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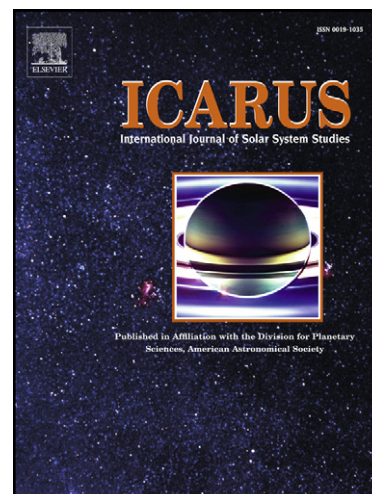
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1 **The 3-5 MHz global reflectivity map of Mars by**  
2 **MARSIS/Mars Express: implications for the current**  
3 **inventory of subsurface H<sub>2</sub>O.**

4

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7

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12

13

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17 10 figures

18

19 **Abstract**

20

21 We extracted the surface echo power from two years of MARSIS  
22 measurements. The retrieved values are calibrated to compensate for changes in the  
23 distance of the spacecraft to the surface and for the attenuation of the signal by the  
24 ionosphere. The results are used to build the first global map of surface echo power at  
25 3-5 MHz. The surface echo power variations are primarily caused by km-scale surface  
26 roughness. Then, we derive the values of dielectric constant of the shallow subsurface  
27 materials by normalizing the surface echo power map using a simulation of MARSIS  
28 signal from the MOLA topography. As a result, we obtain a map that characterizes the  
29 dielectric properties of the materials down to a few decameters below the surface.  
30 Dielectric properties vary with latitude, with high values in mid-latitudes belts (20-  
31 40°) and lower values at both equatorial and high latitudes. From the comparison of  
32 MARSIS reflectivity map to GRS observations, we conclude that the reflectivity  
33 decrease observed poleward of 50-60° corresponds to the onset of water-ice  
34 occurrence within the regolith. Assuming homogenous ground composition and  
35 texture at the scale of the MARSIS resolution cell, our inferred volume of ground  
36 water ice is of  $10^6$  km<sup>3</sup>, equivalent to a polar cap. Low reflectivity areas are also  
37 observed in equatorial regions. From radar studies alone, equatorial low dielectric  
38 constant values could have different interpretations but the correlation with GRS  
39 hydrogen distribution rather points toward a water related explanation.

40

## 41 **1. Introduction**

42           The Martian surface has been scrutinized for decades by a variety of imaging  
43 and spectroscopic techniques, sensitive to the properties of the first micrometers to  
44 millimeters of the surface. Meanwhile, subsurface investigations remained limited to  
45 indirect studies until the arrival of the Mars Odyssey mission in 2001. Data from the  
46 Gamma Ray Spectrometer (GRS) were used to map the average chemical  
47 composition of the first meter below the surface (e.g. Boynton et al., 2007). The  
48 spatial distribution of hydrogen (inferred to be present in the form of ground ice) in  
49 both hemispheres is one of the major results of this instrument (Boynton et al., 2002;  
50 Feldman et al., 2002; Mitrofanov et al., 2002). In order to infer and map the properties  
51 of the Martian regolith and crust below this depth of 1 m, “ground penetrating”  
52 geophysical techniques are required.

53           Whereas low frequency radar has been used in the past to probe the subsurface  
54 of Mars using terrestrial ground based instruments, the Mars Advanced Radar for  
55 Subsurface and Ionospheric Sounding instrument (MARSIS) on board Mars Express  
56 (Picardi et al, 2005) was the first radar sounder to operate from an orbiting probe  
57 around Mars. It has then been joined by the Shallow Subsurface Radar (SHARAD) on  
58 board Mars Reconnaissance Orbiter (Seu et al., 2007). MARSIS is a decameter radar  
59 sounder that operates in 1 MHz-wide frequency bands centered at 1.8, 3, 4 and 5  
60 MHz, whereas SHARAD operates at higher frequency (20 MHz) and with a larger  
61 bandwidth (10 MHz), which leads to smaller penetration depth but higher vertical  
62 resolution. These two instruments measure 2D “radargrams” that represent cross-  
63 sections of the Martian subsurface displaying discrete or continuous reflections of the  
64 radar waves at interfaces between materials of different dielectric constant (e.g. Plaut

65 et al. 2007; Watters et al., 2007; Grima et al. 2009). Penetration depth of MARSIS can  
66 reach 4 kilometers in clear water ice (Plaut et al. 2007), and this instrument was able  
67 to map the bedrock/ice interface below Mars polar caps.

68 Here, instead of looking at individual radargrams, we build a global map of  
69 Mars by extracting the surface echo power from each frame (pulse) of each  
70 radargram. This so-called “reflectivity map” gives important information on the  
71 composition and physical properties of the upper part of the Martian crust at a global  
72 scale. Radar reflectivity maps of the Moon at different frequencies have already been  
73 measured in the past from ground-based instruments and used to infer important  
74 properties of the Lunar terrains (Evans 1962; Evans and Pettengill 1963). In  
75 particular, Campbell and Hawke, 2005 show that reflectivity values measured at the  
76 wavelength:  $\lambda = 70$  cm can be influenced by the composition of terrains buried more  
77 than 50 m deep in some cases, demonstrating the ability of this method to probe at  
78 large depth. In the case of Mars, some measurements had also been obtained using  
79 ground based radio telescopes at  $\lambda = 3.5$  to 70 cm (Simpson et al., 1992; Harmon et  
80 al., 1999) or spacecraft as Mars-3 and Mars-4 spacecraft (Krupenio et al. 1977) and  
81 Viking orbiter 2 at  $\lambda = 13.1$  cm (Simpson et al., 1979) but were rather limited in terms  
82 of spatial resolution and / or geographic extent. The reflectivity values obtained from  
83 these measurements have been used to estimate the dielectric constant of surface  
84 materials (Pettengill et al., 1973; Downs et al., 1973, 1975; Simpson et al., 1982).  
85 Spatial variability of the value of the dielectric constant has been interpreted in terms  
86 of variations of bulk density (Krupenio et al. 1977) and/or compositional variations  
87 (Campbell and Ulrich, 1969) of subsurface materials.

88 In this article, we present the method used to extract the surface echo power,

89 and the corrections required to build the global dielectric map from MARSIS  
90 measurements. We show that the obtained map provides unique information on the  
91 nature of the surface geological material, and more generally on Mars geology and  
92 climatology.

93

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## 94 2. Methods

95

### 96 2.1. Extraction of the surface echoes

97

98 In a first step, the surface echo power values are extracted from the data.

99 MARSIS radargrams are composed of about a thousand frames, with each frame

100 corresponding to a vertical sounding of the Martian surface (figure 1). Radar echoes

101 appear when the transmitted waves meet abrupt changes in dielectric constant. The

102 first echo in the radargram corresponds mostly to the surface echo, because lateral

103 echoes (so called clutter) and subsurface echoes arrive later due to a longer distance

104 between reflectors and spacecraft. Furthermore, the intensity of the surface echo is

105 generally much higher than that of clutter or subsurface echoes. Using these

106 characteristics, we assume that the first echo returned corresponds to surface

107 reflection and we define a selection criteria  $C$  to localize the surface echo expressed

108 as:

$$109 \quad C(i) = \frac{|S(i)|^2}{\text{mean}(|S(i-1:i-30)|^2)} \quad (1)$$

110 where  $S$  is the MARSIS signal in a given frame (512 samples) and  $i$  is the temporal

111 index inside the frame. Equation 1 computes the contrast between a bin and the signal

112 average over the 30-bin window preceding it.

113  $C$  is maximum when  $|S(i)|^2$  is maximum and  $\text{mean}(|S(i-1:i-30)|^2)$ , equal to

114 the power level of the noise galactic and thermal), is minimum. This condition is only

115 satisfied for the surface echo, i.e. the only echo that can precede the surface echo is a

116 noise. An example of detection is given in figure 2 for orbit 2787. Once the surface



117 signal has been identified, the amplitude of the surface echo can easily be extracted as  
118 shown on figure 1.

119 We have systematically applied this process to the MARSIS data, between  
120 orbit 2300 and orbit 5200. Respectively about 0.6, 1.3 and 0.8 millions of  
121 measurements were extracted from the frequency bands centered at 3, 4 and 5 MHz.  
122 Band 1, centered at 1.8 MHz, is not used in this study because the number of  
123 measurements was too small.

124

## 125 **2.2. Estimation of surface echo power**

126

127 The surface echo power largely depends on the attenuation of the  
128 electromagnetic waves between the spacecraft and the surface, which is mainly due to  
129 range attenuation and ionospheric absorption. In the following sections, we describe  
130 the method used to correct these effects.

131

### 132 **2.2.1. Range attenuation**

133

134 Intensity of the surface echoes can be estimated by using a simplified radar  
135 equation applying Snell's law for a vertically incident electromagnetic wave on a flat  
136 surface. When we assume altitude of the satellite as R, the peak power of surface echo  
137 can be estimated as:

$$138 \quad P_r = \frac{P_t G^2 \lambda^2}{4(4\pi R)^2} r_{0,1} \quad (2)$$

139 where  $P_t$ ,  $G$ ,  $\lambda$  and  $r_{0,1}$  are the peak power of transmitted RF pulses, the antenna

140 gain, the wavelength and the reflectivity of the surface, respectively. Thus, the  
 141 intensity of a surface echo is mainly dependent on the propagation range  $R$ , the  
 142 observation wavelength  $\lambda = 2\pi c / \omega$  and antenna gain  $G$  ( $G=1.64$  for a matched dipole  
 143 antenna condition).

144 The received power decreases with  $R^2$  (equation 2). The Mars Express orbit is  
 145 elliptical and the spacecraft altitude during MARSIS passes varies between about 250  
 146 and 1000 km. It is therefore necessary to compensate for the power losses due to  
 147 altitude change, and we simply normalize the surface echo power by the squared  
 148 altitude of the spacecraft.

149

### 150 **2.2.2. Ionospheric absorption**

151

152 As described by Safaeinili et al. (2003), the plasma layer (ionosphere)  
 153 attenuates the MARSIS radar waves. The attenuation of electromagnetic waves per  
 154 meter of ionospheric plasma is given by:

$$155 \quad A = 4.61 \times 10^{-5} \frac{n_e(z)\nu(z)}{\omega^2 + \nu(z)^2} \quad (3)$$

156 where  $n_e$  is the electron density ( $\text{m}^{-3}$ ) at altitude  $z$  (m),  $\nu$  the electron collision  
 157 frequency ( $\text{rad s}^{-1}$ ) and  $\omega$  is the pulsation of the radar wave (radian frequency).

158 For MARSIS data, we have a two-way ionospheric propagation and the total  
 159 attenuation of the radar waves therefore becomes:

$$160 \quad \int Adz = 2 \times 4.61 \times 10^{-5} \int \frac{n_e(z)\nu(z)}{\omega^2 + \nu(z)^2} dz \quad (4)$$

161 The main parameters controlling absorption (equation 4) are the electron

162 density and collision frequency profiles that depend on the neutral density. As these  
163 profiles are related to the ionization due to the solar EUV (extreme ultra violet)  
164 radiation, the absorption changes on Mars have a first order dependence on solar  
165 zenith angle. Safaeinili et al. (2003) have described in detail this phenomenon for  
166 different states (day/night, etc...) of the Martian ionosphere.

167 An accurate estimation of absorption requires knowledge of the exact state of  
168 the ionosphere and the neutral atmosphere (i.e. the density and collision profiles) for  
169 each frame. These profiles are not measured and therefore we cannot precisely  
170 calculate absorption. However, the correction of the dispersion effect (variation of  
171 wave velocity with frequency) due to the ionosphere (Safaeinili et al. 2007, Mougnot  
172 et al. 2008) provides a value of the total electron content (TEC) and, as TEC is the  
173 integral of the electron density, this value is a good proxy to estimate absorption.

174 Figure 3a shows that the surface echo power on MARSIS data decreases as  
175 TEC increases. In this figure, we used the entire set of signal measurements for the 4  
176 MHz band. For a given value of TEC, the surface echo power varies over a 10dB  
177 range due to the variable reflectivity of the different Martian terrains. This trend is  
178 consistent with equation 4, which shows that electron density enhances absorption.

179 The signal decreases until a threshold is reached at about 30 dB below the  
180 maximum power. This threshold corresponds to the MARSIS noise level. Table 1  
181 summarizes, for each frequency band, the values of SZA (Solar Zenith Angle) or TEC  
182 at which the signal becomes lost in the noise. We have chosen to remove the data  
183 outside the limits defined in Table 1. Furthermore, we did not use observations that  
184 show a very low signal power. In these cases, very high attenuation is probably due to  
185 the increased electron density created by precipitation and radiation during solar flares

186 (Espley et al. 2007).

187 The next step consists in evaluating the behavior of the surface echo power as  
188 a function of the TEC values. We compute the mean surface echo power for the  
189 overall data set with a bin for TEC of  $2 \times 10^{13}$  electrons per square meter. The result is  
190 presented in figure 3b. We use this curve of mean surface echo power as function of  
191 TEC to normalize the data to compensate for absorption. The normalized data plotted  
192 in figure 4 are constant on average as function of TEC.

193

### 194 **2.3 Surface echo power**

195

196 After extracting the surface echo power from the echo histories and correcting  
197 them for range dependence and ionospheric absorption, we plot the global map of the  
198 surface echo power in figure 5a. We averaged the data from multiple measurements  
199 with a bin size of 0.5 degrees. MARSIS is a nadir-looking radar and the Mars Express  
200 polar orbit does not allow us to sound the surface beyond about  $87^\circ\text{N}$  and  $87^\circ\text{S}$ . The  
201 surface echo power values for the different frequency bands are very similar and have  
202 thus been combined on this map to provide better coverage of the Martian surface.  
203 However for local or regional studies, it could also be useful in the future to compare  
204 the reflectivity measured in several frequency bands to study materials and/or  
205 structures that could change the surface reflectivity as a function of frequency (e.g.,  
206 Mougnot et al. 2008).

207 Several parameters might affect the surface echo power: surface roughness,  
208 slope distribution and the dielectric constant of the surface materials. Most  
209 backscattering models separate the effect of the dielectric constant (material

210 chemistry) from those of topography (roughness, slopes) (Ulaby et al. 1986; Picardi et  
211 al. 2004). The reflectivity of the surface  $r_{0,1}$  (equation 2) can be expressed by the  
212 equation:

$$213 \quad r_{0,1} = \Gamma_s(\epsilon) f_s(rms_s, \lambda) \quad (5)$$

214 where  $\Gamma_s$  is the Fresnel reflectivity terms and  $f_s$  is the backscattering term (geometric  
215 scattering). The function  $f_s$  is only dependent on the surface topography (i.e. the  
216 roughness and the slopes).

217         The surface roughness at MARSIS wavelength is due to slight variations of  
218 the surface heights over horizontal scale of the tens to hundred meters (figure 5c). The  
219 slope distribution effect is due to surface altitude variation on a scale of a few hundred  
220 to thousands meters. The surface echo is made from nadir return, which is a coherent  
221 specular reflection of the emitted wave at the surface interface. This nadir echo is  
222 mainly reflected on the first Fresnel zone, which has a diameter varying between 5  
223 and 16 km (depending on the spacecraft altitude and emitted wavelength). The slope  
224 distribution inside the first Fresnel zone results in a scattering of the signal to off-  
225 nadir directions, which is related to a decrease of surface reflectivity.

226         As expected for a decametric radar waves, we observe in figure 5 that the  
227 roughness at kilometer-scale (Kreslavsky and Head, 2000) is highly correlated to the  
228 surface reflectivity due to slope distribution effect. The roughest terrains on Mars,  
229 such as the Olympus Mons aureole, Valles Marineris, the Argyre crater rim display  
230 very low surface echo power. Highlands Plateau, which is a heavily cratered region in  
231 the southern hemisphere, presents globally a lower reflectivity compared to the  
232 smoother northern plains (Vastitas Borealis). The smoothest terrains as the volcanic  
233 plateau in Tharsis region or Amazonis Planitia are characterized by a very high  
234 reflectivity.

235           Only few localized terrains that are smooth at kilometer-scale present low  
236 reflectivity, such as the linear dune regions around the north polar cap (mostly in  
237 Olympia Planitia), which are very rough at small scales and smoother at large scales  
238 (Kreslavsky and Head, 2000). An image (figure 6) provided by the High Resolution  
239 Imaging Science Experiment (HiRISE) on board Mars Reconnaissance Orbiter  
240 (MRO) illustrates this small-scale roughness, showing dunes spaced in the range of  
241 50-100 m, which corresponds exactly to the range of MARSIS wavelength. In this  
242 particular case, the coherent part of the signal vanishes and only the incoherent part  
243 (clutter) remains, as observed on the radargrams of orbit #3674 between frames 80  
244 and 130 (figure 6). This region is typically an area where the roughness effect on the  
245 signal is really strong compared to the effect of slope distribution.

246           However, only few regions are really rough at 10-100 meters scale and it  
247 seems that the main effect on surface reflectivity is due to slope distribution  
248 (kilometer-scale roughness). In addition, due to the lack of a global DEM at the  
249 required spatial resolution, we cannot model the effect of small scale (< 100 m)  
250 roughness. Thus, we have decided to neglect the effect of this roughness and to only  
251 take into account the larger-scale slope distribution. Locally, this assumption could be  
252 wrong (i.e. dune fields shown in figure 6). In such a case, we expect errors on the  
253 retrieved reflectivity values up to 3 dB (Campbell and Shepard, 2003). However,  
254 comparisons of the final dielectric map (figures 7 and 8) with the MOLA pulse width  
255 map (Neumann et al., 2003), which describes roughness at 1-20 m and SHARAD  
256 reflectivity map (unpublished data) which is sensitive to 1-15 m roughness do not  
257 show any systematic correlation at global scale. Therefore, we do not think that  
258 neglecting the < 100 roughness has any major effect for this global scale study.

259

## 260 **2.4 Simulation of MARSIS data**

261

262 In order to obtain maps showing the dielectric properties of surface materials,  
263 we have to correct for the effects of relief on the signal and then calibrate the resulting  
264 reflectivity. A method for simulating MARSIS echo histories due to local topography  
265 has been developed to help in the interpretation of the observational data. For this  
266 simulation, the surface is modeled using the facet method (Nouvel 2002; Nouvel et al.  
267 2004), which is an extension of the Kirchhoff model. Such a model can be used  
268 because of the low surface roughness at radar wavelengths and offers a significant  
269 gain in calculation time. The synthetic faceted surface is generated from MOLA data  
270 (Smith et al. 2001).

271 All MARSIS orbits have been simulated and we have extracted the surface  
272 echo power in the same way as for actual MARSIS data. The dielectric constant of the  
273 surface is kept constant in the simulation. Thus, the simulation makes possible the  
274 estimation of the contribution of large-scale scattering resulting from slope  
275 distribution. We can now consider the simulation as a reference for the surface echo  
276 power in order to correct the roughness and topographic effects.

277 The result of all the simulations is shown in figure 5b as a global map of the  
278 expected surface echo power of the Martian surface. Note that the Olympia Planitia  
279 region is characterized by a high-simulated surface echo power compared to the real  
280 power (see figure 5a). As the facet size in the simulation is about 460 m (due to the  
281 limited MOLA resolution), the topography of the dunes is not captured by the digital  
282 elevation model. This shows the limit of the simulation that is unable to simulate the  
283 small-scale scattering resulting from surface roughness over a horizontal scale of tens  
284 of meters or less.

285 The generation of a simulated map makes it possible to normalize the surface  
 286 echo power extracted from the MARSIS data. Finally we estimate the Fresnel  
 287 reflectivity  $\Gamma_s(\varepsilon)$  by dividing the MARSIS data by the simulated data (see equation 5).  
 288 The result of this correction is shown in figure 7.

289 The Fresnel reflection coefficient  $\Gamma_s(\varepsilon)$  at normal incidence at the plane interface  
 290 between two media with refractive indexes  $n_i$  and  $n_j$  respectively is defined as  
 291 followed:

$$292 \quad \Gamma = \frac{n_i - n_j}{n_i + n_j} \quad (6)$$

293 The relation between the dielectric constant  $\varepsilon$  and refractive index  $n$  is  $n = \sqrt{\varepsilon}$ . The  
 294 reflectivity of an interface is given by  $R = |\Gamma|^2$ . For the surface interface, the equation  
 295 6 can be reduced to:

$$296 \quad \Gamma = \frac{1 - n_j}{1 + n_j} \quad (7)$$

297 where  $n_j$  is the refractive index of the surface materials and  $n_i$ , the refractive index of  
 298 the atmosphere taken as 1.

299 Using equation 7, we easily convert  $\Gamma_s(\varepsilon)$  into dielectric constant.

300 As the received power is still not calibrated in an absolute way, we use the regions of  
 301 the North and South Polar Layered Deposits (NPLD and SPLD) as a reference. We  
 302 adjusted the result in figure 7 by multiplying the whole map by a constant. This  
 303 constant is evaluated such that the estimated water ice reflectivity values in polar  
 304 layered deposit is matched (Plaut et al. 2007, Grima et al. 2009). We use:  $\varepsilon' = 3.1$  and  
 305  $\Gamma = 0.275$ . We can test the validity of this method by comparing our estimates of  
 306 dielectric constant to the values determined in the Ascræus Mons region by Carter et  
 307 al. (2009). By analyzing SHARAD radar signal propagation between the surface and



308 shallow interfaces, they found values of permittivity ranging from 6.2 to 17.3 in the  
309 Northern volcanic flows, with an average of 12.2, while in the southern volcanic flow,  
310 values between 7.0 and 14.0 were estimated with an average of 9.8. In the same area,  
311 the value we estimate for this area using surface power echo is of  $11_{-4}^{+11}$  ( $n = 80$ ). This  
312 result is in fair agreement with Carter et al. (2009).  
313

### 314 **3. Results and discussion**

315

#### 316 **3.1 Global geographic variations**

317

318 Examination of the global dielectric map (figure 7) reveals significant spatial  
319 variations over the planet. The most obvious feature is the latitude-dependent pattern  
320 in reflectivity. The equatorial region displays generally low values, the mid-latitudes  
321 generally high values, and the high latitudes generally low values again. The low-  
322 reflectivity pattern in the equatorial regions is interrupted by very high reflectivity on  
323 the Tharsis volcanic plateau, including Solis, Sinai and Daedalia plana, and the area  
324 between the Tharsis Montes and Olympus Mons. In the mid-latitude bands, the largest  
325 continuous patch of high reflectivity is on the northern side of the Elysium Mons  
326 shield. The northern high latitudes are generally lower in reflectivity than the southern  
327 high latitudes, with the exception of the area north of Alba Patera. The south residual  
328 polar cap has a very low reflectivity due to interferences within the thin layer of CO<sub>2</sub>  
329 ice, causing much weaker surface reflections compared to reflections from a pure  
330 water ice surface as described in details by Mouginot et al. 2009. Bands of apparent  
331 low reflectivity in the Terra Cimmeria region are probably uncorrected artifacts  
332 related to the interaction of the remnant crustal magnetics and the ionosphere  
333 (Safaeinili et al. 2007, Mouginot et al. 2008).

334 Two main parameters are likely to control the dielectric constant of the layer  
335 involved in the reflection process: the composition (chemistry, mineralogy, water  
336 content and physical state) and the density of the constituent materials. Because water  
337 ice presents a low dielectric constant (typically ~3.1) compared to igneous rocks (~8)  
338 at MARSIS wavelengths, the presence of a significant amount of ice in the layer

339 involved in the reflection process will lead to a decrease in surface reflectivity,  
340 compared to a dry, dense rock layer. However, low-density materials can also lead to  
341 such low values of the effective dielectric constant and the use of complementary  
342 datasets is crucial to build robust interpretations. We undertook systematic  
343 comparisons between the MARSIS surface dielectric map, visible albedo maps,  
344 topographic maps, the water concentration map derived from neutron spectroscopy,  
345 and the thermal inertia map derived from TES (Thermal Emission Spectrometer on  
346 Mars Global Surveyor) data.

347

### 348 **3.2 Latitudinal variations and the onset of ground-ice**

349

350 Examination of the MARSIS global map (figure 8a, b) and the corresponding  
351 longitudinally averaged profile (figure 9) reveals a strong latitudinal dependence of  
352 dielectric constant. For both hemispheres, the highest values (6 - 10) are observed at  
353 tropical latitudes. These values are in agreement with laboratory measurements on dry  
354 igneous rocks (Campbell and Ulrichs, 1969; Heggy et al., 2007) and are consistent  
355 with igneous basaltic to granitic rocks. Variations of materials density are probably  
356 responsible for the observed variability of the dielectric constant within igneous  
357 terrains.

358 Poleward of around 50-60° in both hemispheres, terrains show a steep  
359 decrease of dielectric constant down to values of 3 – 4. Climate related processes are  
360 likely responsible for the observed latitudinal variations (Head et al., 2003), and the  
361 comparison with other observations of the Martian surface strongly suggests that this  
362 dielectric constant decrease corresponds to the onset of water-ice occurrence within  
363 the regolith. For the current average water vapor atmospheric content, the frost point

364 at Mars surface is around 200 K. Average temperatures below the frost point are  
365 reached for latitudes in excess of 30 to 45° in the Northern hemisphere and 40° in the  
366 Southern hemisphere. Permanent stability of water ice is possible at depth, where the  
367 sub-surface is insulated from the diurnal and seasonal temperature fluctuations. As  
368 ice is not stable on the surface, we expect the ground ice to be overlaid by a layer of  
369 dry regolith whose thickness depends on latitude, soil physical properties, and the  
370 seasonal evolution of surface humidity (Mellon and Jakosky, 1991; Schorghofer and  
371 Aharonson, 2005).

372 The MARSIS dielectric transition is not associated to a systematic change in  
373 surface albedo or thermal inertia (figure 10), which implies that the surface  
374 geological material does not change much upon crossing this transition. There is a  
375 general agreement between the latitudinal distribution of near-surface (~top meter)  
376 ground ice detected by GRS and the dielectric decrease observed by MARSIS  
377 (figures 8 and 10). In the case of the southern hemisphere there is a good agreement  
378 between the latitude of the ground-ice determined by MARSIS and GRS, with an  
379 average latitude of 48-50°, almost constant with regard to longitude. This ground-ice  
380 extent is also in good agreement with the calculated stability limit for the currently  
381 observed atmospheric conditions (Mellon and Jakosky, 1991; Schorghofer and  
382 Aharonson, 2005). This suggests that the icy layer is in equilibrium with the current  
383 climatology.

384 In the case of the Northern hemisphere, the transition between low and high  
385 reflectivity terrains is generally shifted equatorward compared to the ground-ice limit  
386 detected by GRS and shows some longitudinal variations (figure 8). Most of this  
387 longitudinal variability is likely related to spatial variations of the terrains' physical  
388 properties and not to distribution of water ice. Particularly high dielectric values

389 observed at high latitudes in the north Tharsis region are likely caused by a higher  
390 density of the near-surface materials. Excluding these regions, the MARSIS ground-  
391 ice limit appears shifted equatorward by 6-7° compared to the GRS ground-ice limit.  
392 Recent evidences for the presence of ground ice at latitude as low as 43°N have been  
393 reported in Arcadia Planitia. Young craters were found to have excavated bright  
394 materials with the diagnostic spectral features of water-ice (Byrne et al., 2009). The  
395 location of these craters is consistent with the MARSIS limit of ground-ice  
396 occurrence. The bright ice deposits excavated by these young craters as well as the  
397 ground ice sampled by the Phoenix Lander both point to the presence of layers of  
398 nearly pure ice below the regolith.

399 Preliminary MARSIS estimates of ice fraction for the Northern latitudes (50 –  
400 100 % by volume, see section 3.3) would also imply the presence of water ice in  
401 excess of the regolith porosity. In the Southern hemisphere, estimated values do not  
402 allow us to firmly constrain the origin of ground ice. Such a mode of occurrence is  
403 inconsistent with pore-filling ice emplaced by direct condensation from atmospheric  
404 water vapor. It requires the deposition of an ice-rich material on the surface, such as  
405 frost or snow, then buried under a layer of dry regolith formed either by sublimation  
406 of dusty ice or by wind transport. Morphologic evidences support this interpretation  
407 for both Northern and Southern ground-ice (Head et al., 2003). Repeated freeze thaw  
408 (due to long-term variations in obliquity) of an initially ice-saturated regolith can also  
409 result in the migration of thin films of adsorbed water along mineral grain surfaces in  
410 response to the presence of a temperature gradient. On Earth, this process can result in  
411 the formation of massive ice deposits/lenses (Washburn 1980; Williams and Smith  
412 1989), which could also explain the Phoenix and GRS observations. However, this  
413 mechanism is unlikely to explain ice concentrations of 50-100% to depth >60 m into

414 the subsurface. Beyond the in situ enrichment of ice by repeated freeze/thaw (or the  
415 deposition of massive mantles of ice) in response to obliquity variations, there is also  
416 the potential survival of water discharged by the outflow channels (Carr 1990) – or  
417 the survival of frozen relic of an early ocean (Clifford and Parker, 2001).

418

### 419 **3.3 Estimation of the volume of water-ice seen by MARSIS**

420

421 The aim of this paragraph is to propose a rough first order estimate of the  
422 water-ice content of the ground based on reflectivity values measured by MARSIS in  
423 a simple and ideal case. The accurate inversion of reflectivity values in terms of  
424 ground ice content is indeed a challenging task that requires a complete physical  
425 modeling of the radar waves reflection process and the knowledge of various  
426 properties of the ground. If some of these properties remain unconstrained (layering,  
427 density...), it is likely that different models of subsurface composition and structure  
428 can lead to similar values of reflectivity, i.e. the inversion process does not lead to a  
429 unique solution. Our future work on the retrieval of subsurface composition and  
430 texture from MARSIS absolute reflectivity values, based on physical modeling, will  
431 be focused on these issues but is beyond the scope of this first study. Here, we  
432 calculate the order of magnitude of the amount of water ice required in the subsurface  
433 to account for the values of reflectivity measured by MARSIS assuming a  
434 homogeneous composition at the scale of a MARSIS resolution cell and fixed  
435 parameters for the texture of the ground. By doing that, we aim to provide to the  
436 reader an idea of the quantity of ice that can be probed by our method and set a  
437 starting point for future studies that will refine the rough value estimated here. In the  
438 framework of the simple investigated hypothesis, two values have to be estimated to

439 calculate the quantity of water ice: the thickness of the layer probed by the reflection  
 440 of the electromagnetic waves on the surface and the ice/rock ratio in this  
 441 homogeneous layer.

442

### 443 **3.3.1 Thickness of the probed layer**

444

445 The reflectivity coefficient we determine is related to the permittivity of the  
 446 surface materials. If one supposes that the subsurface is an infinite homogenous half-  
 447 space, MARSIS is directly probing the specular reflection of the radar wave. The  
 448 depth involved in the reflection process is thus given by the skin effect. The loss is  
 449 given by:

$$450 \quad \alpha = 0.091 f \sqrt{\epsilon} \tan(\delta)$$

451 Where  $f$  is the frequency (in MHz) and  $\tan(\delta)$  the loss tangent.

452 From the value of  $\alpha$  one might calculate the skin depth by using:

$$453 \quad d = \frac{10 \log_{10} e}{\alpha}$$

454 Loss tangent for Mars surface rocky materials are expected to vary between 0.004 and  
 455 0.03 (Picardi et al., 2004) at 5 MHz and were measured to be within the 0.01-0.03  
 456 range (at 20 MHz) for lava flows west of Ascræus Mons (Carter et al., 2009). Using  
 457 0.004 and 0.03 as likely extreme values for the loss tangent, we find the  
 458 corresponding values for the skin depth to be 780 and 100 m respectively.

459 The MARSIS radar signal has a temporal resolution of  $\Delta t = 1 \mu\text{s}$  that is equivalent to  
 460 a propagation length in the media of  $2L = c\Delta t / \sqrt{\epsilon}$ . Because this distance is smaller

461 than the skin thickness  $d$ , the probed depth in the surface echo is controlled by the  
 462 temporal resolution. For  $\varepsilon = 6$  and 3, this resolution is of 60 to 80 m respectively. It is  
 463 important to note that this probed thickness is considerably larger than for other usual  
 464 remote-sensing methods like visible-NIR imaging and spectroscopy (a few  $\mu\text{m}$ ) or  
 465 neutrons and gamma-ray spectroscopy (a few tens of cm).

466

### 467 3.3.2 Ice/rock ratio

468

469 In the simple case of a two components mixture examined here, mixing ratios  
 470 between rock and ice can be roughly estimated using the Maxwell-Garnett rules  
 471 (Maxwel Garnett, 1904). Maxwell-Garnett rules assume an asymmetric binary  
 472 mixture with a matrix or environment of dielectric constant:  $\varepsilon_e$  and spherical  
 473 inclusions of dielectric constant:  $\varepsilon_i$ . The medium is assumed to be isotropic. Let  $f$  be  
 474 the volumetric fraction of inclusions in the mixture. The Maxwell-Garnett mixing  
 475 formula allows us to calculate the effective dielectric constant,  $\varepsilon_{\text{eff}}$ , of the binary  
 476 mixture:

477

$$\varepsilon_{\text{eff}} = \varepsilon_e + 3f\varepsilon_e \frac{\varepsilon_i - \varepsilon_e}{\varepsilon_i + 2\varepsilon_e - f(\varepsilon_i - \varepsilon_e)} \quad (10)$$

478 The inversion of dielectric constant value measured by MARSIS in terms of  
 479 ice / rock ratio requires the knowledge of the dielectric constants of the pure end-  
 480 members. If the value of  $\varepsilon' = 3.1$  for pure water ice at MARSIS wavelength is  
 481 commonly accepted (Petrenko, 1999) and is not likely to be subject to strong spatial  
 482 or temporal variations, the choice of a value for the rocky component of the mixture is  
 483 more problematic. Usually the dielectric constants for volcanic rocky materials are



484 comprised between 6 and 11 (Campbell and Ulrichs, 1969). The mean value found in  
485 tropical regions by MARSIS is 6.5. Thus, we assume that the dielectric constant of the  
486 dry rock chosen for our calculations are the same as the one measured by MARSIS in  
487 equatorial regions. The chosen value is relatively low and even if the nature of the soil  
488 changes in certain regions, we can give a lower estimate of the ice / rock mixing ratio.  
489 We obtain values of the ice / rock volume mixing ratio of the order of 50 % in the  
490 Southern and between 50 and 100 % in the Northern. Although the variability is high,  
491 it appears that the average amount of ice present in the subsurface is higher in the  
492 Northern hemisphere than in the Southern hemisphere. Furthermore, other  
493 observations point to systematic differences between the two hemispheres (see section  
494 3.3).

495 The total amount of water stored in the two Martian polar layered terrains  
496 (Plaut et al., 2007, Smith et al., 2001) is estimated to be  $2.8 \cdot 10^6 \text{ km}^3$ . The atmosphere  
497 contains the equivalent of  $\sim 3 \text{ km}^3$  of condensed water (Plaut et al., 2001). The total  
498 amount of water contained in the subsurface is certainly the main uncertainty in the  
499 current inventory of water on Mars. While neutron spectroscopy permitted an  
500 estimation of the amount and extent of ice in the top meter of the regolith, low  
501 frequency radar techniques now offer the unique opportunity to probe the regolith and  
502 quantify the amount of ice at decameters to kilometers scale. In the framework of the  
503 ideal hypothesis of a homogeneous subsurface considered in this paragraph, we can  
504 estimate the total amount of ground water ice necessary to explain the low values of  
505 reflectivity measured by MARSIS. Considering the measured extent of both Northern  
506 and Southern ground-ice, the average ice / rock ratio obtained from inversion of  
507 MARSIS dielectric measurements and a probed thickness of 60-80 meters, we  
508 estimate the potential lower limit of the total volume of ice currently stored at high

509 latitude in the ground to be  $\sim 10^6$  km<sup>3</sup>, of the order of magnitude of the volume of one  
510 of the polar caps. As already mentioned at the beginning of this paragraph, future  
511 studies dedicated to the physical modeling of the radar reflection process should now  
512 be undertaken to examine the influence of the heterogeneity of the subsurface on the  
513 inversion of dielectric values in terms of amount of water ice in the subsurface. The  
514 case of meters-thick lenses of pure water ice close to the surface should be  
515 investigated in priority as this ground structure has been mentioned to interpret recent  
516 datasets, especially in-situ observations by the Phoenix Lander (Smith et al., 2009).

517

### 518 3.4 Tropical and equatorial minima

519

520 Examination of the MARSIS surface dielectric map in the equatorial to  
521 tropical regions reveals the occurrence of a nearly continuous low reflectivity belt  
522 between  $-30^\circ$  and  $+30^\circ$  latitude. Extremely low values are reached in the Medusae  
523 Fossae area ( $\epsilon \approx 3$ ) (figure 8 and 10). Other minima are encountered in the areas of  
524 Meridiani Planum ( $\epsilon \approx 4$ ) and from Isidis basin to the highlands south of Elysium  
525 Planitia ( $\epsilon = 3.5$ ). The low values of dielectric constants can have various origins:  
526 different composition of rocks, low density of surface material, presence of water ice  
527 buried beneath a desiccated regolith... Radar measurements alone do not allow  
528 discriminating between these different possibilities. When compared to other datasets,  
529 it appears that, as in the polar regions, the best match for the MARSIS reflectivity  
530 equatorial pattern is obtained from comparison with the GRS WEH map. Indeed, the  
531 low reflectivity regions roughly correspond to regions that were shown to be enriched  
532 in hydrogen by the GRS instrument suite, with WEH values up to 12%. The  
533 interpretation of high hydrogen enrichments detected by GRS in the top first meter of  
534 the regolith is still debated. Different explanations have been proposed, like a high  
535 abundance of nominally hydrated minerals (Fialips et al., 2005), an interaction of the  
536 regolith with atmospheric water vapor (Feldman et al., 2005), or the presence of  
537 transient ground ice (Jakosky et al., 2005).

538 The spatial agreement between the low-reflectivity values on the MARSIS  
539 map and the H-enriched areas on the GRS map strongly points toward a water-related  
540 explanation. In particular, the Medusae Fossae formation terrains, when sounded by  
541 MARSIS (Watters et al., 2007), were found to have a bulk real dielectric constant  
542  $\epsilon = 2.9 \pm 0.4$ , in agreement with our near-surface estimate. The authors proposed two

543 hypotheses to explain these low values: low-density volcanic deposits or the presence  
544 of ice, deposited during a high-obliquity/high-humidity climatic excursion that is  
545 currently sublimating at some depth and hydrating the overlying regolith. Spatial  
546 correlation between MARSIS and GRS measurements supports this last hypothesis.  
547 Unfortunately, the new MARSIS results do not generally allow arguing in favor or  
548 one or another of the hypotheses proposed to explain high WEH values measured by  
549 GRS. Indeed, the presence of low density / highly hydrated sedimentary materials,  
550 such as the ones observed in-situ by the Opportunity rover in Meridiani Planum  
551 (Squyres et al., 2004) could explain the observed low radar reflectivity because of  
552 their low dielectric constant (Campbell and Ulrichs, 1969) as well as the presence of  
553 ground ice as already discussed for high latitudes regions.

554         If the MARSIS radar map does not allow us to choose unambiguously  
555 between the pre-cited mechanisms, its main implication is that anomalous surface  
556 hydrogen enrichments detected by GRS correspond to anomalies of dielectric  
557 constant, possibly related to the presence of H<sub>2</sub>O in an unconstrained state, extending  
558 down to at least a few decameters below the surface. This is a new and important  
559 constraint on the nature and origin of the equatorial anomalies that should be taken  
560 into account in future interpretations of these regions.

561

562

563

## 564 **Conclusion**

565

566 A global 3-5 MHz dielectric map of the Martian surface is built from two  
567 years of measurements by the MARSIS instrument. Using MOLA topographic data,  
568 we corrected the effect of kilometer-scale roughness and slopes to retrieve the spatial  
569 variations of the dielectric constant of near-surface materials. Many parameters are  
570 susceptible to influence the dielectric constant of the top decameters below the surface  
571 that are probed by the radar wave reflection process. With the use of complementary  
572 datasets, we were able to interpret the geographic variations of radar reflectivity in  
573 term of surface geological material. From the comparison of the MARSIS map with  
574 the hydrogen abundance map derived from GRS measurements, we show that low  
575 dielectric values at latitudes poleward of 50-60° are likely to be due to high amount of  
576 water ice in the ground.

577 The thickness probed by the MARSIS surface reflectivity is of the order of a  
578 few decameters, resulting in a significant increase of the total quantity of ground ice  
579 compared to GRS observations, only representative of the first meter below the  
580 surface. Assuming a homogenous composition of the ground at the scale of a  
581 MARSIS resolution cell as a simple first-order hypothesis, measured values of  
582 dielectric constant lead to a minimum estimate of ice stored in the Martian sub-  
583 surface to be of the order  $\sim 10^6$  km<sup>3</sup>, equivalent to a polar cap. Future studies focused  
584 on the physical modeling of the reflection process over more realistic models of the  
585 subsurface structure and texture will be necessary to refine this rough estimation.

586 At least in the Northern hemisphere the amount of ice appears to be in excess  
587 of porosity. Refined analysis of the frequency dependence of the surface reflectivity,  
588 together with laboratory measurements of the dispersion relation of ice-rock mixture

589 might help in confirming this observation that would have strong implication for the  
590 mechanism of ground ice emplacement.

591           Puzzling anomalies in hydrogen at equatorial latitudes first revealed by the  
592 GRS instrument correspond to low reflectivity areas on the MARSIS map. If low  
593 reflectivity alone does not permit to discuss further the origin of the hydrogen  
594 anomalies, presence of highly hydrated minerals or shallow buried ice, it tells us that  
595 the anomalies identified by GRS in the near subsurface actually extend in depth to the  
596 first decameters below the surface.

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598

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608

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783 **Figure Captions**

784

785 **Figure 1:** Top: a typical MARSIS pulse measured over the South Polar Layered  
786 Deposits (orbit #2682, pulse 718) in dB. The surface echo is recorded at about  $t = 130$   
787  $\mu\text{s}$  after the opening of the receiver's window. Another strong echo, attributed to the  
788 reflection on the bedrock below this ice, is received at  $t = 170 \mu\text{s}$ . Bottom: the criteria  
789  $C$  as defined in equation 1 calculated for the same MARSIS pulse. The position of the  
790 surface echo is indisputably highlighted by the high value of  $C$ ,

791 **Figure 2.** Top to bottom: the radargram of orbit #2787, the position of the surface  
792 echo detected by our algorithm, the power reflected by the surface in dB.

793 **Figure 3** (a) Left: raw reflectivity as function of the Solar Zenith Angle (SZA). (b)  
794 Right: mean reflectivity as function of the Total Electron Content. Both graphics have  
795 been plotted using the entire set of data at 4 MHz.

796 **Figure 4.** Reflectivity corrected for absorption as function of the total electron  
797 content. The graph has been plotted using the entire set of data at 4 MHz.

798 **Figure 5. A:** Reflectivity map at 3-5 MHz of the Martian surface as seen by  
799 MARSIS. Red corresponds to high reflectivity and blue to low reflectivity. Grey  
800 regions correspond to a lack of data. The map is in cylindrical projection. The spatial  
801 resolution is 0.5 bin per degree.

802 **B:** Reflectivity map based on simulated radargrams. Grey regions correspond to a  
803 lack of data. The map is a cylindrical projection. The resolution is 0.5 bin per degree.

804 **C:** Roughness map from Kreslavsky and Head (2000)



805 **Figure 6.** The image on the left has been provided by HIRISE on board MRO  
806 (PSP\_001736\_2605, credit: NASA/JPL/University of Arizona). This image with 25  
807 cm resolution per pixel shows in detail the dunes in Olympia Undae. The radargram  
808 on top corresponds to a part of orbit #3674 of MARSIS/MEX over this region. The  
809 bottom image is the MOLA topography corresponding to the MARSIS track.

810 **Figure 7.** Reflectivity map corrected for roughness effect. As described in the text,  
811 the reflectivity has been calibrated using a reference making it possible to provide the  
812 corresponding dielectric constant. The map is a cylindrical projection. The spatial  
813 resolution is one bin per degree. Grey background corresponds either to a lack of data  
814 or a removal of data corrupted by artefacts (effects of magnetic field, high surface  
815 roughness).

816 **Figure 8:** (A) MARSIS dielectric map. Same as figure 7.

817 (B) Global interpolation of the dielectric map (A) using 8-order  
818 spherical harmonics adjustment.

819 (C) Global map of hydrogen concentration in the top meter of the  
820 regolith obtained by the neutrons spectrometer of the GRS / Mars  
821 Odyssey instruments suite (data from Feldman et al., 2004).

822 Concentration is expressed as Water Equivalent Hydrogen  
823 abundance in weight percent.

824 All maps are gridded in Robinson projection.

825 **Figure 9:** Longitudinal averages of MARSIS surface real dielectric constant and GRS  
826 neutrons spectrometer WEH concentration. Red crosses correspond to the dielectric  
827 constant values measured by MARSIS and plotted on figure 8.A whereas the red

828 curve corresponds to the interpolated MARSIS map (figure 8.B). WEH values are  
829 calculated from the map plotted in figure 8.C.

830 **Figure 10:** A: Visible color map (Viking), B: Thermal inertia map (TES), C:  
831 MARSIS reflectivity map. The comparison between these maps shows that the limits  
832 displayed on the MARSIS reflectivity map do not correspond to systematic changes  
833 of albedo and thermal inertia, indicating that the surface material remains unchanged  
834 while subsurface material is different.

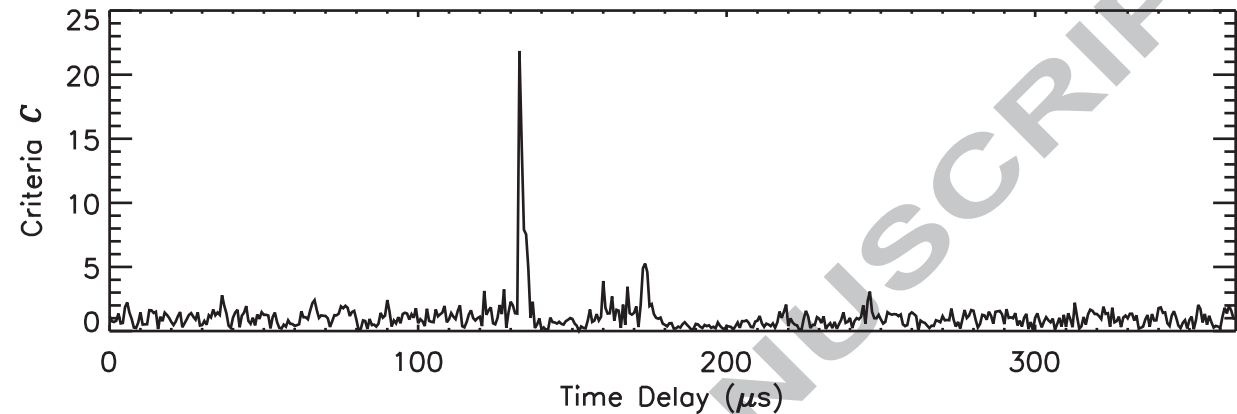
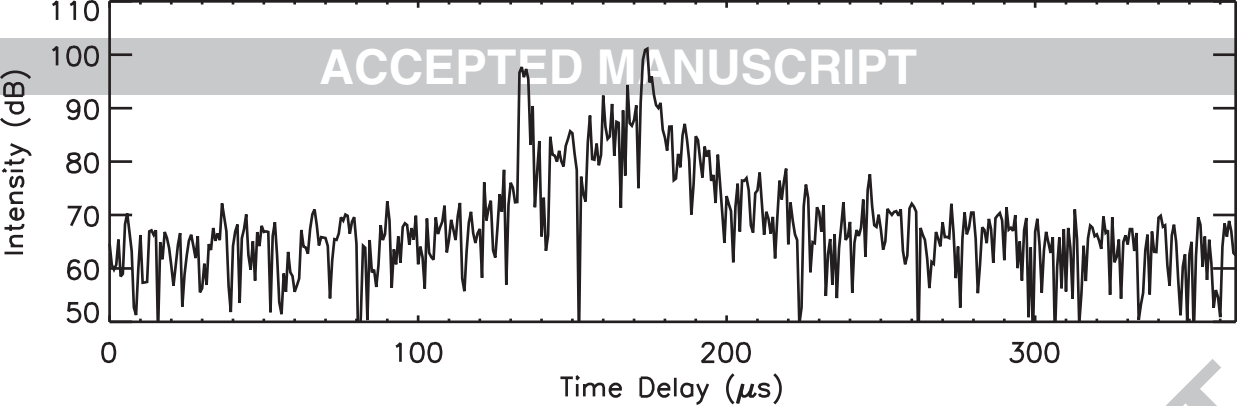
835 **Tables**

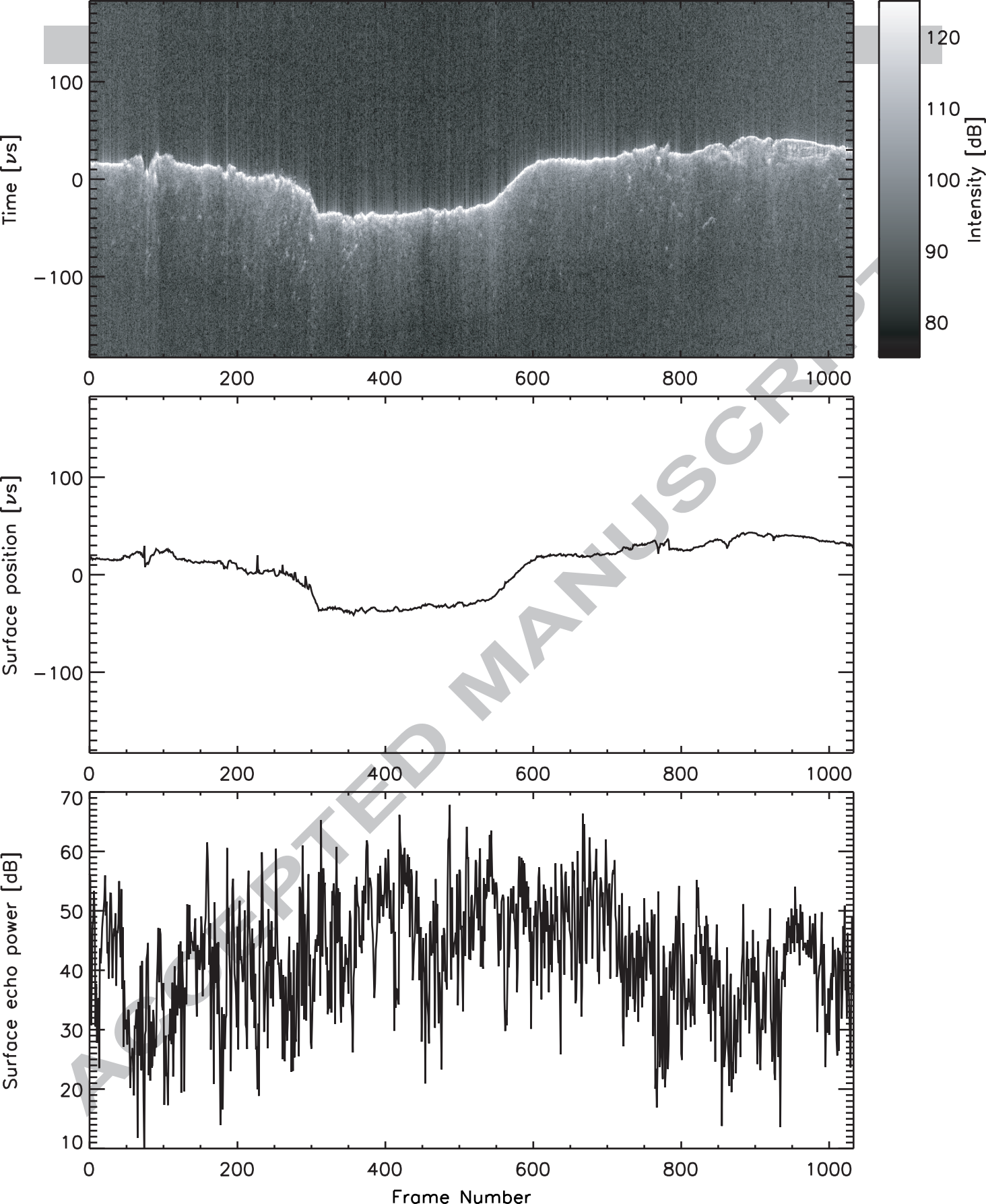
Central Frequency	3 MHz	4 MHz	5 MHz
Total electron Content ( $10^{15}\text{m}^{-2}$ )	3	7.5	7.5
Solar Zenith Angle	85°	70°	60°

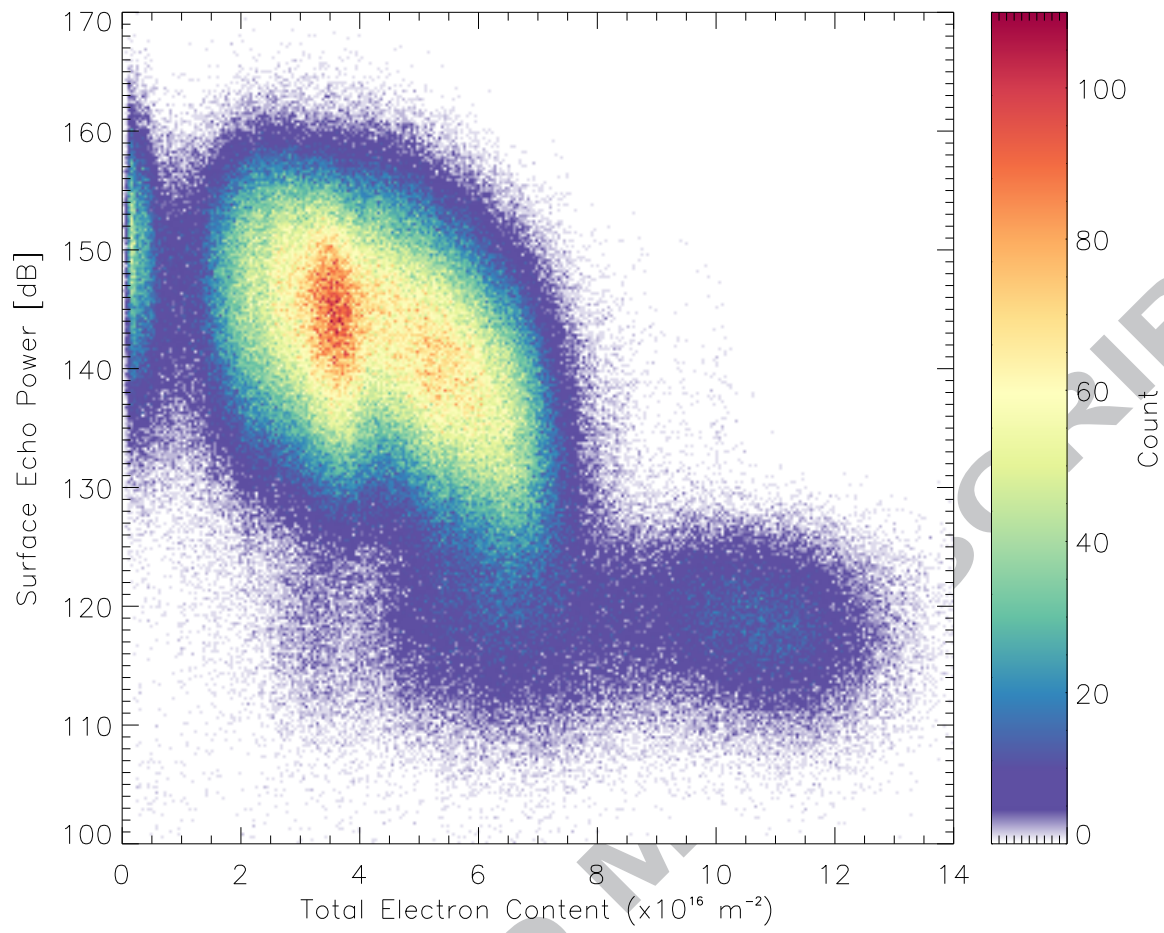
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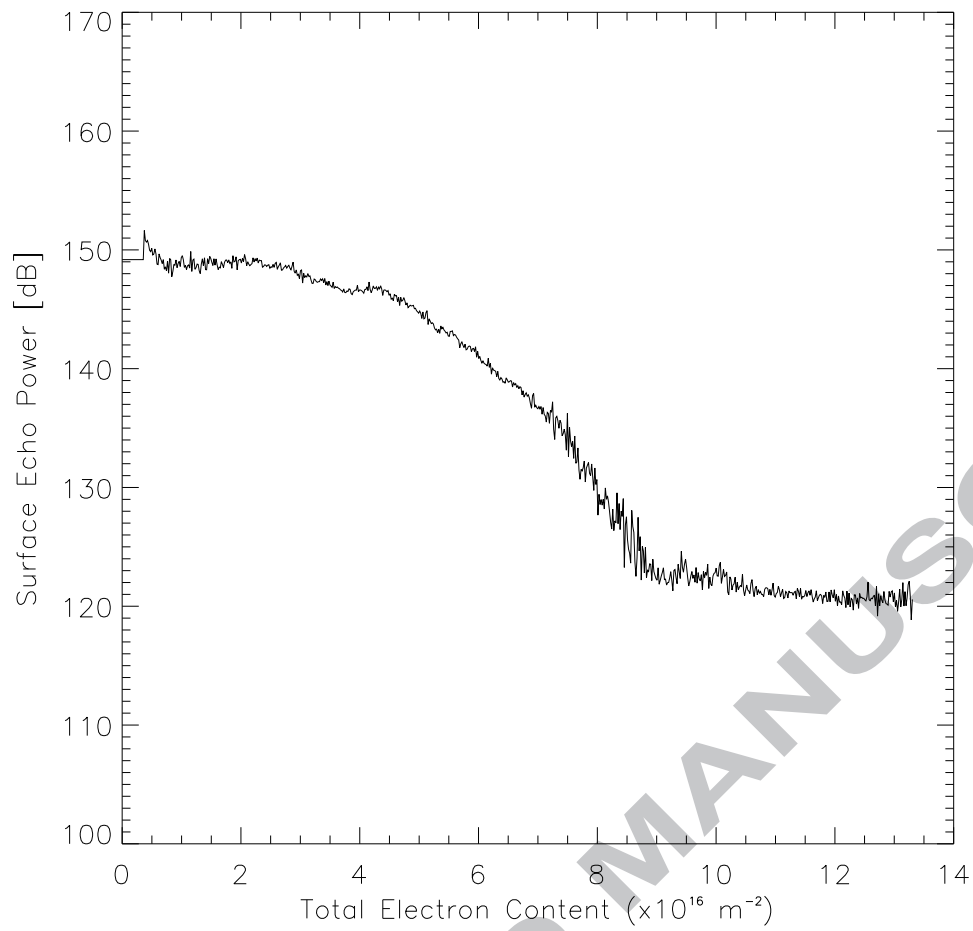
837 **Table 1.** Summary of the limits used to select the data. Measurements are kept when  
838 the total electron content is below the limit indicated by the first line and when the  
839 solar zenith angles are above the limit indicated by the second line.

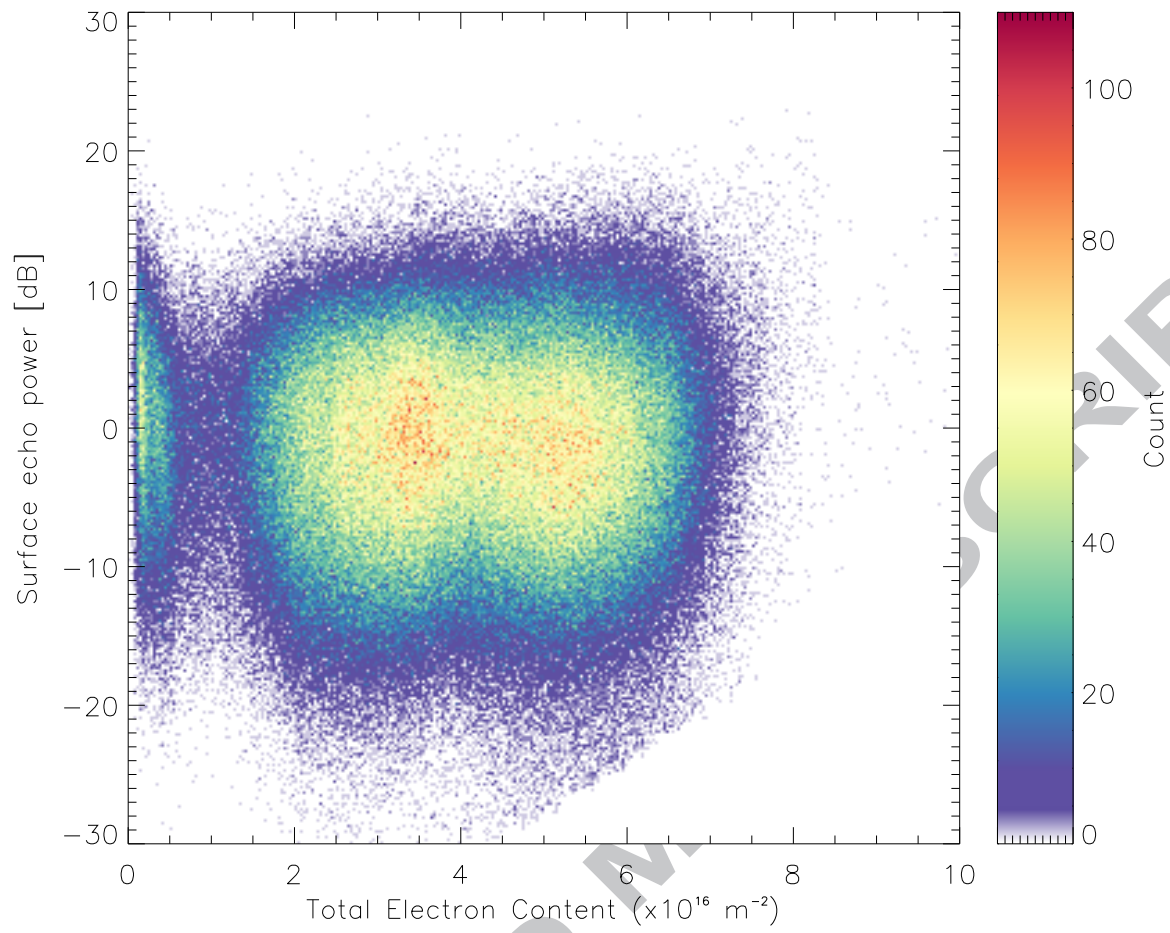
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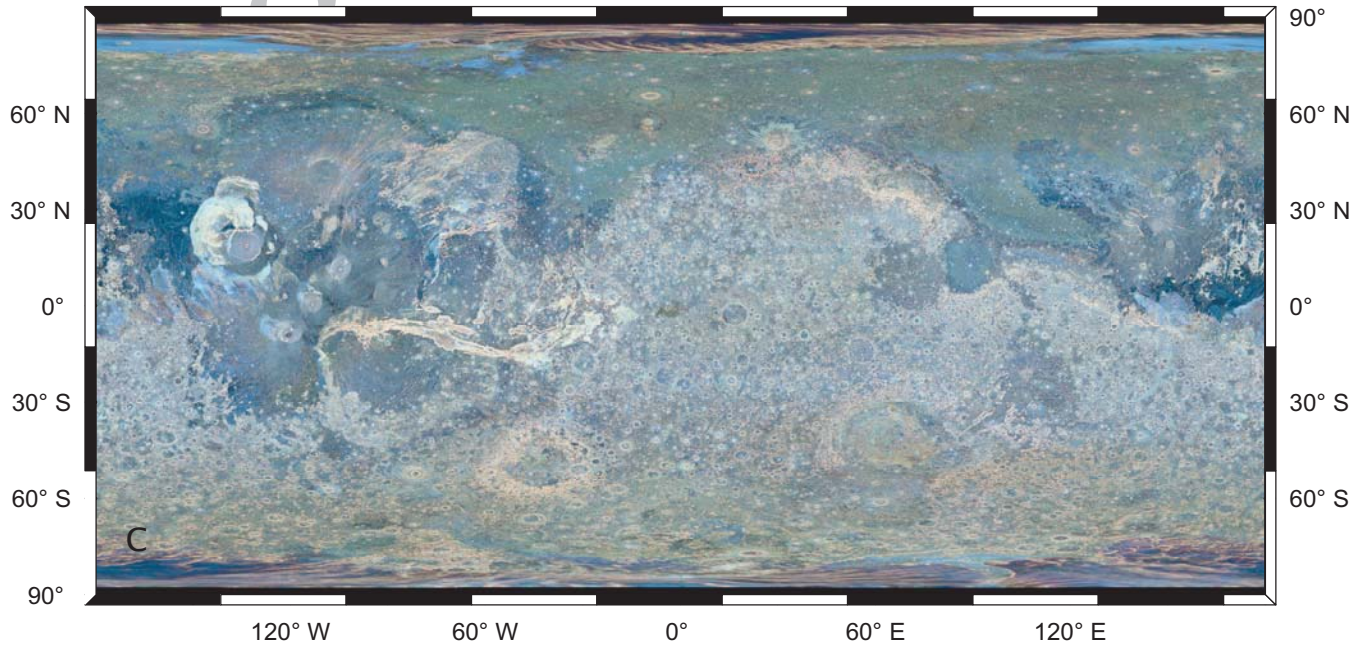
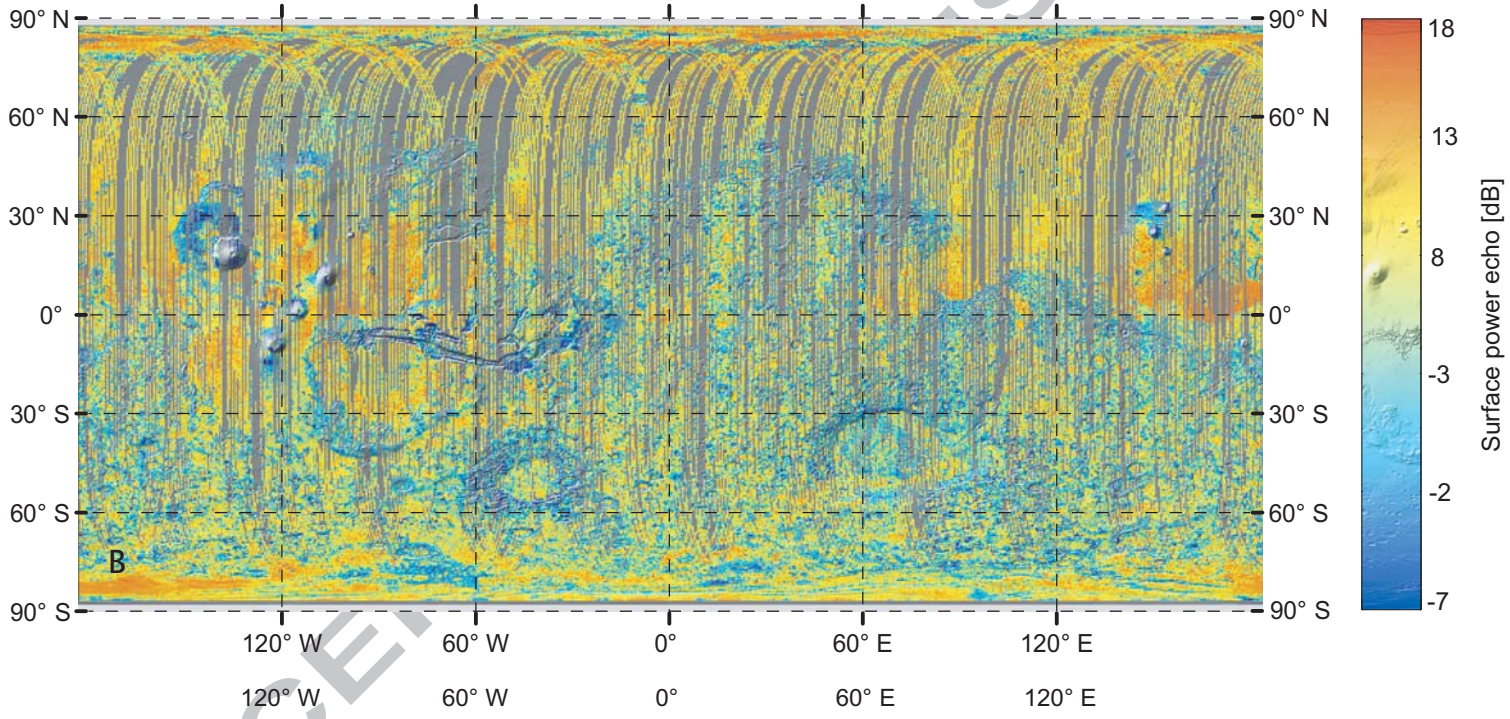
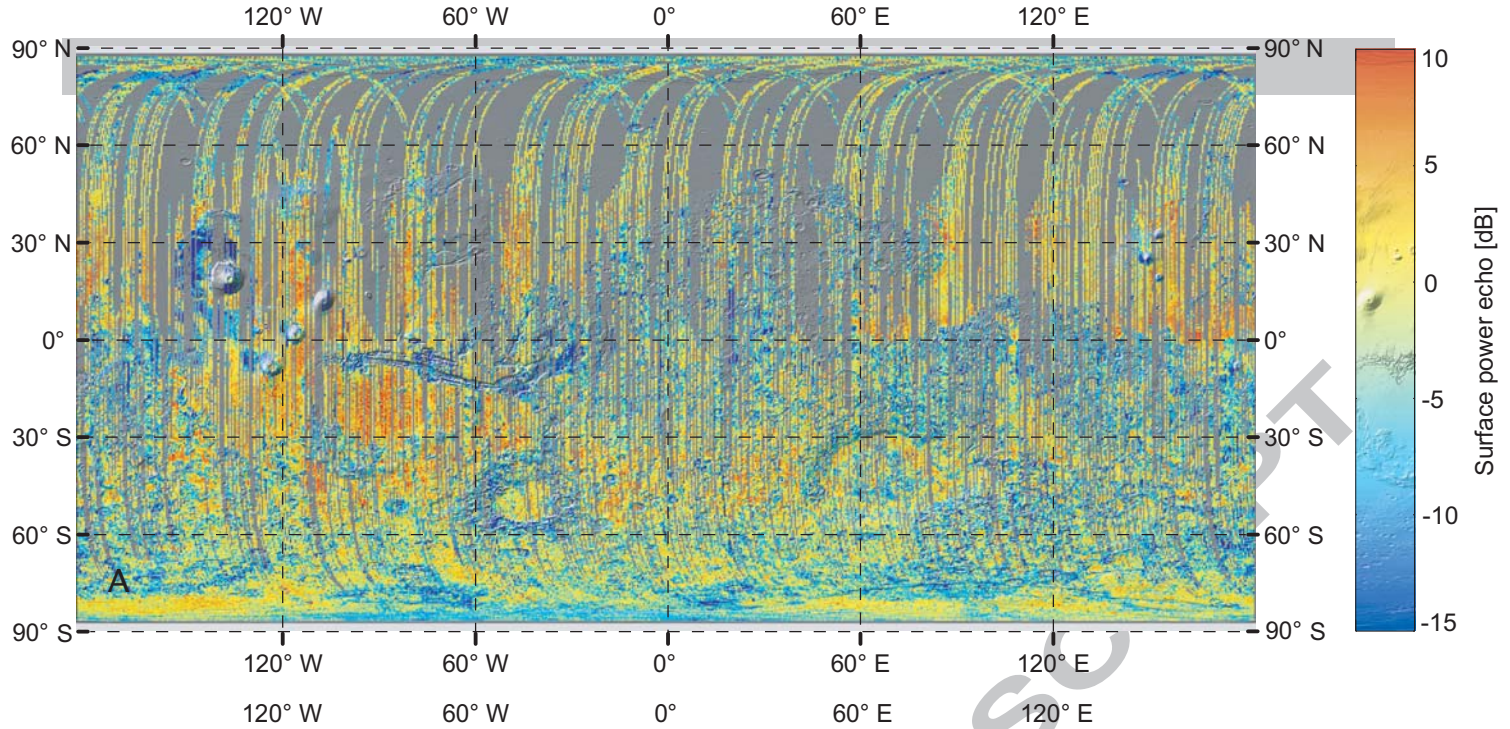


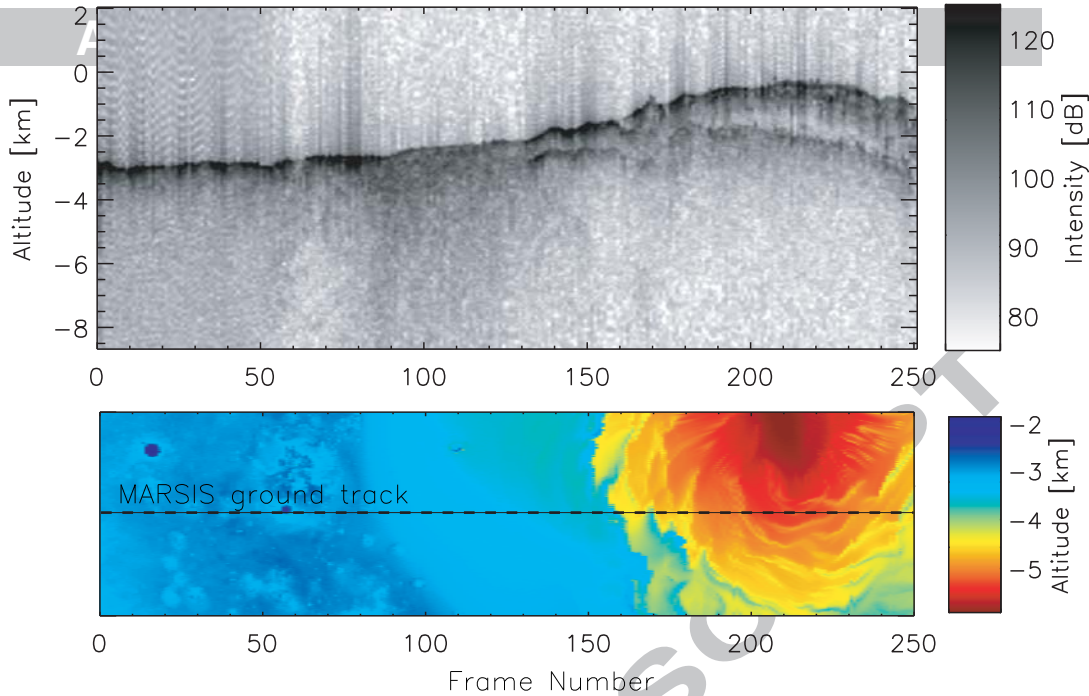


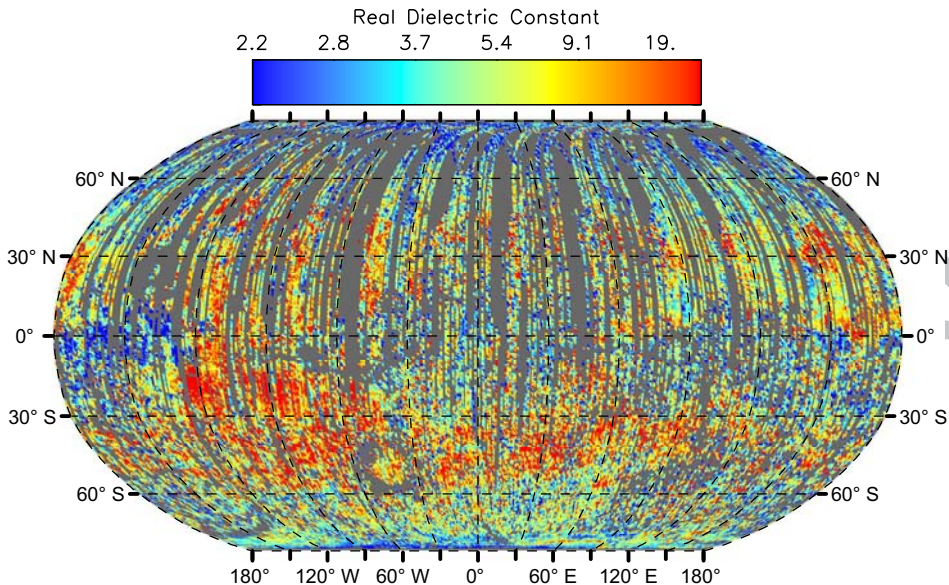


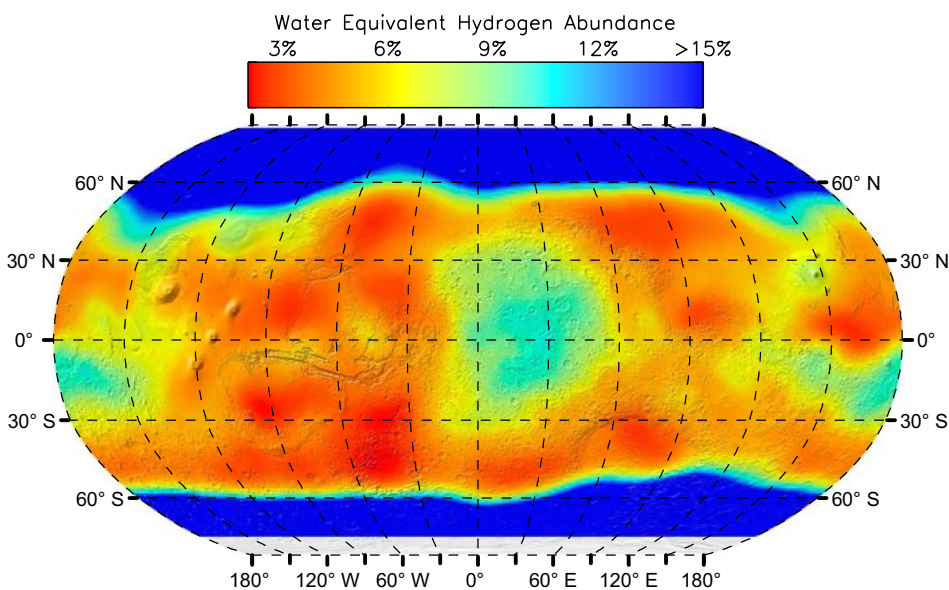
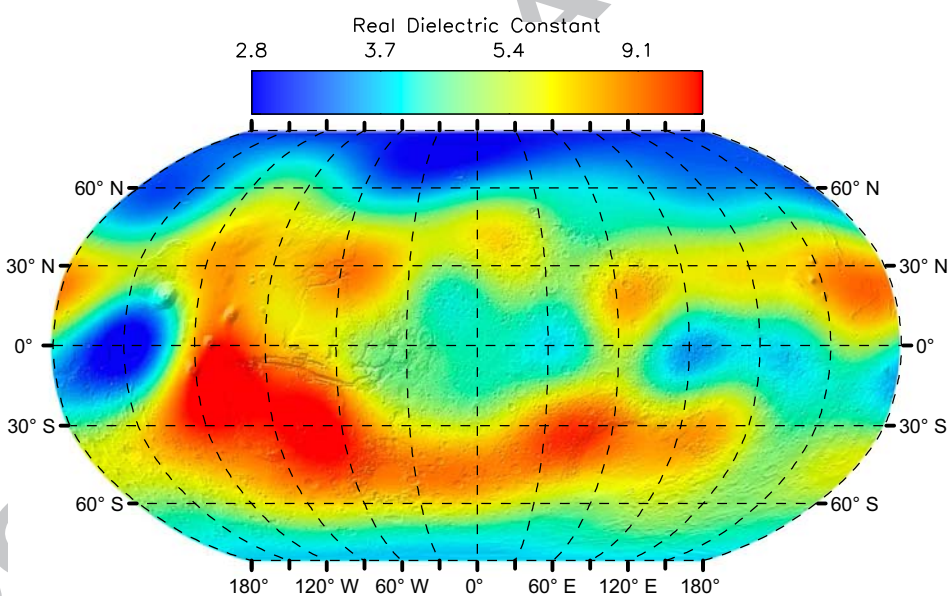
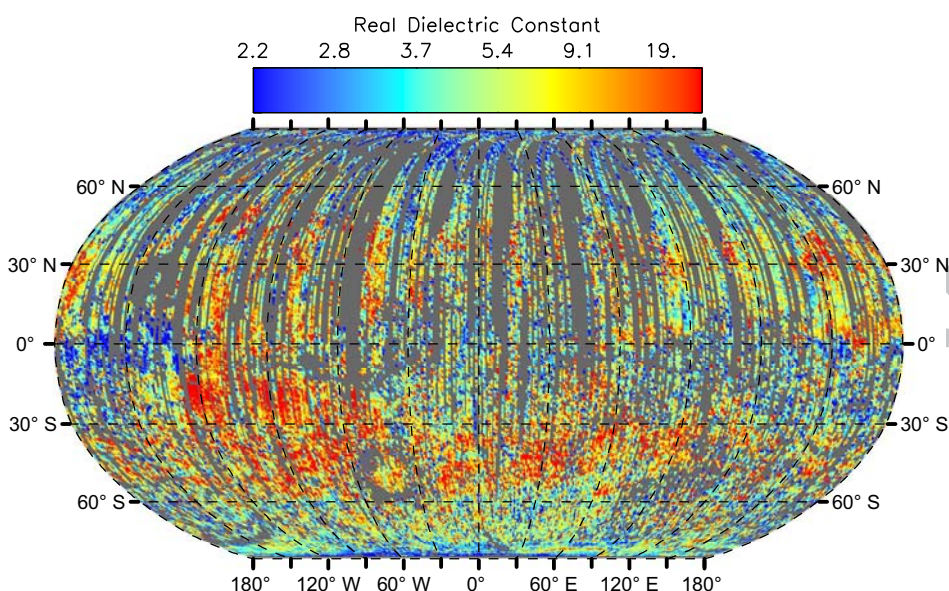


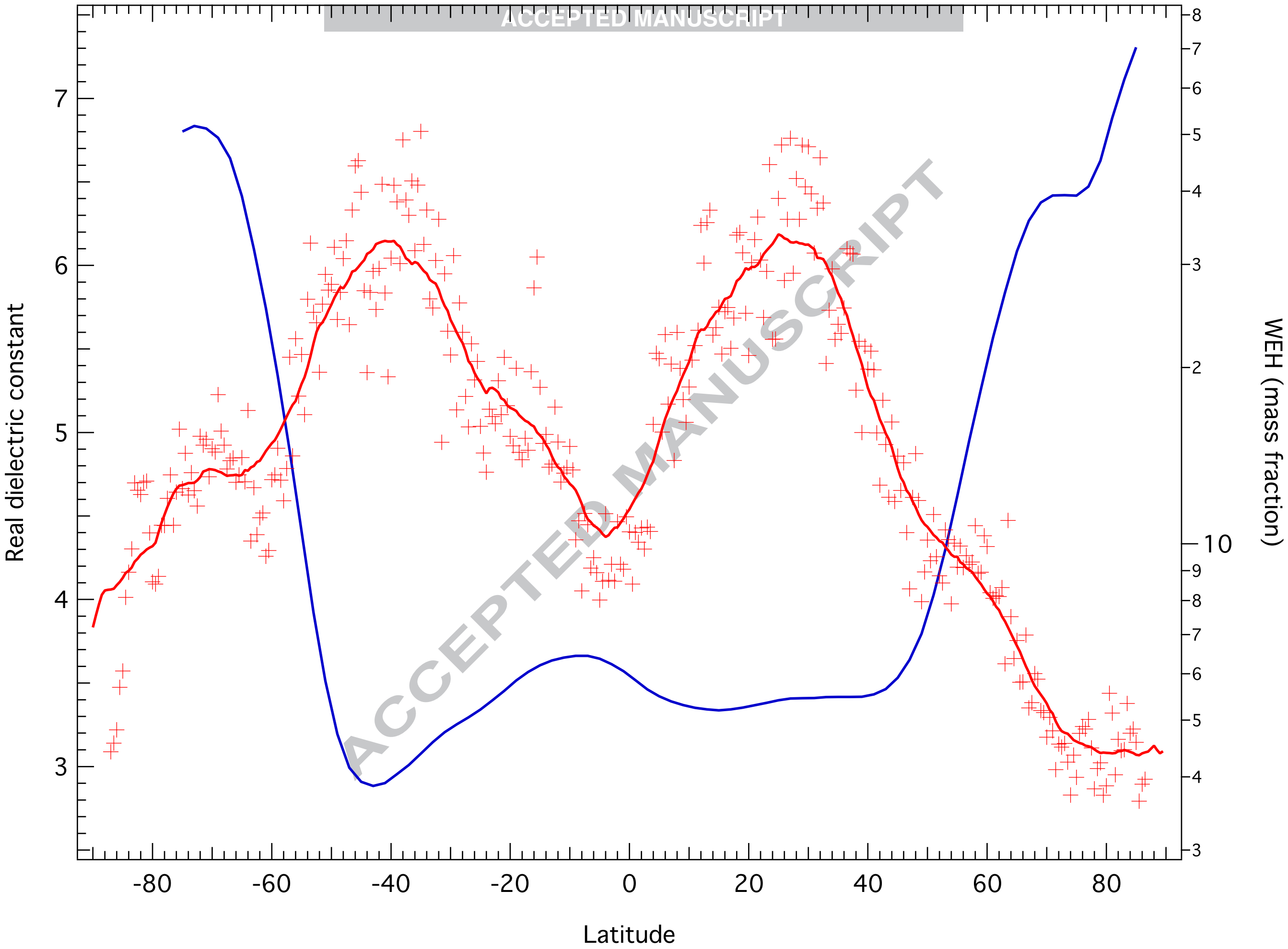








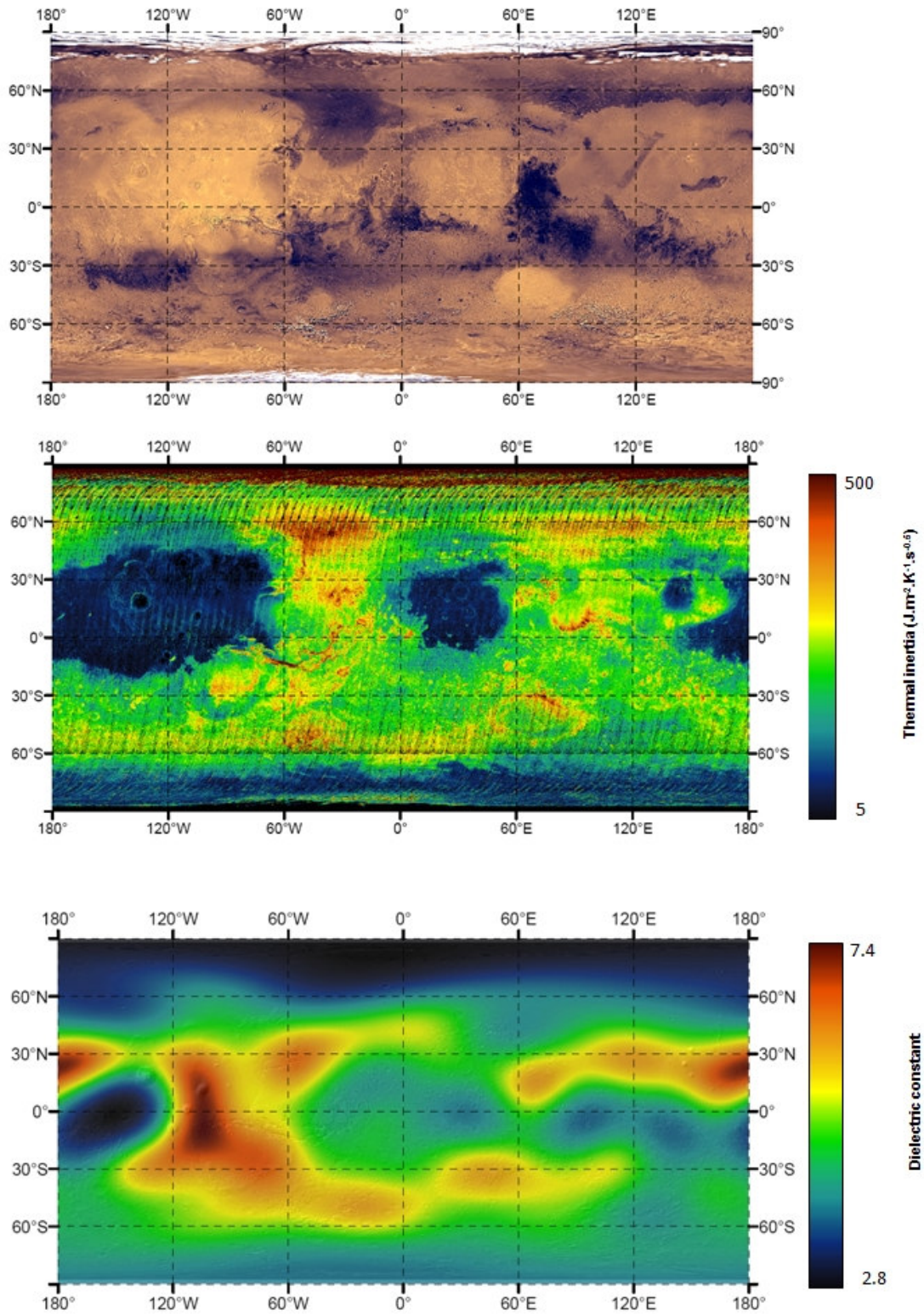




Real dielectric constant

WEH (mass fraction)

Latitude



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Manuscript Draft

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Title: The 3-5 MHz global reflectivity map of Mars by MARSIS/Mars Express: implications for the current inventory of subsurface H<sub>2</sub>O.

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Abstract: We extracted the surface echo power from two years of MARSIS measurements. The retrieved values are calibrated to compensate for changes in the distance of the spacecraft to the surface and for the attenuation of the signal by the ionosphere. The results are used to build the first global map of surface echo power at 3-5 MHz. The surface echo power variations are primarily caused by km-scale surface roughness. Then, we derive the values of dielectric constant of the shallow subsurface materials by normalizing the surface echo power map using a simulation of MARSIS signal from the MOLA topography. As a result, we obtain a map that characterizes the dielectric properties of the materials down to a few decameters below the surface. Dielectric properties vary with latitude, with high values in mid-latitudes belts (20-40°) and lower values at both equatorial and high latitudes. From the comparison of MARSIS reflectivity map to GRS observations, we conclude that the reflectivity decrease observed poleward of 50-60° corresponds to the onset of water-ice occurrence within the regolith. Assuming homogenous ground composition and texture at the scale of the MARSIS resolution cell, our inferred volume of ground water ice is of 106 km<sup>3</sup>, equivalent to a polar cap. Low reflectivity areas are also observed in equatorial regions. From radar studies alone, equatorial low dielectric constant values could have different interpretations but the correlation with GRS hydrogen distribution rather points toward a water related explanation.