



Publication Year	2020
Acceptance in OA	2022-06-20T09:54:05Z
Title	Similarities and Differences of Global Dust Storms in MY 25, 28, and 34
Authors	WOLKENBERG, PAULINA MARIA, GIURANNA, Marco, Smith, M. Â. D., GRASSI, Davide, Amoroso, M.
Publisher's version (DOI)	10.1029/2019JE006104
Handle	http://hdl.handle.net/20.500.12386/32409
Journal	JOURNAL OF GEOPHYSICAL RESEARCH (PLANETS)
Volume	125

1 **Similarities and differences of global dust storms in MY 25, 28 and 34**

2
3 **P. Wolkenberg¹, M. Giuranna¹, M. D. Smith², D. Grassi¹ and M. Amoroso³**

4 ¹ Istituto di Astrofisica e Planetologia Spaziali – Istituto Nazionale di Astrofisica (IAPS – INAF)
5 via del Fosso del Cavaliere, 100, 00133 Rome, Italy.

6 ² NASA, Goddard Space Flight Center, USA

7 ³ Agenzia Spaziale Italiana (ASI), Sede di Matera, Italy

8
9
10
11 Corresponding author: Paulina Wolkenberg (paulina.wolkenberg@inaf.it)

12 ORCID 0000-0001-6769-3719

13
14
15 **Key Points:**

- 16
- 17 • The three global dust storms considered here, have a similar duration of expansion phase.
 - 18 • Global dust storm in Mars Year 34 is characterized by a shorter decay phase in comparison
19 with the decay phases in MY 28 and MY 25 storms.
 - 20 • Relatively large dust opacities are observed over Hellas and southern polar edge regions
21 long before the onset of MY 28 storm.

22 **Abstract**

23 To better understand the dust cycle on Mars during years with planet-encircling dust storms, we
24 analyze the last three events that took place in Mars Year (MY) 25, MY 28, and MY 34. Global
25 dust storms that occurred in MY 25 and MY 34 (June 2018) were taking place during equinox
26 while the MY 28 storm had an onset after perihelion. Before the expansion phase of the MY 25
27 and MY 34 storms we find similar regions (northern rim of Hellas, Arabia Terra and Utopia
28 Planitia) where dust is present. Possible precursor dust storms over Hellas and the southern polar
29 cap edges were observed during MY 28 as a component of background dust activity. These features
30 are not found in equinoctial dust storms on this scale. Dust during the MY 25 and MY 34 storms
31 encircled the entire planet by the similar season ($L_s = 193^\circ$). The MY 34 storm is characterized by
32 a shorter decay phase compared to the events in MY 25 and MY 28. Dust opacity is correlated
33 with atmospheric temperatures at 0.5 mbar and nighttime surface temperatures, while daytime
34 surface temperatures are anti-correlated with dust opacity.

35 **Plain Language Summary**

36 On Mars, one of the most variable atmospheric components is dust. We study it to better
37 understand the cycle of its presence and absence in the Martian atmosphere. One of the particular
38 phenomena of this cycle is the global dust storm which enshrouds the entire planet. Using data
39 measured by two spectrometers, Thermal Infrared Emission Spectrometer (TES) onboard the Mars
40 Global Surveyor and Planetary Fourier Spectrometer (PFS) onboard the Mars Express orbiters, we
41 investigate similarities and differences of the three most recent global dust storms that have
42 occurred on Mars. We find similarities between the two global dust storms that took place in 2001
43 and 2018 (Mars Year 25 and 34). They encircled the entire planet in the equinox season. The
44 storms in Mars Year (MY) 25 and MY 28 dissipated over a very long timescale compared to the
45 global dust storm of MY 34. Local dust storms in the northern rim of Hellas, Arabia Terra and
46 Utopia Planitia were observed before the onset of MY 25 and MY 34 global events.

47 **1 Introduction**

48 Global dust storms occur on Mars at irregular intervals, approximately every 3–5 Mars Years.
49 They mostly originate in the southern hemisphere near perihelion from the combination of multiple
50 local and regional dust lifting events (Haberle et al., 2017, Smith, 2002). However, global dust
51 storms have also been observed during the equinox seasons. Two seasons are recognized in the
52 dust cycle through the whole Martian year. Elevated dust loadings mostly occur in the atmosphere
53 from 130° to 360° of L_s (Haberle et al., 2017). The clear season (non-dusty, $0 - 130^\circ$) are described
54 by low background level of $9\text{-}\mu\text{m}$ dust opacity between 0.1 – 0.2 (Smith, 2008). During this period
55 no large dust storms are observed with exception of regions near the retreating north polar cap
56 edges (Smith, 2008). Observations of dust provide information on its location and time dependence
57 in the Martian atmosphere and on the surface. During the locations and seasons with high dust
58 loading, the thermal state of atmosphere and the atmospheric circulation are significantly affected
59 (Haberle et al., 2017; Cantor et al., 2001).

60
61 Dust lifted into the atmosphere works as a positive feedback for some aspects of the atmospheric
62 circulation (Haberle et al., 1993; Cantor et al., 2001). Haberle (1986) proposed an explanation for
63 a mechanism for the origin of global dust storms. Two regimes of Hadley circulation in the
64 southern and northern hemispheres compete against each other during the dust storm season
65 depending on the dust supply in each hemisphere. Dust can be transported from the southern

66 hemisphere to the northern via the Hadley circulation. This circulation can be intensified by dust,
67 and as a result global dust storms can develop, thereby suppressing the northern circulation built
68 by mid-latitude storm systems. On the other hand, in northern hemisphere baroclinic wave activity
69 can raise dust into the atmosphere when there is sufficient dust on the surface. Northern dust haze
70 can diminish the intensity of the Hadley circulation and in turn the surface stress in the southern
71 hemisphere precluding the emergence of global dust storms. In this paper, we analyze the last three
72 global dust storms (GDS), particularly the most recent one that occurred in MY 34 (Earth year
73 2018).

74

75 **2 Datasets**

76

77 **2.1 Planetary Fourier Spectrometer (PFS) dataset**

78

79 Data presented in this work (Wolkenberg et al., 2019) were retrieved from observations performed
80 by the Planetary Fourier Spectrometer (PFS) onboard the Mars Express in the long-wavelength
81 channel (LWC). They are part of a large dataset composed of observations beginning in 2004 (L_s
82 = 330° , MY 26) that continues to the present day (Giuranna et al., 2019). PFS measures radiation
83 in its LWC that is dominated by the thermal emission by the planet. A complete description of PFS
84 and its radiometric performance can be found in Formisano et al. (2005) and Giuranna et al.
85 (2005a, 2005b). In the LWC we observe some absorption features due to atmospheric constituents,
86 mainly CO_2 and aerosols (dust and water ice). Using a retrieval algorithm (Grassi et al., 2005) we
87 are able to derive information on the atmospheric parameters from these absorption bands.
88 Temperature profiles are obtained from an analysis of radiation in the main CO_2 absorption band
89 at 667 cm^{-1} ($15 \mu\text{m}$). Total dust and water ice content in the vertical column are retrieved from
90 wide absorption bands at 1075 cm^{-1} ($9 \mu\text{m}$) and 825 cm^{-1} ($12 \mu\text{m}$), respectively. All these
91 atmospheric parameters along with surface temperatures are retrieved simultaneously (Giuranna
92 et al., 2019). In this work we use mostly daytime PFS measurements to be consistent with TES
93 results.

94

95 **2.2 Thermal Emission Spectrometer (TES) dataset**

96

97 The Thermal Emission Spectrometer (TES) onboard the Mars Global Surveyor measured
98 radiances in the spectral range from $200 - 1600 \text{ cm}^{-1}$ which corresponds to LWC of PFS. A detailed
99 characterization of the TES instrument is found in Christensen et al. (2001). Hence, the TES dataset
100 includes similar atmospheric parameters such as temperature profiles (Conrath et al., 2000), and
101 dust and water ice aerosol optical depth (Smith et al., 2000a, 2000b, Bandfield and Smith, 2003).
102 The same absorption bands as for PFS were used to retrieve atmospheric parameters. The surface
103 temperatures are estimated from brightness temperatures at $\sim 1300 \text{ cm}^{-1}$ by using the improved
104 algorithm presented by Smith (2004). Thus, a direct comparison of the TES and PFS retrieval
105 datasets is possible (Wolkenberg et al., 2011). TES performed measurements at two local times
106 (LT) during day at 2 PM and during night at 2 AM. We use the daytime observations.

107

108 **3 Uncertainties of dust opacities**

109

110 Uncertainties of dust opacities derived from TES observations come from two main sources.
111 Random and systematic errors of instrument and its calibration are considered as a first source of

112 uncertainty (e.g., Pankine, 2015; 2016). However, we use only daytime dust opacities when spectra
113 have a sufficient contrast between surface and atmosphere. For those measurements, an
114 improvement of calibration has no contribution to the shapes and relative strengths of the aerosols
115 absorption features (Pankine, 2016). The other source takes into account errors in the retrieval
116 algorithm. The final, total uncertainties of dust opacities are around 0.05 or 10% of the total optical
117 depth (Smith, 2004). The PFS dust opacity uncertainties are based on the standard deviations of
118 the final covariance matrix of atmospheric parameters to be retrieved (Wolkenberg et al., 2018).
119 They found that the population of standard deviations of dust opacities were dependent on surface
120 temperatures. Thus, this population was divided into two groups with surface temperatures less
121 and larger than 220 K. For these two subsets the standard deviations were estimated using
122 histograms. For surface temperatures larger than 220 K the typical standard deviations of dust
123 opacities were 0.02 to 0.06, while for surface temperatures less than 220 K the standard deviation
124 was 0.11.

125

126 **4 Basic characteristics of three global dust storms**

127

128 We consider the dusty season as the time interval from $L_s \sim 135^\circ$ to 360° in each Martian year. Dust
129 opacity can be elevated during different time periods of the dusty season. We distinguish a period
130 of early-season and pre-solstice activity: $L_s \sim 135^\circ - 236^\circ$, a period of solstitial activity near the
131 South Pole: $L_s \sim 250^\circ - 300^\circ$, and a period of post-solstice activity: $L_s \sim 308^\circ - 336^\circ$ (Haberle et al.,
132 2017). Global dust storms are events where enhanced dust loading encircles the entire planet. They
133 take place during the dusty season with an irregular frequency averaging once every 3-6 Mars
134 Years. In this work, we analyze the last three global dust storms, which occurred in 2001 (MY 25),
135 2008 (MY 28), and 2018 (MY 34). The global dust storms that occurred during MY 25 and 34
136 began at $L_s \sim 185^\circ$ during the period of early-season pre-solstice activity. On the other hand, the
137 MY 28 dust storm began much later in season ($L_s \sim 265^\circ$) during the period of solstitial activity.
138 However, the MY 28 dust activity was not only constrained to near the South Pole. Elevated dust
139 loading was distributed throughout the entire planet as for north as 40°N during this storm. The
140 next section is dedicated to a presentation of the seasonal and spatial dust activity observed during
141 the last three global storms.

142

143

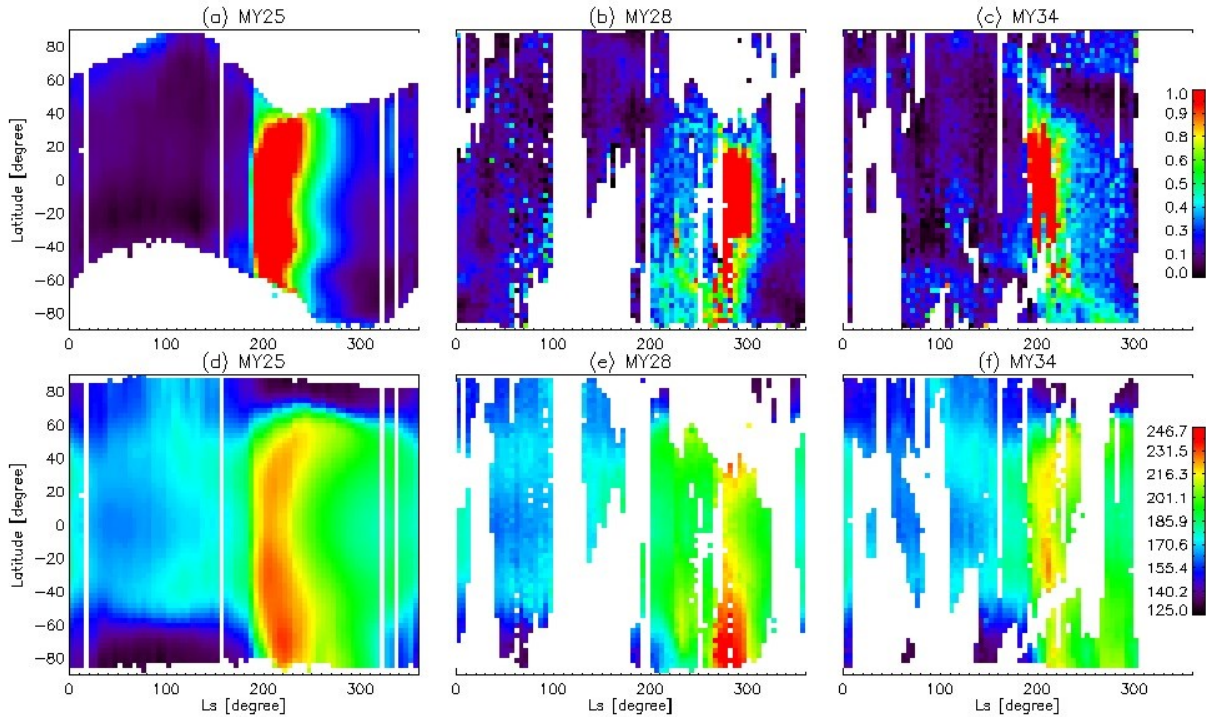
144 **5 Results**

145

146 In this study we mainly focus on analysis of the column-integrated dust optical depths in three
147 Martian years when global dust storms occurred. The column-integrated dust optical depth is
148 normalized to 610 Pa, according to the formula $\tau = 610 * \tau_0 / P_{\text{surf}}$, where τ_0 is the retrieved column-
149 integrated dust optical depth and P_{surf} is the surface pressure [Pa]. For the purpose of characterizing
150 the similarities and differences between the three global dust storms we use both the TES and PFS
151 datasets.

152

153 **5.1 Intercomparison of zonal mean opacities and temperatures for the three Global Dust** 154 **Storms**



155

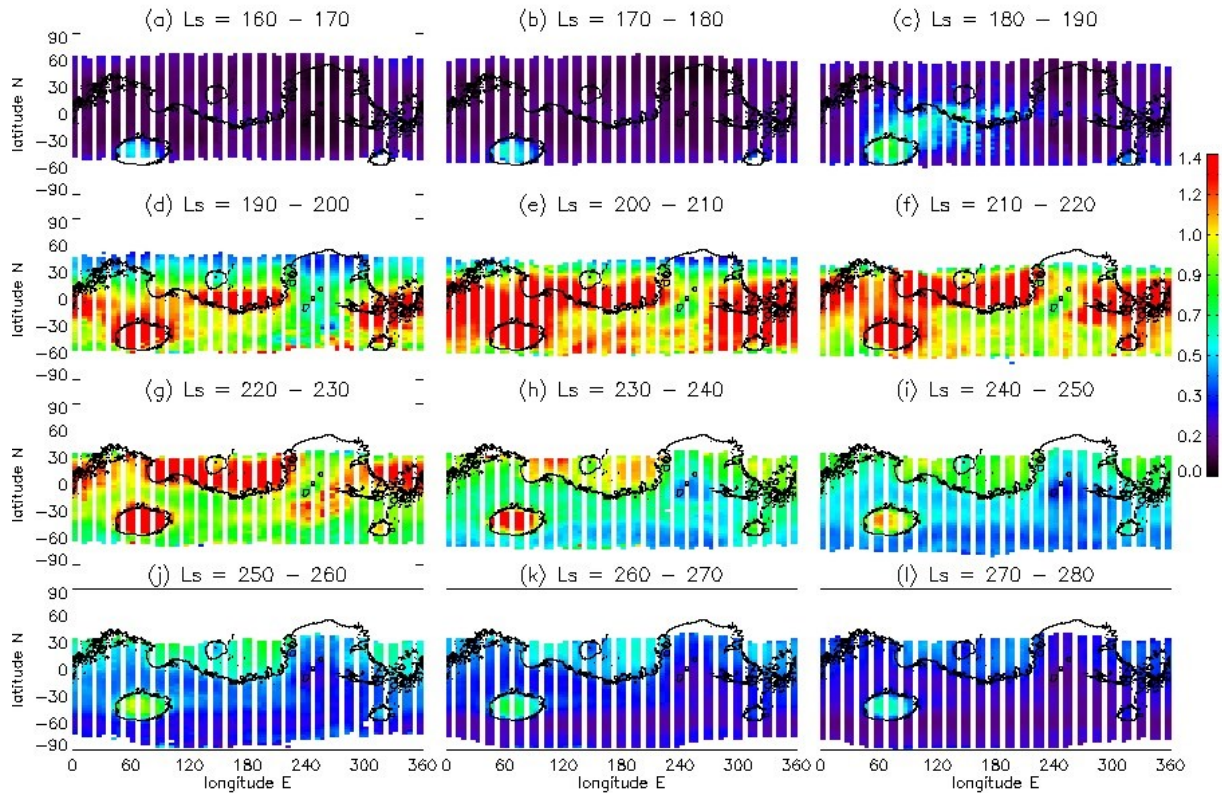
156 **Figure 1.** Zonally averaged dust opacities as a function of season and latitude for MY25 (a), MY
 157 28 (b) and MY34 (c) scaled by surface pressure. Zonally averaged atmospheric temperatures at
 158 0.5 mbar for MY25 (d), MY 28 (e) and MY34 (f). Data from MY 25 are retrieved from TES
 159 observations while data from MY 28 and MY 34 are retrieved from PFS observations. PFS dust
 160 opacities are provided for all available LTs. Atmospheric temperatures at 0.5 mbar were retrieved
 161 from TES and PFS measurements performed at LT 14 and from 8 – 18 LT, respectively.

162 Figure 1 presents seasonal and latitudinal variability of zonally averaged dust opacities and
 163 atmospheric temperatures at 0.5 mbar (~ 25 km altitude) for the three specific Martian years 25,
 164 28, and 34. Zonally averaged dust opacities have been binned by 5° in Ls and by 3° in latitude.
 165 Global dust storms were observed during MY 25, MY28 and MY 34 with dust opacities exceeding
 166 1 at 1075 cm^{-1} over a large fraction of the planet. The global dust storm in MY 25 had the longest
 167 duration compared to other global storms considered here (Fig. 1a). The MY 25 storm was
 168 apparently triggered from local dust storms (Fig. 2a) that developed near the northwestern rim of
 169 the Hellas basin at $L_s = 177^\circ$ (Strausberg et al., 2005). Probably, the local dust storms were induced
 170 by ice-cap thermal contrasts and slope-flows (Strausberg et al., 2005; Cantor, 2007). Then, a rapid
 171 expansion to the east was initiated at $L_s = 185^\circ$ (Smith et al., 2002; Strausberg et al., 2005). No
 172 expansion to the west was observed (Strausberg et al., 2005). When dust approached the western
 173 flanks of Tharsis, new lifting centers arose in southeastern Tharsis (Daedalia and Solis Planitia)
 174 (Strausberg et al., 2005). The longest-lived lifting center (86 sols) was found in Syria Planum –
 175 Claritas Fossae (Cantor, 2007). By $L_s = 193^\circ$ dust encircled the entire planet with lifting centers
 176 dominated in Syria Planum/Solis Planum /Daedalia Planum (Smith et al., 2002, Strausberg et al.,
 177 2005). Lifting centers over Hellas Planitia slowly diminished at this same time. The expansion
 178 phase (core of dust storm) came to an end at around $L_s = 200^\circ$ after which the fastest clearing was
 179 observed at high southerly latitudes with the slowest clearing in the northern hemisphere (Smith

180 et al., 2002). High dust opacities were still found at around $L_s = 210^\circ$ in $0 - 20^\circ\text{N}$ (Smith et al.,
181 2002). For this storm, the decay phase was characterized by a latitudinal gradient in the rate of
182 decrease in dust optical depth (Smith et al., 2002). This means that the decay phase appears at
183 different seasons and latitudes (Strausberg et al., 2005). For example, the decay phase can start at
184 southern latitudes earlier than at northern latitudes. The decay phase began at $L_s = 200^\circ$ and had a
185 duration of about 97 sols ($L_s = 263^\circ$) although clearing continued beyond that date. However, some
186 dust lifting centers were still active as late as $L_s = 214^\circ$ (Strausberg et al., 2005). The atmospheric
187 clearing continued until $L_s = 304^\circ$ when nominal seasonal levels of dust opacity were reached
188 (Cantor, 2007). The decay phase duration defined by Cantor (2007) is different than the duration
189 required for clearing back to the climatological averages.

190 The next planet-encircling event occurred in MY 28 (Fig.1b). In contrast to the global storm of
191 MY 25, this event began at a much later seasonal date near $L_s = 265^\circ$, after perihelion. However,
192 we observed significant dust activity over the southern hemisphere before the onset of this global
193 storm. Some possible precursor dust storms were observed over Hellas and the southern polar cap
194 edge at $L_s = 200^\circ - 235^\circ$ (Wolkenberg et al., 2018) which are illustrated by Fig.1b. The onset of
195 this storm starts at $L_s = 265^\circ$ over regions from the South Pole to 40°N . A series of storms
196 contributed to forming the global dust event in MY 28 (Smith, 2009). The dust storm in MY 25
197 had dust opacity larger than 1 as far south as 60°S , while the region where dust opacity exceeded
198 unity for the dust storms of MY 28 and 34 was more constrained to mid-latitudes. During MY 28,
199 dust opacity exceeded 1 over a latitudinal belt between 40°S to 40°N until around $L_s = 305^\circ$. Then
200 the decay phase began, with opacity still larger than 0.5 until $L_s = 310^\circ$ (Smith, 2009). The most
201 recent global dust storm began in MY 34 at $L_s \sim 190^\circ$ (Fig.1c). The core of this GDS (global dust
202 storm) extended from 50°N to around 40°S . Dust opacity gradually increased toward the South
203 Pole with season. After $L_s = 210^\circ$, dust activity was pronounced southward of 40°S and it gradually
204 diminished at those latitudes until $L_s = 300^\circ$ (Fig.1c). This phase was characterized by opacity \sim
205 0.5. Large dust opacity was still observed over south polar regions ($>60^\circ\text{S}$) at $L_s = 260^\circ$ while dust
206 was less visible over mid-latitude regions at that time. Panels d, e and f of Fig.1 present the zonal
207 means of atmospheric temperatures at 0.5 mbar (~ 25 km altitude) as a function of season and
208 latitude. We observe increases of atmospheric temperatures at 0.5 mbar during the periods of high
209 dust activity in all three years. The atmospheric response to dust is largest at around 0.5 mbar and
210 at higher altitudes (Wilson and Richardson, 2000, Zurek et al., 1992, Wolkenberg et al., 2018).
211 The temperature increase at 0.5 mbar during a GDS in the equatorial region and mid-latitudes is
212 the highest in MY 25. The GDS in MY 28 began just after perihelion when the solar heating is
213 greatest. Meanwhile, the dust storm of MY 34 began during northern autumn, which is
214 significantly earlier than perihelion. Comparing atmospheric temperatures at this level during the
215 GDS against temperatures for years without global dust events, we observe an increase of around
216 20 - 30 K. While the atmospheric temperature increases observed in the southern hemisphere are
217 primarily caused by direct solar heating of the dust, the corresponding increases observed in the
218 northern hemisphere for all three years are instead due to the adiabatic compression of air from the
219 descending branch of the intensified Hadley circulation caused by the response to the dust storm
220 (Kass et al., 2017). In MY 34 we observed a second atmospheric temperature increase at a later

221 season after the GDS over southern polar regions at around perihelion, but it was much weaker
 222 than the one during the GDS.



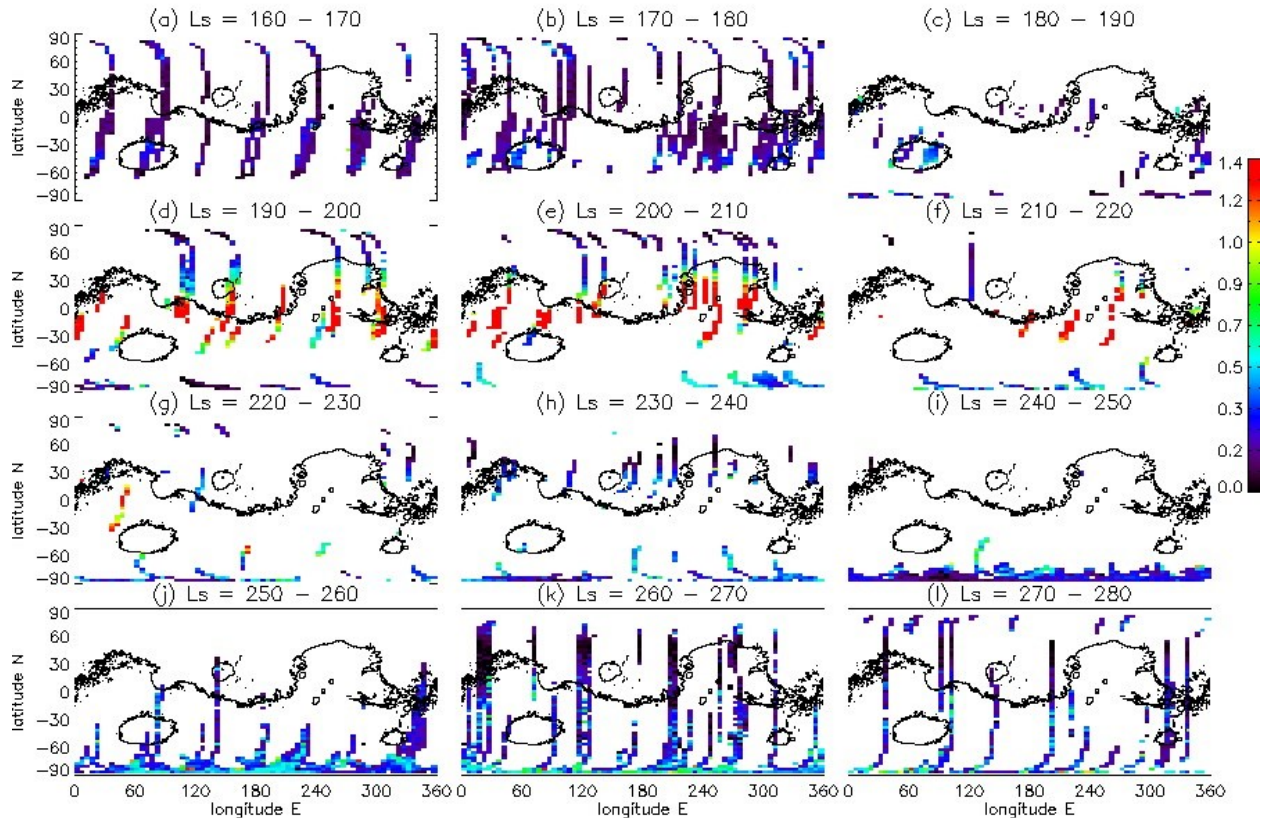
223
 224 **Figure 2.** Spatial dust distributions (TES) from $L_s = 160^\circ$ to 280° in MY 25 at 14 LT (local time).

225 5.2 Spatial analysis of Global Dust Storms

226 We present the spatial distribution of dust from $L_s = 160^\circ$ to 280° in MY 25 obtained from TES
 227 measurements (Fig.2) to compare with MY 34 and MY 28 GDSs. Some local dust storms occurred
 228 over the northwest rim of Hellas at the beginning of the considered interval. By $L_s = 190^\circ$ dust
 229 started expanding toward northeast reaching equatorial regions and west side of Tharsis. An abrupt
 230 explosion of dust was observed over Valles Marineris, equatorial regions and Hellas in the next
 231 interval. Then, dust expanded to larger areas with the exception of Tharsis. Less dust activity was
 232 observed over Tharsis due to its high elevations. By $L_s = 220$ dust progressively began to diminish
 233 in the atmosphere but it was mostly observed in the northern hemisphere. Dust was mostly
 234 constrained to lowlands in the northern hemisphere and appeared in regions featuring large
 235 topographic differences along the dichotomy boundary. Dust from regions around Valles
 236 Marineris moved westward to the southern part of Tharsis in the next L_s interval. Dust opacity
 237 over Hellas was always larger than over other regions for the entire period considered. Eventually
 238 by $L_s = 280^\circ$ dust opacities were less than 0.4 over the northern regions with Hellas opacities
 239 around 0.5 – 0.6.

240 We also studied the spatial distribution of dust aerosol after $L_s = 160^\circ$ in MY 34 to investigate
 241 possible precursor storms. Figure 3 presents the evolution of the spatial distribution of atmospheric

242 dust from $L_s = 160^\circ$ to 280° . Data are binned by 10° in L_s , by 5° in longitude and by 3° in latitude.
 243 Similar to what was observed during MY 25, some signs of the onset of MY 34 storm were
 244 observed by PFS just after the northern autumn equinox at around $L_s = 190^\circ$ (Fig. 1c, 3c).

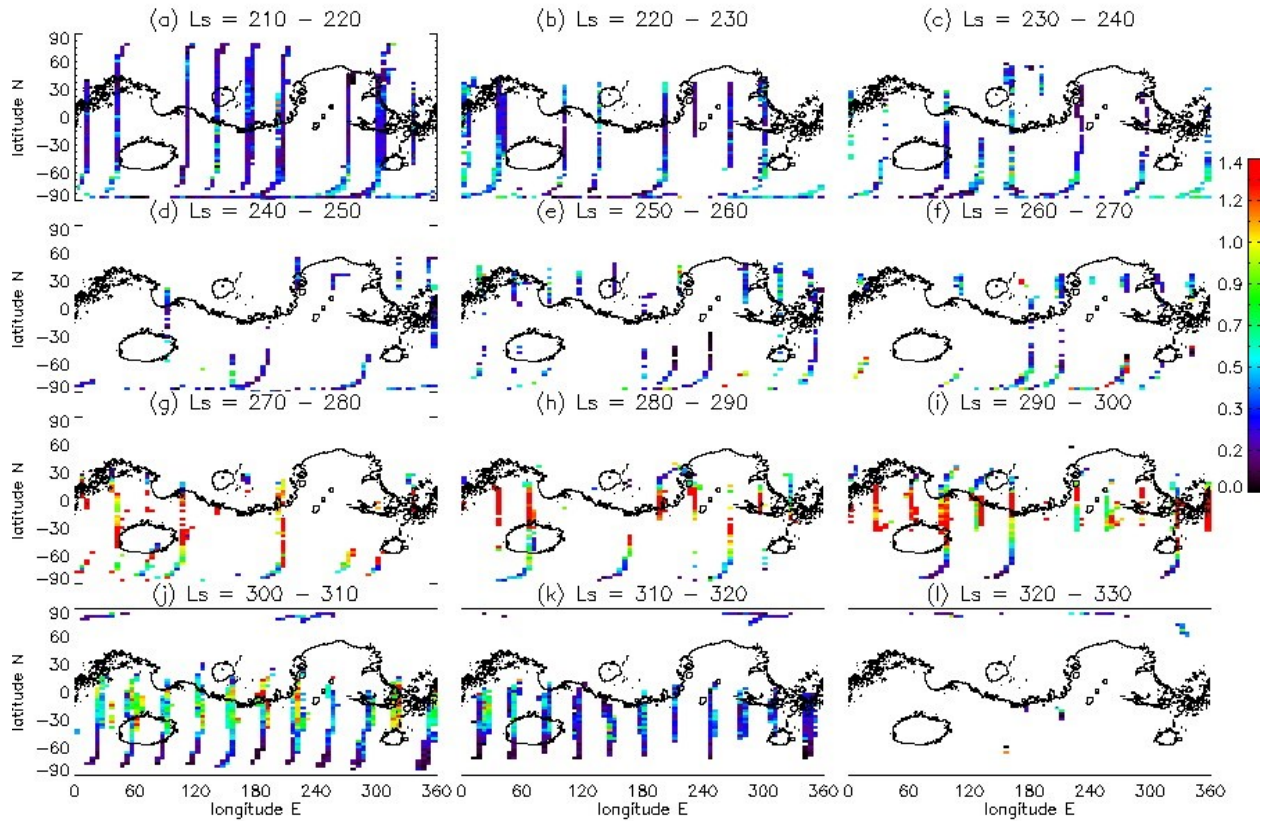


245

246 **Figure 3.** Spatial dust distributions (PFS) from $L_s = 160^\circ$ to 280° in MY 34 from 8 – 18 LT.

247 The ExoMars Trace Gas Orbiter (EMTGO) observed the onset of this GDS at $L_s \sim 188^\circ$ (Vandaele
 248 et al., 2019). The Atmospheric Chemistry Suite (ACS) instrument onboard the EMTGO spacecraft
 249 observed a local dust storm over Chryse Planitia in the L_s range 184° to 187° (Ignatiev et al., 2019).
 250 According to results of assimilated Mars Climate Sounder (MCS) data (Montabone et al., 2018),
 251 the MY 34 storm began its activity over Chryse Planitia and Arabia Terra (Meridiani Planum).
 252 Fig. 3a showed an increase of dust opacity up to ~ 0.4 over Aonia Terra (50°S ; 280°E), the northern
 253 rim of Hellas and Noachis Terra (50°S ; 10°E). Relatively large dust loads were also observed by
 254 ACS-EMTGO below 15 km at $L_s = 168.75^\circ$ in the latitude range: $39^\circ\text{S} - 43^\circ\text{S}$ which corresponds
 255 to latitudes of the Hellas region (Vandaele et al., 2019). Dust activity was then observed westward
 256 of Argyre (Aonia Terra) with opacity ~ 0.5 (Fig. 3b). Acidalia ($30^\circ\text{N} - 60^\circ\text{N}$, $300^\circ\text{E} - 360^\circ\text{E}$) and
 257 Utopia ($30^\circ\text{N} - 60^\circ\text{N}$, $80^\circ\text{E} - 140^\circ\text{E}$) Planitia are regions where local dust storms were imaged at L_s
 258 $= 181^\circ$ by Mars Color Imager (MARCI) (Malin et al., 2018a; 2018b; 2018c). The PFS instrument
 259 was not able to observe these local dust storms seen by MARCI well because of sparse data
 260 coverage during this period ($L_s = 180^\circ - 190^\circ$ - Fig. 3c). However, in the L_s interval from $L_s = 160^\circ$
 261 to 180° (Fig. 3a,b) dust activity was indeed observed by PFS where local dust storms have been
 262 observed by MARCI. There are spots where PFS dust opacities are higher than the background
 263 along the few PFS ground-tracks passing across the Utopia Planitia and Acidalia Planitia. In

264 addition, in the L_s interval ($170^\circ - 180^\circ$, Fig.3b) the atmosphere over Hellas was dustier compared
 265 to the previous L_s interval with pronounced local dust activity over the northwestern rim and the
 266 southeastern region of Hellas. Dust was also distributed westward of Argyre over a wide region
 267 extending from 280°E to 200°E at around 50°S (Sirrenium Terra). Thus, we observe as many as
 268 four regions where precursor dust storms could have originated before $L_s = 181^\circ$.



269

270 **Figure 4.** Spatial dust distributions (PFS) from $L_s = 210^\circ$ to 330° in MY 28 from 8 – 18 LT.

271 These regions are Acidalia and Utopia Planitia imaged by MARCI, Hellas basin, and Aonia –
 272 Sirrenium Terra. These areas might be locations of independently developed precursor dust storms
 273 in MY 34. In spite of sparse PFS data, in the following L_s interval (Fig.3c) we note that the dust
 274 optical depth increased to around 0.7 in Xanthe Terra (southern region of Chryse Planitia, 330°E ;
 275 10°N) and Arabia Terra (20°E ; 20°N). These regions are located near the meridional corridors
 276 (Chryse Planitia) where dust is often transported from the north to the south (Wang and
 277 Richardson, 2015). We also observed larger opacity in the eastern half of the Hellas basin
 278 compared to the western half. According to images from MARCI (Malin et al., 2018c; 2018d),
 279 substantial dust lifting also originated in the southern hemisphere by around $L_s = 188^\circ$ along the
 280 receding southern polar cap. MARCI observed that storms started in the southern hemisphere,
 281 merging with those that originated along and north of equator (Malin et al., 2018c; 2018d). By
 282 around $L_s = 193^\circ$ dust became distributed globally (Malin et al., 2018d; 2018e, Guzewich et al.,
 283 2018). PFS also observed a rapid increase of dust opacity (up to 3 or more) from $L_s = 190^\circ - 200^\circ$
 284 between 45°S to 45°N (Fig. 3d). Dust opacities continued to increase during the next L_s interval
 285 (Fig. 3e) while MARCI started observing a decay phase at around $L_s = 205^\circ$ (Malin et al., 2018f;

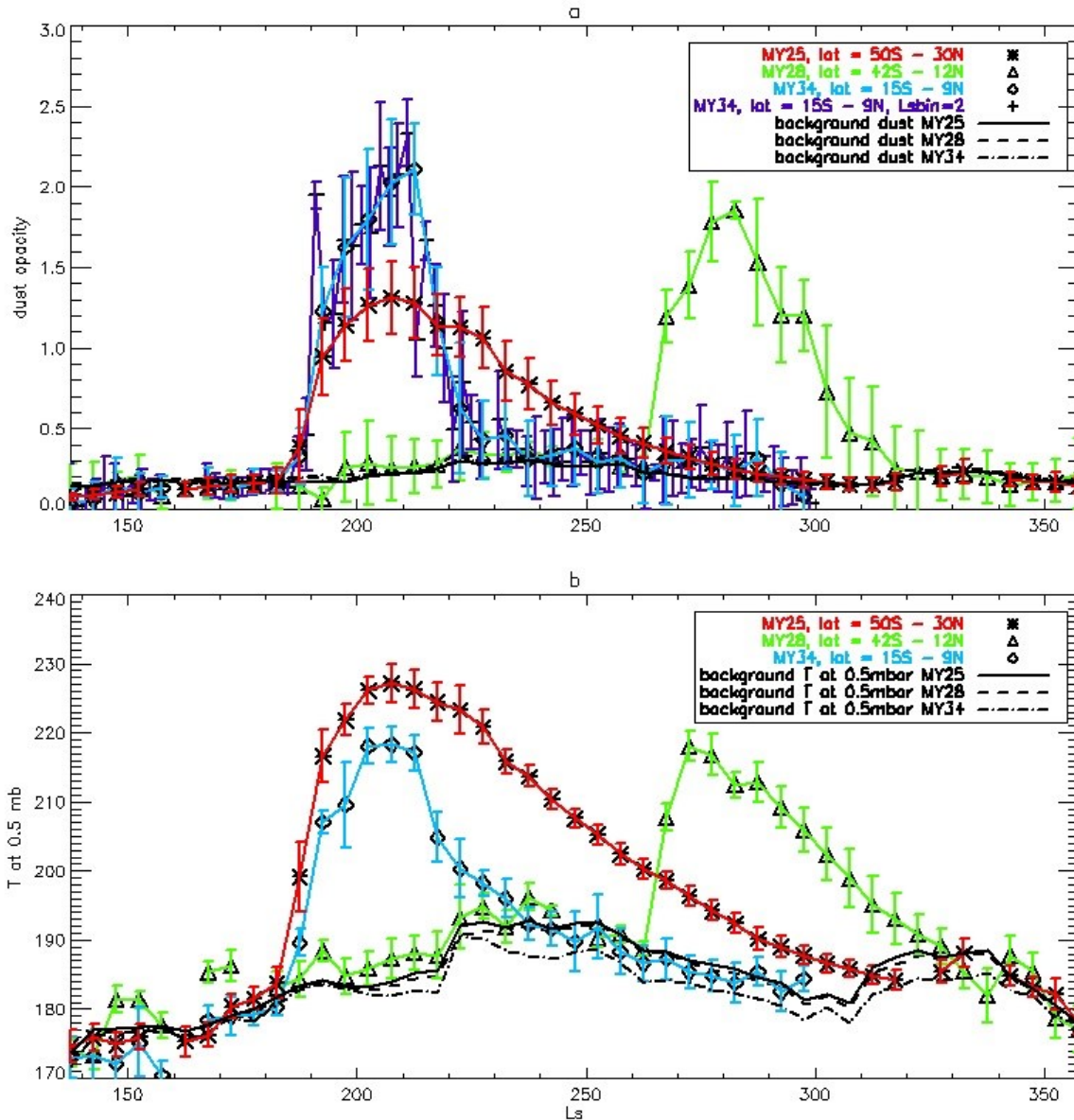
286 2018g). We are unable to distinguish the starting point of the decay phase for the global dust storm
 287 in MY 34 from the PFS observations in Figure 3. However, despite the poor spatial coverage, PFS
 288 observed significant opacities (around 1.5) at Tharsis and northwest of Hellas from $L_s = 210^\circ -$
 289 230° (Fig. 3f,g). The atmospheric dust then progressively decreased by settling down on the
 290 surface (Fig. 3h). It is worth noting that dust activity also occurred close the southern polar regions
 291 during the period from $L_s = 200^\circ$ until $L_s = 280^\circ$ (Fig. 3j,k,l).

292 Figure 4 presents the spatial distribution of dust for the L_s interval from 210° to 330° in MY 28.
 293 As was seen in Fig.1b we observed a significant increase in dust opacity up to 0.6 over southern
 294 polar regions but also over Elysium Planitia in the northern hemisphere before onset of the MY28
 295 GDS at $L_s = 265^\circ$. The dust storm then developed into a global event from $L_s = 270^\circ$ until 300° .
 296 Dust opacity around 0.9 was still observed after $L_s = 300^\circ$ as gradually decreased with season.
 297 Comparison with the MY34 GDS spatial distributions (Fig.3) shows that dust in the MY 28 storm
 298 during its decay phase was not distributed toward southern polar regions. Instead, we observed a
 299 quite clear atmosphere at $L_s = 300^\circ - 320^\circ$ over latitudes higher than 60°S in MY 28. A similar
 300 behavior of spatial dust distributions was also observed in the MY 25 storm during its decay phase.
 301 Another characteristic feature was that the atmosphere over the Tharsis region contained more dust
 302 in MY 28 and 34 (Fig.4 h, i, 3 d, e, f) than in MY 25 (Fig.2 d, e, f) at maximum dust activity at L_s
 303 $= 280^\circ - 300^\circ$ and $L_s = 190^\circ - 220^\circ$, respectively.

304 **5.3 Time series of dust opacities**

305 Cores of storms were found and defined as locations with zonally averaged dust opacity larger
 306 than 1 in each bin. Zonally averaged dust opacity exceeding unity in MY 25 was observed for the
 307 region from 50°S to 20°N for $L_s = 190^\circ - 221^\circ$. Cores of storms in MY 28 and in MY 34 with dust
 308 opacity larger than 1.75 and 1.9 in each bin were constrained to nearly 42°S to 12°N and to 15°S
 309 to 9°N , respectively. We selected these values because they were within 1 standard deviation from
 310 the maxima of Gaussian functions fitted to each dust storm for L_s range from $190^\circ - 221^\circ$ in MY
 311 34 and 28 and from $270^\circ - 297^\circ$. Peaks of dust opacity in MY 25 and MY 34 were observed at
 312 similar L_s (Fig. 5a). The MY 34 maximum was shifted in season due to the latitudinal range taken
 313 into account to plot the variation of dust distribution with season. Dust opacity peaked at the same
 314 L_s like in MY 25 when the range of latitudes for the core was extended to 12°N . The MY 25 and
 315 MY 34 storms started at similar times. We observe an increase in atmospheric dust opacity
 316 suddenly in all MYs due to dust lifting (Fig.5a). The curve of the seasonal variation of dust opacity
 317 is asymmetric with respect to maximum of dust opacity in MY 25 (Fig.5a), with much faster
 318 growth than decay. On the contrary a longer expansion with respect to decay phase was observed
 319 in MY 34. The opacity changed rapidly from 0.25 before the storm to values larger than 1 in just
 320 a couple of days in MY 25. Dust opacity larger than 1 was observed from $L_s = 190^\circ$ until 228° in
 321 MY 25 (Fig. 5a). This is longest seasonal interval when dust opacity exceeds 1 among the other
 322 global dust storms considered here (MY 28 and MY 34). The seasonal dust distribution in MY 34
 323 was also plotted with 2° bin in L_s (blue line) to show dust variation in more details. A separate and
 324 large peak was observed at L_s around 190° which was not evident when the seasonal binning was
 325 5° . A detailed analysis of observations available in this L_s range revealed that large dust opacities

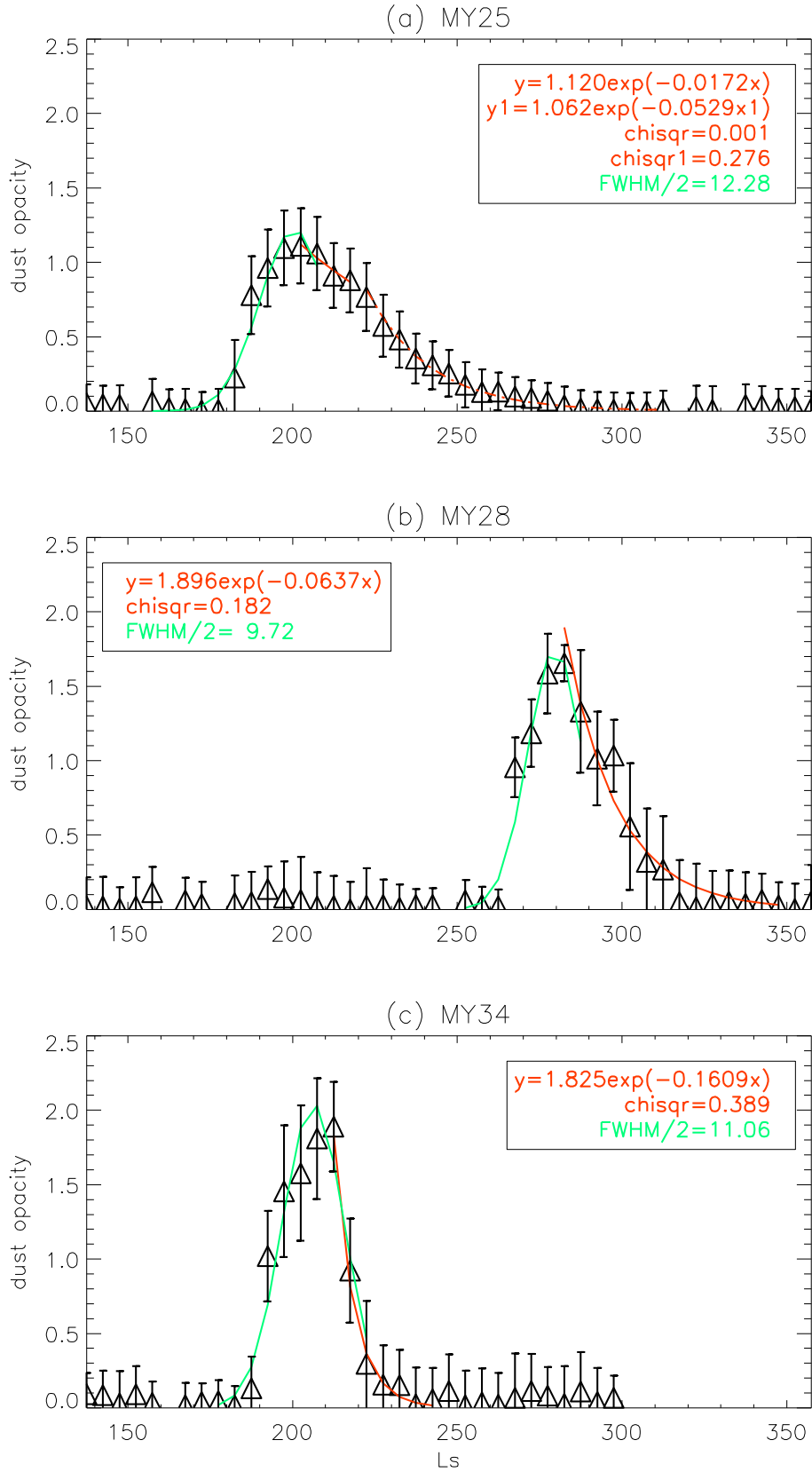
326 around unity were measured over regions of Arabia Terra (5-20°N, 25-27°E) near the Cassini
 327 crater.



328

329 **Figure 5.** Seasonal variations of zonally averaged: (a) dust opacities and (b) atmospheric
 330 temperatures at 0.5 mbar for MY 25, 28 and 34 with standard deviations in each bin. Dust opacity
 331 was binned by 5° in Ls, by 3° in latitude and by 5° in longitude then they were averaged zonally.
 332 The same treatment of data was used to calculate the background level of dust opacity (solid line
 333 for MY25, dashed line for MY 28 and dotted-dashed line for MY 34). The background level of

334 dust opacity is different for three MYs because the different ranges of latitudes are taken into
335 account.



337 **Figure 6.** Time series of dust opacities (triangle symbols) with standard deviations in MY 25 (a),
 338 MY 28 (b) and MY 34 (c). Fit of Gaussian (green line) and exponential decay function (red line)
 339 to dust opacities without background dust opacities is plotted. The decay phase in MY 25 is fitted
 340 by two exponential decay functions (red solid and dashed line). ‘y’ and ‘y1’ mean the model values
 341 of Gaussian functions in MY 25, MY 28 and 34. The numerical value in the exponential function
 342 gives the decay constant. ‘Chisqr1’ and ‘chisqr’ are the chi-squared quality factors of the fit for
 343 exponential decay functions in all MYs.

344 The background level of dust activity was calculated for each bin in L_s and latitude taking into
 345 account all PFS MYs without GDSs. We selected data of dust opacities from PFS MYs including
 346 MY 27 and MY 29 – MY 33. Then we calculated the values of dust opacity for each bin with
 347 latitude, L_s and longitude ($3^\circ \times 5^\circ \times 5^\circ$), respectively. Then we zonally averaged them. We plotted
 348 the background level of dust opacities independently for each storm because we selected different
 349 core regions for each storm. Along with the observed increases in dust opacity, atmospheric
 350 temperatures at 0.5 mbar were observed to grow immediately after the onset of the three global
 351 storms (Fig.5b). The timing of the MY 25 maximum temperature value was closely aligned with
 352 the timing of the maximum dust opacity. For the other dust storms the atmospheric temperatures
 353 peaks occurred somewhat earlier in season than dust opacity maxima. We defined the
 354 climatological values of atmospheric temperatures at 0.5 mbar as a mean of atmospheric
 355 temperatures for all MYs from PFS data excluding MY 28 and MY 34. They are plotted in Fig.5b.
 356 The procedure of calculation is the same as it was for background dust level.
 357

358 It is worth noting that the maximum of dust opacity in MY 34 exceeded 2.1 when the retrievals
 359 are binned and is largest among the three storms (Fig.5a). A similar behavior of dust storm (large
 360 value at peak) in MY 34 was also observed over the Curiosity landing site (Guzewich et al., 2019).
 361 We also observed some possible precursor dust storms as a component of ‘background dust
 362 activity’ in MY 28 beginning at $L_s = 200^\circ$ which is better illustrated in Fig.1b. The core of MY28
 363 storm starts at $L_s = 270^\circ$. If we assume that these possible precursor dust storms could be
 364 interpreted as representing the background level of dust activity from $L_s = 200^\circ$ to 270° in MY 28
 365 then the MY 28 storm can be similar to the others. Indeed, the increases of dust activity were found
 366 from $L_s = 200^\circ$ to 270° and $L_s = 310^\circ$ to 340° (Fig.5a). Figure 5a shows that possible precursor
 367 dust storms were component of background level of dust opacities in MY 28.
 368

369 Figure 6 presents the seasonal distribution of zonally averaged dust opacity after subtraction of
 370 background level of dust activity. This way, the clear behavior of storms was presented. To better
 371 characterize or find similarities and differences between storms, exponential decay and Gauss
 372 functions were fitted to seasonal zonally averaged dust opacities to define the decay and expansion
 373 phases, respectively (Fig.6). The duration of expansion phase could be described by the half of
 374 FWHM (full width at half maximum) derived from the fit of Gaussian function to the dust
 375 opacities.
 376

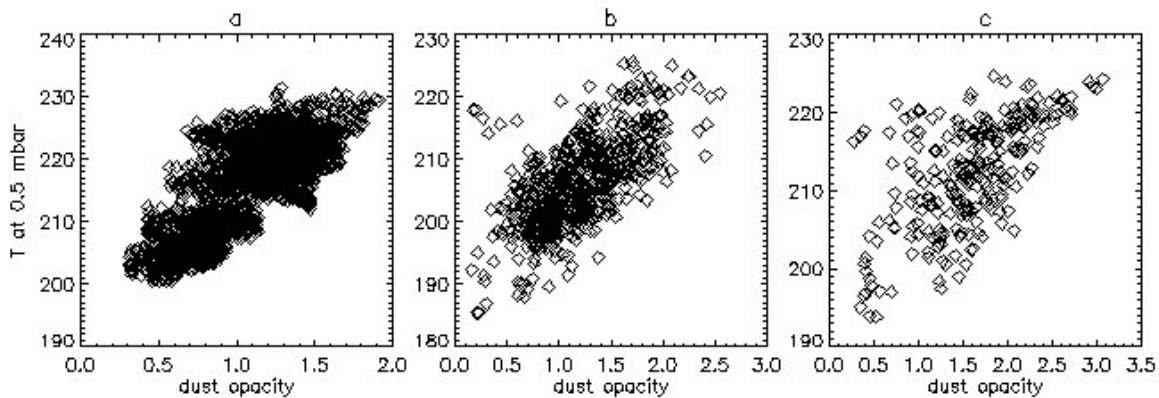
377 It turned out that expansion phases of three storms were very similar. They differed only 1° of L_s
 378 between each other. The MY 34 storm had a medium expansion phase around 11° of L_s (19 sols)
 379 compared to MY 28 with 10° of L_s (15 sols). The longest expansion time with 12° of L_s was in
 380 MY 25 (21 sols). In contrast to expansion phases, each global dust storm had a unique value of
 381 decay phase. The coefficient in the exponential expression is the decay constant (multiplicative
 382 factor at $x (L_s)$ - Fig.6). There was no particularly long decay phase of dust opacity in the MY 34

383 storm as it occurred in MY 25 and MY 28 storms. Similarities in the decay phases were found in
384 MY 25 and MY 28. However, we divided the decay phase into two seasonal intervals in MY 25
385 to describe it better. The decay constants were 0.0529 ± 0.0183 [$1/19^\circ$ of L_s] in MY 25 for the
386 second part of decay phase and 0.0637 ± 0.0072 in MY 28. The first part of the decay phase in
387 MY 25 lasted 10° of L_s (16 sols) with the decay constant equaling 0.0172. The lifetime of the storm
388 (58° of L_s , or 94 sols) was longer when we also consider the duration of the first part of the decay
389 phase. We compared the decay constants for the second part of the decay phase in MY 25 with
390 that in MY 28. After 20° (32 sols), 17° (29 sols) of L_s respectively, the population of dust opacity
391 decreased exponentially (by $1/e$). On the other hand, the storm in MY 34 had the shortest decay
392 phase. The decay constant was estimated to be 0.16 and the mean lifetime was around 6° of L_s (10
393 sols). This means that after around 24° of L_s (38 sols) the dust opacity reached the background
394 level. The settling down to the surface was two and half times faster in MY 34 than in the other
395 storms. This result is unexpected because the two dust storms in MY 25 and 34 started in the same
396 season should have similar development and decay phases taking into account the similar
397 atmospheric conditions. We also evaluated a skewness of these three GDS series, which gave
398 results similar to the Gaussian function fits. We obtained for global dust storms in MY 25: + 0.91,
399 MY 28: + 1.36 and MY 34: + 1.76. These positive values mean that each GDS time series is
400 asymmetric with respect to their maxima with each series having a long decay tail. A larger
401 positive value for the skewness indicates a faster decline in dust opacity. The largest value of the
402 skewness is for the GDS of MY 34. This is in agreement with the decay constant obtained from
403 the fit of the exponential decay function, which also indicated that the fastest decay phase was for
404 the MY34 GDS.

405 Only the time of expansion phase is similar. If the dust aerosols were similar sizes for the three
406 storms they should settle out of the atmosphere at about the same rate for all the storms unless
407 strong vertical winds, cloud formation, or something else occurs. Thus the reason of fast settling
408 down could be other mechanisms associated with the atmospheric circulation. One-dimensional
409 (vertical) simulations of aerosol transport showed that atmospheric motions were key factors in
410 storm decays (Murphy et al., 1990). Moreover, an assumption of disk-like particles showed a good
411 agreement between measured dust optical depths over Viking 1 lander locations and one-
412 dimensional models. However, this model was not able to maintain the initial size distribution for
413 particles in the $1\text{-}10\mu\text{m}$ range, which were predicted by Mariner IRIS results of dust storm in 1971.
414 Thus a two-dimensional model was developed in which spherical particles were more consistent
415 with measured dust optical depth rate over Viking 1 lander location. Moreover, the assumption of
416 spherical particles implied a latitudinal variation of seasonal dust opacity decreases (decay
417 constant) and evolution of particle size distribution. The evolution of particle size distribution and
418 dust opacity declines (decay constants) were much more slowly in the ascending branch of Hadley
419 circulation than in the descending branch (Murphy et al., 1990). The aerosol models showed that
420 the atmospheric circulation played a key role during the decay phases of dust storms in controlling
421 the settling down rate of dust grains of different sizes.

422 Previous works on decay constants provided similar values for some regions (Fenton et al., 1997;
423 Cantor, 2007). Fenton et al. (1997) found 42 sols as a decay constant for regions at 55°S - 65°S in
424 dust storm 1971 (MY 9) observed by the Mariner orbiter. The decay constant over Claritas Fossae
425 region during the MY25 storm was smaller, around 30 sols (Cantor, 2007). However, Cantor
426 (2007) divided the decay phase into two time intervals. He found smaller decay constants during
427 the second interval of the decay phase for other regions. Likely, the small decay constant during

428 the second interval of the decay phase was estimated for the “background” dust occurring every
 429 year. Cantor (2007) derived decay constants corresponding to both the global storm and
 430 “background” dust activity. During the second part of the decay phase, the main dust activity due
 431 to the global storm decreased the level of background activity and thus the decay constants became
 432 smaller. The background dust occurring every year contains the typical behavior of dust activity
 433 during the dust season ($L_s = 135^\circ - 360^\circ$) including regional and local dust storms. Our results
 434 showed larger decay constants (shorter mean lifetimes) compared to those presented by Fenton et
 435 al. (1997) for the equatorial region. Likely, this was due to the fact that the “background” dust
 436 activity was subtracted from our data. Then the exponential function was fitted to the data (Fig.6).



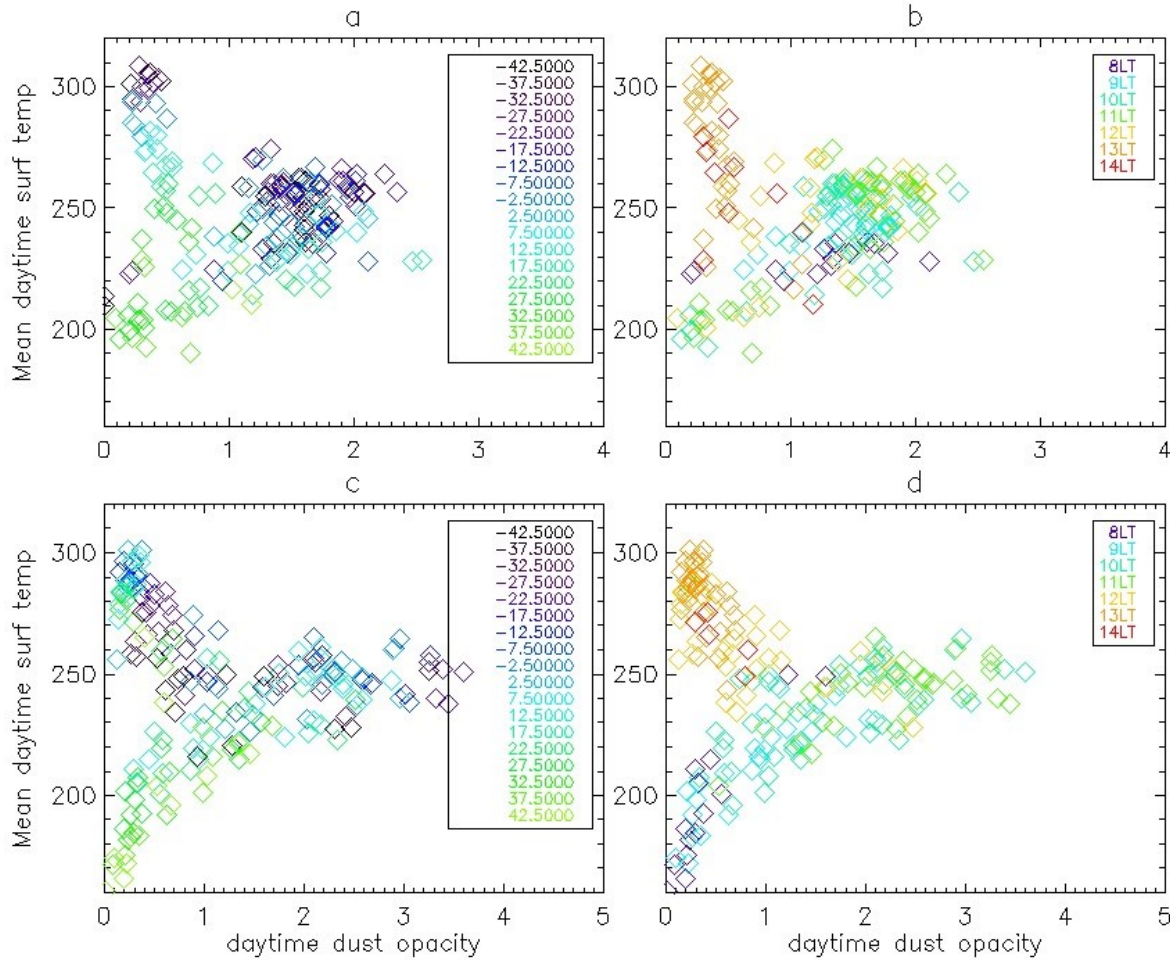
437

438 **Figure 7.** Scatter plots of dust opacities as a function of atmospheric temperatures at 0.5 mbar for
 439 (a) MY 25, (b) MY 28, (c) MY 34.

440 **5.4 Impact of Global Dust Storm on atmospheric and surface temperatures**

441 Dust can be investigated indirectly by using atmospheric temperatures at 0.5 mbar. We plot dust
 442 opacity as a function of atmospheric temperatures at 0.5 mbar for three global dust storms (Fig.7).
 443 As expected, a strong correlation of dust with atmospheric temperatures at 0.5 mbar is observed.
 444 Atmospheric temperatures at this level increase with dust opacity for each MY. Fig.7a presents the
 445 relationship between dust and atmospheric temperatures at 0.5 mbar in MY 25 during global dust
 446 storm. Each point represents data binned in latitude, longitude and L_s ($3^\circ \times 5^\circ \times 10^\circ$), respectively.
 447 We selected daytime data only in the latitudinal range around the equator (12°N to 12°S) and from
 448 the $L_s = 180^\circ - 240^\circ$ range when dust opacity is relatively large. The TES spectrometer performs
 449 measurements at two observation local times (LT), at about 2:00 and 14:00 LT. The atmospheric
 450 temperature range is from 200 K to 230 K for dust opacity variations from 0.3 to 2 in MY 25. The
 451 heating rate of the atmosphere can be a characteristic feature of this dust storm depending on the
 452 chemical composition and the size distribution of the dust aerosol. For MY 28, data are taken from
 453 the $L_s = 260^\circ - 330^\circ$ interval and the local time range is from 9:00 to 18:00 to be comparable with
 454 the data from MY 25 (Fig.7b). Atmospheric temperatures at 0.5 mbar increase along with dust
 455 opacities over southern latitudes from the equator to 42°S . This region is more exposed to the sun
 456 than northern latitudes during the season of the global dust storm (after perihelion). The maximum
 457 atmospheric temperatures at 0.5 mbar are around 220 – 225 K and correspond to dust opacities

458 around 2.5. In MY 34 again a clear trend of atmospheric temperatures at 0.5 mbar with dust is
 459 observed (Fig.7c). The data are selected from the $L_s = 180^\circ - 240^\circ$ range and around the equator
 460 (12°S to 12°N). The atmospheric temperatures increase from 195 K to 225 K with dust opacities
 461 from 0.3 to 3.2 respectively.

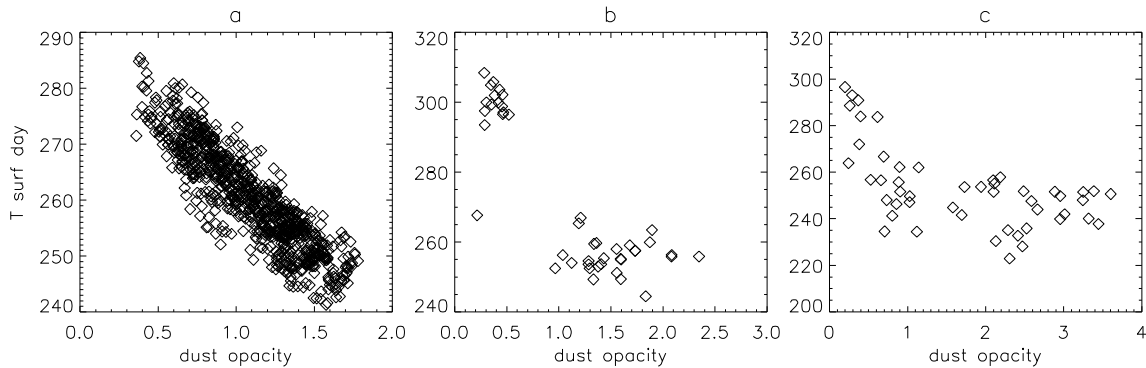


462

463 **Figure 8.** Scatter plots with daytime dust opacities as a function of mean daytime surface
 464 temperatures for MY 28 (a,b) and for MY 34 (c,d) in L_s intervals from $250^\circ - 330^\circ$ and $170^\circ -$
 465 240° , respectively. Colors in panels (a) and (c) represent latitudes from 42.5°S to 42.5°N . Colors
 466 in panels (b) and (d) represent local time. Data are available for LTs from 8 to 14.

467 We also investigate the relationship between mean daytime surface temperatures and daytime dust
 468 opacities for mid-latitude regions in MY 28 and MY34 (Fig.8). Each point corresponds to data
 469 zonally averaged, binned in latitude, L_s and LT ($5^\circ \times 10^\circ \times 1\text{h}$). The local time range is from 8 to 14
 470 LT. For both years we observe a special behavior of dust opacities with surface temperatures
 471 during the day. Temperature around 250 K is found at maximum dust opacities in both years. It
 472 decreases with dust opacity increases when the temperature is higher than 250 K while it increases
 473 with dust opacities for temperature less than 250 K. We plot latitude bins and corresponding LT
 474 bins in different colors to analyze the spatial distribution of dust opacities with surface
 475 temperatures (Fig.8). In MY 28, mean daytime surface temperatures higher than 250 K are mostly

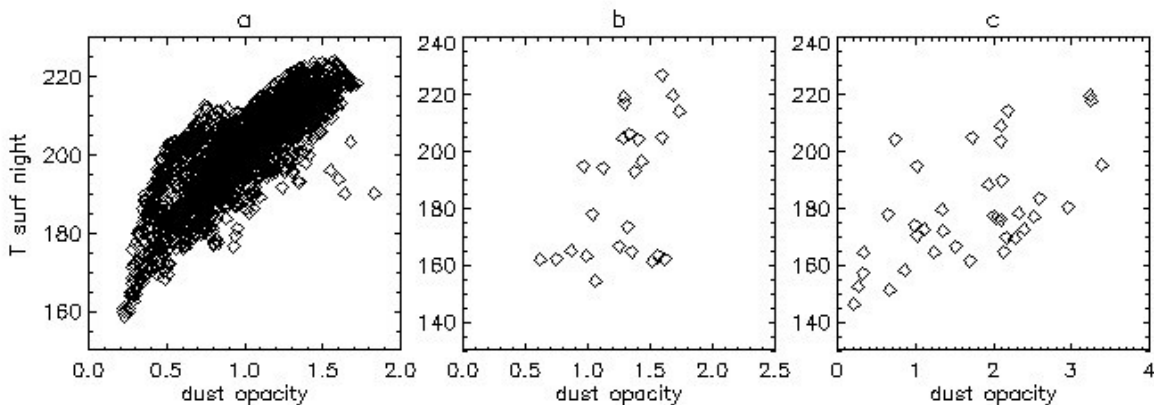
476 measured in the southern hemisphere and over the equator. In this region we have observations
 477 mainly at two LTs: 10 - 11 LT and after 13 LT (Fig.8b). Before noon the dust opacities are very
 478 large with surface temperatures around 250 K while after 12 LT surface temperatures increase with
 479 decreasing dust opacity (< 0.5) in the atmosphere. Lower surface temperatures than 250 K are
 480 found in the northern mid-latitude region. Observations there are performed at morning times from
 481 8 to 11 LT. The lower surface temperatures there are expected according to season and latitude.
 482 Surface temperatures over northern latitudes increase with dust opacities approaching to around
 483 250 K. A similar trend of daytime dust opacities as a function of mean daytime surface
 484 temperatures is also observed in MY 34 (Fig.8 c,d). Southern mid-latitudes and the region over the
 485 equator are characterized by a decrease of surface temperatures with increasing dust opacity.
 486 Observations over these regions mostly take place from 11 to 14 LT. The surface temperature
 487 increases along with dust opacity over the northern hemisphere in the morning at 8 – 10 LT. As it
 488 was for MY 28, the maximum of dust opacity is observed at around 250 K. For both years we also
 489 note that surface temperatures are relatively constant whenever dust opacity is greater than about
 490 1 - 2.



491
 492 **Figure 9.** Scatter plots of dust opacity as a function of daytime surface temperatures for global
 493 dust storm seasons in MY 25 (a), MY 28 (b) and MY 34 (c). (a) Data are binned in latitude,
 494 longitude and Ls ($5^\circ \times 7.5^\circ \times 10^\circ$), respectively for the Ls range from 180° to 240° and the latitude
 495 range from 10°S to the equator. (b) Data are averaged zonally and binned in latitude and Ls
 496 ($5^\circ \times 10^\circ$) in the season from Ls = 240° to 350° and the latitude range from 45°S to 15°S . (c) Data
 497 are averaged zonally and binned in latitude and Ls ($5^\circ \times 10^\circ$) in the season from Ls = 180° to 260°
 498 and the latitude range from 45°S to 5°N .

499 Comparing the two global dust storms that occurred during MY 28 and MY 34 (Fig.8), we observe
 500 a similar distribution of surface temperatures with respect to LT. Southern and equatorial latitudes
 501 were mostly observed during the afternoon while northern latitudes were observed during the
 502 morning. However, the two global dust storms began at different seasons. Therefore, the sun
 503 illumination conditions were different. For example, it is clear that in MY 28 surface temperatures
 504 are warmer at 8 LT than at 10 LT because they are measured for regions further southward from
 505 the equator. Southern regions are more illuminated than northern regions near perihelion, thus
 506 surface temperatures and dust opacities are larger at 8 LT (42°S - 35°S) than at 10 LT (35°N –
 507 42°N). This trend between surface temperatures and dust opacity (Fig.8) helps explain the radiation
 508 budget in the Martian atmosphere in the presence of high dust loads. During the day the dust is
 509 heated by the solar radiation, and contemporaneously this layer of dust prevents the heating of the

510 lower atmosphere and surface below. This leads to the decrease of surface temperature and the
 511 cooling of the atmosphere close to the surface dependent on dust loads (Wolkenberg et al., 2018).
 512 For example, in Fig. 8a,b in MY 28 we observe almost the same region around $7.5^{\circ}\text{N} - 12.5^{\circ}\text{N}$ at
 513 around 11 - 12 LT for different dust opacities from 0.5 to 2. For these values we observe different
 514 surface temperatures from 280K to 250K, respectively. Surface temperatures decreased to 250 K
 515 at 11 - 12 LT for dust opacities around 2. Simultaneously, we observe at the same LT and region
 516 surface temperatures around 280 K but for opacity around 0.5. For dust opacities larger than 1-2
 517 at noon, surface temperatures are constant. For locations with a relatively small amount of dust
 518 less than 0.3, surface temperatures are around 300 – 310 K during the afternoon. When dust
 519 increases, surface temperatures decrease to around 250 K before noon. Afternoon surface
 520 temperature increases when dust decreases over the same region (southern latitudes). Atmospheric
 521 dust prevents the solar radiation from penetrating to the layers close to the surface, and as a result
 522 surface temperature decreases by the reduction of solar heating. The decrease of daytime surface
 523 temperatures with dust opacity is also observed in MY 25 from the TES data (Fig.9a). We note a
 524 similar behavior in MY 28 (Fig.9b) and MY 34 (Fig.9c). In MY 28 there are a lot of gaps,
 525 particularly between 0.5 – 1 of dust opacity (Fig.9b). However, the drop of daytime surface
 526 temperatures is clearly observed until 1.5 of dust opacity (Fig.9b).



527

528 **Figure 10.** Scatter plots of daytime dust opacity against nighttime surface temperature. (a) Data
 529 are binned in latitude, longitude and Ls ($5^{\circ}\times 7.5^{\circ}\times 10^{\circ}$), respectively for the Ls range from 190° to
 530 250° and the latitude range from 5°N to 40°N in MY 25. (b) Data are averaged zonally and binned
 531 in latitude and Ls ($5^{\circ}\times 10^{\circ}$) in the season from Ls = 240° to 350° and the latitude range from 45°S
 532 to 10°N for 18-9LT in MY 28. (c) Data are averaged zonally and binned in latitude and Ls ($5^{\circ}\times 10^{\circ}$)
 533 in the season from Ls = 180° to 260° and the latitude range from 45°S to 45°N for 18-9LT in MY
 534 34.

535 For dust opacities larger than 1.5 a characteristic plateau is observed in MY 28 and MY 34 where
 536 daytime surface temperatures no longer decrease with increasing dust opacity (Fig. 9c). However,
 537 this could be due to observations being at different LTs. In MY 25 this plateau is not evident
 538 because measurements were performed at a LT near 14:00.

539 We also investigate the relation between daytime dust opacities and nighttime surface temperatures
 540 (Fig.10). In each Martian year considered in Fig.10 we observe an increase in nighttime surface

541 temperature with dust opacity. However, the rate of these increases appears to be different in each
542 Martian year. In these nighttime observations the atmosphere (and dust) are warmer than the
543 surface. The more dust there is in the atmosphere, the more warm downward radiation from the
544 dust is received at the surface increasing the surface temperature. In the clear atmosphere the
545 surface cools off during the night through the emission of thermal infrared radiation with little
546 compensating downward radiation from the atmosphere.

547 **6 Summary and conclusions**

548 We report on seasonal and spatial variations of dust opacities in MY 34 obtained from
549 measurements performed by PFS onboard Mars Express spacecraft. Additionally, we also present
550 the evolution of dust opacities in MY 25 obtained from TES measurements and in MY 28 from
551 PFS observations to find similarities and differences between the global dust storms that occurred
552 in those three years. The two GDSs during MY 25 and MY 34 were equinoctial events encircling
553 the whole planet at about $L_s = 193$. Similarities are also found in the onset of MY 25 and 34 storms
554 such as a local dust storm over the northern rim of Hellas (dust opacity > 0.4) that was observed
555 in both MY 25 and MY 34. In MY 25, dust storm area expanded east over the next several days.
556 In MY 34 we also observed dust over on the eastern side of Hellas basin. Some similar dust
557 occurrences in Arabia Terra and Utopia Planitia were also found before the expansion phase in
558 MY 25 and MY 34. A peak of dust activity at $L_s = 190^\circ$ in the expansion phase was observed in
559 Arabia Terra near Cassini crater in MY 34. However, there were differences as well. For example,
560 PFS observed dust over regions further to the north than TES did in MY 25. The MY 28 dust storm
561 began after perihelion near the southern summer solstice contrary to the dust events in MY 25 and
562 34 which occurred much earlier during the season. Hence, the specific features of the MY 28 storm
563 were different than the others. We observed some possible precursor dust storms mainly over the
564 Hellas region and the southern polar cap edge long before the onset. These features were not
565 observed before the equinoctial dust storms on this scale. However, these possible precursor dust
566 storms in MY 28 were a component of background dust activity. After subtraction of background
567 dust opacities, the separate analysis of each global dust storm was possible. These studies revealed
568 that the time of expansion phase of each GDS was similar. They differed between each other by
569 1° of L_s . However, the data binning by 2° in L_s showed a separate peak of dust opacity near unity
570 in the expansion phase of MY 34 GDS which could imply a different expansion phase. In contrast
571 to the expansion time, each GDS had a unique duration of the decay phase. However, the removal
572 of background activity showed that the two GDSs in MY 28 and MY 25 had similar decay
573 constants (0.06 and 0.05, respectively) although they started in different season at different
574 atmospheric conditions. By contrast, GDS in MY 34 was characterized by two and half times faster
575 decay constant than other both storms. This implied a shorter mean lifetime of this storm compared
576 to storms in MY 25 and MY 28. The spatial distribution of the MY 28 storm was more limited to
577 the regions over the southern hemisphere than the storm of MY 34. During the expansion phase of
578 the global storms, dust in MY 25 and MY 34 was more distributed toward the north (50°N) than
579 it was during in MY 28.

580 A clear relation between atmospheric temperatures at 0.5 mbar and dust opacities is presented. As
581 has been found earlier we confirm that atmospheric temperature increases at 0.5 mbar were mainly
582 due to dust opacity increases. We also observed a large decrease in daytime surface temperatures
583 with increasing dust opacity. However, in MY 34 daytime surface temperatures remained nearly
584 constant reaching a characteristic plateau when dust opacity exceeded 2. A similar situation was

585 observed in MY 28 but the trend was less obvious because the maximum dust opacity was $\sim 2 -$
 586 2.5. In MY 28 and MY 34 we observed what appears to be a drop of dust opacities from 10-11 LT
 587 to 13-14 LT at southern low-latitudes. Contemporaneously, we noted an increase of daytime
 588 surface temperatures at those LTs. It is not clear if this afternoon increase of surface temperatures
 589 was due to the decrease in dust with solar radiation better able to reach the surface. Finally, during
 590 all three global storms we observed that nighttime surface temperatures increased with increasing
 591 dust opacity (measured during the day).

592 The physical mechanisms that govern the initiation, growth, and decay of global dust storms on
 593 Mars are still not well understood. While there are broad similarities in the behavior of these global
 594 storms, each storm exhibits its own unique characteristics depending on many factors including
 595 the season in which it begins and the locations where dust raising is initiated. The new observations
 596 of the MY 34 global storm along with a comparison of previous global storms observed in MY 25
 597 and MY 28 presented here provide the crucial observational input that models will be able to use
 598 to better understand the evolution of global dust storms and the Mars dust cycle in general.

599

600 **Acknowledgements**

601 The PFS-MEx and TES-MGS datasets used in this study are publicly available (Wolkenberg et al.,
 602 2019).

603

604 **References**

605

- 606 Bandfield, J. L. and Smith, M. D. (2003). Multiple emission angle surface–atmosphere separations
 607 of Thermal Emission Spectrometer data. *Icarus*, 161, 47–65.
- 608 Cantor, B. A., James, P. B., Caplinger, M. and Wolff, M. J. (2001), Martian dust storms: 1999
 609 Mars Orbiter Camera observations. *Journal of Geophysical Research*, 106(E10), 23653 –
 610 23687.
- 611 Cantor, B. A. (2007). MOC observations of the 2001 Mars planet-encircling dust storm. *Icarus*,
 612 186, 60 – 96.
- 613 Christensen, P.R., et al. (2001). The Mars Global Surveyor Thermal Emission Spectrometer
 614 experiment: investigation description and surface science results. *Journal of Geophysical*
 615 *Research*, 106, 23823–23871.
- 616 Conrath, B. J., Pearl, J. C., Smith, M. D., Maguire, W. C., Christensen, P. R., Dason S. and
 617 Kaelberer, M. S. (2000). Mars Global Surveyor Thermal Emission Spectrometer (TES)
 618 observations: Atmospheric temperatures during aerobraking and science phasing. *Journal*
 619 *of Geophysical Research*, 105, 9509–9519.
- 620 Fenton, L. K., Perl, J. C, Martin, T. Z. (1997). Mapping Mariner 9 Dust Opacities. *Icarus*, 130,
 621 115 – 124.

- 622 Formisano, V., et al. (2005). The Planetary Fourier Spectrometer (PFS) onboard the European
623 Mars Express mission, *Planetary and Space Science*, 53, 963-974.
- 624 Giuranna, M., et al., (2005a). Calibration of the Planetary Fourier Spectrometer Short Wavelength
625 Channel, *Planetary and Space Science*, 53, 975-991
- 626 Giuranna, M., et al., (2005b). Calibration of the Planetary Fourier Spectrometer Long Wavelength
627 Channel, *Planetary and Space Science*, 53, 993-1007
- 628 Giuranna, M., Wolkenberg, P., Grassi, D., Aronica, A., Aoki, S., Scaccabarozzi, D., Saggin, B.,
629 Formisano, V., (2019). The current weather and climate of Mars: 12 years of atmospheric
630 monitoring by the Planetary Fourier Spectrometer on Mars Express, *Icarus*,
631 <https://doi.org/10.1016/j.icarus.2019.113406>
- 632 Grassi, D., Ignatiev, N.I., Zasova, L.V., Maturilli, A., Formisano, V., Bianchini, G.A., Giuranna,
633 M. (2005). Methods for the analysis of data from the Planetary Fourier Spectrometer on
634 the Mars Express mission. *Planetary and Space Science*, 53(10), 1017–1034.
- 635 Guzewich, S. D., Lemmon, M., Smith, C.L., Martínez, G., de Vicente-Retortillo, Á., Newman, C.
636 E., et al. (2019). Mars Science Laboratory observations of the 2018/Mars year 34 global
637 dust storm. *Geophysical Research Letters*, 46. <https://doi.org/10.1029/2018GL080839>
- 638 Haberle, R. M. (1986). Interannual variability of global dust storms on Mars. *Science*, 234, 459-
639 61.
- 640 Haberle, R. M., Pollack, J. B., Barnes, J. R., Zurek, R. W., Leovy C. B., Murphy J. R., et al. (1993).
641 Mars Atmospheric Dynamics as Simulated by the NASA Ames General Circulation Model
642 1. The Zonal-Mean Circulation. *Journal of Geophysical Research*, 98(E2), 3093 – 3123.
- 643 Haberle, R. M., Clancy, R. T., Forget, F., Smith, M. D., Zurek, R. W. (2017). *The atmosphere and*
644 *climate of Mars*. Cambridge University Press, United Kingdom.
- 645 Heavens, N. G. (2010). *The impact of mesoscale processes on the atmospheric circulation of Mars*.
646 Dissertation (Ph.D.), <http://resolver.caltech.edu/CaltechTHESIS:04222010-152158923>,
647 California Institute of Technology.
- 648 Ignatiev et al. (2019), Thermal structure and dust clouds during the 2018 dust storm from ACS-
649 TIRVIM onboard ExoMars/TGO, EGU2019-14988.
- 650 Kass, D. M., Kleinböhl, A., McCleese, D. J., Schofield, J. T. and Smith, M. D. (2016). Interannual
651 similarity in the Martian atmosphere during the dust storm season. *Geophysical Research*
652 *Letters*, 43, 6111–6118. doi:10.1002/2016GL068978.
- 653 Malin, M. C., Cantor, B. A., & Britton, A. W. (2018a). MRO MARCI weather report for the week
654 of 21 May 2018–27 May 2018, Malin Space Science Systems Captioned Image Release,
655 MSSS-532. Retrieved from http://www.msss.com/msss_images/2018/05/30/
- 656 Malin, M. C., Cantor, B. A., & Britton, A. W. (2018b). MRO MARCI weather report for the week
657 of 28 May 2018–3 June 2018, Malin Space Science Systems Captioned Image Release,
658 MSSS-533. Retrieved from http://www.msss.com/msss_images/2018/06/06/
- 659 Malin, M. C., Cantor, B. A., & Britton, A.W. (2018c). MRO MARCI weather report for the week
660 of 4 June 2018–10 June 2018, Malin Space Science Systems Captioned Image Release,
661 MSSS-534. Retrieved from http://www.msss.com/msss_images/2018/06/13/

- 662 Malin, M. C., Cantor, B. A., & Britton, A. W. (2018d). MRO MARCI weather report for the week
663 of 11 June 2018–17 June 2018, Malin Space Science Systems Captioned Image Release,
664 MSSS-535. Retrieved from http://www.msss.com/msss_images/2018/06/20/
- 665 Malin, M. C., Cantor, B. A., & Britton, A. W. (2018e). MRO MARCI weather report for the week
666 of 18 June 2018–24 June 2018, Malin Space Science Systems Captioned Image Release,
667 MSSS-536. Retrieved from http://www.msss.com/msss_images/2018/06/27/
- 668 Malin, M. C., Cantor, B. A., & Britton, A. W. (2018f). MRO MARCI weather report for the week
669 of 2 July 2018–8 July 2018, Malin Space Science Systems Captioned Image Release,
670 MSSS-538. Retrieved from http://www.msss.com/msss_images/2018/07/11/
- 671 Malin, M. C., Cantor, B. A., & Britton, A. W. (2018g). MRO MARCI weather report for the week
672 of 16 July 2018–22 July 2018, Malin Space Science Systems Captioned Image Release,
673 MSSS-540. Retrieved from http://www.msss.com/msss_images/2018/07/25/
- 674 Montabone, L., Spiga, A., Kass, D. M., Kleinboehl, A., Forget, F., Millour, E., Martian Year 34
675 Column Dust Climatology from Mars Climate Sounder Observations: Reconstructed Maps and
676 Model Simulations, submitted to JGR-Planets, this special issue.
- 677 Murphy, J. R., Toon, O. B., Haberle, R. M. and Pollack, J. B. (1990). Numerical Simulations of
678 the Decay of Martian Global Dust Storms. *Journal of Geophysical Research*, 95, B9,
679 14629–14648.
- 680 Pankine, A.A., (2015). The nature of the systematic radiometric error in the MGS TES spectra.
681 *Planetary and Space Sciences*, 109 – 110, 64 – 75.
682 <http://dx.doi.org/10.1016/j.pss.2015.01.022>
- 683 Pankine, A.A., (2016). Radiometric error and re-calibration of the MGS TES spectra. *Planetary*
684 *and Space Sciences*, 134, 112 – 121. <http://dx.doi.org/10.1016/j.pss.2016.10.015>
- 685 Smith, M. D., Pearl, J. C., Conrath, B. J. and Christensen, P. R. (2000a). Mars Global Surveyor
686 Thermal Emission Spectrometer (TES) observations of dust opacity during aerobraking
687 and science phasing. *Journal of Geophysical Research*, 105, 9539–9552.
- 688 Smith, M. D., Bandfield, J. L. and Christensen P. R. (2000b). Separation of atmospheric and
689 surface spectral features in Mars Global Surveyor Thermal Emission Spectrometer (TES)
690 spectra. *Journal of Geophysical Research*, 105, 9589–9608.
- 691 Smith, M. D., Conrath, B. J., Pearl, J. C. and Christensen, P. R. (2002). Thermal Emission
692 Spectrometer Observations of Martian Planet-Encircling Dust Storm 2001A. *Icarus*, 157,
693 259 – 263.
- 694 Smith, M. D. (2004), Interannual variability in TES atmospheric observations of Mars during
695 1999–2003, *Icarus*, 167, 148–165.
- 696 Smith, M. D. (2008), Spacecraft Observations of the Martian Atmosphere, *Annual Review of Earth*
697 *and Planetary Sciences*, 36, 191–219.
- 698 Smith, M. D. (2009). THEMIS observations of Mars aerosol optical depth from 2002 – 2008,
699 *Icarus*, 202, 444 – 452.
- 700 Strausberg M., Wang, H., Richardson, M. I., Ewald, S. P., Toigo, A. D. (2005). Observations of
701 the initiation and evolution of the 2001 Mars global dust storm, *Journal of Geophysical*
702 *Research*, 110 (E02006). doi:10.1029/2004JE002361

- 703 Vandaele A.C., Korablev, O., Daerden, F, Aoki, S., Thomas, I. R., Altieri, F., et al. (2019). Martian
704 dust storm impact on atmospheric H₂O and D/H observed by ExoMars Trace Gas Orbiter.
705 *Nature*, 568, 521–525. <https://doi.org/10.1038/s41586-019-1097-3>
- 706 Wang H. and Richardson, M. I. (2015). The origin, evolution, and trajectory of large dust storms
707 on Mars during years 24-30 (1999 – 2011). *Icarus*, 251, 112 – 127.
- 708 Wilson, R. J. and Richardson, M. I. (2000). The Martian Atmosphere During the Viking Mission,
709 I, Infrared Measurements of Atmospheric Temperatures Revisited, *Icarus*, 145, 555-579.
710 doi:10.1006/icar.2000.6378
- 711 Wolkenberg, P., Smith, M. D., Formisano, V., Sindoni, G. (2011). Comparison of PFS and TES
712 observations of temperature and water vapor in the martian atmosphere. *Icarus*, 215, 628
713 – 638.
- 714 Wolkenberg, P., Giuranna, M., Grassi, D., Aronica, A., Aoki, S., Scaccabarozzi, D., Saggin, B.
715 (2018). Characterization of dust activity on Mars from MY27 to MY32 by PFS-MEX
716 observations. *Icarus*, 310, 32 – 47. doi:10.1016/j.icarus.2017.10.045
- 717 Wolkenberg, Paulina; Giuranna, Marco; Smith, Michael D.; Grassi, Davide (2019), “TES and PFS
718 observations during MY 25, MY 28 and MY 34”, Mendeley Data, v1,
719 <http://dx.doi.org/10.17632/5s5c8j5g2f.1>
- 720 Zurek, R. W., Barnes, J. R., Haberle, R. M., Pollack, J. B., Tillman, J. E. and Leovy, C. B. (1992).
721 *Dynamics of the atmosphere of Mars*. edited by H. H. Kieffer et al., pp. 835–933,
722 University of Arizona Press, Tucson.

723

724

725

726