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Authors	Olsen, K.S., Forget, F., Madeleine, J.-B., Szantai, A., Audouard, J., Geminale, A., ALTIERI, FRANCESCA, BELLUCCI, Giancarlo, Oliva, Fabrizio, Montabone, L., Wolff, M.J.
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Retrieval of the water ice column and physical properties of water-ice clouds in the Martian atmosphere using the OMEGA imaging spectrometer

K. S. Olsen^{a,b,*}, F. Forget^a, J.-B. Madeleine^a, A. Szantai^a, J. Audouard^b,
A. Geminale^c, F. Altieri^c, G. Bellucci^c, F. Oliva^c, L. Montabone^{a,d},
M. J. Wolff^d

^a*Laboratoire de Météorologie Dynamique (CNRS/UPMC/IPSL), Paris, France.*

^b*Laboratoire Atmosphères, Milieux, Observations Spatiales (CNRS/IPSL), Guyancourt, France.*

^c*Istituto di Astrofisica e Planetologia Spaziali (IAPS/INAF), Rome, Italy.*

^d*Space Science Institute, Boulder, USA.*

Abstract

Using spectral images recorded by the OMEGA instrument on Mars Express (Observatoire pour la Minéralogie, l'Eau, les Glaces et l'Activité), we are able derive physical properties of aerosols in water-ice clouds on Mars for a distribution of pixels over an observed cloud formation. These properties, mean effective radius, r_{eff} , and optical depth (at $0.67\text{ }\mu\text{m}$), τ_i , were used to estimate the water ice-column (WIC), and we found an empirical relationship between the WIC and an ice cloud index (ICI). The overall mean of retrieved

*Corresponding author

Email addresses: kevin.olsen@latmos.ispl.fr (K. S. Olsen), francois.forget@lmd.jussieu.fr (F. Forget), jean-baptiste.madeleine@lmd.jussieu.fr (J.-B. Madeleine), szantai@lmd.polytechnique.fr (A. Szantai), joachim.audouard@latmos.ispl.fr (J. Audouard), anna.geminale@iaps.inaf.it (A. Geminale), francesca.altieri@iaps.inaf.it (F. Altieri), giancarlo.bellucci@iaps.inaf.it (G. Bellucci), fabrizio.oliva@iaps.inaf.it (F. Oliva), lmontabone@spacescience.org (L. Montabone), mjwolff@spacescience.org (M. J. Wolff)

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r_{eff} is $\sim 2.2\mu\text{m}$, with a standard deviation of $0.8\mu\text{m}$, and cloud formations with r_{eff} between 4.4 and $5.4\mu\text{m}$ are observed. The optical depth varies between 0.2 and 2.0 . The OMEGA spectra are primarily sensitive to water ice mass due to absorption, and we find that the ICI, very easy to compute, is a good proxy for the mass of the water-ice column (WIC) along the optical line of sight. Our retrieval of physical properties is limited in time (to before 2010) by the exhaustion of coolant for one of the OMEGA channels, and in space (to equatorial observations between 140°W – 90°E) by the availability of surface albedo measurements. However, we used the ICI to compute WIC values for the entire OMEGA data set, which has near-global coverage for Mars years 26–32, and we present a climatology of the WIC derived from the OMEGA data, which features enhancements on the order of 1.2 – $1.6\text{pr.}\mu\text{m}$ over the aphelion cloud belt, and 1.5 – $2.5\text{pr.}\mu\text{m}$ over the polar hoods. The data set analyzed is for observations between 140°W and 90°E , and between 35°S and 35°N . No restriction is placed on season, but the majority of cloudy observations were during the aphelion period from $\text{Ls } 35^\circ$ to 135° . This work was motivated by the ability of the OMEGA instrument to observe the distribution of water-ice cloud physical properties, and by the availability of new a priori data sets, especially multi-spectral, aerosol-free surface albedo retrieved from a subset of the OMEGA data featuring a cloud-free sky. The main limitations of the retrieval algorithm are linked to the uncertainties on surface albedo, the dust opacity, and the quantity of water-ice suspended in the atmosphere, which can lead to spectral fits with lower accuracy or unrealistic results. We present distributions of each retrieved parameter, goodness of fit, ICI, and cloud mass, and our investigation of relationships

between each parameter. Our approach was to maximize the amount of data analyzed, apply stringent data quality cuts and take a statistical approach to interpretation.

Keywords: Mars atmosphere, Mars Express, OMEGA, water ice, aerosols, clouds, retrievals, ICI, Climatology

1. Introduction

Mars has an active hydrological cycle driven by strong diurnal temperature variations and periodic polar sublimation and condensation activities. Seasonal variations in water content are the result of exchange between different reservoirs, such as polar ice and the atmosphere. Clouds play several important roles in the Martian hydrological cycle, notably influencing inter-hemisphere transport. In this study, we present the retrieval of physical properties of water-ice particles in Martian clouds: the mean effective radius and optical depth. Observing these properties and understanding their temporal and spatial distribution can lead to improved understanding of several aspects of the Martian hydrological cycle. Clouds and their formation are associated with the changing meteorological state of the atmosphere, temperature, aerosol and water content, and air parcel dynamics. In turn, they affect the radiative state of the atmosphere by modifying its absorptive and reflective properties, as well as its chemical balance through heterogeneous processes (e.g., [Haberle et al., 2017](#), and references therein). Water is considered essential for the emergence of life and its presence on Mars has guided remote sensing and in situ research activities since the planet was first visited. In order to improve our knowledge of the history and habitability of

20 Mars, we seek to better understand the current climate by identifying and
21 quantifying the water reservoirs, and the exchange of water mass between
22 them, which cloud processes are an integral part of.

23 The physical properties of water-ice clouds on Mars have previously been
24 measured by instruments on Mariner 9 (Curran et al., 1973; Zasova et al.,
25 2001), Phobos 2 (Petrova et al., 1996), Viking 1 and 2 (Christensen and
26 Zurek, 1984; Tamppari et al., 2003), Mars Pathfinder (Smith and Lem-
27 mon, 1999), Mars Global Surveyor (Clancy et al., 2003; Wolff and Clancy,
28 2003), Phoenix (Whiteway et al., 2009), Mars Express (Zasova et al., 2006;
29 Madeleine et al., 2012; Fedorova et al., 2014), and Mars Reconnaissance Or-
30 biter (Smith et al., 2013; Guzewich et al., 2014) These studies are generally
31 in agreement and observed aerosols with a mean effective radius between 2–
32 3.5 μm . Observations have also been made using the Hubble Space Telescope
33 (James et al., 1996) and from terrestrial observatories (see e.g. Parker et al.,
34 1999; Glenar et al., 2003).

35 This work follows that of Madeleine et al. (2012) who presented a retrieval
36 technique for estimating the mean effective radius, r_{eff} , and optical depth,
37 τ_i (at 0.67 μm), by fitting portions of spectra recorded by the OMEGA in-
38 strument (Observatoire pour la Minéralogie, l’Eau, les Glaces et l’Activité)
39 on Mars Express. This paper was motivated by the availability of new prior
40 data sets, which are critical to the accuracy and reliability of the retrieval. Of
41 particular importance are the surface albedos at each wavelength used in the
42 fitting and at the precise locations observed in each pixel. Our objective was
43 to exploit the spatial extent of OMEGA spectral images to investigate the
44 distribution of water ice aerosol physical properties within cloud formations.

45 [Madeleine et al. \(2012\)](#) examined water-ice cloud formations in the aphe-
 46 lion cloud belt using an ice cloud index (ICI) and analyzed clouds over the
 47 Tharsis plateau, seeking diurnal and seasonal trends in particle size and opac-
 48 ity. However, their study was limited to only 14 Mars Express orbits, and the
 49 retrieval only used a small subset of OMEGA spatial pixels to represent the
 50 cloud formations. Our approach differs from that of [Madeleine et al. \(2012\)](#)
 51 in that we have attempted to automate the analysis and retrieve cloud prop-
 52 erties from as many pixels as possible, and evaluate the results statistically.

53 A new method to retrieve multi-spectral surface albedo from OMEGA im-
 54 ages using principal component analysis has been used to recreate OMEGA
 55 images where each pixel contains the surface albedo as a function of wave-
 56 length ([Geminale et al., 2015](#)). We have used the ICI to find OMEGA pixels
 57 containing clouds that overlap with the available cloud-free, multi-spectral
 58 surface albedo maps over the aphelion cloud belt region (between 5° S and
 59 35° N). We have performed retrievals of r_{eff} and τ_i on over 200,000 pixels from
 60 94 cloudy OMEGA spectral images that overlap 60 surface albedo maps

61 The majority of OMEGA observations analyzed have very narrow longitu-
 62 dinal ranges, which makes studying the distribution of r_{eff} difficult. However,
 63 r_{eff} and τ_i are inversely related and can be used to compute the mass, M ,
 64 of the water-ice column, which can more directly inform us about daily, sea-
 65 sonal, or geographical trends in their formation. The position of the curve
 66 resulting from plotting τ_i against r_{eff} varies depending on the ICI of the pixels.
 67 These are, in fact, curves of equal mass in the $r_{\text{eff}} - \tau_i$ parameter space. The
 68 OMEGA spectra are primarily sensitive to absorption by water-ice aerosols,
 69 and therefore to the mass of water-ice along the line of sight. The slope of

70 the relationship $\tau_i \propto 1/r_{\text{eff}}$ exhibits an empirical dependence on ICI, and,
71 therefore, we find that the easily-computed ICI is a proxy for water-ice col-
72 umn mass. We observe that the relationship between the two parameters
73 is strongly obeyed, and that within a single cloud formation there can be
74 large variations in the spatial distribution of retrieved r_{eff} values, with corre-
75 sponding variations in τ_i . The cloud mass, however, has a more homogeneous
76 spatial distribution and is similar to that seen in the visible channel of the
77 OMEGA pixels and the ICI maps.

78 The water-ice column (WIC) is defined as M/ρ and expressed in precip-
79 itable (pr.) μm , with ρ being the density of water-ice. We have used the ICI
80 to estimate the WIC for the entire OMEGA data set to produce a multi-year,
81 near-global climatology of water ice aerosols.

82 After an introduction to the OMEGA instrument and its data product in
83 Section 2, we will introduce the retrieval algorithm in Section 3 and the new
84 prior data sets in Section 4. The subset of OMEGA data used in this study
85 is given in Section 5. Results are presented in Section 6 and will examine
86 the parameter spaces of the retrieved, fixed, and computed variables; the
87 spatial distributions of those variables; seasonal and diurnal trends; the link
88 between water-ice column mass and ICI; and efforts to constrain the retrieval
89 to obtain better performance in the presence of very thin clouds.

90 2. OMEGA

91 The OMEGA instrument, on Mars Express, records three-dimensional
92 image cubes where each x - y position contains the reflectance spectrum be-
93 tween 0.35–5.1 μm . The Mars Express orbit is an ellipse with periareion

94 and apoareion of 298 km and 10,107 km, and OMEGA records images that
 95 are 16, 32, 64, or 128 pixels wide, depending on the orbital characteristics
 96 at the time, and that may be thousands of pixels long. The spectra are
 97 recorded by a 352 detector array in three channels: visible (0.35–1.05 μm),
 98 C (1–2.77 μm), and L (2.65–5.1 μm). Spatial resolution of the image cubes
 99 depends on the orbital parameters at the time of observation, but they are
 100 typically 20–30 km wide for 32 pixels, or 5–7 km wide for 16 pixels (these two
 101 cases account for 80% of our data set), but can be several hundred kms wide
 102 for 128 pixels. The signal-to-noise is at least 100 and the spectral resolution
 103 is generally between 0.013–0.02 μm (Bibring et al., 2004).

104 Mars Express was launched in June 2003, arrived at Mars in December
 105 2003, began returning OMEGA data in January 2004, and continues to do so,
 106 having completed over 14,000 orbits. Unfortunately, instrument degradation
 107 over time has led to the death of several pixels, and in September 2010 the
 108 coolant supply used by the C channel detectors was exhausted. OMEGA
 109 continues to record cubes with the visible and L channels, but our analysis
 110 requires the C channel, and so data is limited to 2004–2010, covering three
 111 Mars years (MY), in this study.

112 **3. Retrieval Algorithm**

113 The inversion method used here, as described in Madeleine et al. (2012),
 114 fits a computed spectrum at seven wavelengths which cover the 1.5 μm , 2 μm ,
 115 and 3.1 μm water absorption bands. The steps in our analysis are: identifying
 116 OMEGA observations with sufficient cloud cover that overlap the OMEGA
 117 surface albedo data set, selecting pixels for analysis, matching cloudy pixels

118 to surface albedo pixels, and performing spectral fitting to retrieve r_{eff} and
119 τ_i .

120 The OMEGA albedo cubes are divided into groups by quadrangle (Bat-
121 son et al., 1979), and pixel matching was done independently for each group.
122 A bounding box is defined for each group by their latitude and longitude
123 extrema. An OMEGA image cube is selected for pixel matching if a portion
124 of its spatial coverage is within the bounding box, and if the cloud cover
125 is greater than 10%. Cloud coverage is defined as the percentage of pixels
126 in a cube with an ICI of ≤ 0.72 . The ICI is defined as the ratio of mea-
127 sured reflectances at 3.4 and 3.52 μm (Langevin et al., 2007), which indicates
128 whether the shape of the 3.1 μm water absorption band reflects the presence
129 of ice. Note that with this definition a lower value means thicker clouds, and
130 the quantity $1 - \text{ICI}$ is also used elsewhere. The threshold value of 0.72 is a
131 qualitative convention adopted by past studies of the ICI (Madeleine et al.,
132 2012; Szantai et al., 2017b, 2019).

133 For each pixel in a cloudy OMEGA cube, a set of criteria must be met for
134 it to be selected for analysis. The ICI must have been successfully retrieved
135 and be ≤ 0.72 . We want to avoid areas with strongly changing brightness
136 features, so the ICI is compared to those of the surrounding pixels. We
137 required that each of the eight surrounding pixels had a successfully retrieved
138 ICI and that the difference between the candidate pixel and each of the
139 surrounding pixels be less than 6%. Smoothing will be implemented by
140 taking the average spectrum with surrounding pixels. To avoid introducing
141 a bias by introducing pixels several times, we also required that no more
142 than three of the surrounding pixels overlap with those of the previously

143 selected pixel. Finally, the closest pixel in the set of albedo cubes is found.
144 The distance between a candidate pixel and the nearest albedo pixel must
145 be within 1.5 km.

146 The spectrum to be fitted is the mean of the selected pixel and the eight
147 pixels surrounding it. This reduces instrumental noise, but increases the foot-
148 print of the observation being analyzed. The surface albedo is the most criti-
149 cal parameter in the computed spectrum, but the footprint of the observation
150 used to retrieve albedo can already be very large and will not perfectly over-
151 lap the footprint of the cloudy pixel being analyzed. The averaging smooths
152 the spectrum, and reduces the effects of small-scale topographical features
153 on reflectance. Our requirement on the similarity of the ICIs of all pixels
154 in the mean improves the likelihood that we are analyzing a homogeneous
155 cloud mass.

156 Spurious behaviour has been observed in portions of the OMEGA spectral
157 images during different time periods. These pixels are excluded. No orbits
158 after 8485 (MY 30, Ls 135°) are considered due to the loss of coolant in the
159 C-channel detectors. A known misalignment exists between the L- and C-
160 channels during different time periods. This is corrected in both the cloudy
161 OMEGA cubes and the albedo cubes.

162 The wavelengths in the OMEGA spectra used for the retrieval were cho-
163 sen to avoid interfering gaseous absorption. A transmission spectrum of the
164 Martian atmosphere in this wavelength region is characterized by several
165 gaseous absorption features (see [Madeleine et al. \(2012\)](#)) which can be ex-
166 cluded from the fitting to reduce the number of parameters fit and avoid
167 steps taken to correct for gaseous absorption, which both result in reduced

168 uncertainties. The detectors for two of the wavelengths previously used be-
 169 came faulty during later orbits, so we use adjacent wavelengths. This version
 170 of the retrieval method computes spectral reflectances at 1.18, 1.49, 1.73,
 171 2.23, 2.43, 3.40, and 3.52 μm (previous pixels included 1.51 and 2.46 μm).

172 Reflectances are computed using the DISORT radiative transfer code
 173 (Stamnes et al., 1988), and the a priori discussed in Section 4: the surface
 174 albedo at each wavelength retrieved from a cloud-free, overlapping OMEGA
 175 cube, climatological dust opacity, surface temperature and pressure, and the
 176 atmospheric temperature profile. Details about the forward modelling of the
 177 computer spectrum and sources of error are given in Madeleine et al. (2012).
 178 The cloud layer is assumed to be at 25 km, and dust is assumed to be uni-
 179 formly mixed. Single scattering parameters for dust are calculated using the
 180 T-Matrix code (Mishchenko et al., 1996), refractive indices from Wolff et al.
 181 (2009), and Gamma size distribution with $r_{\text{eff,d}} = 1.5 \mu\text{m}$, $\nu_{\text{eff,d}} = 0.3$, as
 182 recommended by Wolff et al. (2009). For water ice, single scattering param-
 183 eters are deduced from Mie theory, refractive indices are taken from Warren
 184 and Brandt (2008), and a log-normal distribution of effective variance, ν_{eff} ,
 185 of 0.1 is used. The computed spectrum is then fitted to the measured spec-
 186 trum, with r_{eff} and τ_{i} as free parameters. Minimization is done using a
 187 Levenberg-Marquardt least squares routine. An example OMEGA spectrum
 188 and corresponding surface albedo and best fit are shown in Fig. 1a. Fig. 1b
 189 shows the water absorption features present in the OMEGA data by com-
 190 paring a spectrum obtained in the presence of clouds, and the corresponding
 191 cloud-free spectrum used to retrieve surface albedo.

192 The retrieval of water-ice cloud properties from the OMEGA image cubes

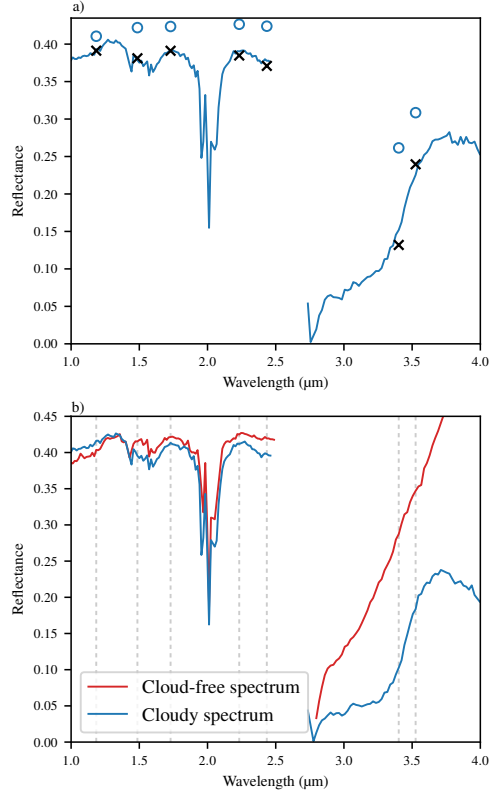


Figure 1: a) An example OMEGA spectrum and best fit for a single selected pixel. Shown are the spectrum, in blue, from the C channel (1–2.6 μm) and L channel (2.6–4 μm); the a priori albedo, blue circles; and the best fit results, black $\times$. b) Comparison of two OMEGA observations of the same location for periods when water ice clouds were present, blue, and when the region was cloud free, red, and the spectrum was used to retrieve surface albedo.

193 is very difficult and not always successful. The majority of clouds observed
 194 are too optically thin for significant water-ice absorption features to appear
 195 in the spectra. The OMEGA instrument has limitations for this sensitive
 196 application, such as a narrow spectral range, coarse spectral resolution, and
 197 a noise level that is significant relative to the water-ice features. The retrieval

198 depends on several a priori parameters which are uncertain. Errors made in
 199 each a priori parameter will propagate to the retrieved values of r_{eff} and τ_i .
 200 Of critical importance are the retrieved albedo values. Our retrieval is most
 201 sensitive to this parameter. Note, however, that both the surface albedo
 202 retrieval and the cloud properties retrieval use the same source data for the
 203 dust optical depth. Changing the a priori dust optical depth, atmospheric
 204 temperature, or surface temperature results in small changes in the results.
 205 Varying the albedo, however, can result in non-convergence.

206 In Fig. 2, we show the distribution of χ^2 values (see [Madeleine et al.](#)
 207 [\(2012\)](#) for details) resulting from a range of r_{eff} and τ_i values for two spectra,
 208 for which we found retrieved r_{eff} values of $3.8\text{ }\mu\text{m}$ (panel a) and $7.6\text{ }\mu\text{m}$ (panel
 209 b). The higher retrieved r_{eff} value is associated with a lower optical depth.
 210 The curves visible in both panels show allowed r_{eff} and τ_i combinations for
 211 an equal water ice column mass. The OMEGA instrument is sensitive to
 212 the mass of water ice and any errors made in the retrieval of r_{eff} and τ_i will
 213 still reflect the mass of the column. Combinations of r_{eff} and τ_i that lie far
 214 from this curve (e.g., large particle sizes and optically thick clouds) cannot be
 215 used to accurately model an OMEGA spectrum. The relationship between
 216 r_{eff} and τ_i is discussed in sections [6.2](#) and [6.3](#).

217 We also see that the regions of low χ^2 , where good spectral fits and likely
 218 solutions are found, tend to be large areas where several possible solutions
 219 can produce reasonable fits. For optically thin clouds, with low τ_i values,
 220 these regions can be broad, but still follow a curve of equal mass in the
 221 parameter space, as shown in Fig. 2b. For retrievals with $\tau_i \lesssim .4$ a wide
 222 range of r_{eff} values produce reasonable fits, and the minimization can find

best-fit solutions with r_{eff} values greater than our expectations of 2–5 μm , which are based on past observations such as those shown in Fig. 4c.

A large effort was made to understand what causes these high- r_{eff} retrievals and whether they were reliable and properly representative of the state of the atmosphere. This effort is discussed in Section 7. This type of solution is highly degenerative and occurs only for thin clouds when the water-ice absorption in the spectrum is weakest. Small errors in our signal (e.g., noise) or our model (e.g., errors in a priori) can lead to large errors in r_{eff} . Often, these results will occur within a cloud formation alongside retrieved r_{eff} values of $\sim 3 \mu\text{m}$, such as those shown in Fig. 6c, while we do not expect such extreme variation in r_{eff} within a single formation. However, we also find cloud formations predominantly made up of large particles, as shown in Section 7 (Fig. 12a). We have identified conditions that preferentially cause high- r_{eff} results, but we do not discount this data, as it may be real in some cases, and there are exceptions to any rule for data rejection. Instead, we present the full set of retrievals, pushing the limitations of our data and method, and urge caution when interpreting retrievals of $r_{\text{eff}} \gtrsim 10 \mu\text{m}$, which may be viable in some instances.

4. Prior Information

The retrieval algorithm requires accurate prior knowledge of the surface albedo at each fitted wavelength, the surface temperature the surface pressure, the vertical profiles of atmospheric pressure and of temperature, and the opacity of atmospheric dust. To select pixels for analysis, we also require the ICI. To aid studies of the cloud cover on Mars, pre-computed ICI maps

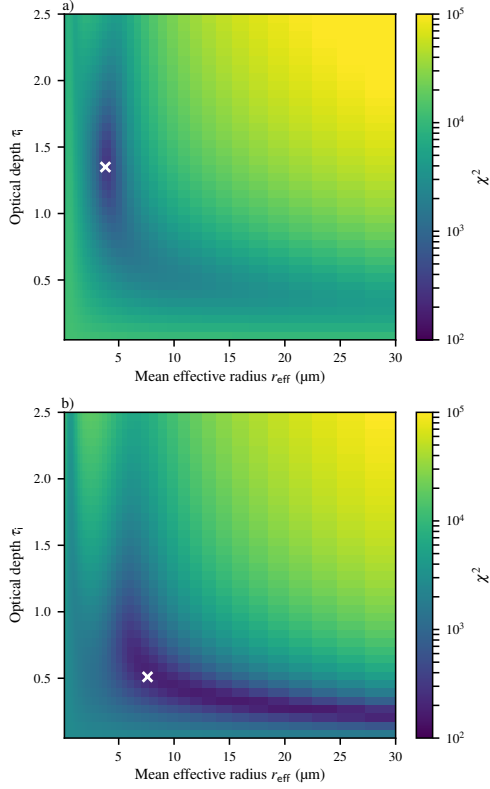


Figure 2: The distribution of χ^2 values for two OMEGA spectra with features that result in retrievals of: a) $r_{\text{eff}} = 3.8 \mu\text{m}$ and $\tau_i = 1.35$; and b) $r_{\text{eff}} = 7.6 \mu\text{m}$ and $\tau_i = 0.51$. The χ^2 minimization results are shown with white $\times$.

247 have been created for the entire OMEGA data set, along with a database
 248 of parameters and statistics (such as percent of pixels with water ice cloud
 249 cover) for each OMEGA cube (Szantai et al., 2017a).

250 To obtain the temperature of the surface and atmosphere for each pixel,
 251 we used the latest version of the LMD Mars general circulation model (LMD-
 252 GCM) (Forget et al., 1999) accessed through the Mars Climate Database
 253 (MCD) V5.2 (Millour et al., 2015). Madeleine et al. (2012) computed the ICI

254 for each analyzed pixel at the time of the retrieval and used an older version
 255 of the MCD (V4.3) to obtain temperature information. The more recent
 256 version of the MCD reflects significant improvements in the modelling of the
 257 main atmospheric cycles on Mars, such as dust, water, and CO₂, due to large
 258 improvements in the LMD-GCM, such as the inclusion of radiatively active
 259 clouds ([Madeleine et al., 2012](#)), semi-interactive dust transport, and new
 260 dust radiative properties ([Madeleine et al., 2011](#)). Other improvements were
 261 made to the cloud micro-physics ([Navarro et al., 2014](#)), convective boundary
 262 layer scheme, CO₂ cycle, and the representation of non-local thermodynamic
 263 equilibrium cooling in the upper atmosphere.

264 To get the atmospheric dust content, [Madeleine et al. \(2012\)](#) scaled the
 265 measurements from the nearest Mars Exploration Rover on the day of the
 266 observation, which would have been thousands of km away, assuming that
 267 dust was uniformly mixed horizontally and vertically. A new climatologi-
 268 cal database of dust opacities has been developed using observations from
 269 the Mars Global Surveyor Thermal Emission Spectrometer (TES), the Mars
 270 Odyssey Thermal Emission Imaging System (THEMIS), and the Mars Re-
 271 connaissance Orbiter Mars Climate Sounder (MCS), and including observa-
 272 tions from the Martian surface when available ([Montabone et al., 2015](#)). In
 273 our analyzed data set, there are no directly coincident dust measurements in
 274 the database, so we used the complete-coverage reconstructed maps (kriged
 275 data rather than gridded). In order to use the data in our retrieval algo-
 276 rithm, several steps were taken: the dust column provided is at a constant
 277 pressure level, and we correct to the surface pressure and elevation of each
 278 pixel location; the data provided are absorption optical depths and we re-

quire extinction optical depths, so they are converted by multiplying by a factor of 1.3 (Wolff and Clancy, 2003); the data provided are at $9.3\text{ }\mu\text{m}$, and we require the optical depth at $0.88\text{ }\mu\text{m}$, obtained by multiplying by a factor of 2.0 as recommended by Montabone et al. (2015). These approximations introduce non-negligible uncertainties into the dust content used by both the surface albedo retrievals and cloud properties retrievals.

In order to retrieve cloud properties from OMEGA spectra, Madeleine et al. (2012) first searched for a cloud-free OMEGA spectrum at the same location. A strict set of criteria were applied to the selection of a cloud-free spectrum, restricting the slope of the ground, difference in dust content between the cloud and cloud-free spectra, and the difference in their observation angles. The cloud-free spectrum was first used to retrieve the surface albedo using spectral fitting between the observed spectrum and a spectrum computed using the DISORT radiative transfer code (Stamnes et al., 1988) using the surface albedo as the free parameter. A priori information for the cloud-free spectrum came from the same sources as for the cloudy spectrum.

Geminale et al. (2015) have analyzed a subset of OMEGA image cubes to retrieve surface albedo for each wavelength of the OMEGA spectra. Their retrieval method uses a combination of principle component analysis and target transformation to remove the gaseous atmospheric contribution to the OMEGA spectra in the spectral range between $0.4\text{--}4\text{ }\mu\text{m}$. Their aim is to remove the spectral components produced by the atmosphere to improve studies of Martian surface mineralogy. The covariance matrix for a set of OMEGA spectra is used to derive a set of eigenvectors which are transformed into a set of spectral end-members that can be used to reproduce the

304 observed spectra. A spectrum of the surface reflectance is constructed from
 305 a linear combination of spectral end-members with those attributed to the
 306 gaseous atmospheric contribution removed. Moreover, the dust contribution
 307 has been removed by the spectrum considering that the observed reflectance
 308 factor at a given wavelength is a function of surface reflectance and dust
 309 optical depth (Vincendon et al., 2007). Taking advantage of this relationship
 310 and using dust opacities from Montabone et al. (2015), it is possible to sim-
 311 ulate reflectance factors using a multiple-scattering radiative transfer code
 312 (Ignatiev et al., 2005). The surface reflectance factor is determined as the
 313 value that corresponds to the best fit between the observed reflectance factor
 314 and simulated one for each wavelength. Finally, the surface thermal contri-
 315 bution is removed in the spectral range between 3–4 μm (Audouard et al.,
 316 2014). It should be taken into account that the surface spectra retrieval
 317 strongly depends on the assumptions made on dust (grain size distribution,
 318 radius, and variance) and on the dust optical depth values. Indeed, dust
 319 properties could depart from the ones assumed in the model and this can
 320 be even more important when dust optical depth is high. This is the same
 321 conclusion reached by Madeleine et al. (2012).

322 In their original publication, Geminale et al. (2015) analyzed two OMEGA
 323 image cubes, but have since expanded their data set. We used 60 multi-
 324 spectral albedo cubes in this study. A comparison of the albedos retrieved
 325 by Madeleine et al. (2012) and by Geminale et al. (2015) is discussed in Sec-
 326 tion 6 and shown in Fig. 4. Using the ICI as an indicator, and considering
 327 the fraction of pixels with an ICI less than 0.72, 55 of the 60 OMEGA image
 328 cubes are cloud free, and four of them have cloud cover over less than 0.3%

329 of their area. The remaining one cube has a cloud fraction of 9%, but none
330 of the overlapping pixels used from this cube bore clouds. Dust affects the
331 accuracy of the albedo retrieval and can impact cloud properties retrievals.
332 This is discussed in Section 7.

333 5. Data Set

334 The OMEGA data were produced in groups and cover the Mars quad-
335 rangles MC09 (Tharsis) to MC13 (Syrtis Major). This region stretches from
336 140°W to 90°E and from 35°S to 35°N. The coverage of the albedo data is
337 presented in Fig. 3, which shows the surface albedo at 1.51 μm . From these
338 60 albedo cubes, we found 94 cloudy OMEGA image cubes with overlap-
339 ping pixels suitable for analysis. Table 1 gives a summary of the number of
340 pixels analyzed over each region, and shows that the highest number were
341 found over the Tharsis region, where the extreme topography of the Tharsis
342 volcanoes drives cloud formation. In total, we analyzed 209,936 pixels, with
343 158,190 coming from the MC09 and MC17 regions.

344 No restriction was placed on Ls, but the majority of cloudy observations
345 found come from the aphelion period, between Ls 35° and Ls 135°. The
346 distribution of observations has two peaks near Ls 50° and Ls 100°. There
347 are four observations between Ls 200°–208°, one observation at Ls 12.5°, and
348 another at Ls 351.2°.

349 A set of data quality cuts were applied to the data to remove poor fits, ex-
350 treme outliers, and physically unrealistic values. The data have been archived
351 conforming to the Planetary Data System (Version 4) for distribution as part
352 of the UPWARDS project through the ESA Planetary Science Archive at

Table 1: Summary of analyzed pixels from the OMEGA data set showing the locations, Mars quadrangles (MC), number of OMEGA observations used (Obs.), the total number of pixels analyzed, and the number of analyzed pixels passing data quality cuts.

Region	MC	Area	Obs.	pixels	After cuts
Tharsis	09	0°–30°N 90°–135°W	65	142498	100891
Oxia Palus	11	0°–30°N 0°–45°W	20	30771	9203
Arabia	12	0°–30°N 0°–45°E	4	10926	7202
Syrtis Major	13	0°–30°N 45°–90°E	10	10049	6959
Phoenicis Lacus	17	-30°–0°N 90°–135°W	12	15692	5517

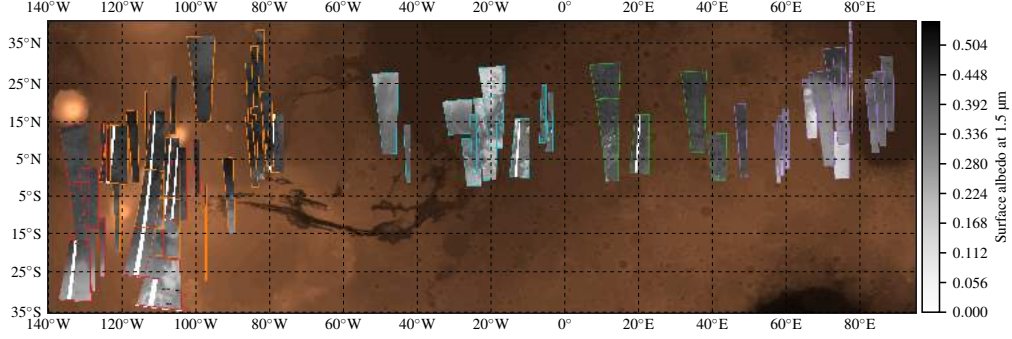


Figure 3: A map of available retrieved surface albedo data, showing the surface albedo at 1.51 μm . The background is a relief of MOLA surface elevation. The outlines indicate the Mars quadrangle, from left to right: MC17 (red), MC09 (orange), MC11 (blue), MC12 (green), and MC13 (violet).

353 open.esa.int/esa-planetary-science-archive/. The data quality re-
 354 quirements applied are: $\chi^2 < 2000$, $0.5 < r_{\text{eff}} < 25$, and $\tau_i > 0.15$. The
 355 χ^2 requirement affects 49306 (23%) pixels, the requirement imposed on r_{eff}
 356 affects 60736 (29%) pixels, and the τ_i requirement affects 6519 (3%) of pix-
 357 els. Overall, 62% of analyzed pixels passed the data quality cuts. The total
 358 number of pixels passing the cuts for each region is given in Table 1. The
 359 fitting of 30056 pixels failed to converge, only two thirds of which resulted in
 360 χ^2 outside our criterion. Fitting fails to converge when our computed spec-
 361 trum does not properly model the measured spectrum, which can occur for a
 362 variety of reasons that include: errors made in the measurement (e.g., detec-
 363 tor degradation over time), a water-ice signal below the instrument noise, or
 364 topographic features impacting reflectance (e.g., crater rims). Our investiga-
 365 tion into results with good fits, but large retrieved values of r_{eff} is discussed
 366 in Section 7.

367 6. Results

368 Armed with the new multi-spectral albedo data set, the first thing we
 369 examined was the effect the new a priori had on retrievals by comparing
 370 them with those presented in [Madeleine et al. \(2012\)](#). We also examined the
 371 sensitivity of the retrieval to each piece of a priori information by comparing
 372 retrievals after updating only one parameter at a time. What we observed,
 373 for 13 of the 14 locations examined in [Madeleine et al. \(2012\)](#) (the fourteenth,
 374 orbit 1034_6 (MY 27, Ls 111.9°), had errors in its ICI analysis and so was
 375 omitted here), was that changing the a priori dust, surface temperature, or
 376 atmospheric temperature vertical profiles caused changes in the retrieved r_{eff}
 377 on the order of tenths of μm , with the dust having the largest impact. Two
 378 of the locations, in orbits 0887_5 (MY 27, Ls 93.4°) and 1023_6 (MY 27,
 379 Ls 110.6°), proved challenging to reproduce the results in [Madeleine et al.](#)
 380 [\(2012\)](#) with any changes to the a priori. [Madeleine et al. \(2012\)](#) showed that
 381 orbit 0887_5 had the highest sensitivity to initial conditions in their study,
 382 resulting in the highest uncertainty, while orbit 1023_6 had the lowest water-
 383 ice column (WIC), implying the thinnest layer of cloud cover analyzed. We
 384 confirm that this pixel also had the highest ICI value within the data set.
 385 Finally, changing the a priori surface albedo had a critical impact on the
 386 results, causing differences of a few tenths of μm to a few μm in the retrieved
 387 r_{eff} .

388 A comparison of the albedo data retrieved by [Madeleine et al. \(2012\)](#) and
 389 [Geminale et al. \(2015\)](#) is shown in Fig. 4, which shows a correlation plot
 390 of all the spectral points for the 13 pixels re-analyzed, as well as the mean
 391 difference at each spectral point between the albedos retrieved by [Madeleine](#)

et al. (2012) and Geminale et al. (2015). We observe a small bias in the results, with those from Geminale et al. (2015) having higher albedos in the C channel region, and much lower albedos in the L channel region. The L channel has the strongest water-ice absorption feature, is the most difficult region of the spectrum to fit, and has a significant impact on the results. It should be noted that the results of (Geminale et al., 2015) also benefit from improvements made to the sources of surface temperature, dust optical depth, and an updated version of the L channel calibration.

Fig. 4 also compares the retrieved r_{eff} results using the new a priori information to those reported in Madeleine et al. (2012) by reproducing Fig. 11a in that paper, which shows r_{eff} as a function of Ls. We find that after updating the a priori, we no longer observe the trend in Fig. 4c discussed in Madeleine et al. (2012), and that some of our results now deviate strongly, illustrating how sensitive and difficult this retrieval is.

In the original analysis of Madeleine et al. (2012), they carefully examined the sensitivity of the retrieval to each piece of a priori information to quantify the uncertainty of their results, and we refer the reader there to further explore the retrieval sensitivity. They obtained uncertainties for retrieved r_{eff} between 0.4–1 μm , as indicated in Fig. 4c, and between 0.09–0.13 for τ_i . In all cases, the new apriori data produce better fits for the spectra analyzed by Madeleine et al. (2012). Because of that, and our belief that the new surface albedo is more accurate than before, the uncertainties will be on the same order, but less than those presented by Madeleine et al. (2012). Uncertainties for individual pixel retrievals had not been computed for the entire 200,000 pixel data set at the time of writing due to the unavailability of

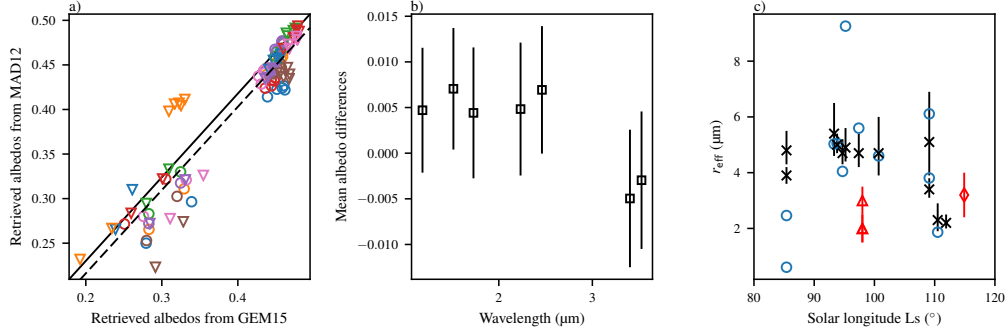


Figure 4: A comparison of the inputs and retrievals presented in [Madeleine et al. \(2012\)](#) and here. Panel a) shows the correlation plot between retrieved aerosol-free surface albedos from [Madeleine et al. \(2012\)](#) and [Geminale et al. \(2015\)](#) (labelled MAD12 and GEM15, respectively). Data shown are for the OMEGA orbits analyzed in [Madeleine et al. \(2012\)](#). Each observation has seven points for each wavelength fit in Fig. 1, and is given a unique colour and symbol combination. Panel b) shows the mean albedo differences at each retrieval wavelength for the same data as in panel a). Panel c) shows the retrieved mean effective radius as a function of solar longitude. Results found for single pixels by [Madeleine et al. \(2012\)](#) are shown in black, and our results, which are the mean of the analysis of many pixels, are shown in blue. Also shown are measurements made by [Zasova et al. \(2001\)](#) (red triangles) and [Wolff and Clancy \(2003\)](#) (red diamond).

417 rigorously quantified uncertainties in the surface albedo data set, which our
 418 calculation is critically dependant on. However, a very good estimate of the
 419 uncertainties is determined by considering the ensemble of analyzed pixels,
 420 shown in Fig. 5. The standard deviation of retrieved r_{eff} in the primary
 421 peak centred near $2\mu\text{m}$ is $0.81\mu\text{m}$. The corresponding standard deviation
 422 for retrieved τ_1 is 0.51 , indicating that the particle size has less variability
 423 than the optical depth.

424 The objective of this analysis is two-fold: to circumvent the pitfalls of

425 fitting a single spectrum from a cloud with uncertain a priori information
 426 by analyzing a great number of spectra and considering the entire sample
 427 of their results; and to explore the variability of water-ice cloud physical
 428 properties within a cloud formation.

429 We have analyzed 94 OMEGA image cubes, with the number of pixels
 430 used ranging from a few hundred to several thousand. Fig. 5 presents his-
 431 tograms of r_{eff} , τ_i , ICI, and χ^2 for the entire data set, after applying data
 432 quality cuts outlined in Section 5. The r_{eff} results are characterized by hav-
 433 ing two distinct peaks: a tall, narrow peak centred around 2.21 μm (with a
 434 standard deviation of 0.81); and a wide, low peak centred around 12.4 μm .
 435 (standard deviation of 6.2). Note that the appearance of a minor peak near
 436 5 μm is due to two observations, 0887_5 and 0898_5, with large samples of
 437 analyzed pixels and means r_{eff} of 5.2 and 4.7 μm , respectively. These obser-
 438 vations were both analyzed by [Madeleine et al. \(2012\)](#) who found similarly
 439 large r_{eff} values (5.4 and 4.7 μm). They were recorded on Ls 93.5° and 94.8°
 440 in MY 27. Two distinct populations of water ice aerosols were also observed
 441 by [Wolff and Clancy \(2003\)](#) and were interpreted to represent distinct cloud
 442 types.

443 These results are in agreement with previous observations and analyses
 444 (e.g. [Toon et al., 1977](#); [Clancy et al., 1995](#); [Pollack et al., 1995](#); [Wolff and](#)
 445 [Clancy, 2003](#)), which found water ice aerosol sizes to be around 1–2 μm . [Fe-](#)
 446 [dorova et al. \(2014\)](#); [Guzewich et al. \(2014\)](#); [Clancy et al. \(2003\)](#) observed an
 447 altitude-dependence in water ice aerosols formed over the aphelion cloud belt,
 448 with smaller particles (1–2 μm) forming at higher altitudes, and larger par-
 449 ticles (3–4 μm) forming closer to the surface. Smaller particle sizes (0.1 μm)

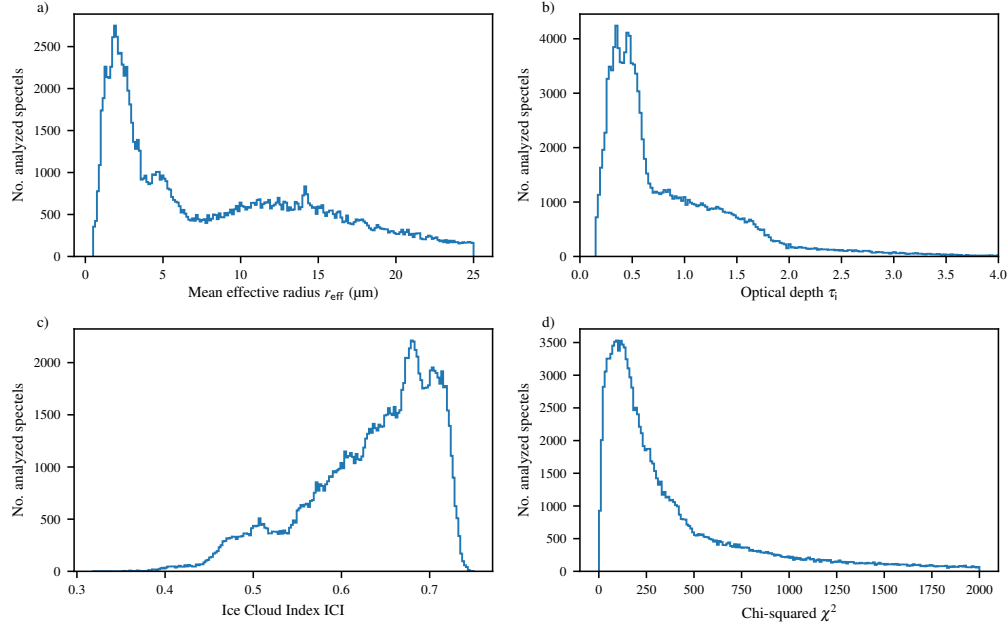


Figure 5: Histograms of: a) retrieved r_{eff} ; b) retrieved τ_i ; c) ICI; and d) χ^2 of the best fit for all analyzed pixels in our data set, after applying data quality cuts.

450 were also observed by [Rannou et al. \(2006\)](#). Note that τ_i is generally less
 451 easy to directly compare since it is not always reported at the same wave-
 452 length. [Madeleine et al. \(2012\)](#) compares the 0.67 μm τ_i retrieved with the
 453 method used here to the 0.4 μm τ_i reported by [Benson et al. \(2003\)](#), and note
 454 agreement within uncertainty.

455 The τ_i results also exhibit a less distinct pair of peaks, with the r_{eff} results
 456 above 7 μm corresponding to the τ_i results less than 0.6. The relationship
 457 between r_{eff} , τ_i , and ICI is discussed in sections [6.2](#) and [6.3](#), and the results
 458 with $r_{\text{eff}} > 7 \mu\text{m}$ are discussed in Section [7](#).

459 6.1. Spatial Distributions

460 Our approach to cloud properties retrievals has yielded a data set that
461 allows us to examine the distribution of cloud properties within cloud for-
462 mations. At this time, however, our data set is limited in size, but also by
463 the viewing geometry of the OMEGA instrument. Figs 6 and 7 provide two
464 examples of the mapping of retrieved cloud properties, and show observa-
465 tions from OMEGA orbits 0937_5 (MY 27, Ls 99.7°) and 3272_3 (MY 28, Ls
466 85.0°), respectively. From left to right they show a visible image derived from
467 OMEGA, the ICI, retrieved r_{eff} , and retrieved τ_{i} , with backgrounds showing
468 the topography and overlapping retrieved surface albedo. The visible images
469 are useful to identify the presence of cloud cover and distinct surface fea-
470 tures. A quick comparison of panels a) and b) in both figures highlights the
471 usefulness of the ICI to identify the presence of clouds.

472 The topography and visible images in panels a) also help evaluate aspects
473 the retrieved surface albedo. The surface albedo does not reflect changes in
474 elevation, but rapid changes in topography, such as craters visible in Fig. 6
475 and a large cliff band in Fig. 7, appear visible in the albedo data. Since
476 albedo values in these regions reflect physical changes in the surface and in
477 the illumination conditions rather than constant radiometric properties, they
478 prevent accurate cloud properties retrievals. In Fig. 6 there are three large
479 craters. Retrievals were not possible near 9°N, and retrievals resulted in very
480 high estimates of r_{eff} along the crater edges near 10 and 15°N. In Fig. 7, the
481 cliff band near 5°N affects the OMEGA spectra in such a strong way that
482 this feature is visible in the ICI measurements. Cloud properties retrievals
483 in this region result in unrealistically high τ_{i} values.

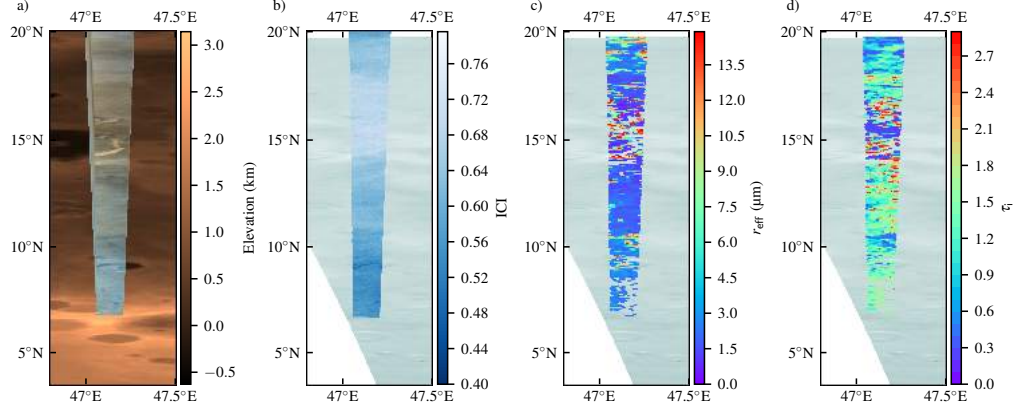


Figure 6: Spatial distributions of retrieved water-ice cloud properties for OMEGA orbit 0937.5 (Ls 99.7, and 16:15 LT). The foregrounds of each panel show the: a) visible image generated from the OMEGA observation, b) the ice cloud index, c) retrieved mean effective radius, and d) retrieved optical depth. The background of panel a) shows MOLA surface elevation, and the backgrounds of panels b), c), and d) show the surface albedo at 1.51 μm retrieved from a co-located cloud-free OMEGA observation.

These figures also illustrate the limitations of the retrieval when cloud thickness, as measured by the ICI, is low. Both figures feature regions with very thin clouds, indicated by high ICI values, where the retrievals result in unrealistic values. This occurs between 15 and 18°N in Fig. 6, where the retrieval found optically thin clouds with very large particles, and between 1 and 3°N in Fig. 7, where the retrieval found optically thick clouds when none were present.

6.2. Cloud Mass

The WIC and water-ice column mass can be computed from the product of the retrieved parameters, r_{eff} and τ_i (Mateshvili et al., 2007; Madeleine et al., 2012). Therefore, r_{eff} and τ_i are inversely related to one another

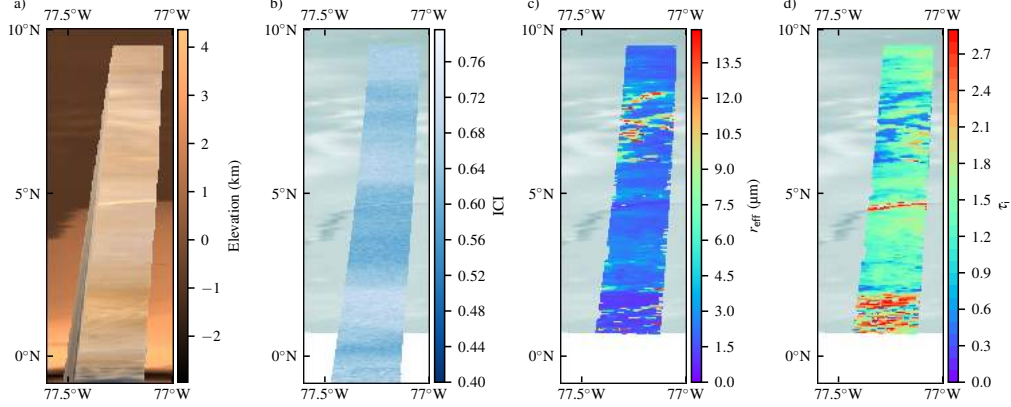


Figure 7: Spatial distributions of retrieved water-ice cloud properties for OMEGA orbit 3272.3 (Ls 85.0, and 14:11 LT). Panels are as in Fig. 6.

through the mass, extinction efficiency, and density of the scattering water-ice aerosols:

$$\tau_i = \frac{3MQ_{\text{ext}}}{4\rho r_{\text{eff}}}, \quad (1)$$

where M is the water-ice column mass along the line of sight, ρ is the density of water ice, and Q_{ext} is the extinction efficiency. A database of Q_{ext} values were pre-computed for range of r_{eff} values using method of [Bohren and Huffman \(1983\)](#) for a log-normal distribution of scattering particles with an effective variance of 0.1, as in [Wolff and Clancy \(2003\)](#) and refractive indices from [Warren and Brandt \(2008\)](#). M is directly proportional to WIC ($\text{WIC} = M/\rho$) which was derived from OMEGA retrievals by [Madeleine et al. \(2012\)](#).

Eq. 1 allows us to calculate the column mass of water-ice for each pair of retrieved values. Fig. 8 shows four examples of the spatial distribution of column mass. This is more homogeneous than the individual parameters

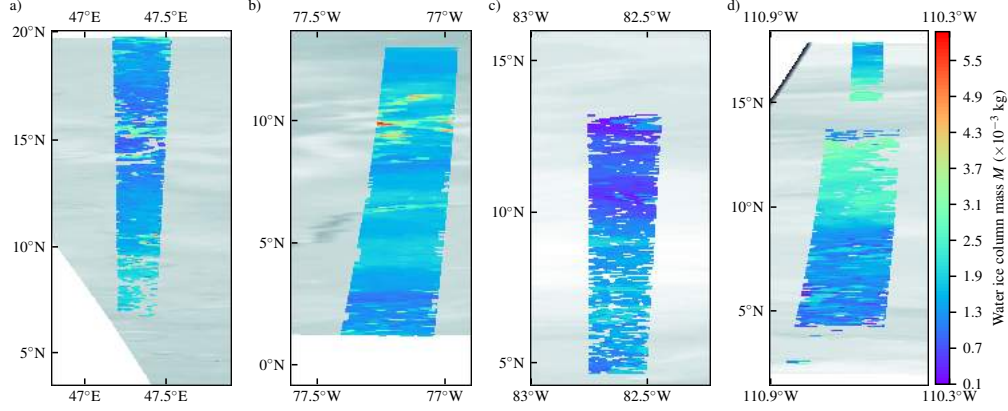


Figure 8: Spatial distributions of retrieved water-ice cloud column mass for OMEGA orbits: a) 0937_5 (Ls 99.7, LT 16:15, MY 27), b) 3272_3 (Ls 85.0, LT 14:11, MY 28), c) 2957_5 (Ls 46.3, LT 16:54, MY 28), and d) 3276_4 (Ls 85.5, LT 14:22, MY 28).

and well correlated to the distribution of ICI maps. For example, compare
 Figs 6b and 7b to Fig. 8a and 8b. With the exception of areas around crater
 edges, we find that the mass distribution is aligned with the ICI distribution,
 with greater mass correlated to thicker clouds, as expected.

6.3. ICI and Cloud Mass

The relationship given in equation 1 is extremely well obeyed by the cloud
 properties retrievals from OMEGA observations. This implies that even when
 a retrieval results in an unrealistic value of r_{eff} , the retrieval is still well enough
 constrained to obtain a reasonable estimate of water-ice cloud column mass
 because the cloud spectral signature is primarily controlled by absorption.
 Fig. 9a shows the relationship between retrieved r_{eff} and τ_1 values. The colour
 scale is the ICI and the shapes exhibited by monochromatic bands are curves
 of equal mass. What we observe is that a retrieval resulting in high mean

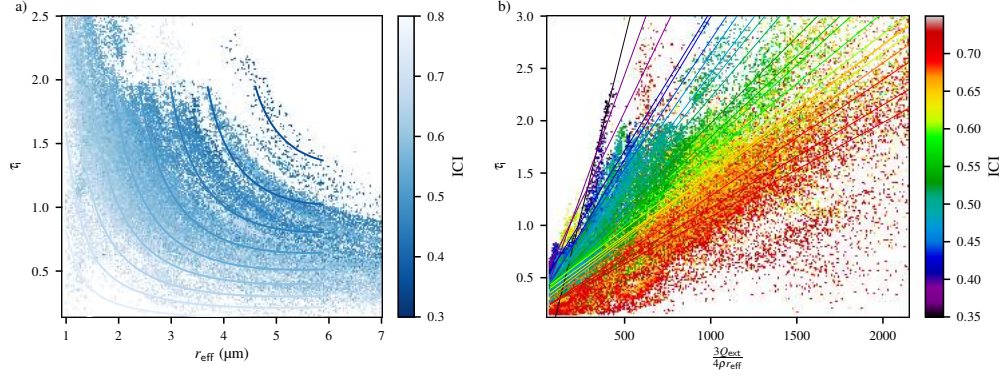


Figure 9: The parameter space of retrieved water-ice cloud properties. Panel a) shows τ_i as a function of r_{eff} , with the ICI represented with the colour scale. The contours show ICIs calculated from synthetic spectra generated for the range of r_{eff} and τ_i values shown and represent curves of equal mass. Panel b) shows the same data as in panel a) linearized using equation 1. The data was binned according to ICI (each bin represented by a colour) and the slope found. Best fit lines for each ICI are shown, coloured to match the data.

516 effective radius is always associated with a very low optical depth. We also
 517 see a trend in the position of the inverse relationship with ICI, and therefore
 518 infer that column mass, M , is be related to ICI. This relationship between
 519 r_{eff} and τ_i is also exhibited in the χ^2 distributions shown in Fig. 2.

520 To confirm this relationship we created a model of the $r_{\text{eff}}-\tau_i$ parameter
 521 space. For a grid of r_{eff} and τ_i values covering the domain and range of Fig. 9a,
 522 and a synthetic spectrum was computed at each location. The ICI was found
 523 for each synthetic spectrum and the model results are shown in Fig. 9a as
 524 contours of ICI. A trend in the contour location with ICI is clearly seen, and
 525 the colour scale agrees with the retrieval results. However, the relationship is
 526 not exact since the synthetic spectra also depend on surface albedo and the
 527 atmospheric dust loading. The model was recreated for several albedo and

dust configurations, and each different scenario alters the contour locations. The spread in the data that have similar ICI values in Fig. 9a is due to variations in those physical parameters.

Fig. 9b presents the data from Fig. 9a in a linearized form, showing the relationship between τ_i and $1/r_{\text{eff}}$. The data were grouped into ICI bins (22 bins, widths of 0.019), which are represented by the colours in Fig. 9b, and the slope and intercept of each group were found, with the slope being directly proportional to column mass as in equation 1. The resulting column masses as a function of ICI are shown in Fig. 10. Fig. 10 also shows the results of the same analysis performed for the synthetic data shown in Fig. 9, as well as the distribution of individual retrievals (restricted to r_{eff} values below $7\text{ }\mu\text{m}$).

What we have found is that the ICI, which is very straight-forward to compute (ratio of reflectances at 3.4 and $3.52\text{ }\mu\text{m}$), is a good proxy for the column mass of the water-ice aerosols present, or the WIC. Though estimating M or WIC from the resulting relationship shown in Fig. 10 will not be as accurate as performing a full retrieval of r_{eff} and τ_i , it is much less computationally expensive and has applications for other data sets. For example, full retrievals of r_{eff} and τ_i cannot be made for the entire OMEGA data set due to a lack of retrieved surface albedo data, a lack of suitable overlapping cloud free observations to use for the albedo retrieval, and the relatively short life of the C-channel over the mission duration. This relationship allows us to generate maps of the distribution of M or WIC for the entire OMEGA data set between 2004 to the end of life, when water-ice clouds are present, as determined by the ICI. The primary limitation in this method is that the

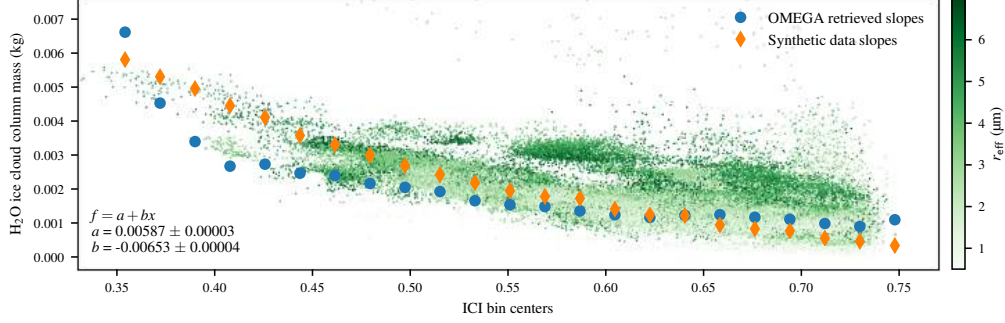


Figure 10: The relationship between water-ice cloud column mass and the ICI. The entire data set, limited to retrievals with $r_{\text{eff}} < 7 \mu\text{m}$, is shown in green, with the colour scale indicating r_{eff} value. The slopes from the binned data in Fig. 9b are shown in blue, as a function of ICI bin centre. The same analysis was performed for the synthetic data shown in Fig. 9a, and its results are shown in orange. Least squares regression results for the $r_{\text{eff}} - \tau_1$ data set are indicated.

553 distribution of calculated M values will be narrow and will not fully reflect
 554 anisotropies in less well constrained parameters such as atmospheric dust
 555 opacity and surface albedo.

556 Regression in the data shown in Fig. 10 is not very well constrained at
 557 low ICI and future work will aim to increase the sample size of retrievals
 558 from this type of spectra. The slope and intercept given in Fig. 10, -0.00653
 559 and 0.00587 , respectively, are the regression statistics for the individual M
 560 and ICI values shown. The same analysis has been done for binned data
 561 (resulting in $b = -0.009$ and $a = 0.007$), but this is strongly affected by the
 562 lack of data below an ICI of 0.4 . We also fit various curves to the data, but
 563 this leads to unrealistically high and poorly constrained results at low-ICI.

564 We have used this relationship to estimate the WIC for the entire available

OMEGA data set covering Mars years 26 to 32 (455 million pixels containing clouds). Results were binned and averaged according to Ls and latitude. Fig. 11a) shows the climatology of WIC estimated from the OMEGA data using our empirical relationship between ICI and M . Clearly visible is the aphelion seasonal cloud belt which peaks near Ls 90°. This matches water ice aerosol climatologies derived from other instruments, such as the TES (Smith, 2004), THEMIS (Smith, 2009), and the Mars Express Ultraviolet and Infrared Atmospheric Spectrometer (SPICAM) (Mateshvili et al., 2009; Willame et al., 2017). These climatologies have been made for water ice optical depth, are generally in agreement with one another, and feature two prominent features: the aphelion clouds between Ls 30° and 180°, and between 30°S and 30°N; and the polar hoods, along the northern and southern limits of coverage. The climatologies derived from TES and THEMIS have very good coverage, but view less of the polar hoods than what we show in Fig. 11, while those derived from SPICAM are very sparse. In the fringes of the polar hoods observed by TES and THEMIS, the opacity of the water ice clouds is equal to those over the aphelion belt, while we have observed clouds with greater mass in the polar hoods than in the aphelion belt. This was also observed by SPICAM, though with limited coverage.

The magnitude of the WIC within the aphelion cloud belt is between 1.2–1.6 pr. μm , but climbs above 2.5 pr. μm in the polar hoods, where the WIC is generally between 1.5–2.5 pr. μm . It must be noted that the polar hoods are only covered on the fringes of the spatial domain of OMEGA observations, fewer data points are used in the climatology, resulting in higher standard errors for these points. Madeleine et al. (2012) reported WIC values

590 between 2–3.5 pr. μm for single OMEGA pixels over the aphelion belt, and
591 noted that this result was higher than previous measurements presented by
592 [Benson et al. \(2003\)](#) and [Mateshvili et al. \(2007\)](#), who reported WIC values
593 of 0.07–2.1 pr. μm and 1.35–1.8 pr. μm , respectively.

594 The standard deviations of the climatology pixels are distributed between
595 0.1–0.3 pr. μm over the majority of the covered area, as shown in Fig. 11b).
596 Higher standard deviations, between 0.2–0.7 pr. μm , are seen in regions where
597 clouds may be present, such as the aphelion belt. Outliers, with standard
598 deviations between 1–2 pr. μm , appear on the southern edge of the OMEGA
599 limit near the south polar hood (L_s 30°–160°). These pixels correspond to
600 WIC enhancements in Fig. 11 (dark red and orange), and suffer below average
601 sample sizes ($< 5,000$ compared to a mean of 50,000). The enhancements
602 seen in the north polar hood ($L_s > 280^\circ$) have nominal sample sizes ($>$
603 10,000) and low standard deviations (0.25–0.75 pr. μm). The standard errors
604 for the WIC climatology pixels shown in Fig. 11 are distributed between
605 0.0002–0.0015 pr. μm .

606 A detailed analysis of the ICI distribution and climatology has already
607 been performed by [Szantai et al. \(2017b, 2019\)](#), with a focus on examining
608 the extent of cloud cover and the diurnal cycle of water-ice clouds. That
609 work compared the ICI and the percentage of cloudy pixels for an OMEGA
610 observation to the integrated water ice optical thickness derived from TES
611 ([Smith, 2004](#)) and the integrated water ice column from the MCD ([Forget](#)
612 [et al., 1999](#); [Navarro et al., 2014](#)). OMEGA water ice and cloudiness indica-
613 tors were temporally averaged onto maps (latitude and longitude) and were
614 compared to spatially and temporally collocated TES and MCD data.

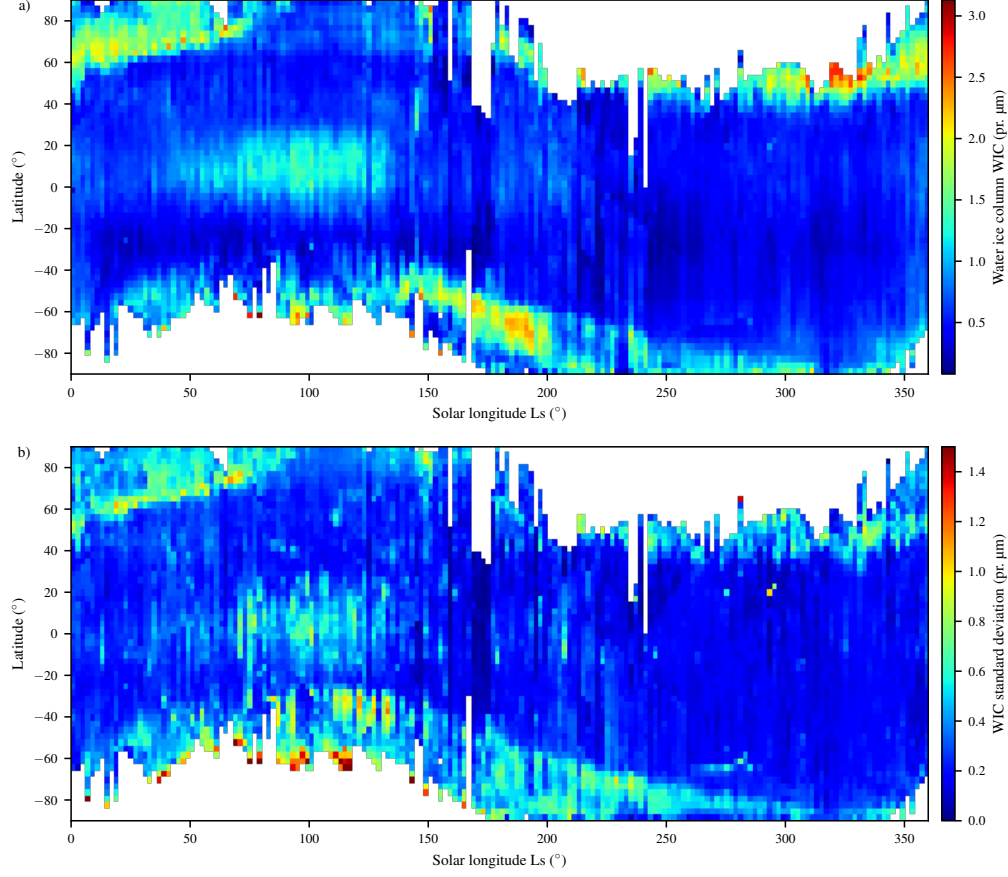


Figure 11: a) Water-ice column (WIC) climatology estimated from OMEGA ICI data. Data shown are for all years (MY 26–32) restricted to local daytime(06:00 to 20:00). Data are binned in a grid of 2° Ls and 3° latitude. Panel b) shows the corresponding standard deviation of the WIC computed for each bin.

615 7. Constraining r_{eff}

616 A very large number of the analyzed pixels resulted in higher than ex-
 617 pected values of r_{eff} (e.g., with respect to [Wolff and Clancy, 2003](#); [Fedorova](#)
 618 [et al., 2014](#)), as seen in Fig. 5a. A large effort was dedicated to understand-

ing these results, to determine whether they were real and truly reflected the state of the atmosphere, to diagnose what elements of the spectral modelling and fitting were responsible, and to establish whether the retrieval could be modified or constrained to reduce their occurrence.

While some results are related to surface features, as seen in Figs 6 and 7, it is difficult to quantify their contribution to Fig. 5a without a method for identifying cratering and cliffs in the OMEGA images. The slope of the surface at a given pixel centre can be computed from the MOLA data, but due the spatial resolution, the maximum slopes found within our data set are only around $10\text{--}11^\circ$, very few pixels lie on slopes greater than a few degrees, and we found no correlation between pixels with steep slopes and high- r_{eff} retrievals or χ^2 .

To investigate these high- r_{eff} results, we looked for correlations between each parameter in the retrieval results (r_{eff} , τ_i , and χ^2) and any other properties of the pixels. For every property we investigated, we found poor correlations between it and our results. There are no clear rules that can be applied to the data to reject a pixel's result. We can obtain a good fit and a reasonable result when a pixel is abnormal in some aspect, while having a failed retrieval, either a very poor fit or a very high r_{eff} , when all aspects are nominal.

The properties of an OMEGA pixel that we investigated are: the surface elevation, the surface slope, the surface temperature, the ICI, the mean retrieved surface albedo (at spectral points used, below $2.5\text{ }\mu\text{m}$), the mean level of the cloudy spectrum (at spectral points used, below $2.5\text{ }\mu\text{m}$), the difference between the mean spectrum levels, the H_2O band depth at $1.5\text{ }\mu\text{m}$

644 (Langevin et al., 2007), the dust opacity in the cloudy spectrum, the dust
 645 opacity in the cloud-free spectrum used to retrieve surface albedo, the ratio
 646 of dust opacities between the cloudy and cloud-free spectra, the spatial res-
 647 olution of the pixel, the spatial resolution of the cloud-free pixel, the ratio
 648 of the spatial resolutions, the maximum of the digital number (raw OMEGA
 649 signal), phase angle, and the incidence angle. No trends were found for the
 650 majority of the listed properties, but some provided insight into the limita-
 651 tions of our retrieval. We found correlations between high- r_{eff} results and
 652 the ICI, which was expected, and surface elevation, which was coincidental.
 653 The following four properties, which are all related to one another, may con-
 654 tribute to unreliable retrievals: the dust opacity in the cloud-free pixels, the
 655 mean retrieved surface albedo, the difference between the mean albedo and
 656 the mean spectrum level, and the ratio of dust opacities between the cloudy
 657 and cloud-free pixels.

658 Large ICI values infer that cloud cover is thin, which limits the infor-
 659 mation content of the data in our retrieval. As the amount of water-ice in
 660 the atmosphere decreases, so too does our ability to accurately determine its
 661 properties. Therefore, we did find that the occurrence of high- r_{eff} retrievals
 662 tends to increase with the ICI, but the variability of retrieval results for thin
 663 clouds is still very large, and we must understand what other factors are at
 664 play.

665 The appearance of a correlation between high- r_{eff} retrievals and surface
 666 elevation is coincidental. We observed an increase in high- r_{eff} values at el-
 667 evations near zero, but that was only because the majority of observation
 668 occur at low elevations. Observations made at higher elevations tend to be

669 of thicker clouds due to cloud formation processes near the Tharsis volcanoes
670 (e.g., [Pearl et al., 2001](#); [Benson et al., 2003](#); [Smith, 2004](#)), where the majority
671 of our data comes from.

672 The correlation between r_{eff} and the ratio of dust opacities in the cloudy
673 and cloud-free spectra is also coincidental. In both cases, the dust opacity is
674 frequently low and their ratio is close to unity. Thus we see many high- r_{eff}
675 retrievals with this ratio, but we see the same thing for all values of r_{eff} .

676 We searched for aspects of the a priori multi-spectral surface albedo,
677 which we established that our retrievals are critically dependent on, that
678 systematically lead to unrealistic results. We found that as the dust opac-
679 ity of the cloud-free pixels used in the surface albedo retrieval increases, the
680 retrieved albedo values can as well. Indeed, the presence of dust in the atmo-
681 sphere decreases the reflectance factor of bright surfaces and increases that of
682 dark surfaces ([Vincendon et al., 2007](#)). Therefore, when dust is removed from
683 the observed spectra, the brightness factor increases for bright regions and
684 decreases for dark regions. To retrieve albedo, OMEGA observations were
685 chosen to have low dust opacities and to be cloud-free, but dust is always
686 present to a small extent and is generally between 0.04–0.45. Dust opaci-
687 ties used in the albedo retrieval come from the climatology of [Montabone](#)
688 [et al. \(2015\)](#) and tend to be fairly homogeneous over the area covered by an
689 OMEGA observation. There is more variability in the levels of the resulting
690 multi-spectral surface albedo among the pixels in an OMEGA image cube
691 than in the dust opacities used, due to intrinsic variations in the properties
692 of the surface. High levels of dust optical depth do not necessarily imply
693 high values for the retrieved surface albedos.

694 7.1. *High- r_{eff} Case Studies*

695 To further explore these high- r_{eff} results, we looked at several distinct
 696 cases, or groupings, within the data. Two such groups were the set of
 697 OMEGA image cubes with the highest percentage of high- r_{eff} results, and
 698 the sets of retrievals with the highest, and lowest, resulting values of r_{eff} .

699 In the first case we calculated the ratio of retrievals with $r_{\text{eff}} < 7 \mu\text{m}$
 700 to those with $r_{\text{eff}} > 7 \mu\text{m}$, and took the ten OMEGA observations that
 701 had the lowest ratio and that had more than 1000 pixels analyzed (without
 702 applying data quality cuts). These observations are: 0902_5 (MY 27, Ls
 703 95.3°), 0931_5 (MY 27, Ls 98.9°), 0942_5 (MY 27, Ls 100.3°), 0942_6 (MY
 704 27, Ls 100.3°), 1022_6 (MY 27, Ls 110.4°), 1030_7 (MY 27, Ls 111.4°), 1055_6
 705 (MY 27, Ls 114.7°), 3305_3 (MY 28, Ls 89.0°), 3316_3 (MY 28, Ls 90.4°), and
 706 8371_1 (MY 28, Ls 111.7°). These ten observations collectively have 41019
 707 analyzed pixels, but account for about half of the results with $r_{\text{eff}} > 7 \mu\text{m}$
 708 in Fig. 5a. Histograms of the retrieved r_{eff} values for these observations are
 709 shown in Fig. 12a, and they infer that the cloud formations they represent
 710 are predominantly made up of very large water-ice particles. Fig. 12b shows
 711 a histogram of the combined data from Fig. 12a, compared with the entire
 712 data set (same as Fig. 5a).

713 We compared the properties of this group of pixels to the whole data set
 714 and found that most properties, such as ICI, phase angle, incidence angle,
 715 surface temperature, elevation, H₂O band depths, spatial resolutions, etc.,
 716 were average. However, two properties stood out: the dust opacity in the
 717 cloudy observations, and the dust opacity in the cloud-free observation used
 718 to retrieve surface albedo. Fig. 12c shows a histogram of the dust opacities

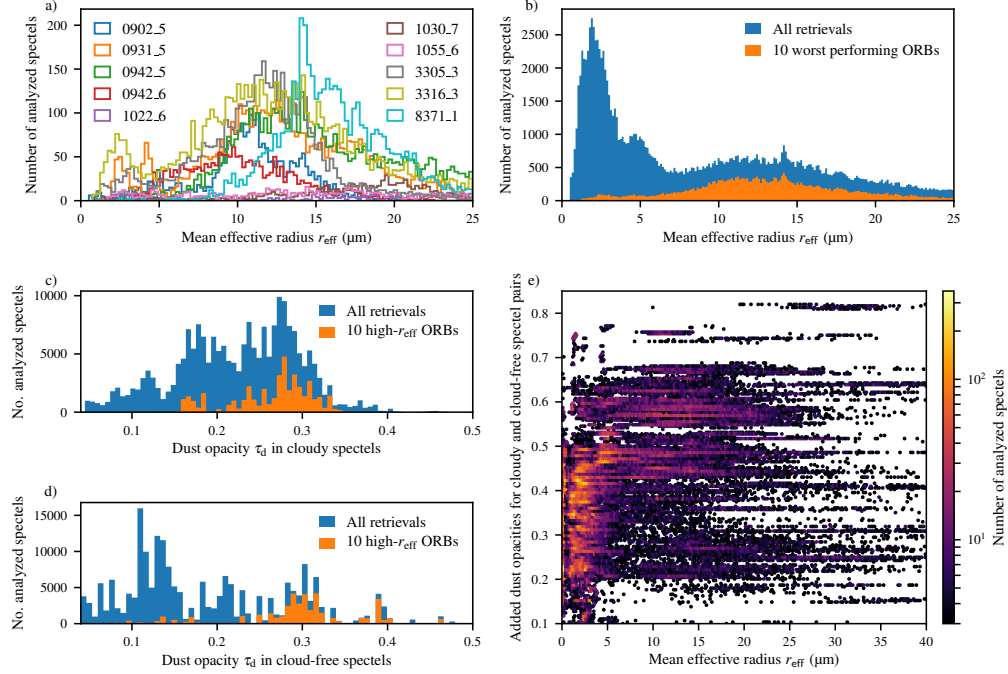


Figure 12: Results from a case study of the ten OMEGA observations returning the highest proportions of high- r_{eff} retrievals. Panel a) shows histograms of retrieved r_{eff} for each of the ten observations, and panel b) compared histograms of r_{eff} after combining all ten observations in panel a) to the entire data set. Panel c) and d) are the same as panel b), but comparing a) the dust opacity used for the cloudy pixels used in the r_{eff} retrievals, and b) the dust opacity in the cloud-free pixels used in the surface albedo retrieval. Panel e) shows the parameter space between the retrieved r_{eff} and the sum of the dust opacities used in the r_{eff} retrieval and the surface albedo retrieval.

719 from the cloud-free pixels used for albedo retrievals used here, and Fig. 12d
 720 shows a histogram of the dust opacities from the cloudy pixels. These fig-
 721 ures also show the contributions from the ten observations identified in this
 722 investigation. The dust present in both the cloudy and cloud-free spectra is
 723 above average

724 What we learned from this analysis was that we can use surface albedos
725 retrieved when the dust opacity is elevated, and we can successfully retrieve
726 water-ice cloud properties when the dust opacity is elevated, but we cannot
727 perform retrievals when the dust opacity is elevated in both the cloudy and
728 cloud-free OMEGA observations. A generalization of this observation is in-
729 ferred from Fig. 12e, which shows the relationship between the retrieved r_{eff}
730 and the sum of the dust opacities used for the albedo and cloud properties
731 retrievals. While high- r_{eff} retrievals may occur in any dust situation, when
732 the sum of the dust opacities is above 0.5, the retrieval favours $r_{\text{eff}} > 5 \mu\text{m}$.

733 In a second case study, we compared groups of OMEGA pixels that were
734 distinguished by the magnitude of their retrieved r_{eff} values. Within each
735 grouping, we compared different properties and calculated the mean spec-
736 trum and mean multi-spectral albedo for hundreds of pixels. In general, we
737 found that the level of the baseline of the spectra tended to be related to
738 high- r_{eff} results. The baseline is affected by the intrinsic surface albedo as
739 well as dust suspended in the atmosphere. If the albedo was measured with a
740 dust-free atmosphere, then dust in the cloudy observation can be decoupled
741 from the water ice signature. However, if dust absorption contaminated the
742 albedo measurement, then dust in the cloudy spectrum is not fully corrected
743 for, resulting in compensation in the r_{eff} and τ_{i} retrieval.

744 Three groups of particular interest were pixels with: $1 \mu\text{m} < r_{\text{eff}} < 3 \mu\text{m}$,
745 representing retrievals with expected values; $16 \mu\text{m} < r_{\text{eff}} < 25 \mu\text{m}$, repre-
746 senting retrievals with un-physically large results; and $r_{\text{eff}} > 40 \mu\text{m}$, repre-
747 senting failed retrievals (see Fig. 5a). A striking feature of the last group is
748 in the ICIs data. It is made up predominantly of pixels with very thin clouds

749 and high-ICI. In this region, the retrieval is unreliable as the water-ice infor-
 750 mation content in the spectra is minimal. By comparing these groups, we
 751 are able to see that there is a strong correlation between the shape of the
 752 spectrum and the result.

753 Fig. 13a shows averaged spectra representing the following criteria: low
 754 r_{eff} , with $2.3\mu\text{m} \leq r_{\text{eff}} \leq 2.8\mu\text{m}$ and $1.1 \leq \tau_i \leq 1.7$; and high r_{eff} , with
 755 $40\mu\text{m} \leq r_{\text{eff}} \leq 48\mu\text{m}$, $0 \leq \tau_i \leq 1$, and $\chi^2 < 2000$. The means of their
 756 corresponding multi-spectral surface albedo are also shown. The difference is
 757 striking between the two groups. Because this large difference is also reflected
 758 in the albedo data, which are taken from different OMEGA observations over
 759 the same location, we believe that these data are not anomalous and that the
 760 spectra reflect aspects of the surface not accounted for by the retrieval. This
 761 could be strong shadowing, or steep surfaces around crater rims (Vincendon
 762 et al., 2007).

763 To further investigate these data, we estimated a baseline for the C-
 764 channel portion of the spectra by computing the average value of the points
 765 used in the retrieval below $2.5\mu\text{m}$. Fig. 13b shows the relationship between
 766 the mean baseline and the retrieved r_{eff} . There are three large clusters: the
 767 data with expected r_{eff} values, and the majority, lies below 7μ ; there is a
 768 large group between $7\text{--}25\mu$, explored above and in Fig. 12; and there is
 769 a group covering the entire r_{eff} range, but with baselines below 0.37. This
 770 group of pixels, with low baseline and high- r_{eff} results, tends to have lower
 771 than average maximum digital number (the raw OMEGA signal strength is
 772 lower than average), and lower than average surface elevation. These pixels
 773 are most likely affected by shadowing on the surface, which our model does

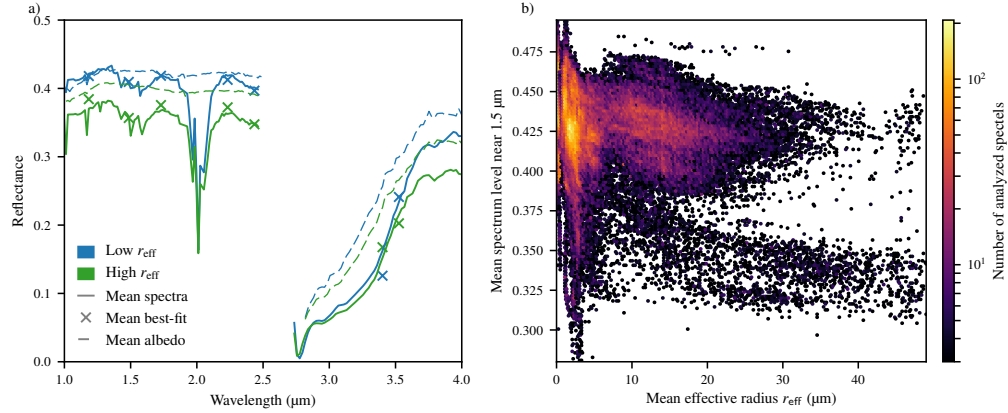


Figure 13: Case study for extremely large retrieved r_{eff} . Panel a) shows the means of 1000 randomly selected spectra that resulted in retrieval results that satisfied: low r_{eff} (blue) $2.3 \mu\text{m} \leq r_{\text{eff}} \leq 2.8 \mu\text{m}$, $1.1 \leq \tau_1 \leq 1.7$; and high r_{eff} (green) $40 \mu\text{m} \leq r_{\text{eff}} \leq 48 \mu\text{m}$, $0 \leq \tau_1 \leq 1$, and $\chi^2 < 2000$. Also shown are the means of the multi-spectral albedo data used in each retrieval, and the mean best-fit results. Panel b) shows the parameter space of the retrieved r_{eff} and the baseline of the C-channel spectra, which is defined as the mean of the spectral points at 1.18, 1.49, 1.73, 2.23, and 2.43 μm .

not accommodate for.

7.2. Mitigation

We made several attempt to alter the retrieval method to try to get more realistic results from all the analyzed pixels. The objective was to affect the the χ^2 minimization for retrievals with low values of τ_1 in order to favour a position in the phase space with equal mass, but lower r_{eff} (e.g., see Fig. 2b). While some methods returned some desirable results, they came at the cost of impacting our already reliable retrievals, made with a robust method, in negative ways. In the absence of a precise method of identifying low-quality results, and the desire to maintain a consistent data product, we have not

784 incorporated any retrievals derived from these modifications into our data
 785 product. Instead, we provide our understanding of the limitation of the
 786 retrieval and the causes of high r_{eff} results, which may be real in some cases.

787 We varied the number of parameters, and the position of the spectral
 788 points we used in the retrieval. These tests resulted in surprising results
 789 that reflected water vapour as well as ice aerosols. We included a third free
 790 parameter, such as the dust opacity, or a multiplicative or additive parameter
 791 to modify the albedo data. The spectra, especially when using a limited
 792 number of spectral points, do not support this many degrees of freedom.
 793 The computation of the χ^2 was modified with a parameter to favour a r_{eff} of
 794 2 or 2.2, but this had a negative impact on retrievals which already returned
 795 $r_{\text{eff}} < 7$. We attempted to remove the signature of CO₂ gas in the spectra
 796 prior to the retrieval, but this resulted in worse fits, higher χ^2 , and more
 797 convergence failures.

798 We also tried to use the ICI to constrain the r_{eff} retrievals by applying the
 799 relationship presented in Section 6.3 to the degenerate data. For retrievals
 800 with a r_{eff} above a threshold (e.g., 7 or 9 μm), we assume that the τ_{i} retrieval
 801 is more accurate than that of r_{eff} . We use the ICI to compute the empirical
 802 column mass, then use equation 1 to compute a new value for r_{eff} from the
 803 retrieved τ_{i} and the estimated M . While this method works and does not
 804 have an impact on nominal r_{eff} retrievals, there is a lack of variability in
 805 the r_{eff} and τ_{i} parameter space after applying such a simple estimation to
 806 constrain r_{eff} . The variability we see in Fig. 9a, for example, reflects physical
 807 aspects of the OMEGA spectra. This method imposes a forced smoothing
 808 on the high- r_{eff} data, making it difficult to re-combine the constrained data

809 with the original data.

810 8. Conclusions

811 In this manuscript we have presented a novel, statistical approach to the
812 retrieval of physical cloud properties from OMEGA spectral images. We
813 have applied that retrieval to a large subset of OMEGA data motivated
814 by the availability of improved prior information, especially maps of multi-
815 spectral surface albedo derived from co-located, cloud-free OMEGA observa-
816 tions. The average water-ice aerosol size in low-latitude cloud formations is
817 $2.2\ \mu\text{m}$ (standard deviation of 0.83). One of our objectives was to examine the
818 spatial distribution of cloud properties in single cloud formations. However,
819 this is challenged by the limitations of the shape of the OMEGA observations
820 (often only 16 or 32 pixels wide) and the size of the subset of OMEGA data
821 analyzed (limited by the lifetime of the OMEGA C-channel detector and the
822 amount of processed albedo data). We also found that applying the retrieval
823 over many pixels of varying cloud thickness reaches the limitations of the
824 technique and the water-ice information content in the spectra.

825 The retrieved parameters, water-ice optical depth, τ_1 , and mean effective
826 radius, r_{eff} , closely conform to an inverse relationship (see equation 1 and
827 Fig. 9a). We found that for optically thin clouds, it is difficult to get an
828 accurate retrieved r_{eff} value, and that there is some degeneracy in the re-
829 trieval parameter space (see Fig. 2b). A large effort was made to understand
830 retrievals with higher than expected r_{eff} . We believe that these results may
831 be real in some cases, but that several aspects may lead to unreliable results,
832 such as the impact of dust in both of the co-located OMEGA observations

833 used to retrieve surface albedo and cloud properties, or OMEGA spectra
834 with uncharacteristically low reflectances.

835 We have found an empirical relationship between the ice cloud index,
836 ICI, and the column mass of water ice, M , or the water ice column, WIC.
837 The temporal and spatial extent of the ICI data is much larger than those
838 of the cloud properties retrievals, and work is being done to study the cli-
839 matology of clouds using the ICI data, and to compare them to other data
840 sets. We have applied our empirical relationship to the entire OMEGA data
841 set and presented a climatology, binned by latitude and solar longitude, of
842 derived WIC from daytime OMEGA observations covering Mars years 26–32
843 (Fig. 11). The primary features seen are the aphelion cloud belt centred on
844 LS 90°, and the polar hoods. These are both observed in other water-ice
845 aerosol climatologies. WIC values range from 1.2–1.6 pr. μm over the aphe-
846 lion belt, and 1.5–2.5 pr. μm over the poles. Future work will expand the
847 retrieved surface albedo and cloud properties data sets, focusing on increas-
848 ing the number of results for pixels with $\text{ICI} < 0.35$ to better constrain our
849 empirical relationship and the derived WIC climatology.

850 The results of this work have been uploaded to the ESA Planetary Science
851 Archive at open.esa.int/esa-planetary-science-archive/ in a format
852 conforming to the Planetary Data System (Version 4).

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 Climate Database.

Audouard, J., Poulet, F., Vincendon, M., Bibring, J.-P., Forget, F.,
 Langevin, Y., Gondet, B., May 2014. Mars surface thermal inertia and
 heterogeneities from OMEGA/MEX. *Icarus* 233, 194–213.

Batson, R. M., Bridges, P. M., Inge, J. L., 1979. Atlas of Mars : the
 1:5,000,000 map series. Scientific and Technical Information Branch,
 NASA, Washington, D.C.

Benson, J. L., Bonev, B. P., James, P. B., Shan, K. J., Cantor, B. A.,
 Caplinger, M. A., Sep. 2003. The seasonal behavior of water ice clouds in
 the Tharsis and Valles Marineris regions of Mars: Mars Orbiter Camera
 Observations. *Icarus* 165, 34–52.

Bibring, J.-P., Soufflot, A., Berthé, M., Langevin, Y., Gondet, B., Drossart,
 P., Bouyé, M., Combes, M., Puget, P., Semery, A., Bellucci, G., Formisano,
 V., Moroz, V., Kottsov, V., Bonello, G., Erard, S., Forni, O., Gendrin, A.,
 Manaud, N., Poulet, F., Poulleau, G., Encrenaz, T., Fouchet, T., Mel-
 chiori, R., Altieri, F., Ignatiev, N., Titov, D., Zasova, L., Coradini, A.,

- 881 Capacionni, F., Cerroni, P., Fonti, S., Mangold, N., Pinet, P., Schmitt, B.,
882 Sotin, C., Hauber, E., Hoffmann, H., Jaumann, R., Keller, U., Arvidson,
883 R., Mustard, J., Forget, F., Aug. 2004. OMEGA: Observatoire pour la
884 Minéralogie, l'Eau, les Glaces et l'Activité. In: Wilson, A., Chicarro, A.
885 (Eds.), Mars Express: the Scientific Payload. Vol. 1240 of ESA Special
886 Publication. pp. 37–49.
- 887 Bohren, C. F., Huffman, D. R., Mar. 1983. Absorption and Scattering of Light
888 by Small Particles. Wiley-VCH Verlag GmbH & Co. KGaA, Weinheim,
889 Germany.
- 890 Christensen, P. R., Zurek, R. W., Jun. 1984. Martian north polar hazes
891 and surface ice - Results from the Viking Survey/Completion mission. J.
892 Geophys. Res. 89, 4587–4596.
- 893 Clancy, R. T., Lee, S. W., Gladstone, G. R., McMillan, W. W., Rousch, T.,
894 Mar. 1995. A new model for Mars atmospheric dust based upon analysis
895 of ultraviolet through infrared observations from Mariner 9, Viking, and
896 PHOBOS. J. Geophys. Res. 100, 5251–5263.
- 897 Clancy, R. T., Wolff, M. J., Christensen, P. R., Sep. 2003. Mars aerosol
898 studies with the MGS TES emission phase function observations: Opti-
899 cal depths, particle sizes, and ice cloud types versus latitude and solar
900 longitude. J. Geophys. Res. 108 (E9), 5098.
- 901 Curran, R. J., Conrath, B. J., Hanel, R. A., Kunde, V. G., Pearl, J. C., Oct.
902 1973. Mars: Mariner 9 Spectroscopic Evidence for H₂O Ice Clouds. Science
903 182, 381–383.

904 Fedorova, A. A., Montmessin, F., Rodin, A. V., Korablev, O. I., Määttänen,
905 A., Maltagliati, L., Bertaux, J.-L., Mar. 2014. Evidence for a bimodal
906 size distribution for the suspended aerosol particles on Mars. *Icarus* 231,
907 239–260.

908 Forget, F., Hourdin, F., Fournier, R., Hourdin, C., Talagrand, O., Collins,
909 M., Lewis, S. R., Read, P. L., Huot, J.-P., Oct. 1999. Improved general
910 circulation models of the Martian atmosphere from the surface to above
911 80 km. *J. Geophys. Res.* 104, 24155–24176.

912 Geminale, A., Grassi, D., Altieri, F., Serventi, G., Carli, C., Carrozzo, F. G.,
913 Sgavetti, M., Orosei, R., D’Aversa, E., Bellucci, G., Frigeri, A., Jun. 2015.
914 Removal of atmospheric features in near infrared spectra by means of prin-
915 cipal component analysis and target transformation on Mars: I. Method.
916 *Icarus* 253, 51–65.

917 Glenar, D. A., Samuelson, R. E., Pearl, J. C., Bjoraker, G. L., Blaney, D.,
918 Feb. 2003. Spectral imaging of martian water ice clouds and their diurnal
919 behavior during the 1999 aphelion season ($L_s = 130^\circ$). *Icarus* 161, 297–318.

920 Guzewich, S. D., Smith, M. D., Wolff, M. J., Dec. 2014. The vertical distri-
921 bution of Martian aerosol particle size. *J. Geophys. Res.* 119, 2694–2708.

922 Haberle, R. M., Clancy, R. T., Forget, F., Smith, M. D., Zurek, R. W., Mar.
923 2017. *The atmosphere and climate of Mars*. Cambridge Planetary Science.
924 Cambridge University Press.

925 Ignatiev, N. I., Grassi, D., Zasova, L. V., Aug. 2005. Planetary Fourier spec-

926 trometer data analysis: Fast radiative transfer models. *Planet. Space Sci.*
927 53, 1035–1042.

928 James, P. B., Bell, J. F., Clancy, R. T., Lee, S. W., Martin, L. J., Wolff,
929 M. J., Aug. 1996. Global imaging of Mars by Hubble space telescope during
930 the 1995 opposition. *J. Geophys. Res.* 101, 18883–18890.

931 Langevin, Y., Bibring, J.-P., Montmessin, F., Forget, F., Vincendon, M.,
932 Douté, S., Poulet, F., Gondet, B., Jul. 2007. Observations of the south
933 seasonal cap of Mars during recession in 2004-2006 by the OMEGA
934 visible/near-infrared imaging spectrometer on board Mars Express. *J. Geo-*
935 *phys. Res.* 112, E08S12.

936 Madeleine, J.-B., Forget, F., Millour, E., Montabone, L., Wolff, M. J., Nov.
937 2011. Revisiting the radiative impact of dust on Mars using the LMD
938 Global Climate Model. *J. Geophys. Res.* 116 (E15), E11010.

939 Madeleine, J.-B., Forget, F., Spiga, A., Wolff, M. J., Montmessin, F., Vincen-
940 don, M., Jouglet, D., Gondet, B., Bibring, J.-P., Langevin, Y., Schmitt, B.,
941 May 2012. Aphelion water-ice cloud mapping and property retrieval using
942 the OMEGA imaging spectrometer onboard Mars Express. *J. Geophys.*
943 *Res.* 117, E00J07.

944 Mateshvili, N., Fussen, D., Vanhellemont, F., Bingen, C., Dekemper, E.,
945 Loodts, N., Tetard, C., Jul. 2009. Water ice clouds in the Martian atmo-
946 sphere: Two Martian years of SPICAM nadir UV measurements. *Planet.*
947 *Space Sci.* 57, 1022–1031.

948 Mateshvili, N., Fussen, D., Vanhellemont, F., Bingen, C., Dodion, J.,
949 Montmessin, F., Perrier, S., Dimarellis, E., Bertaux, J.-L., Jul. 2007. Mar-
950 tian ice cloud distribution obtained from SPICAM nadir UV measure-
951 ments. *J. Geophys. Res.* 112, E07004.

952 Millour, E., Forget, F., Spiga, A., Navarro, T., Madeleine, J.-B., Montabone,
953 L., Pottier, A., Lefevre, F., Montmessin, F., Chaufray, J.-Y., Lopez-
954 Valverde, M. A., Gonzalez-Galindo, F., Lewis, S. R., Read, P. L., Huot,
955 J.-P., Desjean, M.-C., MCD/GCM development Team, Oct. 2015. The
956 Mars Climate Database (MCD version 5.2). European Planetary Science
957 Congress 10, EPSC2015–438.

958 Mishchenko, M. I., Travis, L. D., Mackowski, D. W., May 1996. T-matrix
959 computations of light scattering by nonspherical particles: a review. *J.*
960 *Quant. Spectrosc. Radiat. Transfer* 55, 535–575.

961 Montabone, L., Forget, F., Millour, E., Wilson, R. J., Lewis, S. R., Cantor,
962 B., Kass, D., Kleinböhl, A., Lemmon, M. T., Smith, M. D., Wolff, M. J.,
963 May 2015. Eight-year climatology of dust optical depth on Mars. *Icarus*
964 251, 65–95.

965 Navarro, T., Madeleine, J.-B., Forget, F., Spiga, A., Millour, E., Montmessin,
966 F., Määttänen, A., Jul. 2014. Global climate modeling of the Martian water
967 cycle with improved microphysics and radiatively active water ice clouds.
968 *J. Geophys. Res.* 119, 1479–1495.

969 Parker, D. C., Beish, J. D., Troiani, D. M., Joyce, D. P., Hernandez, C. E.,

970 Mar. 1999. Telescopic Observations of Mars, 1996-1997: Results of the
971 Marswatch Program. *Icarus* 138, 3–19.

972 Pearl, J. C., Smith, M. D., Conrath, B. J., Bandfield, J. L., Christensen,
973 P. R., Jun. 2001. Observations of Martian ice clouds by the Mars Global
974 Surveyor Thermal Emission Spectrometer: The first Martian year. *J. Geo-*
975 *phys. Res.* 106, 12325–12338.

976 Petrova, E., Keller, H. U., Markiewicz, W. J., Thomas, N., Wuttke, M. W.,
977 Oct. 1996. Ice hazes and clouds in the Martian atmosphere as derived from
978 the Phobos/KRFM data. *Planet. Space Sci.* 44, 1163–1176.

979 Pollack, J. B., Ockert-Bell, M. E., Shepard, M. K., Mar. 1995. Viking Lander
980 image analysis of Martian atmospheric dust. *J. Geophys. Res.* 100, 5235–
981 5250.

982 Rannou, P., Perrier, S., Bertaux, J.-L., Montmessin, F., Korablev, O.,
983 Rébérac, A., Sep. 2006. Dust and cloud detection at the Mars limb with
984 UV scattered sunlight with SPICAM. *J. Geophys. Res.* 111, E09S10.

985 Smith, M. D., Jan. 2004. Interannual variability in TES atmospheric obser-
986 vations of Mars during 1999-2003. *Icarus* 167, 148–165.

987 Smith, M. D., Aug. 2009. THEMIS observations of Mars aerosol optical depth
988 from 2002-2008. *Icarus* 202, 444–452.

989 Smith, M. D., Wolff, M. J., Clancy, R. T., Kleinböhl, A., Murchie, S. L.,
990 Feb. 2013. Vertical distribution of dust and water ice aerosols from CRISM
991 limb-geometry observations. *J. Geophys. Res.* 118, 321–334.

992 Smith, P. H., Lemmon, M., Apr. 1999. Opacity of the Martian atmosphere
 993 measured by the Imager for Mars Pathfinder. *J. Geophys. Res.* 104, 8975–
 994 8986.

995 Stamnes, K., Tsay, S.-C., Jayaweera, K., Wiscombe, W., Jun. 1988. Numer-
 996 ically stable algorithm for discrete-ordinate-method radiative transfer in
 997 multiple scattering and emitting layered media. *Appl. Opt.* 27, 2502–2509.

998 Szantai, A., Audouard, J., Forget, F., Madeleine, J.-B., Pottier, A., Millour,
 999 E., Gondet, B., Langevin, Y., Bibring, J.-P., Jan. 2017a. Construction of a
 1000 4D Water Ice Cloud Database from Mars Express / OMEGA Observations
 1001 - Derivation of the Diurnal Martian Cloud Life Cycle. In: Forget, F.,
 1002 Millour, E. (Eds.), *The Mars Atmosphere: Modelling and observation*. p.
 1003 3208.

1004 Szantai, A., Audouard, J., Forget, F., Madeleine, J.-B., Pottier, A., Millour,
 1005 E., Gondet, B., Langevin, Y., Bibring, J.-P., Sep. 2017b. Martian cloud
 1006 coverage and diurnal cloud life cycle derived from gridded Mars/Express
 1007 OMEGA data. In: *European Planetary Science Congress*. Vol. 11. pp.
 1008 EPSC2017–879.

1009 Szantai, A., Audouard, J., Forget, F., Olsen, K. S., Millour, E., Madeleine,
 1010 J.-B., Pottier, A., Gondet, B., Langevin, Y., Bibring, J.-P., 2019. Martian
 1011 cloud climatology and life cycle extracted from Mars Express OMEGA
 1012 spectral images. *J. Quant. Spectrosc. Radiat. Transfer*, this issue.

1013 Tamppari, L. K., Zurek, R. W., Paige, D. A., Jul. 2003. Viking-era diurnal
 1014 water-ice clouds. *J. Geophys. Res.* 108, 9–1.

- 1015 Toon, O. B., Pollack, J. B., Sagan, C., Apr. 1977. Physical properties of
1016 the particles composing the Martian dust storm of 1971-1972. *Icarus* 30,
1017 663–696.
- 1018 Vincendon, M., Langevin, Y., Poulet, F., Bibring, J.-P., Gondet, B., Jul.
1019 2007. Recovery of surface reflectance spectra and evaluation of the optical
1020 depth of aerosols in the near-IR using a Monte Carlo approach: Application
1021 to the OMEGA observations of high-latitude regions of Mars. *J. Geophys.*
1022 *Res.* 112, E08S13.
- 1023 Warren, S. G., Brandt, R. E., Jul. 2008. Optical constants of ice from the
1024 ultraviolet to the microwave: A revised compilation. *J. Geophys. Res.*
1025 113 (D12), D14220.
- 1026 Whiteway, J. A., Komguem, L., Dickinson, C., Cook, C., Illnicki, M.,
1027 Seabrook, J., Popovici, V., Duck, T. J., Davy, R., Taylor, P. A., Pathak,
1028 J., Fisher, D., Carswell, A. I., Daly, M., Hipkin, V., Zent, A. P., Hecht,
1029 M. H., Wood, S. E., Tamppari, L. K., Renno, N., Moores, J. E., Lemmon,
1030 M. T., Daerden, F., Smith, P. H., Jul. 2009. Mars Water-Ice Clouds and
1031 Precipitation. *Science* 325, 68.
- 1032 Willame, Y., Vandaele, A. C., Depiesse, C., Lefèvre, F., Letocart, V.,
1033 Gillotay, D., Montmessin, F., Aug. 2017. Retrieving cloud, dust and ozone
1034 abundances in the Martian atmosphere using SPICAM/UV nadir spectra.
1035 *Planet. Space Sci.* 142, 9–25.
- 1036 Wolff, M. J., Clancy, R. T., Sep. 2003. Constraints on the size of Martian

1037 aerosols from Thermal Emission Spectrometer observations. J. Geophys.
 1038 Res. 108 (E9), 5097.

1039 Wolff, M. J., Smith, M. D., Clancy, R. T., Arvidson, R., Kahre, M., Seelos,
 1040 F., Murchie, S., Savijärvi, H., Jun. 2009. Wavelength dependence of dust
 1041 aerosol single scattering albedo as observed by the Compact Reconnaissance
 1042 Imaging Spectrometer. J. Geophys. Res. 114, E00D04.

1043 Zasova, L., Grassi, D., Formisano, V., Maturilli, A., Aug. 2001. The Martian
 1044 atmosphere in the region of the great volcanoes: Mariner 9 IRIS data
 1045 revisited. Planet. Space Sci. 49, 977–992.

1046 Zasova, L. V., Formisano, V., Moroz, V. I., Bibring, J.-P., Grassi, D., Ig-
 1047 natiev, N. I., Giuranna, M., Bellucci, G., Altieri, F., Blecka, M., Gnedykh,
 1048 V. N., Grigoriev, A. V., Lellouch, E., Mattana, A., Maturilli, A., Moshkin,
 1049 B. E., Nikolsky, Y. V., Patsaev, D. V., Piccioni, G., Ratai, M., Saggin, B.,
 1050 Fonti, S., Khatuntsev, I. V., Hirsh, H., Ekonomov, A. P., Jul. 2006. Results
 1051 of measurements with the Planetary Fourier Spectrometer onboard Mars
 1052 Express: Clouds and dust at the end of southern summer. A comparison
 1053 with OMEGA images. Cosmic Research 44, 305–316.