clouds.

1 Introduction

Even though the role of water ice clouds in the Martian atmosphere was not well understood for many years after the Viking missions, they are now considered to play a major role in Martian climate and weather (Richardson, 2002; Montmessin et al., 2004; Navarro et al., 2014; Montmessin et al., 2017; Clancy et al., 2017; Vals et al., 2022; Rossi et al., 2022). Clouds perform an important role in the Martian water cycle as they are a major actor in the inter-hemispheric water exchange (Clancy et al., 1996; Montmessin et al., 2004, 2017). Similarly to atmospheric dust particles, water ice crystals also absorb and scatter incoming solar radiation, thus impacting the atmospheric structure and temperature (Wilson et al., 2007, 2008; Haberle et al., 2011; Madeleine et al., 2012; Navarro et al., 2014). In addition, the formation of clouds affects the ability of water (or hydrogen) to be further mobilized and to escape from the planet. However, more observational data are needed to better characterize the properties of water ice clouds in order to understand and model the evolution of the Martian atmosphere (Vals et al., 2018).

69 The formation of clouds in the atmosphere depends on several factors, such as the
70 presence of water vapor, the pressure and temperature conditions, and the availability
71 of condensation nuclei (Michelangeli et al., 1993; Montmessin et al., 2004). However,
72 as water may exist in a supersaturated state in the Martian atmosphere (Maltagliati
73 et al., 2011; Fedorova et al., 2020; Poncin et al., 2022), simply considering the freezing
74 point for the condensation of atmospheric water is problematic. In addition, accu-
75 rately predicting the distribution and the properties of airborne dust particles also
76 remains a challenging aspect in current Global Climate Models (GCM) (Forget et al.,
77 2014; C. Wang et al., 2018; D’Aversa et al., 2022). Thus, accumulating additional
78 observations of the distribution and properties of the water ice crystals in the Mar-
79 tian atmosphere is required to further constrain the present-day clouds cycle, and its
80 relationship to the Martian climate and water cycle.

81 The Atmospheric Chemistry Suite (ACS) instrument is a set of three spectrom-
82 eters onboard the ExoMars Trace Gas Orbiter (TGO) ESA-Roscosmos spacecraft,
83 which has been conducting science operations since March 2018 (Korablev et al., 2018,
84 2019; Vandaele et al., 2019). The Mid-InfraRed (MIR) channel is a high-resolution,
85 cross-dispersion echelle spectrometer dedicated to solar occultation (hereafter "SO")
86 geometry. Each observation covers a ~ 300 nm wide spectral interval selected between
87 2.3 and 4.2 μm , which is set by rotating the secondary grating to one of the 10 posi-
88 tions (Trokhimovskiy et al., 2015; Korablev et al., 2018). The cross-dispersion optical
89 scheme produces 10 to 21 stacked diffraction orders. The number and separation be-
90 tween the displayed orders depend on the secondary grating position. Considering
91 the instantaneous angular field of view of the detector and its displacement during
92 the integration time for each spectrum, ACS-MIR provides a sampling of the Martian
93 atmosphere with a vertical resolution of ~ 2.5 km.

94 Even though the primary objective of ACS and NOMAD is the study of the
95 atmospheric trace gases in the Martian atmosphere, they also provide a unique dataset
96 for the study of the distribution and properties of Martian aerosols (Stcherbinine et
97 al., 2020; Luginin et al., 2020; Liuzzi et al., 2020, 2021; Streeter et al., 2022). In
98 this paper we use ACS-MIR observations acquired in position 12 (i.e., ~ 3.1 – 3.4 μm
99 spectral range) to retrieve the properties of the Martian water ice clouds from the 3 μm
100 water ice absorption band. Indeed, water ice atmospheric particles exhibit a specific
101 signature due to the O-H stretching and bending that enables the distinction between
102 water ice and bound water in dust. And as the depth and shape of the absorption
103 feature depend on both the abundance and the sizes of the ice crystals, we can retrieve
104 information on these quantities (Vincendon et al., 2011; Guzewich et al., 2014; Clancy
105 et al., 2019). The methodology has already been developed and applied to observations
106 conducted during the global dust storm (hereafter "GDS") of Martian Year (MY) 34.
107 These efforts revealed very high altitude water ice hazes (up to 100 km) and large ice
108 crystals (effective radius $r_{\text{eff}} \sim 1.5$ – 2 μm) up to 65 km during the storm (Stcherbinine et
109 al., 2020). ACS observations now cover two MY, one with and one without a GDS, for
110 all latitudes, allowing us to observe the seasonal and spatial evolution of the Martian
111 water ice clouds. In addition, observations acquired during MY 35 during the same L_s
112 period when the GDS occurs in MY 34 also offer a reference to be compared to MY
113 34 GDS observations.

114 First, we briefly describe in Section 2 the ACS-MIR dataset and the methods
115 used in this study. Sections 3 & 4 present the results of the annual monitoring of
116 water ice clouds with their particle sizes and opacity, and Section 5 compares the
117 retrieved vertical profiles of water ice clouds with results of GCM simulations from
118 the Mars Planetary Climate Model (PCM) (Forget et al., 2022). Finally, Section 6
119 summarizes the main points of this study.

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2 Data and methods

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2.1 Dataset

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Atmospheric transmittances are computed through ratios to measurements at 120 km above the surface, which is free from atmospheric absorption in this spectral range. Indeed, we observed in Stcherbinine et al. (2020) (where transmittances were computed relative to 150 km) that the haze top altitude only reaches 100 km during an extreme event like the MY 34 GDS, but without extending over 105 km. Data calibration and geometry calculations are described in Trokhimovskiy et al. (2020) and Olsen et al. (2021).

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Since the publication of Stcherbinine et al. (2020), ACS-MIR has continued to acquire new data in the grating position 12, allowing us to observe the atmospheric 3 μm absorption band by covering the 3.1–3.4 μm spectral range (Korablev et al., 2018). We do not consider the observations acquired in the so-called "partial mode" (i.e., with a smaller number of diffraction orders) around the conjunction of Mars and Earth (when the downlink from the instruments is very limited). A total of 514 observations obtained between $L_s = 163^\circ$ (MY 34) and $L_s = 181^\circ$ (MY 36) were used in this study. The spatial and temporal distribution of these observations is shown in Figure 1. Due to the SO geometry, observations naturally occur in the periods near dawn and twilight.

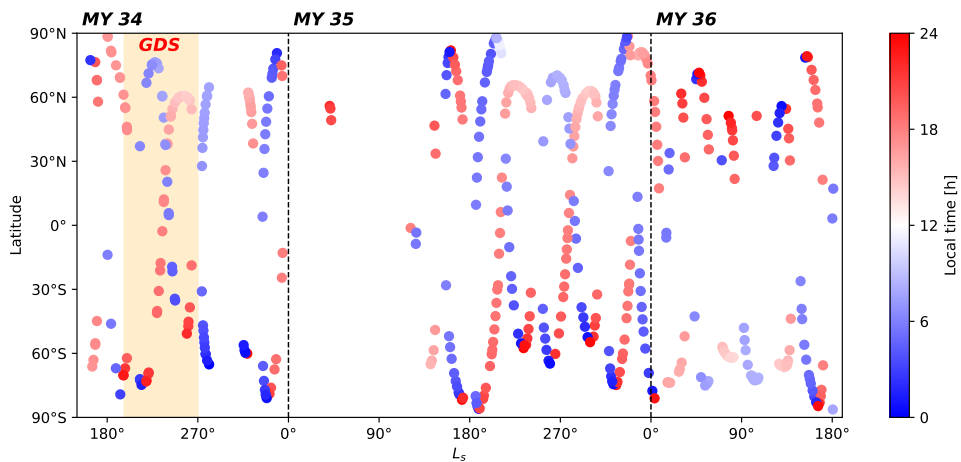


Figure 1. Spatial (latitude) and temporal (L_s and local time) distribution of the 514 ACS-MIR position 12 observations used in this study. The orange region corresponds to the period of the 2018/MY34 global dust storm.

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2.2 Continuum spectra

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Based on the methodology previously described in Stcherbinine et al. (2020), we extract the spectral continuum for each observed altitude from the 20 spectral segments (i.e., diffraction orders) that compose the ACS-MIR observation (Korablev et al., 2018; Trokhimovskiy et al., 2015). For each diffraction order, we use successive iterations of a median filter and a second-degree polynomial fit on the 200 centered points to determine the transmission level of the spectral continuum at the center of the diffraction order, while minimizing the effects of contamination by the gas absorption bands or the broad instrumental curvature that affects the orders. This results (for each observed altitude) in a spectrum with one point per diffraction order, with a

149 spectral resolution of $\Delta\lambda \sim 20$ nm. To avoid detector edges effects, we only consider
 150 the center of each diffraction order to process the extraction. We also ignore the outer
 151 regions of the extracted spectra, that is, the spectels corresponding to 3.10, 3.12, and
 152 3.44 μm .

153 We have updated the uncertainties estimation for the ACS-MIR transmittances.
 154 Thus, as we are considering the center of each diffraction order, we include an uncertainty
 155 of $\pm 5\%$ of the transmission value to account for the uncorrected straylight that
 156 affects the data in position 12 (Trokhimovskiy et al., 2020).

157 We compute the haze top altitude (i.e., the highest altitude at which aerosols
 158 can be detected along the line of sight), defined here as the first altitude for which the
 159 transmission is greater than 0.98 over the entire spectrum, along with the extinction
 160 coefficient k_{ext} for each wavelength (forming an extinction spectrum) and altitude from
 161 the total optical depth (i.e., integrated along the line of sight, $\tau = -\log(\text{Tr})$) using
 162 a vertical profile inversion algorithm based on the onion-peeling method (Goldman &
 163 Saunders, 1979) as described in sections 2.2 and 2.4 of Stcherbinine et al. (2020).

164 2.3 Water ice particle size retrieval method

165 To identify the Martian water ice clouds in our observations, and constrain their
 166 particle sizes, we compare each retrieved extinction spectrum with models of the extinction
 167 coefficient's wavelength dependence for either water ice or dust spherical particles
 168 (Stcherbinine et al., 2020). These theoretical extinction coefficients are computed using
 169 a public domain Mie code (Toon & Ackerman, 1981) and assuming a gamma size
 170 distribution (Hansen & Travis, 1974) with an effective variance of 0.1 (e.g., Wolff et
 171 al., 2017, and references contained within).

172 To avoid the problem of the presence of local minima for r_{eff} during the fitting
 173 process, we generate models on a grid of radii of 0.01 between 0.1 and 8 μm , and
 174 for each one we compute the χ^2 and Δ^2 values defined in equations (2) & (3), which
 175 quantify the deviation between the data and the model. Finally, we have slightly
 176 adjusted the criterion for the detection and characterization of the atmospheric water
 177 ice particles compared to Stcherbinine et al. (2020). This modification results from a
 178 better understanding of detection biases thanks to the new larger dataset, and also
 179 accounts for the new uncertainty estimate (section 2.2). The new criterion is as follows:

$$180 \quad (\chi_{\nu, \text{ice}}^2 < 1) \ \& \ \left(\frac{\Delta_{\text{dust}}^2}{\Delta_{\text{ice}}^2} > 4 \right) \quad (1)$$

181 Where

$$182 \quad \chi_{\nu}^2(r_{\text{eff}}) = \frac{1}{N-2} \sum_{i=1}^N \frac{(\text{data}_i - \text{model}_{r_{\text{eff}}, i})^2}{\sigma_i^2} \quad (2)$$

183 and

$$184 \quad \Delta^2(r_{\text{eff}}) = \frac{1}{N-2} \sum_{i=1}^N (\text{data}_i - \text{model}_{r_{\text{eff}}, i})^2 \quad (3)$$

185 Notation

186 **data_i** The i^{th} spectel of the k_{ext} spectra from the ACS-MIR observation.

187 **model_{r_{eff}, i}** The i^{th} spectel of the model extinction spectra for a particle size of r_{eff} .

188 **σ_i** The uncertainty on the value of data_i .

189 **N** The number of spectral points in the considered spectrum.

190 Specifically, we require that:

- 191 1. the modeled extinction spectrum using water ice provides a good fit to the
192 observational data, including the spectrum uncertainties ($\chi_{\nu, \text{ice}}^2 < 1$).
- 193 2. the water ice model provides a significantly better solution than the best fit that
194 can be obtained with a dust model. So we consider only water ice models with
195 a mean square difference to the data lower than the one of the best dust model by
196 at least a factor 4. This factor 4 has been determined experimentally by visual
197 comparison between the models and the data for many spectra, to retain only
198 those which match within the errorbars.

199 For each spectrum where water ice crystals are detected, we retrieve the opti-
200 mal r_{eff} as the one corresponding to the model associated with the lower $\chi_{\nu, \text{ice}}^2$ that
201 verifies equation (1). Then, the lower (respectively upper) bound for the particle size
202 uncertainties corresponds to the minimum (respectively maximum) value of r_{eff} in the
203 set of models associated with $\chi_{\nu, \text{ice}}^2$ and Δ_{ice}^2 that prove equation (1).

204 3 Cloud monitoring

205 In sections 3.1 and 3.2, we present the results of the analysis of the new data, as
206 described in Stcherbinine et al. (2020). They cover a complete year without a GDS,
207 from $L_s = 140^\circ$ (MY 35) to $L_s = 182^\circ$ (MY 36). This range corresponds to ACS-MIR
208 operating in full-observation mode (cf. Figure 1). Then, we compare this new dataset
209 to the previous one for the GDS year MY 34 in section 3.3.

210 In the following, a "profile" refers to one set of ACS-MIR observations from the
211 surface to the haze top, while a "cloud" corresponds to contiguous water ice detections
212 within a profile. Additionally, as we observe the presence of water ice clouds up to
213 the haze top altitude in most of the profiles, we consider the haze top equivalent to
214 the maximum altitude of the water ice clouds. Also, our water ice clouds detections
215 extend typically over ~ 20 km below their maximum altitude, thus we will only discuss
216 in this section the maximum altitude of the water ice clouds as it is representative of
217 the overall behavior of the water ice clouds (and the maximum altitude of the aerosols
218 in general).

219 3.1 Seasonal variations

220 We show seasonal variations for three distinct latitude ranges in Figure 2 in order
221 to delineate the seasonal variations from the latitudinal dependency of the clouds that
222 will be discussed in paragraph 3.2.

223 As reported in previous studies (e.g., Clancy et al., 2019; Stcherbinine et al.,
224 2020; Luginin et al., 2020; Liuzzi et al., 2020), we find a decrease in the size of the
225 water ice crystals as altitude increases, both from an individual cloud and from a global
226 perspective (cf. Figures 2, 3 & 4). In addition, clouds are usually found at the top of
227 the profiles, capping the layers composed of other types of aerosols (dust or large ice
228 crystals below).

229 We also observe in Figure 2 that if the maximum altitude of the clouds fluctuates
230 with season (Jaquin et al., 1986; Forget et al., 1999; Montmessin et al., 2006; Heavens et
231 al., 2011; Määttänen et al., 2013; Smith et al., 2013), the observed variations depend on
232 the latitude. Indeed, at equatorial latitudes ($45^\circ\text{S} \leq \text{lat} \leq 45^\circ\text{N}$, panel b), we observe
233 the presence of water ice crystals up to 83 km around the perihelion ($L_s \sim 270^\circ$),
234 while the maximum altitude of the clouds does not exceed 50 km at $L_s \sim 80^\circ$. Plus,
235 for latitudes southward 45°S (panel c) we note an increase of the maximum altitude
236 of the clouds that goes from 30 km at $L_s \sim 160^\circ$ to 80 km at $L_s \sim 260^\circ$, followed by a

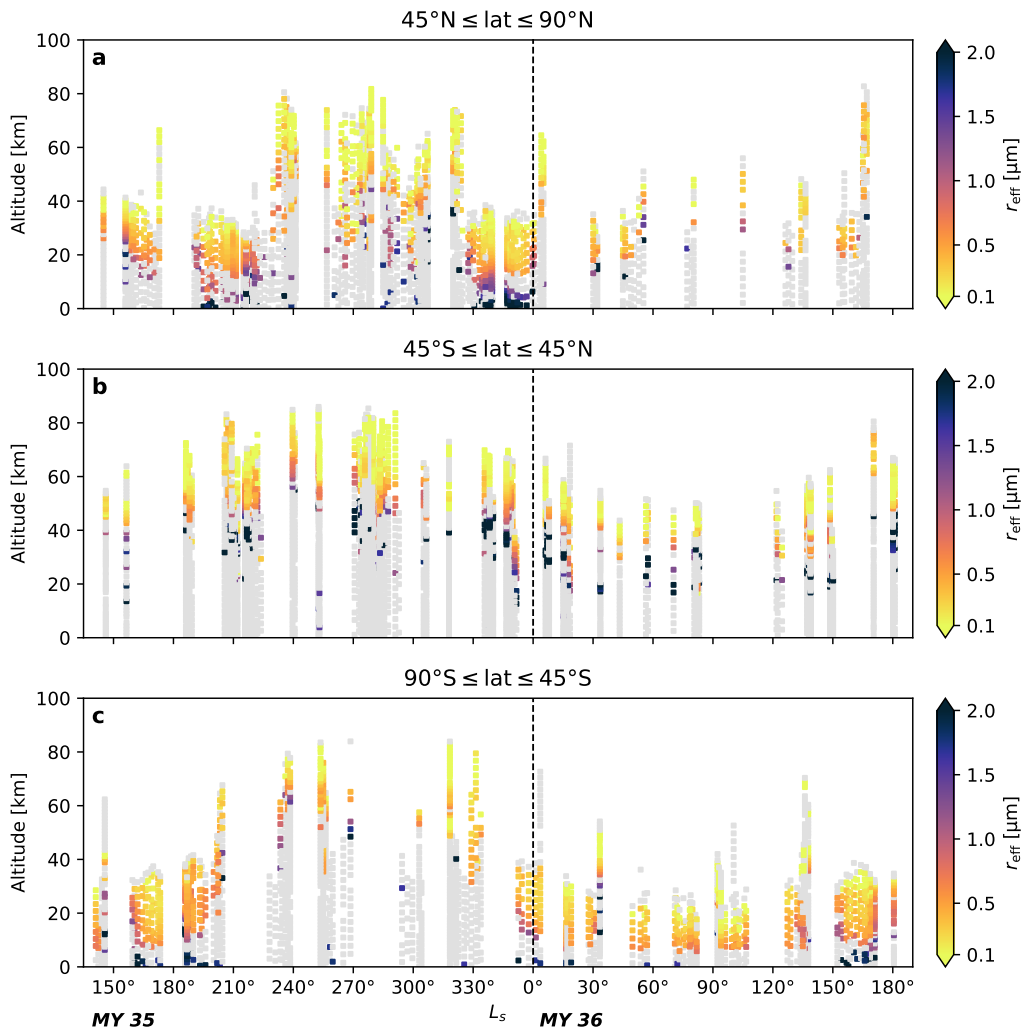


Figure 2. Vertical profiles of water ice clouds in the Martian atmosphere as observed by ACS-MIR over mid-MY 35 to mid-MY 36, with their crystal size determined using the method presented in section 2.3. Each panel represents the profiles grouped by ranges of latitude (North/Equatorial/South). Observations without water ice detections are in gray.

237 decrease from $L_s \sim 330^\circ$ that brings it back to 30 km at $L_s \sim 20^\circ$. Symmetrically, in
 238 the Northern hemisphere, between 45°N and 90°N (panel a) the maximum altitude of
 239 the clouds moves from 35 to 20 km between $L_s = 145^\circ$ and $L_s = 215^\circ$, and from 20 to
 240 40 km between $L_s = 30^\circ$ and $L_s = 100^\circ$ before coming back to 25 km at $L_s = 160^\circ$.

241 *Observation of a regional dust storm*

242 In addition to these variations of the water ice clouds altitude with L_s , we observe
 243 in the Northern hemisphere (for latitudes greater than 45°N ; Figure 2a) a sudden
 244 increase of their maximum altitude between $L_s = 236^\circ$ and $L_s = 323^\circ$, which indicates
 245 the presence of a regional dust storm in this region. This dust storm and its impact on
 246 the altitude of the water ice clouds have also been observed by the NOMAD instrument
 247 and are reported in Streeter et al. (2022). Indeed, from $L_s = 225^\circ$ to $L_s = 240^\circ$ the
 248 maximum altitude of the clouds goes from 35 to 80 km. That is to say, there is an
 249 increase in the altitude of the water ice clouds of about 40 km while the latitude of
 250 the observations remains similar (between 60°N and 65°N , cf Figure 1). The clouds
 251 remain in these high atmospheric layers (with crystals observed up to 80 km) until
 252 $L_s = 325^\circ$, even though we can observe a progressive decrease of the maximum altitude
 253 of the clouds over the storm, which goes from 80 to 65 km between $L_s = 235^\circ$ and
 254 $L_s = 305^\circ$ while the observed latitudes oscillate between 41°N and 63°N . Then, the
 255 maximum altitude of the water ice clouds suddenly decrease between $L_s = 323^\circ$ and
 256 $L_s = 328^\circ$, going from 71 to 28 km. However, as 10° of latitude separate these two
 257 observations (61.5°N for $L_s = 323^\circ$ and 71.5°N for $L_s = 328^\circ$, plus, 71.5°N is above the
 258 northernmost latitude of the observations of the storm), it is challenging to determine
 259 precisely the end date of the dust storm with our observations. For example, the
 260 northern limit of the MY 34 GDS was very abrupt and located under similar latitudes
 261 (Stcherbinine et al., 2020). In addition, 71.5°N is also close to the typical boundary of
 262 the north polar vortex at this period (Ball et al., 2021), which may act as an efficient
 263 barrier to the latitudinal transport of atmospheric particles (Toigo et al., 2020).

264 Regarding the temporal behavior of this dust storm, we observe that it occurs in
 265 the second half of the "dust storm season", which illustrates the diversity of Martian
 266 dust storms and the necessity to observe and study a large number of these events
 267 (H. Wang & Richardson, 2015; Battalio & Wang, 2021). We also observe that this
 268 storm does not push clouds to altitudes as high as what has been observed during the
 269 MY 34 GDS (80 km here vs 100 km during the GDS), and that no water ice crystals
 270 larger than $1.5 \mu\text{m}$ are observed above 40 km, unlike during the GDS where such large
 271 crystals have been detected between 50 and 65 km.

272 **3.2 Latitudinal variations**

273 As mentioned above, and reported in previous studies (Jaquin et al., 1986; Forget
 274 et al., 1999; Montmessin et al., 2006; Heavens et al., 2011; Smith et al., 2013) the ver-
 275 tical structure of Martian aerosols depends on the latitude. To isolate this latitudinal
 276 dependence from the seasonal variation discussed in section 3.1, Figure 3 presents the
 277 vertical profiles of water ice crystals size within the clouds. We adjust several L_s bins
 278 to minimize the impact of the seasonal variations while providing enough coverage in
 279 terms of latitude. If the entire range of latitudes is not covered in each panel, we
 280 can observe on panels a, c & d that the altitude of the water ice clouds (and more
 281 generally of the Martian aerosols) increases when getting closer to the equator. This
 282 also produces a bell-shaped distribution of the altitude of the clouds as a function
 283 of latitude. Indeed, we observe on Figures 3 & 4 that if the water ice crystals with
 284 radius $r_{\text{eff}} \leq 1 \mu\text{m}$ are located between 10 and 40 km in the polar regions ($> 60^\circ\text{N}$
 285 or $< 60^\circ\text{S}$), they are found typically between 30 and 80 km around the equator, and
 286 occasionally up to 85 km.

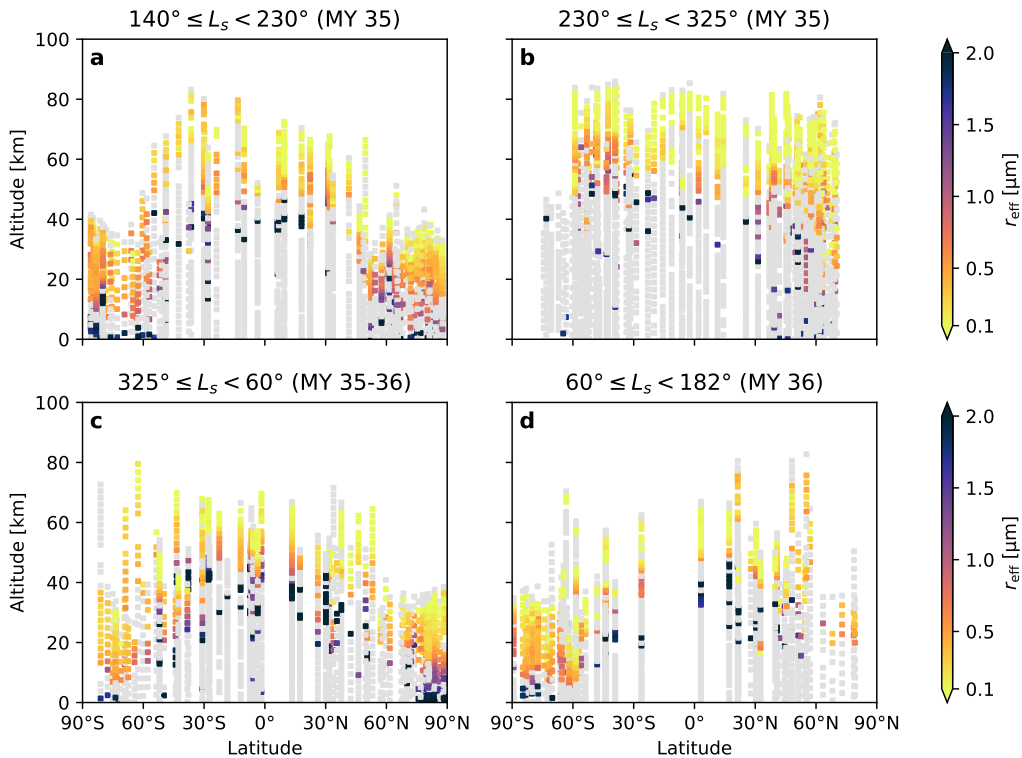


Figure 3. Vertical profiles of water ice clouds in the Martian atmosphere as observed by ACS-MIR between $L_s = 140^\circ$ (MY 35) and $L_s = 182^\circ$ (MY 36) as a function of the latitude of the observation. To highlight the latitudinal variations of the clouds while removing their seasonal dependence (discussed in section 3.2), each panel represents the profiles grouped by L_s ranges. Observations without water ice detection are in gray.

287 Another noteworthy point regarding the size of the water ice crystals that compose
 288 the clouds detected by ACS-MIR: we observe that for a fixed size of particle,
 289 the altitude of the detections still follows the bell-shaped distribution with latitude
 290 already identified for the general distribution of the clouds in the previous paragraph.
 291 Thus, there is no strict correlation between the size of a crystal and its altitude in
 292 the atmosphere. Other parameters have to be considered, such as the latitude of the
 293 cloud. This is consistent with the decrease of the scale height of the atmosphere in the
 294 high latitudes, where the temperatures are lower. As the temperatures decrease with
 295 increasing latitude, one finds a decrease in the cloud altitudes. However, we observe
 296 in Figure 4 that we also do not detect crystals with radius $\geq 1.5 \mu\text{m}$ above 55 km
 297 (outside the MY 34 GDS), there is no crystal with size $\leq 0.2 \mu\text{m}$ at altitudes lower
 298 than 15 km. Plus, except for one atypical cloud observed by 35°S of latitude and 252°
 299 of L_s (MY 35), only crystals with sizes $\leq 0.1 \mu\text{m}$ populate the altitudes higher than
 300 80 km. Thus, we can identify with Figure 4 the range of altitudes that are populated
 301 by water ice crystals for a given radius.

302 Regarding the larger crystals ($r_{\text{eff}} \geq 1.5 \mu\text{m}$), they dominate the lower layers
 303 of the clouds and are observed up to 55 km of altitude around 30°S . However, such
 304 particles are not observed for all profiles in which a water ice cloud has been detected,
 305 unlike small-grained high altitude hazes that are observed at the top of almost every
 306 profile. The localized aspect together with the observed altitudes of these clouds

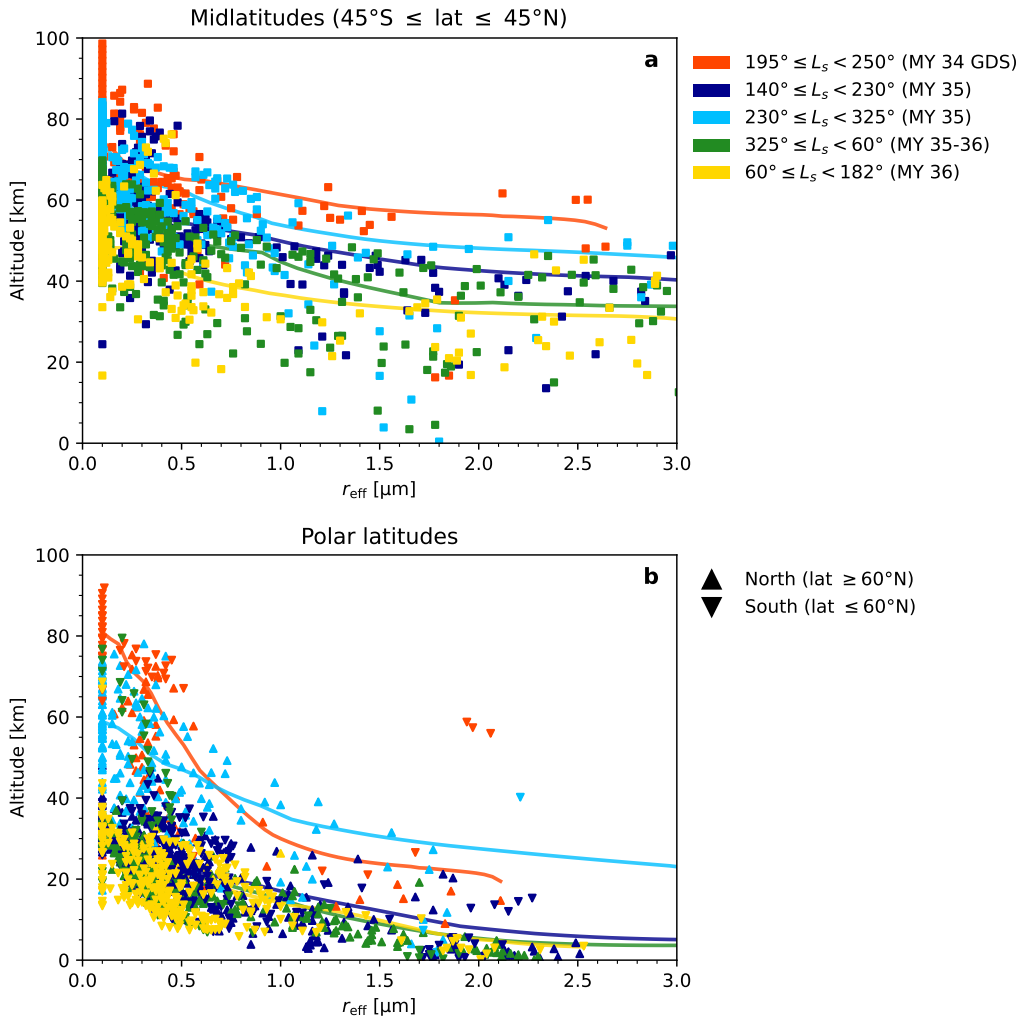


Figure 4. Distribution and size of the water ice crystals within the clouds as a function of their altitude and L_s of observation, for 3 ranges of latitude: midlatitudes (panel a), Northern and Southern polar latitudes (panel b). The solid lines show the average trend for each range of L_s . Observations acquired during the MY 34 GDS are shown in red for comparison with data from non-GDS years (MY 35 & 36).

307 composed of crystals with $r_{\text{eff}} \sim 2 \mu\text{m}$ is consistent with the results obtained using
 308 CRISM limb observations presented in figures 10 to 17 of Smith et al. (2013).

309 Concerning the local dust storm occurring in the Northern hemisphere during
 310 MY 35 (cf. section 3.1), panel b of Figure 3 presents profiles of clouds acquired during
 311 this period. This event allows us to study its specific impact on the clouds in the
 312 Northern latitudes in relation to the rest of the planet. Although TGO's orbit over
 313 this period did not allow ACS to probe latitudes above 65°S and 70°N , we observe
 314 that the maximum altitude of the clouds observed near 70°N extends to 70 km, which
 315 is higher than clouds observed at these latitudes outside of the storm (cf. panels a, c
 316 & d of Figure 3), but lower than the altitudes of the mesospheric clouds between 40°S
 317 and 20°N (Clancy et al., 2019).

318 3.3 Comparison with the 2018/MY34 GDS

319 Stcherbinine et al. (2020) used the early observations provided by ACS-MIR to
 320 study the planetary-scale dust storm that encircled Mars during MY 34. With the
 321 current extended dataset, it is now possible to compare observations acquired during
 322 the MY 34 GDS with the ones obtained for the same range of L_s in MY 35, when
 323 there was no GDS.

324 Figure 5 presents a superimposition of the profiles obtained between $L_s = 190^\circ$
 325 and $L_s = 250^\circ$ during MY 34 (panels a & e) and MY 35 (panels b & f). The latitudes
 326 and local times of the observations are indicated in panels c, d, g & h respectively. This
 327 information should be taken into account when comparing the two years, as the latitude
 328 of the clouds has a significant impact on its profile characteristics (cf. section 3.2).
 329 Even when the two datasets cover the same latitude range, we note in Figures 4 & 5
 330 that the altitude of the clouds for MY 35 does not exceed 85 km (consistent with the
 331 results presented in Clancy et al. (2019)), while small-particle clouds ($r_{\text{eff}} \leq 0.1 \mu\text{m}$)
 332 are observed up to 100 km at the beginning of the GDS. In particular, we observe
 333 at the onset of the GDS ($L_s = 195.5^\circ$, MY 34) a profile in the Southern hemisphere
 334 (62°S) with a cloud that extends up to 92 km along with large water ice crystals
 335 ($r_{\text{eff}} \sim 2 \mu\text{m}$) around 58 km, which contrasts with the observations acquired under the
 336 same conditions during MY 35 with clouds that do not extend over 40 to 50 km.

337 The period around $L_s \sim 240^\circ$ in the Northern hemisphere is also interesting, It
 338 not only corresponds to the end of the MY 34 GDS, but also to the maximum of dust
 339 activity for a regular (i.e., without GDS) year (Lemmon et al., 2015). We also note
 340 the presence of a regional dust storm during this period in MY 35 (cf. section 3.1).
 341 We observe that the maximum altitudes of the clouds are quite similar in both cases,
 342 i.e., between 70 and 85 km. However, we do not observe the large water ice crystals
 343 ($r_{\text{eff}} > 1 \mu\text{m}$) during the MY 35 dust storm that were seen between 50 and 60 km in
 344 MY 34 during the GDS.

345 Regarding the larger water ice crystals ($r_{\text{eff}} \sim 1.5 - 2 \mu\text{m}$), whose detection
 346 between 55 and 65 km during the GDS was surprising (Stcherbinine et al., 2020),
 347 similar sizes can be observed up to 55 km in the equatorial regions in MY 35 (i.e.,
 348 no GDS). However, these larger crystals generally remain confined to altitudes below
 349 50 km. For smaller particle sizes ($r_{\text{eff}} \sim 1 - 1.5 \mu\text{m}$), we observe that while the
 350 detections during the GDS are mostly concentrated around 60 km of altitude, they are
 351 typically found near 50 km during the same period in MY 35 (cf. Figure 4). Similarly,
 352 we also observe in Figure 5 that the detections of water ice crystals with sizes $0.13 \mu\text{m}$
 353 $\leq r_{\text{eff}} \leq 0.5 \mu\text{m}$ that are detected between 75 and 85 km during the GDS (panel a) are
 354 not present in the following year (panel b), where this size of crystals is not observed
 355 above 75 km. Thus, the increase of the average altitude of the clouds during the GDS
 356 does not affect all the ice particles in the same way, rather it depends on their size:

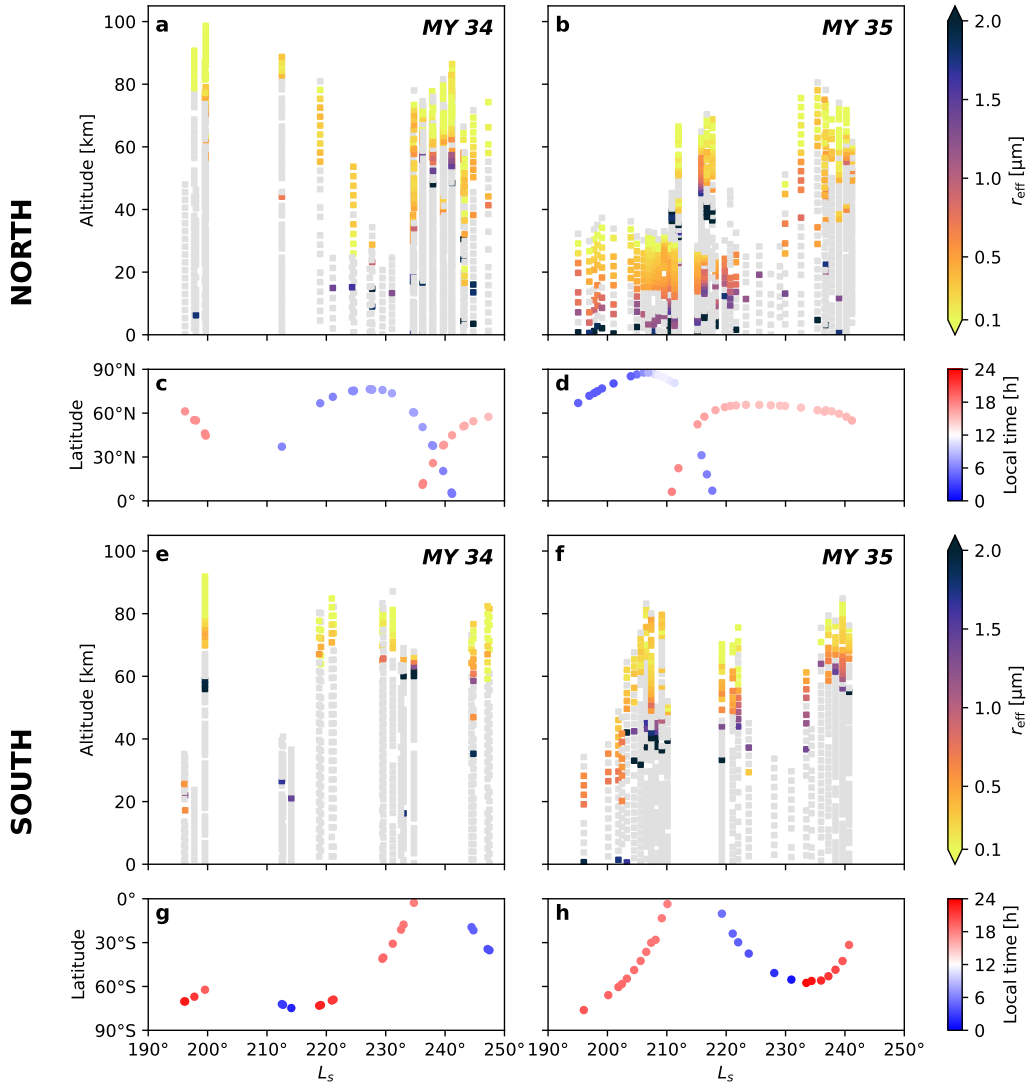


Figure 5. Vertical profiles of water ice clouds in the Martian atmosphere as observed by ACS-MIR between $L_s = 190^\circ$ and $L_s = 250^\circ$ over MY 34 (a & e) and MY 35 (b & f). Panels (c, d, g & h) indicate the latitudes and local times of the observations presented in panels (a, b, e & f) respectively. Observations without water ice detection are in gray.

357 the increase of the maximum altitude of the smallest crystals ($r_{\text{eff}} \leq 0.1 \mu\text{m}$) is 20 km
 358 more during the storm, and only 10 km for those between 0.1 and $0.5 \mu\text{m}$.

359 We also observe that the impact of the MY 34 GDS on the clouds and aerosols is
 360 stronger at high latitudes than in the equatorial region. Indeed, it has been reported
 361 that the altitude of clouds does not vary as a function of latitude during the GDS,
 362 which tends to homogenize the vertical structure of the atmosphere across the planet
 363 (Stcherbinine et al., 2020; Luginin et al., 2020; Liuzzi et al., 2020). This effect is
 364 particularly visible in the Southern hemisphere (panels e to h of Figure 5). Thus, while
 365 the typical maximum altitude of equatorial clouds varies from ~ 80 km to ~ 90 km
 366 during the GDS, it moves from ~ 40 to ~ 80 km for latitudes poleward of 60°S and
 367 60°N . In addition, Figure 4b shows that the detections of $2 \mu\text{m}$ ice crystals at 60 km
 368 during the GDS in the Southern latitudes are significantly higher than all the other
 369 detections in MY 35 that usually do not exceed 20 km of altitude for this particle size.
 370 Although one exception is a cloud with crystals of $r_{\text{eff}} = 2.2 \mu\text{m}$ detected at 40 km
 371 in the Southern hemisphere at $L_s = 321^\circ$ (MY 35), which is still 20 km below the
 372 altitude where such clouds were detected during the GDS.

373 4 Aerosols extinction

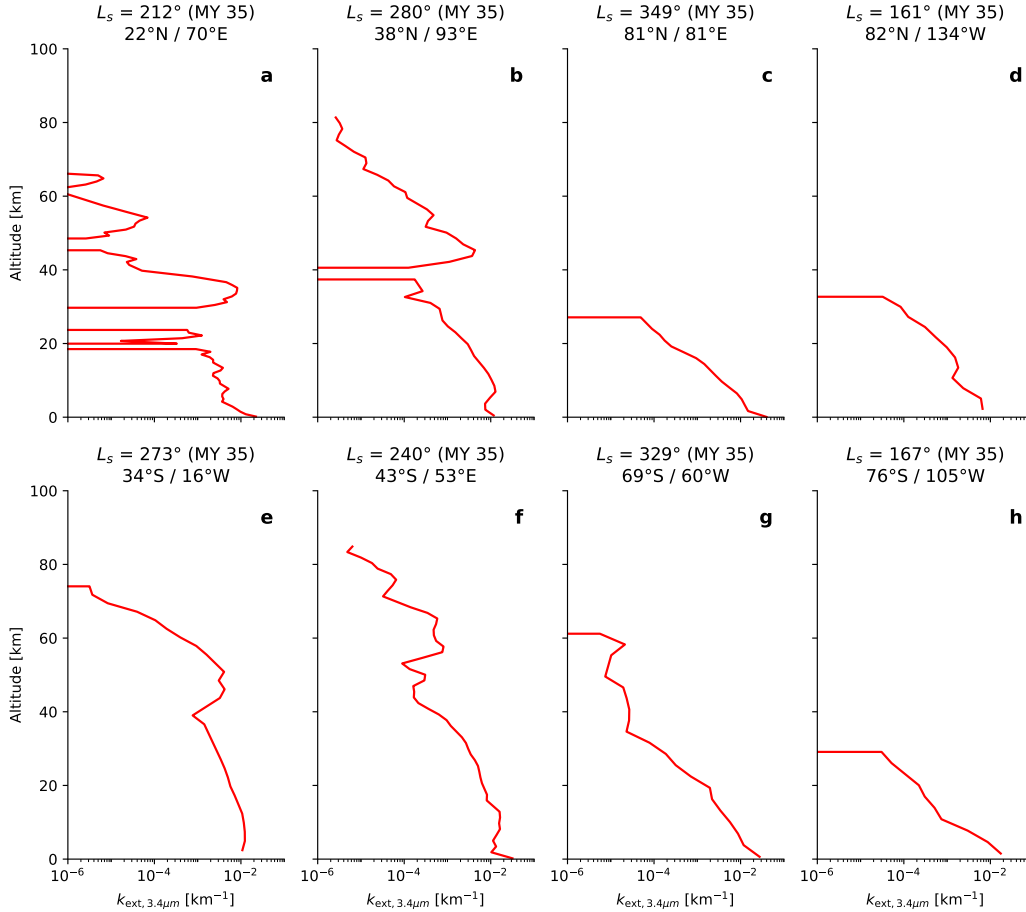


Figure 6. Vertical profiles of the derived extinction coefficient k_{ext} at $\lambda = 3.4 \mu\text{m}$ from 8 different ACS-MIR observations acquired during MY 35 at various latitudes. The light blue areas represent the altitudes where water ice clouds have been identified.

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4.1 Study of individual profiles

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Along with particle size, the extinction coefficient is an important property. It will influence the amount of incoming solar energy that will be thermally transferred to the atmosphere (Gierasch & Goody, 1972; Montmessin et al., 2002; Madeleine et al., 2011, 2012). The SO geometry used by ACS-MIR gives us direct access to the vertical profiles of the atmospheric extinction at each observed wavelength through the vertical inversion technique. In fact, as described in Section 2 and Stcherbinine et al. (2020), this step has already been performed as part of our derivation of the properties of the water ice clouds.

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Figure 6 shows 8 extinction profiles at $\lambda = 3.4 \mu\text{m}$ acquired by ACS-MIR at various latitudes during MY 35, showing altitudes where water ice particles have been detected in each profile. Because $3.4 \mu\text{m}$ is located on the edge of the $3 \mu\text{m}$ band, it is less sensitive to the composition of the atmospheric layer (water ice or dust) and to the size of the water ice crystals (Vincendon et al., 2011). As a result, it is more indicative of the general aerosol extinction of the atmosphere (compared to the extinction at $3.2 \mu\text{m}$ that highlight the presence of small-grained water ice crystals). We observe a difference between polar latitudes (c, d, g & h) and more tropical or equatorial observations (a, b, e & f). Indeed, while the main cloud layers are typically observed above 50 km in the equatorial profiles, we also note the presence of additional layers at lower altitudes (around 10 km). This is particularly clear in profile b, where the extinction goes from $7 \cdot 10^{-3} \text{ km}^{-1}$ at 3 km to $1.3 \cdot 10^{-2} \text{ km}^{-1}$ at 7 km before progressively decreasing to 10^{-4} km^{-1} at 40 km. Then, the extinction increases again up to $4 \cdot 10^{-3} \text{ km}^{-1}$ between 40 km and 52 km, which corresponds to the bottom of the water ice layer that extends from 47 km to 82 km. The presence of multiple high-extinction layers has been previously observed by SPICAM (Spectroscopy for the Investigation of the Characteristics of the Atmosphere of Mars) onboard Mars Express (Fedorova et al., 2009, 2014).

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We also note in profiles b & e that we only detect water ice crystals in the upper part of the detached layers and above, which means that the lower altitudes of these layers are primarily composed of either dust or large water ice crystals. Similarly, we observe in profile a some detections of large ($r_{\text{eff}} = 1.5 - 3 \mu\text{m}$) water ice crystals between 23 km and 36 km of altitude, which suggests the presence of water ice and not only dust in these lower layers. This vertical structure of ice layers capping the dust layers has already been reported with limb observations (Smith et al., 2013). In addition, clouds are detected at the top of the profiles associated with extinction down to a few 10^{-5} km^{-1} . The detection of these very tenuous water ice hazes is made possible by the high sensitivity provided by ACS-MIR and the SO geometry technique.

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4.2 Latitudinal variations

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Figures 7 & 8 present vertical profiles of the extinction coefficient k_{ext} at $3.4 \mu\text{m}$ acquired by ACS-MIR between $L_s = 140^\circ$ (MY 35) and $L_s = 90^\circ$ (MY 36), filtered by latitude and L_s to highlight the seasonal and latitudinal variations of the atmospheric vertical structure (similarly to Figures 2 & 3). We observe in Figure 7, similarly to what has been noted above for the global behavior of the clouds, that there is a strong latitudinal dependence in the atmospheric extinction profiles: a given k_{ext} value is observed at higher altitude close to the equator than in the polar regions. This is in agreement with previous retrievals of the atmospheric dust extinction from Mars Climate Sounder observations (Kleinböhl et al., 2015). This variation is only about 10 km for the lower layers of the atmosphere, but between $\sim 30^\circ\text{S}$ and 55°N we observe in most of our profiles the presence of a second detached atmospheric layer (i.e., atmospheric extinction decreases with altitude then increases again by at least one order of magnitude) around $\sim 30\text{--}55$ km (as previously noted in individual profiles

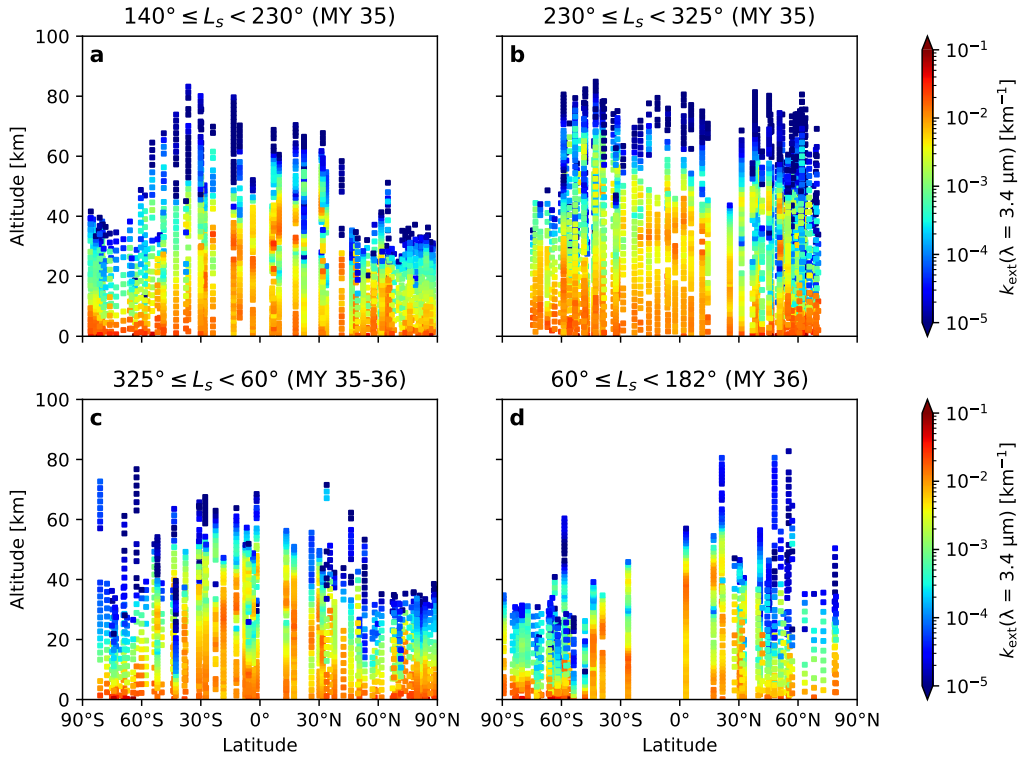


Figure 7. Vertical profiles of the measured extinction of the Martian atmosphere ($k_{\text{ext}} = d\tau/dz$, in km^{-1}) at $3.4 \mu\text{m}$ as observed by ACS-MIR between $L_s = 140^\circ$ (MY 35) and $L_s = 182^\circ$ (MY 36) as a function of the latitude of the observation. To highlight the latitudinal variations of the clouds by getting rid of their seasonal dependency (discussed in section 3.2), each panel represents the profiles grouped by ranges of L_s .

425 from Figure 6). Thus, extinction values of $\sim 10^{-2} \text{ km}^{-1}$ can be observed up to 50 km
 426 (i.e., 40 km above the altitude where they are observed around 60°N/S) in these layers.
 427 This suggests that when a large quantity of aerosols are raised above a certain altitude
 428 in the atmosphere ($\sim 30 \text{ km}$) they do not remain in one single atmospheric layer but
 429 are split in two distinct layers, with a significant decrease of the atmospheric extinction
 430 (by one order of magnitude) between them.

431 4.3 Seasonal variations

432 Along with the latitudinal variations, we also present in Figure 8 the evolution
 433 of the atmospheric extinction profiles with the L_s . We observe that the variations
 434 of altitude for a given extinction value are of lower amplitude than the variations of
 435 the average water ice clouds altitude for the same profiles (cf. section 3.1). Between
 436 90°S and 55°S (panel c), the altitude where $k_{\text{ext}} \sim 5 \cdot 10^{-3} \text{ km}^{-1}$ goes from 5 km
 437 at $L_s = 180^\circ$ to 30 km at $L_s = 320^\circ$, and then back to 5 km at $L_s = 30^\circ$. We
 438 also note that when the haze top altitude increases, for a fixed range of latitudes, it
 439 is largely due to the expansion of low-extinction layers to the higher altitudes rather
 440 than a shift of the vertical structure of the extinction profile, i.e., a given extinction
 441 value remains approximately associated with the same altitude, but layers with lower
 442 k_{ext} values appear at the top of the profiles, which is coherent with NOMAD/IUVS

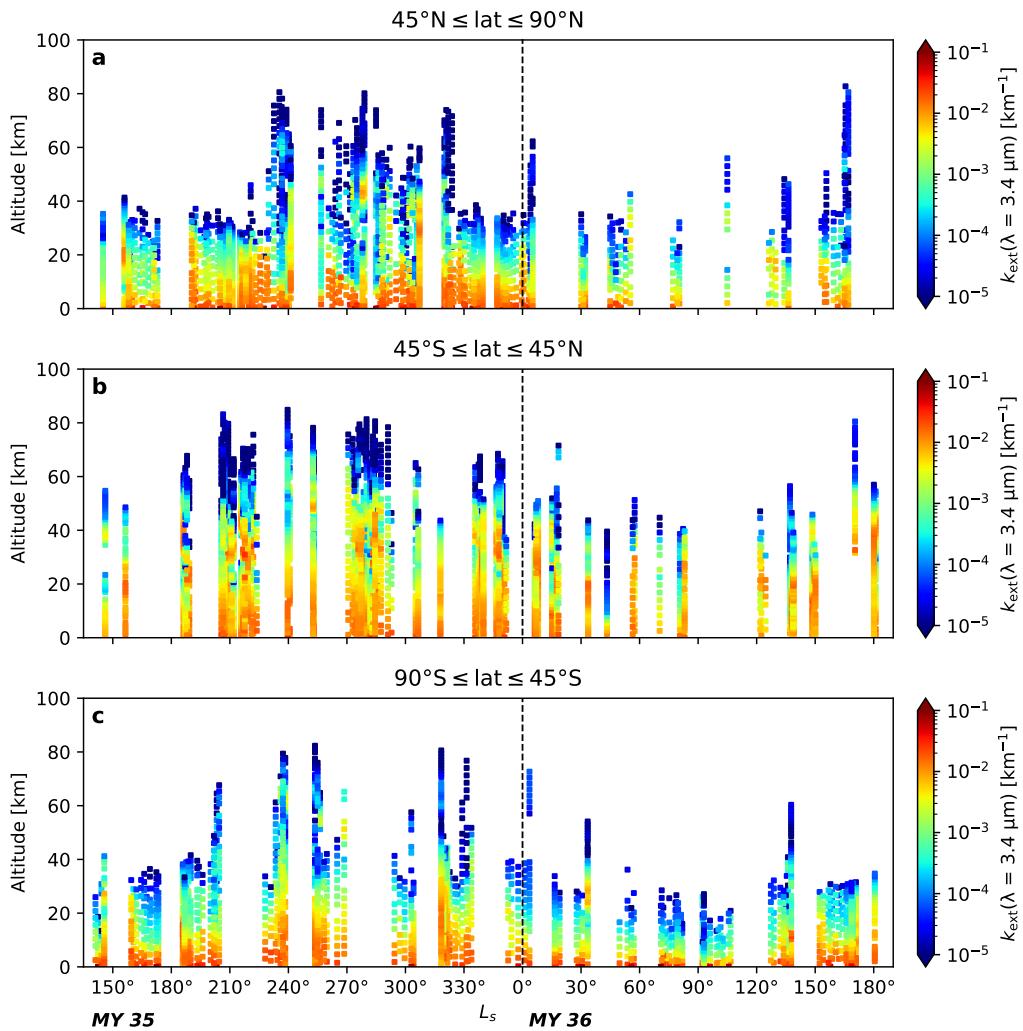


Figure 8. Vertical profiles of the measured extinction of the Martian atmosphere ($k_{\text{ext}} = d\tau/dz$, in km^{-1}) at $3.4 \mu\text{m}$ as observed by ACS-MIR over mid-MY 35 to mid-MY 36. To highlight the seasonal variations of the clouds by getting rid of their latitudinal dependency, each panel represents the profiles grouped by ranges of latitude (North/Equatorial/South).

443 measurements over MY 35 (Streeter et al., 2022) and previous SPICAM observations
 444 from MY 27 to MY 31 (Määttänen et al., 2013).

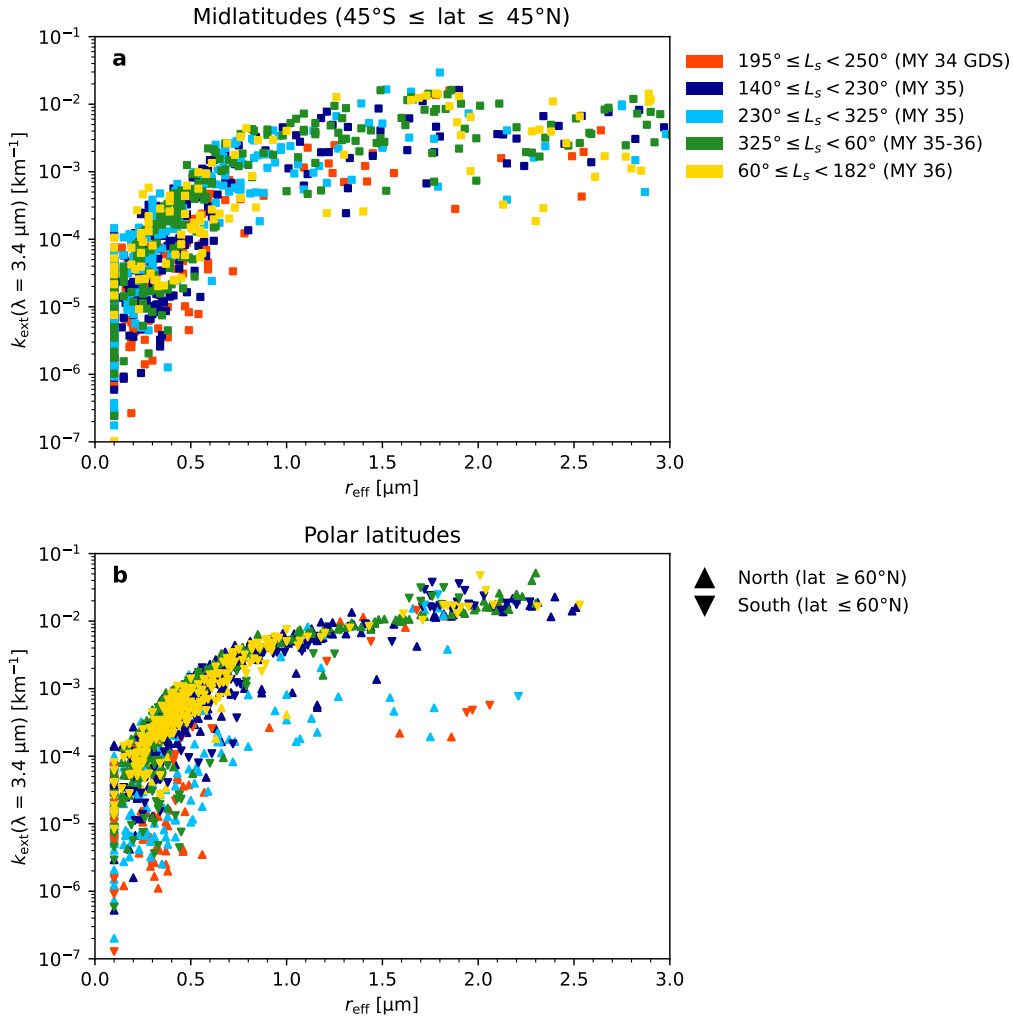


Figure 9. Distribution and size of the water ice crystals within the clouds as a function of their extinction coefficient k_{ext} and L_s of observation, for 3 ranges of latitude: midlatitudes (panel a), Northern and Southern polar latitudes (panel b). Observations acquired during the MY 34 GDS are shown in red for comparison with data from non-GDS years (MY 35 & 36).

4.4 Extinction vs crystal size

Figure 9 presents the extinction coefficient k_{ext} of the clouds at $\lambda = 3.4 \mu\text{m}$ as a function of the retrieved size of the water ice crystals r_{eff} for 3 ranges of latitudes (midlatitudes, northern and southern polar latitudes) and 5 ranges of L_s , including the MY 34 GDS. We observe at all latitudes the presence of 2 regimes depending on the size of the water ice crystals: for $r_{\text{eff}} \leq 1 \mu\text{m}$ the extinction k_{ext} increases with r_{eff} , while for larger particles the extinction of the clouds remains in the same range of values when r_{eff} varies between 1.5 and 3 μm .

453 We can still note that, if we exclude the data from the MY 34 GDS and the
 454 MY 35 dust storm (discussed further in the next paragraph), the points in the polar
 455 regions (panel b) have less scatter in terms of k_{ext} than in the midlatitudes (panel a),
 456 especially for the large particles ($r_{\text{eff}} \geq 1.5 \mu\text{m}$). This may be related to the smaller
 457 variations of the cloud altitudes in these regions (cf. Figure 4), which suggests a more
 458 stable vertical distribution of the properties of the clouds (both k_{ext} and r_{eff}). Also,
 459 for fixed values of r_{eff} and λ the extinction coefficient k_{ext} is solely a function of the
 460 number of particles that have scattered the light along the line of sight. Thus, the
 461 small dispersion in the cloud extinction values in the polar regions indicates that the
 462 number density of the water ice crystals within the clouds is overall constant over the
 463 year.

464 However, we observe in Figure 9b that some detections of particles with r_{eff}
 465 between 0.9 and 2.2 μm are associated with lower extinction values than the ones
 466 from the typical trend: $k_{\text{ext}} \sim 10^{-4}$ – 10^{-3} km^{-1} instead of $k_{\text{ext}} \sim 10^{-2} \text{ km}^{-1}$. A
 467 noteworthy point is that most of these detections correspond to observations that
 468 have been mostly acquired either during the MY 34 GDS or during the regional dust
 469 storm in the Northern hemisphere in MY 35. In addition, even for lower size of ice
 470 crystals, the GDS detections are correlated with lower k_{ext} values. These observations
 471 with lower extinction values are associated with detections at higher altitudes than
 472 those typical for these sizes of crystals (cf. Figure 4b). Thus, while water ice clouds
 473 are observed at higher altitudes during dust storms, their ice crystal number density
 474 is lower than what is typically observed outside these events.

475 5 Comparison with the Mars PCM

476 Our dataset provides precise and systematic monitoring of the cloud altitude,
 477 effective radius, and extinction over two Martian years. Thus, these new observational
 478 constraints can be compared to predictions by numerical simulations such as GCMs.
 479 Here we will compare our results with the Mars PCM (Forget et al., 1999; Montmessin
 480 et al., 2002; Madeleine et al., 2012; Navarro et al., 2014). We will focus on the
 481 observations acquired during the second half of MY 35, a non-GDS year for which we
 482 have access to a climatology of the atmospheric dust distribution (Montabone et al.,
 483 2015, 2020).

484 Even though the size of the ice crystals within the clouds is among the quantities
 485 available from version 5 of the Mars PCM, the model sizes are systematically and
 486 significantly larger than the values retrieved in this study as well as in previous ones
 487 (e.g., Wolff & Clancy, 2003; Vincendon et al., 2011; Clancy et al., 2019; Stcherbinine
 488 et al., 2020; Luginin et al., 2020; Liuzzi et al., 2020). Investigations on this discrepancy
 489 are currently ongoing by the Mars PCM team. Thus, we do not discuss further the
 490 crystals' size here, but we instead focus on the altitude of the clouds, through the
 491 volume mixing ratio of H_2O ice crystals in the atmosphere.

492 5.1 Global comparison for MY 35

493 Figure 10 shows the comparison between the water ice clouds detected by ACS-
 494 MIR (left column) and predicted by the Mars PCM (right column) from $L_s = 140^\circ$
 495 to $L_s = 360^\circ$ (MY 35) for three ranges of latitude (North and South high latitudes,
 496 mid-latitudes) in order to delineate the latitudinal from the seasonal trends.

497 We observe that the seasonal trend shown in the ACS-MIR data for the mid-
 498 southern latitudes (i.e., increase of the altitude of the clouds around the perihelion) in
 499 the panels c & e is well reproduced by the PCM (panels d & f). However, the strong
 500 increase observed in the altitude of the clouds associated with the regional dust storm
 501 in the Northern hemisphere is only reproduced for a few model profiles (panel b).

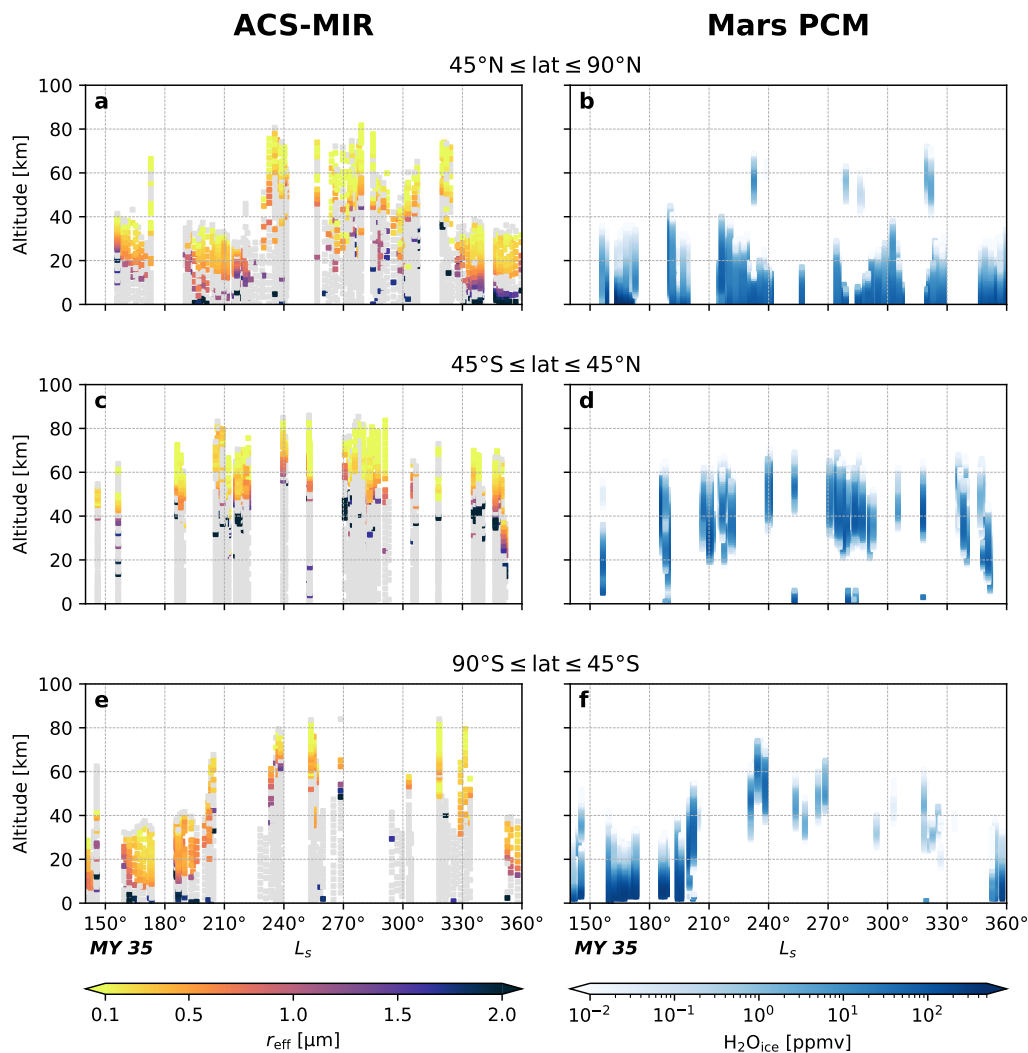


Figure 10. Comparison between the ACS-MIR cloud detections (left) and the water ice clouds predicted by the Mars PCM (right) over the second half of MY 35. Similarly to Figure 8 the profiles are filtered as a function of latitude in each panel to decorrelate between the seasonal and the latitudinal trends.

502 We also observe that in the polar regions, clouds are present in the PCM down
 503 to the surface while they are not detected at low altitudes in the ACS-MIR retrievals,
 504 which is not inconsistent. Indeed, as mentioned in Stcherbinine et al. (2020) our
 505 method is not sensitive to the larger ($r_{\text{eff}} \geq 2 - 3 \mu\text{m}$) ice particles, and to mixed dust-
 506 ice layers dominated by dust. Thus, it is likely that we do not detect some atmospheric
 507 layers with water ice crystals at the bottom of the profiles.

508 We observe another difference between the model and the data: the maximum
 509 altitude of the clouds in the PCM is usually lower than that observed: by ~ 10 km in
 510 the polar regions and up to ~ 20 km in the mid-latitudes. This difference is this time
 511 not linked with an observational bias (we detect more clouds layers than predicted by

512 the model). Further investigations are required to understand why the simulations do
 513 not reproduce precisely observed altitudes.

514 5.2 Comparison on individual profiles

515 Figure 11 presents a close-up view of the comparison between the ACS-MIR
 516 retrievals and the Mars PCM predictions for 6 individual profiles taken between $L_s =$
 517 121° and $L_s = 352^\circ$ in MY 35. For each profile, the information about the altitude of
 518 the water ice cloud in both the ACS-MIR observation and the Mars PCM simulations
 519 are provided, as well as the extinction coefficient at $\lambda = 3.2 \mu\text{m}$.

520 We observe that the modeled extinction values (k_{ext}) at $\lambda = 3.2 \mu\text{m}$ predicted
 521 by the model match the values retrieved by ACS-MIR, at least for the altitudes where
 522 the PCM produces clouds (panels a, c, d & e). However, we still note a shift of about
 523 10 to 20 km in the altitudes where a cloud layer is typically detected versus where
 524 they are predicted by the PCM (panels a & f). Despite the limitations of our method,
 525 it is clear that ACS observes clouds higher than predicted by the PCM. Indeed, by
 526 looking at the ACS-MIR k_{ext} profiles, we suspect that some clouds layers are missing
 527 at the bottom of the clouds (panels a, e & f) as the detached layer that is visible in
 528 the extinction profile extends to a lower altitude than the water ice cloud detection.
 529 However, we note that when a cloud layer is not predicted to extend as high as it
 530 is observed by ACS, the model atmospheric extinction also decreases faster than the
 531 measured one (panels a & f), which strengthens the validity of the differences noted
 532 between PCM predictions and ACS observations for these altitudes. Plus, we also find
 533 that the largest discrepancies are observed for the higher altitude clouds, and seem
 534 likely to be related to the difficulty of the model to reproduce in general the
 535 high-altitude detached layers of water ice detected by ACS (panels b & d).

536 Thus, clouds predicted at lower altitudes in the Mars PCM compared to the
 537 ACS-MIR observations may be the manifestation of the limits of our retrieval method,
 538 while the lack of the upper layers in the PCM is a shortcoming in the current model.

539 6 Conclusion

540 In this paper, we present the results of our study on the identification and char-
 541 acterization of the Martian water ice clouds properties from IR SO data acquired by
 542 the ACS-MIR instrument onboard TGO between $L_s = 163^\circ$ (MY 34) and $L_s = 181^\circ$
 543 (MY 36). Using the methodology previously described in Stcherbinine et al. (2020),
 544 we are able to simultaneously detect the presence of water ice clouds in the Martian at-
 545 mosphere, and constrain the size of their crystals at each observed altitude. As TGO's
 546 orbit allows ACS to span all the latitudes between 80°S and 80°N in about $\sim 20^\circ$ of
 547 L_s , this new dataset allows us to monitor the properties of the water ice clouds as a
 548 function of both the season and the latitude. In addition, the data acquired during
 549 MY 35 give us reference measurements at the same period to be compared to the
 550 observations acquired during the MY 34 GDS to better constrain the effects of such
 551 an event.

552 The main results are summarized below:

- 553 • The Solar Occultation technique used by ACS-MIR provides highly sensitive
 554 measurements that allow us to observe optically thin clouds ($k_{\text{ext}}(\lambda = 3.2 \mu\text{m}) \sim$
 555 10^{-4} km^{-1}). Such clouds are typically much harder to detect through other
 556 observing geometries (on-disk or limb scattering).
- 557 • Where thick water ice clouds appear locally in the Martian atmosphere (Smith
 558 et al., 2013; Wolff et al., 2019; Szantai et al., 2021), our SO observations reveal
 559 the quasi-systematic presence of small water ice crystals within the upper layers

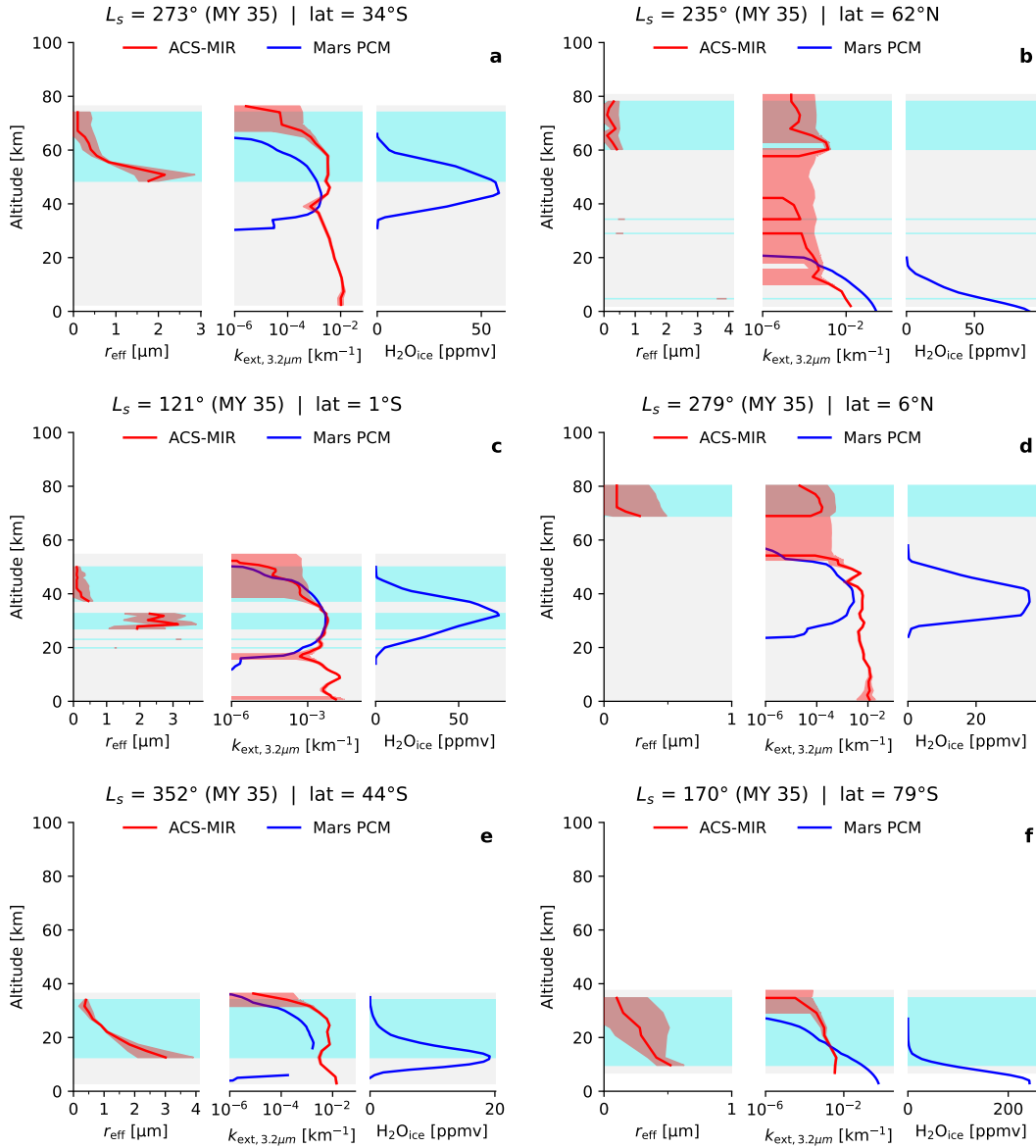


Figure 11. Comparison between the ACS-MIR retrieval (red lines) and the Mars PCM prediction (blue lines) for 6 individual profiles from MY 35. Each panel shows the water ice r_{eff} retrieved by ACS-MIR (left), the atmospheric extinction (k_{ext}) at $\lambda = 3.2 \mu\text{m}$ (center) and the amount of water ice predicted by the Mars PCM (right). The light blue areas represent the altitudes where water ice clouds have been identified in the ACS-MIR data.

- 560 of the atmospheric aerosols (at least near the morning and evening terminators;
561 i.e., at local times $\sim 06:00$ and $18:00$).
- 562 • The decrease of the ice crystals size as the altitude increases, previously noticed
563 in the GDS study (Stcherbinine et al., 2020) and by other TGO studies (Luginin
564 et al., 2020; Liuzzi et al., 2020), remains the observed behavior in this extended
565 2-MY dataset.
 - 566 • The comparison between the perihelion periods of MY 34 and MY 35 allows us
567 to contrast conditions during the MY 34 GDS with those of the same season
568 in the absence of such a dramatic atmospheric event. It can be seen that the
569 global altitude of the water ice clouds has increased by 10 km during the GDS
570 compared to a more typical year, and this altitude increase can be 20 km for
571 the smallest ice crystals ($r_{\text{eff}} \leq 0.1 \mu\text{m}$).
 - 572 • No large-grained clouds ($r_{\text{eff}} \geq 1.5 \mu\text{m}$) have been observed at high altitudes (60
573 to 65 km range) during the non-GDS year, while they were detected during the
574 MY 34 GDS. However, such clouds have been detected at 55 km (a relatively
575 high altitude for such large ice crystals) close to the equator without being
576 associated with a particular dust storm event.
 - 577 • We also note that this elevation of the water ice clouds during a dust storm
578 corresponds to a decrease of the typical extinction values associated with a
579 specific ice crystal sizes.
 - 580 • We observe variations of 20 to 40 km between the average altitude of the water
581 ice clouds during summer and winter for both hemispheres, with water ice clouds
582 often detected up to 80 km during summer seasons. The variations are stronger
583 in the high latitudes compared to those in the equatorial regions.
 - 584 • In addition to the seasonal variations, clouds are also detected typically 20 to
585 40 km higher close to the equator than in the polar regions.
 - 586 • Finally, comparison between the ACS-MIR retrievals and the Mars PCM shows
587 that the water ice clouds are usually predicted at lower altitudes in the model,
588 up to 10 km lower in the polar regions and 20 km in the midlatitudes.

589 In conclusion, we investigated the impact of season and latitude on the altitude
590 and the size of the water ice crystals. These results have been compared to predictions
591 from climate models, showing a good agreement overall, while revealing some regions
592 of discrepancies associated with the cloud altitudes. The quasi-systematic detection
593 of water ice clouds in our observations highlights the importance of considering these
594 clouds in studies of the current Martian atmosphere and climate.

595 Data Availability

596 Raw ACS data are available on the ESA PSA at [https://archives.esac.esa](https://archives.esac.esa.int/psa/#!Table%20View/ACS=instrument)
597 [.int/psa/#!Table%20View/ACS=instrument](https://archives.esac.esa.int/psa/#!Table%20View/ACS=instrument). Derived particles sizes and atmospheric
598 extinction profiles can be found in Stcherbinine (2022).

599 Acknowledgments

600 ExoMars is a space mission of ESA and Roscosmos. The ACS experiment is led by IKI
601 Space Research Institute in Moscow. The project acknowledges funding by Roscosmos
602 and CNES. Science operations of ACS are funded by Roscosmos and ESA. Science
603 support in IKI is funded by Federal agency of science organization (FANO). MJW
604 acknowledges support from Jet Propulsion Laboratory subcontract contract 1551112.

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