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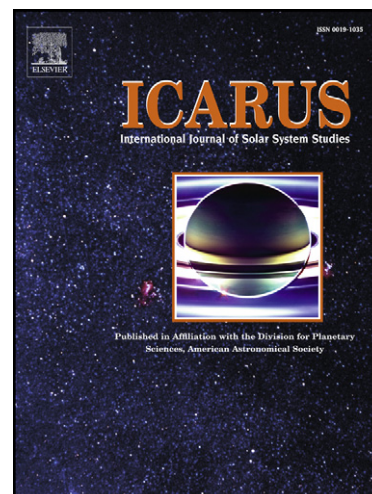
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1 **MARTIAN POLAR AND CIRCUM-POLAR SULFATE-BEARING DEPOSITS:**
2 **SUBLIMATION TILLS DERIVED FROM THE NORTH POLAR CAP**

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49 **Abstract**

50 Previous spectroscopic studies have shown the presence of hydrated minerals in various kinds
51 of sedimentary accumulations covering and encircling the Martian North Polar Cap. More
52 specifically, gypsum, a hydrated calcium sulfate, has been detected on Olympia Planum, a
53 restricted part of the Circumpolar Dune Field. To further constrain the geographical distribution
54 and the process of formation and accumulation of these hydrated minerals, we performed an
55 integrated morphological, structural and compositional analysis of a key area where hydrated
56 minerals were detected and where the main polar landforms are present. By the development of a
57 spectral processing method based on spectral derivation and by the acquisition of laboratory
58 spectra of gypsum-ice mixtures we find that gypsum-bearing sediment is not restricted to the
59 Olympia Planum dunes but is also present in all kinds of superficial sediment covering the
60 surface of the North Polar Cap and the Circumpolar Dune Field. Spectral signatures consistent
61 with perchlorates are also detected on these deposits. The interpretation of landforms reveals that
62 this gypsum-bearing sediment was released from the ice cap by sublimation. We thus infer that
63 gypsum crystals that are now present in the Circum-Polar Dune Field derive from the interior of
64 the North Polar Cap. Gypsum crystals that were initially trapped in the ice cap have been released
65 by sublimation of the ice and have accumulated in the form of ablation tills at the surface of the
66 ice cap. These gypsum-bearing sublimation tills are reworked by winds and are transported
67 towards the Circum-Polar Dune Field. Comparison with sulfates found in terrestrial glaciers
68 suggests that gypsum crystals in the Martian North Polar Cap have formed by weathering of dust
69 particles, either in the atmosphere prior to their deposition during the formation of the ice cap,
70 and/or in the ice cap after their deposition.

71 **Keywords:** Mars, polar caps; Mars, polar geology; mineralogy; Ices, IR Spectroscopy; Mars,
72 surface.

73 1. Introduction

74 Various kinds of sulfates have been discovered in several regions of Mars, both from in-situ
75 ground observations by the Spirit and Opportunity rovers and from orbital observations by the
76 OMEGA and CRISM imaging spectrometers. In equatorial and mid-latitude regions, Mg-, Fe-
77 and Ca-sulfates have been detected in light toned layered deposits and in soils [e.g. Christensen et
78 al., 2004; Gendrin et al., 2005; Arvidson et al., 2006; Squyres et al., 2006; Le Deit et al., 2008,
79 Massé et al., 2008; Roach et al., 2009; Wray et al., 2009]. At higher latitudes, Ca-sulfates have
80 been detected in dark dune fields in the vicinity of the North Polar Cap [Langevin et al., 2005b;
81 Roach et al., 2007] and possibly in the permafrost of the Phoenix landing site [Hecht et al., 2009].
82 Understanding the origin of these various kinds of sulfate-bearing deposits is of importance
83 because they constitute key elements to constrain the evolution of the Martian surface and past
84 climate. Various hypotheses have been suggested so far to explain their formation. On Earth,
85 large accumulations of sulfates commonly form in evaporitic environments [Rouchy and Blanc-
86 Valleron, 2006]; one of the classical hypotheses for the formation of sulfates on Mars is therefore
87 evaporitic concentration and deposition in water bodies [Catling, 1999]. Other classical
88 hypotheses involve in-situ weathering of sulphide deposits or basaltic materials by acid fogs or
89 acid groundwaters [see review in Chevrier and Mathé, 2007]. Recently, Niles and Michalski
90 [2009] suggested that Martian equatorial sulfates have formed by weathering of dust trapped in Figure 1
91 ancient equatorial ice caps.

92 Interestingly, the largest accumulation of sulfates detected so far on Mars is located on
93 Olympia Planum, a crescent-shaped dome, 500 km in radius, located at the present-day border of
94 the North Polar Cap (Fig. 1a and b) [Langevin et al., 2005b; Roach et al., 2007]. This dome
95 corresponds to an ancient part of the North Polar Cap, which has been exhumed by erosional
96 retreat of its upper part [Zuber et al., 1998; Fishbaugh and Head, 2000, 2005]. This dome is now

97 extensively covered by superficial sediment and dark dunes. These correspond to the densest part
98 of the Circum-Polar Dune Field that encircles the whole North Polar Cap [Fishbaugh et al., 2007;
99 Tanaka and Hayward, 2008]. The spectral signature of hydrated Ca-sulfate (most likely gypsum)
100 observed by OMEGA and CRISM in this region is correlated to the dark dunes that cover
101 Olympia Planum (Fig. 1b) [Langevin et al., 2005b; Roach et al., 2007].

102 The presence of gypsum in dunes resting on an erosional relict of the North Polar Cap raises
103 the question of whether a link exists between the processes of formation and/or accumulation of
104 these circum-polar sulfates and the past or present activity of the ice cap. The aim of our study is
105 therefore to evaluate, by an integrated morphological, structural and compositional analysis,
106 whether polar and circum-polar gypsum-rich deposits, such as those observed in Olympia
107 Planum, derive from the North Polar Cap or not.

Figure 2

108 For that purpose, we focus on a key area located at the border of the North Polar Cap,
109 between Olympia Cavi and Gemini Scopuli (Fig. 1a and 2). This area provides a comprehensive
110 view, on the same location, of (1) the Circum-Polar Dune Field, (2) the surface of the North Polar
111 Cap and (3) a natural cross-section that reveals the internal structure of the ice cap.
112 Morphological and structural information on this area is provided by the analysis of HiRISE and
113 CTX images and MOLA topographic data, while compositional constraints are provided by the
114 analysis of CRISM and OMEGA hyperspectral data.

115 In section 2, we briefly describe the North Polar Cap, the associated polar and circum-polar
116 superficial accumulations of dark sediment (including the Circum-Polar Dune Field) and we
117 review current hypotheses for the origin of the circum-polar sulfate deposits. Then, we introduce
118 in section 3 the structure and landforms of the study area and discuss their significance. In section
119 4, we present the results of the spectroscopic compositional analysis. We finally discuss in
120 section 5 the implications for the origin of polar and circumpolar sulfates.

121

122 2. Geological setting

123 2.1 The North Polar Cap

124 The North Polar Cap rests in the lowest part of the Vastitas Borealis topographic basin (Fig.
125 1a and c). It is 1300 km in diameter and reaches a maximum thickness of 3 km at its center
126 [Zuber et al., 1998]. The formation of the whole ice cap took place during the Amazonian [Carr
127 and Head, 2009]. It is a stack of water ice layers containing various amounts of intermixed
128 sediment (dust or sand) [Kieffer et al., 1976; Tsoar et al., 1979; Howard et al., 1982; Fishbaugh et
129 al., 2008]. Two distinct units have been recognized in this stack (Fig. 1c). The first one
130 corresponds to the Basal Unit (BU), which rests directly on the Vastitas Borealis Formation
131 (VBF). High-resolution images reveal that the BU consists of a low albedo, 1 km-thick formation
132 displaying platy interbedded sequences of ice-rich and sediment-rich layers [Byrne and Murray,
133 2002; Edgett et al., 2003; Fishbaugh and Head, 2005]. SHARAD and MARSIS radar soundings
134 have confirmed the existence of this sediment-rich BU, and have revealed that it is largely
135 confined to the major lobe of the North Polar Cap (Fig. 1) [Picardi et al., 2005; Phillips et al.,
136 2008; Putzig et al., 2009]. The second unit corresponds to the upper (and younger) part of the ice
137 cap, which is composed of the North Polar Layered Deposits (NPLD). On high-resolution
138 images, the NPLD appear brighter and more finely layered than the BU. The majority of the
139 NPLD is made of water ice and their layering results from varying fractions of included sediment
140 and/or varying ice grain sizes [Kieffer et al., 1976; Calvin et al., 2009]. Radar soundings have
141 confirmed that the amount of sediment in the NPLD is small, with only ~2% for most layers and
142 ~30% for a few strong radar reflective layers [Picardi et al., 2005; Phillips et al., 2008]. On the

143 basis of morphological and sedimentary interpretation of high-resolution images, Tanaka et al.
144 [2008] subdivided the BU and the NPLD into a full suite of stratigraphic units.

145 The surface morphology of the North Polar Cap is dominated by interior spiral troughs and
146 marginal steep arcuate scarps [Howard, 2000] (Fig. 1a and c). These spiral troughs and arcuate
147 scarps provide natural cross-sections where the internal structure of the North Polar Cap can be
148 observed. Spiral troughs only cut through the NPLD and do not reach the BU [Putzig et al.,
149 2009]. The most widely accepted hypothesis for their formation is ablation due to katabatic winds
150 and sublimation [Howard, 2000; Ivanov and Muhleman, 2000]. Katabatic winds and sublimation
151 are most probably also responsible for the formation of the marginal arcuate scarps [Warner and
152 Farmer, 2008]. The correlation between the location of the arcuate scarps and the spatial extent of
153 the BU has been attributed to preferential undermining of the scarp bases because of the presence
154 of sand in the BU [Howard, 2000; Edgett et al., 2003].

155 During Martian winters, a layer of CO₂ and H₂O frost, approximately one meter in thickness,
156 covers the North Polar Cap. This seasonal frost coverage begins to sublimate in spring and has
157 disappeared by the end of summer [Smith et al., 2001b; Byrne et al., 2008].

158 **2.2 Polar and circum-polar superficial accumulations of dark sediment**

159 In addition to sediment intermixed in the ice of the BU and of the NPLD, dark polar and
160 circum-polar sediment has also been observed in the form of (1) superficial accumulations on the
161 floors of some spiral troughs, (2) extensive superficial mantles (classically named dark veneers
162 [Rodriguez et al., 2007]) covering the surface of the NPLD, and (3) dunes located in the Circum-
163 Polar Dune Field [Rodriguez et al., 2007; Horgan et al., 2009]. The sources of the dark veneers
164 and of the sedimentary accumulations in spiral troughs are sediment-rich layers of the NPLD
165 [Rodriguez et al., 2007; Tanaka et al., 2008].

166 The Circum-Polar Dune Field is the largest dune field on Mars. It extends between 70°N and
167 85°N in latitude and entirely rings the North Polar Cap (Fig. 1b). Most of the dark dunes
168 constituting the Circum-Polar Dune Field have been classified as transverse and barchan dunes
169 [Tsoar et al., 1979]. Based on the shape of these dunes and on their systematic association with
170 arcuate scarps, Thomas and Weitz [1989] inferred that the source of the circum-polar dune
171 material is sand that was initially contained within the North Polar Cap. Herkenhoff and Vasada
172 [1999] thus conjectured that the dune material might be composed of filamentary sublimation
173 residue formed by concentration of dust in sand-size aggregates during sublimation of the ice cap.
174 Alternatively, it has been suggested that the major source for the circum-polar dune material is
175 sand derived from the BU rather than dust derived from the whole ice cap [Fishbaugh and Head,
176 2005; Herkenhoff et al., 2007].

177 **2.3 Sulfates in polar and circum-polar superficial sediment**

178 Spectroscopic studies have revealed unambiguous signatures of a calcium-rich hydrated
179 sulfate (most likely gypsum) on Olympia Planum (Fig. 1b) [Langevin et al., 2005b, Roach et al.,
180 2007]. In this portion of the Circum-Polar Dune Field, the density of dunes is at its highest
181 [Tanaka and Hayward, 2008]. In the inter-dune substrate of Olympia Planum, Roach et al. [2007]
182 detected gypsum signatures, which appear weaker than those found on dunes. These weak
183 signatures indicate either that small amounts of fine-grained gypsum are present within the bulk
184 of the inter-dune substrate or that a thin layer of gypsum-rich particles covers the inter-dune
185 substrate.

186 Poulet et al. [2005], Horgan et al. [2009] and Calvin et al. [2009] detected spectroscopic
187 signatures of hydrated minerals on the entire Circum-Polar Dune Field, on the NPLD dark
188 veneers and on the sedimentary accumulations covering spiral troughs. It is unclear however
189 whether these signatures correspond to gypsum, to another kind of hydrated sulfate or more

190 generally to any other hydrated minerals. If these signatures do correspond to gypsum, as they do
191 in Olympia Planum, their weakness has to be attributed either (1) to low gypsum concentrations,
192 (2) to scattering effects due to texture, or (3) to spatial exposures of gypsum being too small to be
193 resolved with the spatial resolution of the instrument [Calvin et al., 2009].

194 The darkness of the circum-polar sediments suggests that they are not pure gypsum. Poulet et
195 al. [2005] and Horgan and Bell [2009] have also identified signatures of mafic minerals, which
196 could explain their low albedo. Alternatively, Fishbaugh et al. [2007] suggested that secondary
197 oxides may darken these sediments.

198 **2.4 Currently proposed origins for circum-polar sulfates**

199 Because of its softness, gypsum is easily susceptible to physical weathering; therefore the
200 gypsum detected on Olympia Planum has probably formed in-situ or within a short distance from
201 its current location [Fishbaugh et al., 2007]. On this basis, Langevin et al. [2005b] suggested two
202 different hypotheses for the formation of gypsum in Olympia Planum: interaction of Ca-rich
203 minerals with snow containing H_2SO_4 derived from volcanic activity or formation as an evaporite
204 deposit after major meltwater outflows from the ice cap during warm climatic incursions.
205 Fishbaugh et al. [2007] suggested that water from nearby channels percolated through dunes that
206 cover the eastern end of Olympia Planum and attributed the formation of gypsum there to a
207 combination of (1) in-situ aqueous weathering of sulfide- and high-calcium-pyroxene-bearing
208 dune materials and (2) formation of evaporitic gypsum crystals in the pore spaces of these
209 materials. Szyrkiewicz et al. [2010] suggested that gypsum crystals were formed by evaporation
210 of saline waters and were later transported by winds towards Olympia Planum.

211 Alternatively, it has been suggested that gypsum minerals could derive directly from the
212 underlying BU [Roach et al., 2007; Calvin et al., 2009]; this interpretation would be consistent

213 with the possible existence of small amounts of gypsum in the inter-dune substrate [Roach et al.,
214 2007].

215 To further constrain the extent and the origin of polar and circum-polar sulfate deposits, we
216 focus on a key area which encompasses all the features described above (Fig. 2): (1) the surface
217 of the North Polar Cap, (2) two spiral troughs, (3) two marginal arcuate scarps providing natural
218 cross-sections through the NPLD and the BU and (4) a portion of the Circum-Polar Dune Field.
219 This area is located at the border of the North Polar Cap, between Olympia Cavi and Gemini
220 Scopuli, at the latitude of 83°N and the longitude of 118°E. The portion of the Circum-Polar
221 Dune Field comprised in this study area is located outside the gypsum-rich area previously
222 identified by Langevin et al. [2005b] (Fig. 1b), but spectroscopic signatures of hydrated minerals
223 have been detected there [Horgan et al., 2009; Calvin et al., 2009]. Based on a detailed
224 morphological and compositional analysis of this area, we will demonstrate that gypsum is
225 present in all kinds of polar and circum-polar superficial accumulations of dark sediment, and
226 that it derives from the ice cap.

227

228 **3. Structure and landforms of the study area**

229 **3.1. Data and Methods**

230 We investigate the structure and landforms of the study area with complementary data sets,
231 which provide different kinds of information. All these datasets have been incorporated into a
232 geographic information system using the Mars 2000 geographic coordinate system and the polar
233 stereographic projection.

234 Topographic information is provided by the Mars Orbiter Laser Altimeter (MOLA). The
235 selected polar digital terrain model has a relative vertical accuracy of 1 m and an average spatial
236 resolution of 512 pixel / degree [Smith et al., 2001a]. Geomorphological and structural

237 information is provided by (1) images of the Context Camera (CTX) with a spatial resolution of 6
238 m/pixel over a swath that is about 30 km wide [Malin et al., 2007], (2) images of the High
239 Resolution Imaging Science Experiment (HiRISE) with a resolution up to 30 cm/pixel for a
240 swath width of 6 km [McEwen et al., 2007]. Since we aim at observing sediment in the bulk and
241 at the surface of the permanent ice cap, we use CTX and HiRISE images acquired in summer
242 only, in order to minimize the effect of the seasonal CO₂ and H₂O frost coverage.

243 **3.2. Description of landforms in the study area**

244 A CTX image, an interpretative sketch-map and an interpretative cross-section of the whole
245 study area are shown in Fig. 3.

Figure 3

246 The surface of the NPLD occupies the northern half of the image. The outer reaches of two
247 spiral troughs strike NW-SE across the NPLD (Fig. 2 and 3). In the area encompassed by the
248 image, these troughs are 300 to 500 m deep. Unlike other spiral troughs observed elsewhere on
249 the North Polar Cap by Rodriguez et al. [2007] and Horgan et al. [2009], the troughs of the study
250 area are not covered by dark sediment. The internal layering of the NPLD is thus exposed in
251 these troughs. Differences in albedo between the exposed layers indicate either that they contain
252 differing amounts of sediment intermixed with the ice or that they have differing ice grain sizes
253 [Calvin et al., 2009]. Along the northern border of the northernmost trough, the surface of a
254 dissected packet of ice layers forms a tabular outcrop covered by dark sediment and bordered by
255 an escarpment 100 m-high (Fig. 3).

256 In the central-eastern and south-western parts of the image, two arcuate scarps, facing south,
257 cut through the NPLD (Fig. 2 and 3). The northernmost arcuate scarp is ~25 km wide and extends
258 beyond the eastern border of the image; its maximal height is 500 m. The southernmost one is
259 ~20 km wide and extends beyond the western border of the image; its maximal height is 300 m.

260 The average slope gradient is 30° in the steepest parts of both scarps. The two arcuate scarps
261 intersect, hence were probably carved later than, the spiral troughs.

262 These scarps provide vertical sections through the whole stack of NPLD present in this area,
263 and the BU that has been exhumed at their base (Fig. 3). The BU is distinguishable from the
264 NPLD by a break in topographic slope at the base of the scarps (Fig. 3c and d), by its darker tone
265 due to its larger concentration in sediment, and by the fact that the beds of the BU form resistant
266 shelves, and are organized in platy interbedded sequences of ice-rich and sediment-rich layers
267 (Fig. 3c). The fact that the BU crops out at the base of arcuate scarps in this region is consistent
268 with the extension of the BU as inferred from radar soundings [Putzig et al., 2009] (Fig. 1 and 2).

269 South of each scarp, the topographic surface dips gently northwards. In these latter regions,
270 superficial sediment and dunes of the Circum-Polar Dune Field cover the NPLD and the BU,
271 except for the southernmost part of the image where ice is visible (Fig. 3).

272 **3.3. Release of dark superficial sediment towards dune fields by horizontal ablation of ice** 273 **layers at arcuate scarps**

274 On Earth, the formation of steep arcuate marginal scarps is typical of those glaciers where
275 sublimation is the dominant process of ablation. These include equatorial glaciers such as those
276 located on the Kilimanjaro in Africa and polar glaciers such as those located in the Dry Valleys in
277 Antarctica. It has been shown that these steep terrestrial ice walls are erosional forms that move
278 backward by horizontal regressive ablation of the ice under the effect of radiant heating [Fountain
279 et al., 2006; Hoffman et al., 2008; Mölg and Hardy, 2004; Mölg et al., 2003, 2008]. We infer that
280 the steep arcuate scarps observed in the study area have formed by a similar process of horizontal
281 regressive ablation of the North Polar Cap. In the pressure and temperature conditions currently
282 prevailing at these latitudes on Mars, sublimation is the most probable process by which ablation
283 of ice and erosional retreat of the scarps may have occurred [Ivanov and Muhleman, 2000].

284 The substrate of the regions lying at the bottom of the two arcuate scarps comprises the BU
285 and the NPLD (Fig. 3a, b, d). This substrate is covered by a superficial mantle of dark sediment
286 and by dark dune fields. These are separated from the corresponding scarps by distances of 0 to 6
287 km (Fig. 3a and b). The dunes belong to the barchan or barchanoid types defined by McKee
288 [1979] and Hayward et al. [2007], which are indicative of unidirectional winds [Bagnold, 1954].
289 The orientation of the dunes indicates that the dominant wind blows down scarp from the NE
290 (Fig. 3a and b).

291 The northern border of each dune field comprises a belt of large barchans, which is
292 particularly well defined for the northernmost field (Fig. 3a and b). The presence of this upwind
293 belt is classical in terrestrial dune fields and can be explained by the fact that dunes of the first
294 row lose less sand than they gain and turn into mega dunes [Bagnold, 1954; Tsoar et al, 1979]. At
295 a distance of 0 to 900 m from their upwind border belts, both dune fields are constituted by
296 smaller, closely packed, barchanoids. At the down-wind portions of the fields, the end barchans
297 have a tendency to disperse. This organization is consistent with the wind blowing dominantly
298 from the NE down the surface of the North Polar Cap and down the scarps. Frost streaks at the
299 surface of the NPLD also indicate that the wind blows dominantly from the NE (Fig. 3a).

300 At the feet of both scarps, dark streamers striking NE are visible (Fig. 3a and c). These
301 features indicate that dark sediment that has been released from ice layers ablated by the retreat
302 of the scarps was accumulated at their feet in the form of dark sublimation tills covering the
303 NPLD and the BU. This sediment is then transported by the wind towards the dune fields.

304 All of our observations in this region are thus consistent with the systematic association,
305 elsewhere around the North Polar Cap, of dune fields with arcuate scarps and with the
306 corresponding interpretation that the material present in the Circum-Polar Dune Field derives
307 from the North Polar Cap [Tsoar et al., 1979; Thomas and Weitz, 1989; Howard, 2000; Warner

308 and Farmer, 2008]. We infer that this material (1) was released from the ice cap as the arcuate
309 scarps retreated by sublimation, (2) was then mobilized by katabatic winds descending along the
310 surface of the North Polar Cap and (3) was eventually deposited in dune fields around the ice cap.

311 **3.4. Release of dark sediment at the surface of the NPLD by vertical ablation of ice layers**

Figure 4

312 The northern (south-facing) slope of the northernmost spiral trough is interrupted by an
313 escarpment 100 m high (Fig. 3). We interpret this scalloped escarpment as the erosional front of a
314 dissected packet of ice layers (Fig. 4a and d). Above the escarpment is a 3 km wide topographic
315 plateau entirely covered by dark sediment and corresponding to an erosional surface that
316 developed at the expense of this packet of ice layers. Erosional remnants of the uppermost layers
317 of the packet appear in the form of linear tongues and ovoid tabular ridges striking East-West, a
318 few hundred meters in average width, that are entirely covered by dark sediment (Fig. 4a, b and
319 d). At the center of some of these ovoid ridges is a depression where the superficial dark
320 sediment is less abundant (Fig. 4a and b). These remnant buttes indicate that some ice layers,
321 which formerly covered the whole plateau, have now been extensively dissected by erosion.
322 Sublimation is the major process by which near complete ablation of these layers can have
323 occurred in this region [Ivanov and Muhleman, 2000]. This interpretation is supported by the
324 existence of similar landforms in terrestrial glaciers subjected to sublimation, such as on the
325 Kilimanjaro [Mölg and Hardy, 2004; Mölg et al., 2003, 2008].

326 Between the remnant buttes, the plateau has a specific roughness, composed of closely
327 spaced, regular, polygonal hollows (Fig. 4b). These hollows are 20 to 50 m in diameter and are
328 separated from each other by ridges that form a “honeycomb-like network” [Milkovich and Head,
329 2006]. Differences in albedo between the hollow centers and the boundary ridges reveal that the
330 sediment is denser on the ridges than in the hollows. This specific kind of surface texture is
331 common on terrestrial glaciers and snowfields (Fig. 4c). It is known as ablation hollows or

332 suncups [Rhodes et al., 1987; Betterton, 2001]. Terrestrial ablation hollows form through radiant
333 heating of the ice surface due to direct or indirect sunlight, which causes ablation of the ice by
334 melting or sublimation [Rhodes et al., 1987; Betterton, 2001; Milkovich and Head, 2006]. These
335 features grow because hollow centers receive more reflected light than ridges. As surface
336 lowering by ablation proceeds, any sediment particle that was initially present within the bulk of
337 the ice will move in a trajectory perpendicular to the surface and the sediment will thus
338 concentrate on the ridges [Rhodes et al, 1987]. Though they have not unambiguously observed
339 them, Milkovich and Head [2006] postulated that ablation hollows likely exist on the North Polar
340 Cap of Mars. On Earth, these features generally form by surface melting. By contrast, under the
341 pressure and temperature conditions prevailing at these latitudes on Mars, the development of
342 ablation hollows most probably involves sublimation [Ivanov and Muhleman, 2000; Milkovich
343 and Head, 2006]. The difference in size between terrestrial ablation hollows (typically 0.1 – 1 m
344 in diameter) and those described here (20 – 50 m) is attributable to either (1) differences in initial
345 sediment concentration within the ice, (2) differences in conditions of radiant heating, (3)
346 differences in the process of ablation (melting on Earth versus sublimation on Mars) or (4)
347 differences in the time available for their development (one season at most on Earth versus
348 several years at least on Mars). The effect of each these various parameters on the size of ablation
349 hollows is poorly known because of the lack of measurements and physical models on their
350 development. Betterton (2001) however demonstrated that the amount of sediment strongly
351 controls the size of suncups and it seems reasonable to assume that longer development times will
352 help create larger features.

353 Our observations on this dissected plateau thus demonstrate the presence of a superficial
354 accumulation of sediment on an ablation surface that developed at the expense of a packet of ice
355 layers. The concentration of this superficial sediment on ridges forming a honeycomb-like

356 network is best explained by a process, well known on terrestrial glaciers, according to which the
357 sediment that was initially present within the bulk of the ice has concentrated at the surface by
358 vertical ablation of the ice.

359 This interpretation is supported by the fact that this superficial accumulation of sediment can
360 be traced from place to place, along the same sub-horizontal ice layers, throughout the study area
361 (Fig. 3a, b and d). Had this sediment been imported from elsewhere and had it been deposited
362 unconformably from above after the formation of the ablation surface, its extent would not fit that
363 of the ice layers. We thus infer that the superficial sediment covering the plateau has been
364 released from the eroding packet of ice layers itself (Fig. 3d); it has concentrated in the form of a
365 sublimation till on the surface of the ice cap, as ablation by sublimation dissected this sediment-
366 rich packet of ice layers.

367 **3.5. Synthesis: origin of polar and circum-polar accumulations of dark sediment**

368 We have shown that in the region under study, dune fields are fed by sediment released from
369 arcuate scarps that develop by regressive ablation of the North Polar Cap. This sediment may
370 derive either from the BU exhumed at the base of the arcuate scarps or from the NPLD that
371 covered the area before erosional retreat of the arcuate scarps. We have also shown evidence that
372 sediment accumulations at the surface of the NPLD are released by vertical ablation of sediment-
373 rich ice layers of the NPLD.

374 As a conclusion, the superficial sediment present on the NPLD and in the Circum-Polar Dune
375 Field derives from the North Polar Cap. This sediment, which was initially intermixed within the
376 bulk of the ice, now forms ablation tills that have concentrated at the ice surface by sublimation.
377 Katabatic winds are responsible for its re-mobilization and its deposition in the Circum-Polar
378 Dune Field. Previous observations elsewhere on the North Polar Cap are consistent with this
379 interpretation [Tsoar et al., 1979; Thomas and Weitz, 1989; Herkenhoff and Vasada, 1999;

380 Howard, 2000; Fishbaugh and Head, 2005; Rodriguez et al., 2007; Tanaka et al., 2008; Warner
381 and Farmer, 2008].

382

383 **4. Mineralogical analysis**

384 Two different hypotheses remain for the origin of gypsum crystals that have been observed in
385 Olympia Planum and that are possibly present in the Circum-Polar Dune Field as a whole.
386 Similarly to the dark sediment described in part 3, gypsum crystals may have been initially
387 present within the bulk of the ice of the North Polar Cap, may have been released to the surface
388 by ablation, and transported towards dunes [Calvin et al., 2009]. Alternatively, they may have
389 formed as secondary minerals in dunes, at the expense or in the pore spaces of the dark sediment
390 derived from the North Polar Cap [Langevin et al., 2005b; Fishbaugh et al., 2007]. To evaluate
391 the validity of the first hypothesis, we must be able to detect the presence of gypsum in the North
392 Polar Cap from orbital hyperspectral data. For that purpose, we have first investigated the
393 spectral behavior of experimental ice-gypsum mixtures in Martian pressure and temperature
394 conditions. The results of this laboratory study were then used to derive the composition of
395 sediments associated with the North Polar Cap from the available orbital hyperspectral data.

396 **4.1. Spectral behavior of experimental ice-gypsum mixtures**

397 **4.1.1. Experimental procedure**

398 In order to investigate the spectral behavior of ice-gypsum mixtures, we have acquired
399 laboratory reflectance spectra of ice samples containing various volumetric proportions of
400 gypsum. The ice was produced by condensation of liquid water on a cooled metal plate [Dupire et
401 al., 2009]. It was then crushed to obtain grain sizes between 500 and 600 μm . These ice grains
402 were homogeneously mixed with an industrial pure gypsum powder (producer: Merck Chemicals;
403 reference: 102160, commercial name: calcium sulfate dihydrate precipitated, chemical formula:

404 $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$, grain size: less than $10\mu\text{m}$). The samples were then placed in a liquid nitrogen
405 cryostat (MicrostatN, Oxford Instruments) to reproduce the Martian pressure and temperature
406 conditions (7 mbar, 150 K) and their spectra were acquired with a Nicolet 5700 Fourier
407 Transform Infrared Spectrometer, which collects 4149 spectral channels from 1 to $5\mu\text{m}$.

408 4.1.2. Results

Figure 5

409 Fig. 5 displays the laboratory spectra obtained for pure water ice, pure gypsum, a mixture
410 composed of ~50% ice and ~50% gypsum in volume and a mixture composed of ~30% ice and
411 ~70% gypsum in volume. Pure water ice displays specific absorption bands at $1.04\mu\text{m}$ and 1.25
412 μm , two broad absorption bands between 1.50 and $1.66\mu\text{m}$ and between 1.96 and $2.05\mu\text{m}$ and,
413 one band at $2.55\mu\text{m}$. Pure gypsum exhibits one absorption band centered at $1.20\mu\text{m}$ (due to H_2O
414 combinations), a triplet of absorption bands of progressively decreasing intensity at 1.44 , 1.49
415 and $1.53\mu\text{m}$ (due to O-H stretches), and a band at $1.74\mu\text{m}$ (due to an OH combination), a double
416 band near 1.94 and $1.97\mu\text{m}$ (due to H_2O combinations), a broad band centered at $2.2\mu\text{m}$
417 constituted of two narrower components centered at 2.21 and $2.27\mu\text{m}$ (due to H_2O combinations
418 and/or S-O stretching overtones), and a band centered at $2.48\mu\text{m}$ (due to S-O stretching
419 combinations) [Cloutis et al., 2006; 2008]. When gypsum is mixed with water ice, the spectral
420 behavior of the mixture may be described as follows.

421 - The gypsum absorption band at $1.20\mu\text{m}$ is masked by the ice band at $1.25\mu\text{m}$ even at high
422 gypsum concentrations. Therefore the band at $1.20\mu\text{m}$ cannot be used to detect gypsum when it
423 is mixed with ice.

424 - Gypsum absorption bands remain visible at 1.44 , 1.49 and $1.53\mu\text{m}$. However, the bands at
425 1.49 and $1.53\mu\text{m}$ are small and are close to the center of the broad ice band at $1.50\mu\text{m}$.
426 Therefore these two bands cannot be reliably discriminated from the $1.50\mu\text{m}$ ice band. The 1.44

427 μm absorption band is sufficiently far from the center of the $1.50 \mu\text{m}$ ice band, that it remains
428 visible at high gypsum concentrations and disappears progressively with increasing ice content.

429 - The gypsum absorption band at $1.74 \mu\text{m}$ turns into an inflexion at high gypsum
430 concentrations and disappears at high ice concentrations.

431 - The 1.94 and $1.97 \mu\text{m}$ gypsum absorption bands remain visible even at high ice
432 concentrations distorting the $2 \mu\text{m}$ water ice band.

433 - The broad gypsum absorption band at $2.2 \mu\text{m}$ disappears and only its two narrow
434 components at 2.21 and $2.27 \mu\text{m}$ remain on the wings of the ice peak centered at $2.24 \mu\text{m}$. At low
435 gypsum concentrations, these two bands turn into residual inflexions on both sides of the ice
436 peak.

437 - The $2.48 \mu\text{m}$ gypsum absorption band is close to the center of the broader $2.55 \mu\text{m}$ water ice
438 band, and is therefore difficult to isolate.

439 To summarize, laboratory spectra show that, when gypsum is mixed with water ice,
440 diagnostic gypsum absorption bands remain at 1.44 , 1.74 , 1.94 , 2.21 , 2.27 and $2.48 \mu\text{m}$, but all of
441 these bands are very shallow and may be overlapped by those of water ice. These bands are thus
442 predictably difficult to detect in spectra acquired remotely of ice-rich regions on natural planetary
443 surfaces. In addition, currently available Martian data are noisier than laboratory spectra and their
444 spectral resolution is lower. We have therefore developed specific spectral processing methods to
445 detect and to isolate these shallow diagnostic bands on Martian hyperspectral data.

446 **4.2. Analysis of hyperspectral data in the study area**

447 **4.2.1. Data**

448 The mineralogical composition of the study area was investigated from data acquired by the
449 OMEGA and CRISM imaging spectrometers. We selected data acquired in summer (OMEGA

450 cube ORB1056_2 and CRISM cube hrl0000330c) to minimize the contribution of the seasonal
451 CO₂ and H₂O frost coverage.

452 OMEGA (Observatoire pour la Mineralogie, l'Eau, les Glaces et l'Activite) [Bibring et al.,
453 2004], onboard Mars Express, acquires hyperspectral images at a spatial resolution ranging from
454 300 m to 4.8 km per pixel. A spectrum is acquired in 352 spectral channels from 0.38 to 5.2 μm
455 for each pixel of an image, thus producing data cubes. CRISM (Compact Reconnaissance
456 Imaging Spectrometer for Mars) [Murchie et al., 2007] is onboard Mars Reconnaissance Orbiter
457 (MRO). In the targeted hyperspectral mode, CRISM collects 544 spectral channels from 0.36 to
458 3.9 μm at a spatial resolution ranging from 15 to 19 m/pixel.

459 We restricted our analysis to the spectral domain comprised between 1.0 and 2.5 μm . In this
460 wavelength range, the solar reflected light dominates the spectrum, and the thermal emission is
461 negligible [Gendrin et al., 2005]. This range is also suitable for the detection of hydrated minerals
462 commonly identified on Mars, and is particularly diagnostic for gypsum. In order to avoid the
463 effect of the CRISM smile [Murchie et al., 2007], we studied only the center of the image.

464 **4.2.2. Extraction of the spectral information**

465 **4.2.2.1. Data reduction**

466 OMEGA and CRISM spectra are acquired remotely through the atmosphere. In order to
467 extract the spectral contribution of the surface only, the atmospheric spectral contribution is
468 removed by using an empirical atmospheric transmission law derived from the ratio between two
469 spectra acquired at the summit and the base of the Olympus Mons volcano, and scaled to the
470 depth of the CO₂ band [Langevin et al., 2005a; McGuire et al., 2009]. The CRISM Analysis
471 Toolkit (CAT) also corrects the photometric angles [Murchie et al., 2007]. Custom software
472 routines are used to georeference the OMEGA and CRISM images in the Mars 2000 coordinate
473 system with a polar stereographic projection.

474 **4.2.2.2. Denoising**

475 The experimental study of the spectral behavior of gypsum-ice mixtures described in part 4.1
 476 suggests that most bands diagnostic of gypsum are close to the noise level in unprocessed
 477 Martian data. Different kinds of denoising procedures have thus been applied to CRISM and
 478 OMEGA data.

479 The CIRRUS tool, available in the CAT, was applied to the CRISM cube. CIRRUS first
 480 removes isolated noise spikes with the “despiking” tool and then corrects the column bias with
 481 the “destriping” tool [Parente, 2008]. We have also performed a Minimum Noise Fraction (MNF)
 482 transform on OMEGA and CRISM data. This procedure, available in the ENVI software,
 483 segregates the noise from the information in the data. An inverse MNF transform computed only
 484 on the components containing the information can be used to decrease the noise level in the data
 485 cube [Green et al., 1988].

486 **4.2.2.3. Definition of spectral criteria**

487 To identify spectral features (e.g. absorption bands, spectral slopes) that are diagnostic of
 488 minerals, we have computed spectral criteria. These correspond to combinations of reflectance
 489 measured at different wavelengths. In order to further decrease the contribution of noise when
 490 computing these spectral criteria, the reflectance at a given wavelength was taken as the median
 491 of the values of reflectance measured in three adjacent instrumental channels centered on this
 492 wavelength. Spectral criteria were computed for each pixel of a cube, and maps of each spectral
 493 criterion were then produced.

494 We mapped the distribution of water ice with the calculation of the 1.5 μm absorption band
 495 depth. This criterion is defined as follows:

$$496 \quad BD(1.50) = 1 - \frac{R(1.50)}{(0.7) * R(1.37) + (0.3) * R(1.82)}$$

497 where $R(x)$ is the value of reflectance corresponding to the wavelength at $x \mu\text{m}$.

498 On Mars, hydrated minerals such as gypsum have classically been identified with spectral
 499 criteria based on the depth of the 1.4 and 1.9 μm absorption bands [e.g. Pelkey et al., 2007;
 500 Massé et al., 2008]. However, these criteria cannot be used to detect hydrated minerals when they
 501 are mixed with ice because the broad 1.5 and 2.0 μm water ice bands overlap the 1.4 and 1.9 μm
 502 mineral hydration bands. Therefore, we used the criterion defined by Horgan et al. [2009] to
 503 isolate the 1.9 μm hydration band. This criterion is based on the assumption that the water ice
 504 band at 2.0 μm is symmetrical and that the addition of hydrated minerals imposes a slight
 505 asymmetry to this band. This criterion is defined as follows:

$$506 \quad BD(1.934) = 1 - CR_{1934} / CR_{2108}$$

$$507 \quad CR(x) = \frac{R(x)}{\left(\left[\frac{R(2.205) - R(1.842)}{2.205 - 1.842} \right] (x - 1.842) + R(1.842) \right)}$$

508
 509 The same method cannot be used to isolate the 1.4 μm hydration band from the 1.5 μm water
 510 ice band because the broad water ice band centered at 1.5 μm is not symmetrical. Similarly, the
 511 2.21 and 2.27 μm gypsum absorption bands interfere with the ice peak centered at 2.24 μm . In
 512 addition, the 1.74, 2.21 and 2.27 μm gypsum bands appear as inflexions rather than as deep bands
 513 when gypsum is mixed with ice. Therefore all these bands, which are diagnostic of gypsum,
 514 cannot be detected by computing simple spectral criteria. Hence, we have therefore used a
 515 complementary method to monitor these specific shallows bands.

516 4.2.2.4. Spectral derivative method

517 The spectral derivative method has been initially developed to analyze terrestrial
 518 hyperspectral data. It allows one to determine the wavelength position of narrow bands and to

Figure 6

519 resolve overlapping absorption bands [Huguenin and Jones, 1986; Talsky, 1994; Tsai and
520 Philpot, 1998; Louchard et al., 2002, Verpoorter et al., 2007, Verpoorter, 2009; Verpoorter et al.,
521 2010]. This method has the advantage of being less dependent on the shape of the continuum and
522 thus removes background signals caused, for example, by differences in grain sizes or variations
523 in topography and illumination conditions. The method is based on the principle that absorption
524 bands constitute local minima in the spectrum (Fig. 6). They will therefore appear as zeros in the
525 first order derivative of the spectrum and as maxima in its second order derivative. On the other
526 hand, peaks between absorption bands constitute local maxima in the spectrum. They will
527 therefore appear as zeros in the first order derivative and as minima in the second order derivative
528 (Fig. 6). Inflexions will appear as local maxima or minima in the first order derivative and as
529 zeros in the second order derivative (Fig. 6). By the systematic detection of local minima, local
530 maxima and inflexions in a reflectance spectrum, the derivative method thus allows the
531 identification of the exact center of absorption bands, even for shallow or overlapping ones, and
532 independently from background signals. This technique enhances subtle fluctuations in
533 reflectance spectra and separates closely related absorption features.

534 We have adapted this method to detect all local minima in hyperspectral cubes and to produce
535 maps of their geographical distribution [Verpoorter, 2010]. The algorithm first detects the
536 wavelengths of all local minima in the spectra of all pixels of a cube. Then it produces, for each
537 identified wavelength, a map of the distribution of pixels where this wavelength corresponds to a
538 local minimum. Finally, a density map is produced for each identified wavelength by counting, in
539 a moving kernel (3x3), the number of pixels that display a local minimum at this wavelength.

540 **4.3. Results**

541 The histogram in Fig. 7 shows the frequency of all absorption bands detected by the spectral
542 derivative method in the whole CRISM cube. Fig. 8 displays representative CRISM spectra of the

543 morphological units described in part 3. Maps of spectral criteria and maps of absorption bands
544 detected by the spectral derivative method are given in Fig. 9 and 10. Maps of spectral criteria
545 provide quantitative information such as band depths, whereas maps derived from the spectral
546 derivative method show the distribution of pixels where an absorption band is present at a given
547 wavelength. The spectral derivative method enhances all the spectral fluctuations and is thus very
548 sensitive. Therefore, overlapping absorption bands and weak spectral features that would not be
549 detected by computing spectral criteria, are readily identified by the spectral derivative method.
550 On the other hand, strong denoising must be performed before applying the spectral derivative
551 method. In some cases, this denoising can lead to the loss of some of the faintest absorption
552 bands. Our interpretation is therefore based on the joint analysis of both kinds of maps. To finally
553 ensure that a given local minimum detected by the spectral derivative method corresponds to an
554 absorption band rather than to residual noise, we check that pixels displaying this local minimum
555 define a spatially and geologically consistent area.

556 **4.3.1. Absorption bands detected in the study area**

Figure 7

557 Fig. 7 shows the distribution of absorption bands that more than 50% of the pixels in the
558 image display absorption bands centered at 1.50, 1.96 and 2.06 μm . Between 20% and 50% of
559 the pixels display absorption bands at 1.07, 1.22, 1.37, 1.92, 2.14, 2.21 and 2.27 μm . Absorption
560 bands are also detected at 1.30, 1.42, 1.57, 1.64, 1.73, 1.79, 1.86, 2.00 and 2.48 μm on less than
561 20% of the pixels.

562 **4.3.2. Mineralogical interpretation**

Figure 8
Figure 9

563 The absorption bands detected at 1.07, 1.22, 1.30, 1.50, 1.64, 1.96, 2.00 and 2.06 μm are
564 consistent with those of water ice. However, the absorption bands at 1.96, 2.00 and 2.06 μm may
565 be artifacts due to the fact that the CRISM atmospheric correction fails between 1.97 and 2.08
566 μm . The absorption bands at 1.22 and 1.64 μm could also correspond to CRISM artifacts and the

567 absorption bands at 1.07 and 1.30 μm could be attributed to other components. We have thus
568 only used the 1.50 μm absorption band to discriminate the water ice signature. This water ice
569 absorption band appears most distinctly on clean ice layers of the NPLD (Fig. 8, 9a and 9d). The
570 1.50 μm water ice absorption band is also shallower on sediment-rich ice layers of the NPLD and
571 on the polar and circum-polar superficial accumulations of dark sediment (Fig. 8, 9a and 9d).
572 These include the sublimation till covering the surface below arcuate scarps, the arcuate scarps
573 themselves, the sublimation till covering the dissected packet of NPLD ice layers described in
574 part 3.4 and the dunes.

575 The absorption bands detected at 1.42, 1.73, 1.92, 2.21, 2.27 and 2.48 μm are consistent with
576 those of gypsum mixed with ice, as we have identified them experimentally (Fig. 5). These
577 gypsum absorption bands are spatially anti-correlated to those of water ice (Fig. 9). By contrast,
578 they are correlated with the polar and circum-polar superficial accumulations of dark sediment
579 (Fig. 8, 9b, 9c, 9e, 9f, 9g, 9h, 9i, 9j and 10). More specifically, they are present on the
580 sublimation till covering the BU and the NPLD below arcuate scarps (Fig. 8, 10e and 10f), on the
581 dunes (Fig. 8, 10e and 10f), on the dark streamers (Fig. 8, 10e and 10f), on the sediment-rich ice
582 layers of the arcuate scarps (Fig. 8, 10e and 10f), and on the sublimation till covering the
583 dissected packet of NPLD ice layers described in part 3.4 (Fig. 8, 10b and 10c).

584 The 1.42, 1.92, 2.21 and 2.27 μm gypsum absorption bands are well expressed on all of these
585 superficial sedimentary accumulations (Fig. 9c, 9f, 9h and 9i) (corresponding to more than 20%
586 of the pixels in the image, Fig. 7), while the 1.73 μm gypsum absorption band appears
587 dominantly on sediment-rich layers of the arcuate scarps and on dune crests, and the 2.48 μm
588 gypsum absorption bands appear dominantly on dune crests (Fig. 9e and 9j). As we have shown
589 experimentally, this difference can be attributed to the fact that the gypsum/ice mixing ratio is
590 higher in the dunes than in the sublimation tills. This interpretation is consistent with the

591 observation in Olympia Planum that the strongest gypsum signature is located on dune crests
592 rather than on interdunes [Roach et al., 2007; Calvin et al., 2009].

593 An absorption band centered at 2.14 μm is present on 30 % of the pixels in the image and is
594 well visible both on raw and denoised CRISM spectral data (Fig. 7 and 8). This band is spatially
595 correlated with the gypsum absorption bands and is characteristic of the dark polar and circum-
596 polar sediment (Fig. 8 and 9g). Few common terrestrial minerals display an absorption band at
597 this wavelength. This band could correspond to perchlorate, a mineral uncommon on Earth but
598 that has been detected in the Martian permafrost at the Phoenix landing site [Hecht et al., 2009]
599 and that displays a deep diagnostic absorption band at this wavelength [Hanley et al., 2009;
600 Hanley et al., 2010; Morris et al., 2009]. It is interesting to notice that the possible presence of
601 perchlorate could have some important implications on the flow of the North Polar Cap [Fisher et
602 al., 2010]. As an alternative to perchlorate, the sulfite hannebachite ($2\text{CaSO}_3 \cdot \text{H}_2\text{O}$) has also a
603 strong absorption band at 2.14 μm . As perchlorate has already been detected on Mars by the
604 Phoenix lander, we favor perchlorate to interpret the absorption band at 2.14 μm . However, the
605 presence of hannebachite cannot be ruled out since it would be consistent with the chemical
606 model of Halevy and Schrag [2009] showing that SO_2 on Mars prevents the formation of calcium
607 carbonate in favor of this hydrated calcium sulfite.

608 Interestingly, gypsum and possibly perchlorate are present also in the bulk of sediment-rich
609 ice layers of the NPLD. Absorption bands at 1.93, 2.21 (gypsum) and 2.14 μm (perchlorate?) are
610 visible directly on spectra acquired on sediment-rich ice layers (Fig. 8) and spectral criterion
611 $\text{BD}(1.934)$ computed on raw data recognizes the 1.93 μm band on all the sediment-rich ice layers
612 (Fig. 9b). With the spectral derivative method applied on denoised data, we detect the absorption
613 bands at 1.93, 2.14 and 2.21 μm only in the most sediment-rich ice layers close to the dissected

614 packet of NPLD (Fig. 9f, 10b and 10c), because these very shallow, narrow bands are easily
615 erased by the the denoising procedure.

616 Other local minima detected by the spectral derivative method at 1.37, 1.57, 1.79 and 1.86
617 μm are more difficult to interpret. They might correspond to unidentified minerals or to spectral
618 noise (Fig. 7).

619 **4.3.3. Regional validation with OMEGA data**

Figure 11
Figure 12

620 In order to validate the absorption bands detected in the study area and to check the
621 consistency of these detections over a wider portion of the North Polar Cap, we applied the same
622 processing methods to OMEGA data.

623 Spectra acquired on dunes and on sublimation tills covering the North Polar Cap display the
624 same absorption bands, both on CRISM and OMEGA data (Fig. 11). This demonstrates that they
625 have similar mineralogical compositions. The most subtle bands barely show up on OMEGA data
626 however, due to the lower spatial resolution of the instrument, which induces higher spatial
627 mixing. On a regional scale, absorption bands at 1.42, 1.92, 2.21 and 2.27 μm , which are
628 diagnostic of gypsum, and an absorption band at 2.14 μm , which might be attributed to
629 perchlorate, are found to be systematically correlated to sediment-rich areas (Fig. 12b, 12d, 12e,
630 12f and 12g) and anti-correlated with water ice-rich areas (Fig. 12a and 12c). These include the
631 whole portion of the Circum-Polar Dune Field enclosed in the OMEGA cube and a significant
632 number of pixels on the North Polar Cap, corresponding to sublimation tills, to arcuate scarps and
633 to spiral troughs covered by superficial sediment (Fig. 12b, 12d, 12e, 12f and 12g).

634

635 **5. Discussion**

636 The mineralogical interpretation of CRISM and OMEGA hyperspectral data reveals that all
637 kinds of polar and circum-polar superficial accumulations of dark sediment present in the study

638 area (including sublimation tills at the surface of the North Polar Cap and dunes in the Circum-
639 Polar Dune Field) display similar absorption bands (Fig. 8, 9, 10, 11 and 12). This indicates that
640 they have similar mineralogical compositions. They are a mixture of ice and various minerals
641 including gypsum and possibly perchlorate. Sediment with a similar composition is intermixed in
642 the ice of the North Polar Cap. The interpretation of landforms in the study area reveals that the
643 gypsum-bearing sediment present at the surface of the ice cap and in the Circum-Polar Dune
644 Field was released from the ice cap by sublimation. We infer that gypsum crystals that are now
645 present in the Circum-Polar Dune Field derive also from the North Polar Cap (Fig. 13). Dunes are
646 necessarily composed of grains with a size of a few hundreds of microns [Bagnold, 1954]. This
647 does not necessarily mean that the gypsum crystals derived from the ice cap must be so large. As
648 suggested by Herkenhoff and Vasada [1999], the grains in the dunes may be composed of smaller
649 gypsum crystals aggregated with ice and other minerals.

Figure 13

650 Two hypotheses remain for the ultimate origin of the gypsum crystals in the ice cap. (1) Pre-
651 existing gypsum crystals might have been deposited together with ice crystals during the
652 formation of the ice cap or (2) authigenic gypsum crystals might have grown within the ice cap
653 by weathering of sediment trapped in the ice. We have no means to answer this question directly
654 from currently available observations on Mars, but comparisons with terrestrial analogues might
655 be meaningful.

656 Sulfate salt inclusions have been found in Greenland and Antarctica ice cores, with gypsum
657 being the dominant salt species in layers deposited during glacial stages and Na- and/or Mg-
658 sulfates being dominant in layers deposited during interglacial stages [Ohno et al., 2006]. The
659 amount of each salt species in the ice depends on the ion balance and the priority sequence of
660 chemical reactions. This priority sequence has been identified to be (1) calcium sulfate, (2) other
661 sulfates, (3) nitrates, (4) chlorides and (5) carbonates [Iizuka et al., 2008]. Sulfate crystals, and

662 particularly gypsum, are thus common in terrestrial glaciers. On Mars, the sequence of salts
663 formation may be slightly different, but Tosca et al. [2006] demonstrated that gypsum is the first
664 salt to form when SO_4^{2-} is sufficiently abundant.

665 Sulfate crystals found in terrestrial glaciers can form in the atmosphere, prior to their
666 deposition, by the neutralization of volcanic H_2SO_4 on dust containing Ca^{2+} , Na^+ or Mg^{2+} [Iizuka
667 et al. 2006 and 2008]. Postdepositional processes also can affect the abundances and forms of ion
668 species. These postdepositional processes include: (1) reworking and mixing of the surface snow
669 by winds, (2) sublimation, condensation and volatilization in surface snow and (3) molecular
670 diffusion in firn or ice. By this way, sulfate crystals can also form directly in the ice when liquid-
671 phase SO_4^{2-} diffuses to relatively immobile Ca^{2+} , Na^+ or Mg^{2+} [Iizuka et al., 2006; 2008].

672 To conclude, sulfate crystals that are present in the Earth's closest cold-desert analogs for the
673 Martian North Polar Cap, have probably formed both in the atmosphere and in the ice [Iizuka et
674 al., 2006 and 2008]. By the same way, gypsum crystals that are present in the North Polar Cap of
675 Mars may be both: (1) pre-existing gypsum crystals that have formed in the atmosphere and have
676 been deposited together with ice crystals during the formation of the ice cap and (2) authigenic
677 gypsum crystals that have formed in the ice cap by post-depositional processes.

678 A genetic link between sulfates and ice bodies has been suggested previously to explain the
679 formation of ancient sulfate deposits in equatorial regions of Mars [Niles and Michalski, 2009].
680 These authors argue that ice bodies with intermixed sediment, similar to the Polar Layered
681 Deposits that are currently present in the North and South Polar Caps, have formed in equatorial
682 regions of Mars during former periods of high obliquity or polar wander. Within these massive
683 ice deposits, acid weathering of the intermixed sediment would have led to the formation of
684 sulfates. Later, aeolian reworking of the sublimation residue of this mixture of ice and sulfates
685 would have formed the equatorial sulfate-bearing sediments. This model is consistent with recent

686 results of thermodynamical models, which indicate that low-temperature acid weathering of
687 basaltic dust with small amounts of liquid water can lead to the formation of sulfates in a few
688 decades only [Berger et al., 2009]. According to Niles and Michalski [2009], their ice-weathering
689 model resolves many chemical and morphological problems highlighted by previous hypotheses
690 concerning the formation of equatorial sulfates on the Martian surface. In equatorial regions, the
691 ancient accumulations of intermixed ice and sediment required by the model have disappeared.
692 Therefore, the ice-weathering model is difficult to evaluate directly in these regions. Our results
693 however demonstrate that recent polar and circum-polar sulfate deposits derive from the North
694 Polar Cap and thus support the possibility that similar processes may have occurred at lower
695 latitudes in the past.

696

697

698 **6. Conclusion**

699 The development of a spectral processing method based on spectral derivation allows
700 resolving overlapping absorption bands between water ice and sediment in the North Polar Cap
701 and in the Circum-Polar Dune Field. We find that the hydrated minerals detected by Horgan et al.
702 [2009] and Calvin et al. [2009] in the study area correspond to gypsum and possibly perchlorate.
703 Therefore, the presence of gypsum is not restricted to the Olympia Planum dunes but can
704 probably be extended to the whole Circumpolar Dune Field. Gypsum is also found on all the
705 superficial sediment present in the study area including: sublimation tills at the surface of the
706 NPLD and the BU, sediment-rich ice layers and dark streamers released from the polar cap. We
707 thus deduce that circumpolar gypsum has not formed by in-situ weathering of the dunes but was
708 initially present in the North Polar Cap and has been released to the surface by the ablation of the
709 ice. This morphological and mineralogical study also shows that sediment constituting the

710 circumpolar dunes originates both from the NPLD and the BU. The proportion of sediment
711 intermixed with ice is smaller in the NPLD than in the BU [Picardi et al., 2005; Phillips et al.,
712 2008], therefore the majority of the dune material is probably provided by the BU.

713 According to these results we propose the following scenario for the formation of polar and
714 circum-polar sulfate-bearing deposits on Mars (Fig. 13).

- 715 1. Gypsum crystals form in the atmosphere and/or in the ice cap by neutralization of
716 volcanic H₂SO₄ on Ca-bearing sediment.
- 717 2. These gypsum crystals are trapped in the ice.
- 718 3. Erosion of the ice cap by winds and sublimation leads to the formation of spiral troughs,
719 arcuate scarps and other ablation landforms.
- 720 4. As erosion proceeds, gypsum crystals are released from the ice and they concentrate in
721 sublimation tills at the surface of the ice cap.
- 722 5. Gypsum-bearing sublimation tills are reworked by winds and transported towards the
723 Circum-Polar Dune Field.

724 This scenario is consistent with the observation of sulfate crystals, and particularly gypsum,
725 in Greenland and Antarctic ices cores, which form both by predepositional processes in the
726 atmosphere and postdepositional processes in the ice [Ohno et al., 2006; Iizuka et al., 2006 and
727 2008]. It may also provides a basis, derived from the analysis of processes recently active in the
728 polar regions of Mars, to constrain the processes that were responsible for the formation of
729 sulfate-bearing deposits at lower latitudes in the past.

730

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739

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1038 **Figure captions**

1039 **Figure 1:** a) Topographic map of the North Polar Cap (shaded and colored relief image computed
1040 from the MOLA DEM at ~512 m/pixel overlain on a MOC mosaic at ~221 m/pixel). The orange
1041 dotted line represents the extension of the Basal Unit (BU) inferred from radar soundings by
1042 Putzig et al. [2009]. b) Map of gypsum concentration in the Circum-Polar Dune Field, computed
1043 from 11 OMEGA observations by Langevin et al. [2005b]. The North Polar Cap appears in white
1044 and grey, whereas the Circum-Polar Dune Field appears in black. Colors indicate the depth of the
1045 1.927 μm absorption band, interpreted as a proxy for the distribution of gypsum. c) Interpretative
1046 cross-section of the North Polar Cap based on SHARAD radargrams [Putzig et al., 2009]. The
1047 dotted line indicates the top of the Vastitas Borealis Formation (VBF), as interpreted from scarce
1048 radar returns.

1049
1050 **Figure 2:** Context map of the study area (shaded and colored relief image computed from the
1051 MOLA DEM at ~512 m/pixel overlain on a CTX mosaic; location indicated by a grey box in
1052 Figure 1). Topographic contours drawn at 200 m intervals underline the interior spiral troughs
1053 and the marginal arcuate scarps. The Circum-Polar Dune Field appears as a low-albedo unit,
1054 whereas ice appears as light outcrops. The study area (red box) encompasses the surface of the
1055 North Polar Cap, two spiral troughs, two arcuate scarps and a portion of the Circum-Polar Dune
1056 Field. The orange dotted line represents the extension of the Basal Unit (BU) inferred from radar
1057 soundings by Putzig et al. [2009].

1058
1059 **Figure 3:** Structure and morphology of the study area. a) CTX image
1060 P01_001593_2635_XI_83N241W (location indicated by a red box on Figure 2). b) Interpretative

1061 sketch-map of Figure 3a. The interval of MOLA elevation contours is 100m. c) Close-up view of
1062 the northernmost arcuate scarp, showing the NPLD in its upper part, the BU in its lower part and
1063 superficial dark streamers extending from its base towards the associated dune field (location
1064 indicated by a black box on Figure 3a). d) Interpretative cross-section of the study area (location
1065 on Figure 3a and b). The topographic profile is based on MOLA DTM.

1066
1067 **Figure 4:** a) Sublimation landforms observed on a dissected packet of sediment-rich ice layers in
1068 the NPLD (Portion of HiRISE image PSP_009267_2640, location indicated by a black box in
1069 Figure 3a). The extent of the superficial ablation till released by sublimation of the packet of ice
1070 layers is delineated by the white dotted line. b) Close-up view on sublimation till and polygonal
1071 ablation hollows at the surface of the packet of ice layers (location indicated by the white box in
1072 Figure 4a). c) Example of terrestrial ablation hollows with concentration of ablation till on ridges
1073 at the surface of the Vatnajökull glacier in Iceland (20 cm long hammer for scale) (photo M.
1074 Massé). d) Interpretative block diagram illustrating the internal structure of the packet of ice
1075 layers and the surface sublimation landforms produced by downward ablation of the ice layers.

1076
1077
1078 **Figure 5:** Laboratory spectra of experimental water ice - gypsum mixtures acquired at 150 K and
1079 7 mbar. Grey and black dotted vertical bars underline spectral features attributed to water ice and
1080 gypsum respectively.

1081
1082 **Figure 6:** First order ($R'(x)$) and second order ($R''(x)$) derivatives of a portion of a CRISM
1083 spectrum ($R(x)$).

1084

1085 **Figure 7:** Distribution of absorption bands detected by the spectral derivative method in the
1086 CRISM cube hrl0000330c. The most geologically plausible mineralogical interpretation is
1087 indicated above each frequency peak.

1088
1089 **Figure 8:** Representative spectra of various morphological units of the study area. The central
1090 image is a color composition (R: 2.5295 μm , V: 1.5066 μm , B: 1.0800 μm) of the study area
1091 computed from CRISM cube hrl0000330c (location indicated by a grey box on Figure 2). Clean
1092 ice layers appear in blue, sediment-rich ice layers appear in white, superficial sublimation tills
1093 appear in various tones ranging from pale to medium brown and dunes appear in dark brown.
1094 Raw (grey) and denoised (black) spectra derived from this CRISM cube are shown to the left and
1095 right of the image. The reflectance values are given for the denoised spectra. For clarity, each raw
1096 spectrum is offset by a constant reflectance value with respect to the corresponding denoised
1097 spectrum. Grey and black dotted vertical bars underline spectral features attributed to ice and
1098 sediment respectively. To decrease the contribution of noise on the displayed spectra, these have
1099 been averaged over boxes (shown in the image) comprising several pixels located on the same
1100 morphological unit. All the spectra have been masked between 1.97 and 2.08 μm because the
1101 atmospheric correction fails in this wavelength range.

1102
1103 **Figure 9:** Distribution of ice, gypsum and other hydrated minerals in the study area, computed
1104 from the CRISM cube hrl0000330c. a) Map of the spectral criterion BD(1.50). b) Map of the
1105 spectral criterion BD(1.934). c) to j) Maps of selected absorption bands detected by the spectral
1106 derivative method. On c) to j), pixels where the spectrum meets a local minimum at the displayed
1107 wavelength are colored. Colors represent the number of pixels (1 to 9 pixels from blue to red)
1108 where this minimum is detected in a moving kernel of 3x3 pixels.

1109

1110 **Figure 10:** Zooms on the CTX image P01_001593_2635_XI_83N241W (location indicated on
1111 Figure 9a) and on maps of absorption bands detected by the spectral derivative method (Figure
1112 9), showing the concentration of gypsum in a) to c) sublimation tills on the dissected packet of
1113 the NPLD ice layers and in d) to f) arcuate scarp and dunes.

1114

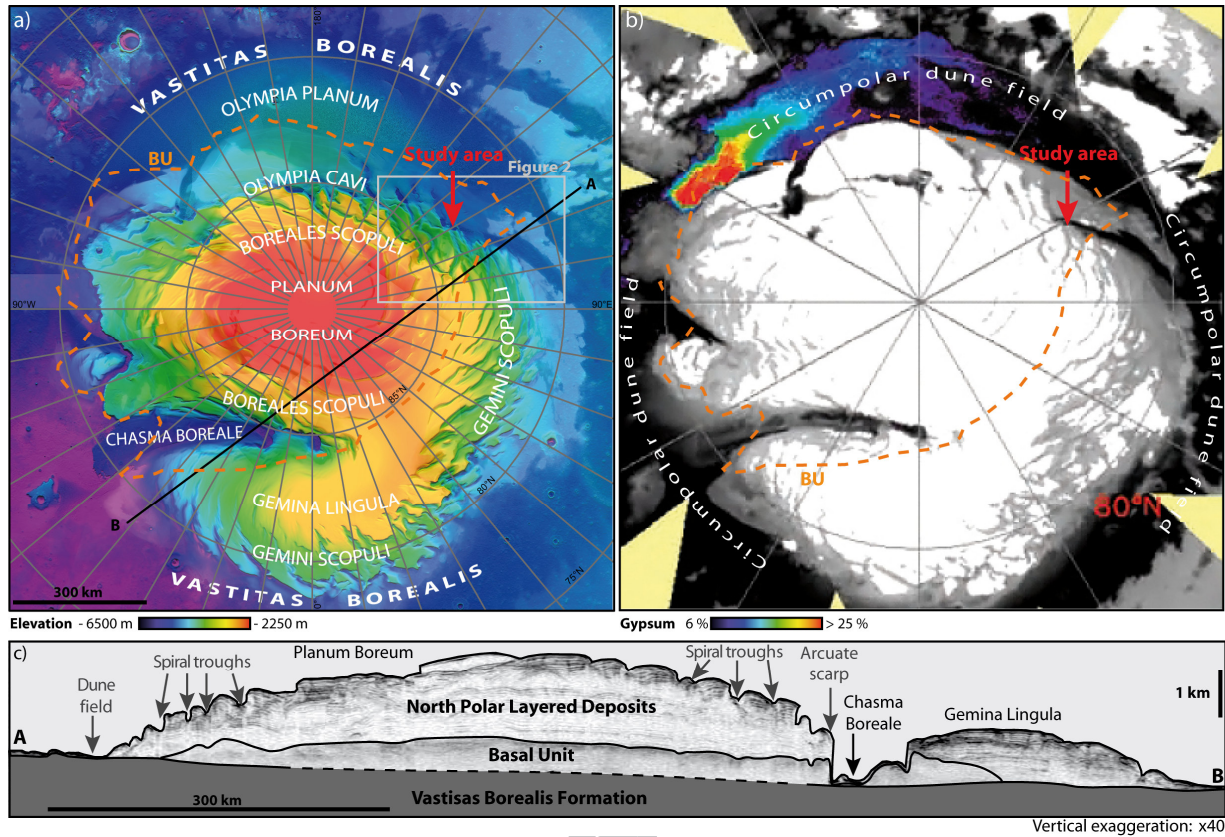
1115 **Figure 11:** Comparison of (denoised) CRISM and (raw) OMEGA spectra acquired on superficial
1116 accumulations of dark sediment in the study area (location indicated on Figure 8). To ensure
1117 consistency of spatial coverage between CRISM and OMEGA data, the displayed CRISM spectra
1118 have been averaged over spatial boxes corresponding to the ground coverage of OMEGA pixels.
1119 Grey and black dotted vertical lines underline spectral features attributed to ice and sediment
1120 respectively.

1121

1122 **Figure 12:** Regional distribution of ice, gypsum and other hydrated minerals around the study
1123 area (shown by the white box in a), computed from the OMEGA cube ORB1056_2. a) Map of
1124 the spectral criterion $BD(1.50)$. b) Map of the spectral criterion $BD(1.934)$. c) to g) Maps of
1125 selected absorption bands detected by the spectral derivative method. On c) to g), pixels where
1126 the spectrum meets a local minimum at the displayed wavelength are colored. Colors represent
1127 the number of pixels (1 to 9 pixels from blue to red) where this minimum is detected in a moving
1128 kernel of 3x3 pixels. h) Location of the OMEGA cube indicated on a shaded and colored relief
1129 image computed from the MOLA DEM at ~256 m/pixel.

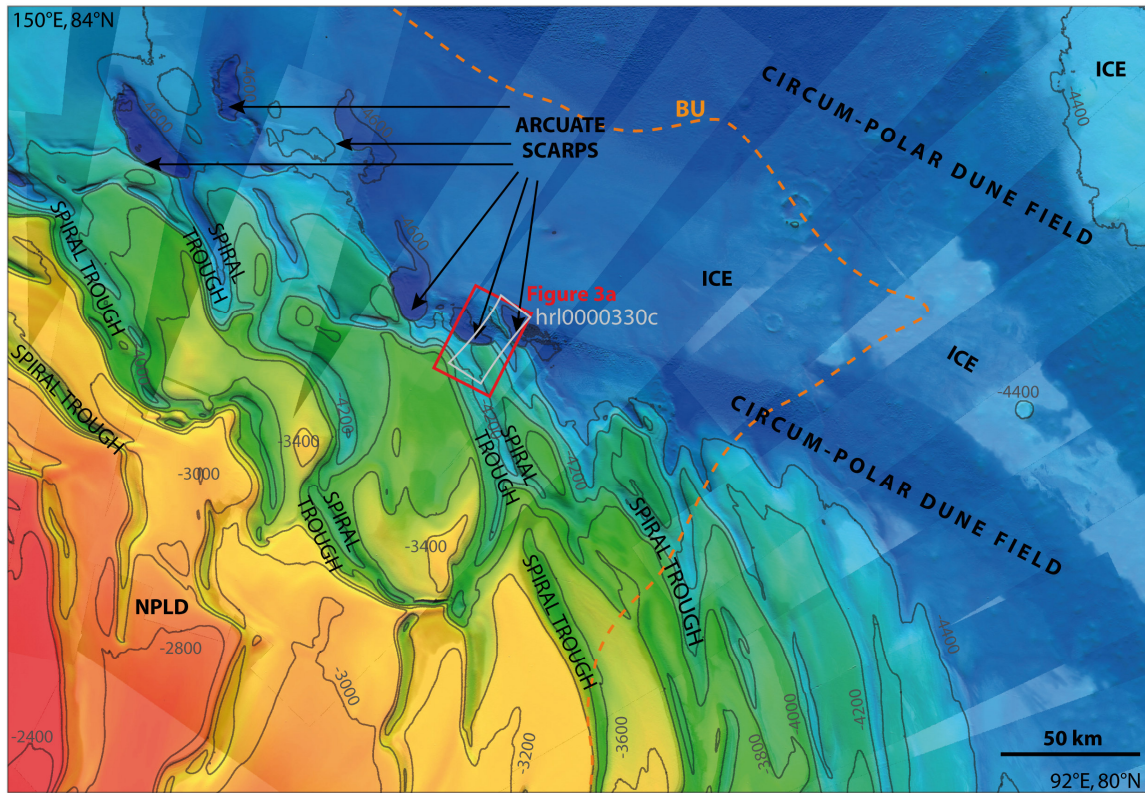
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1131 **Figure 13:** Interpretative scenario for the formation of gypsum in the North Polar Cap.



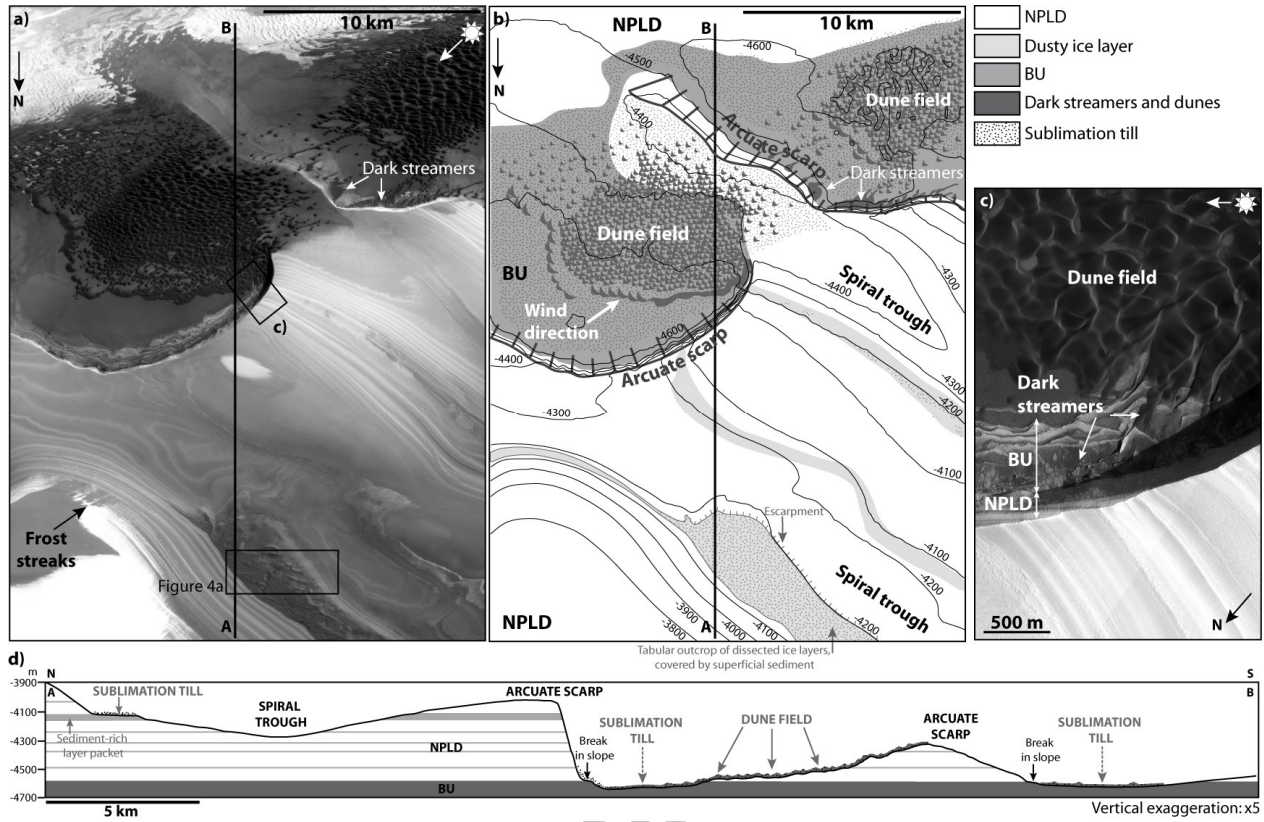
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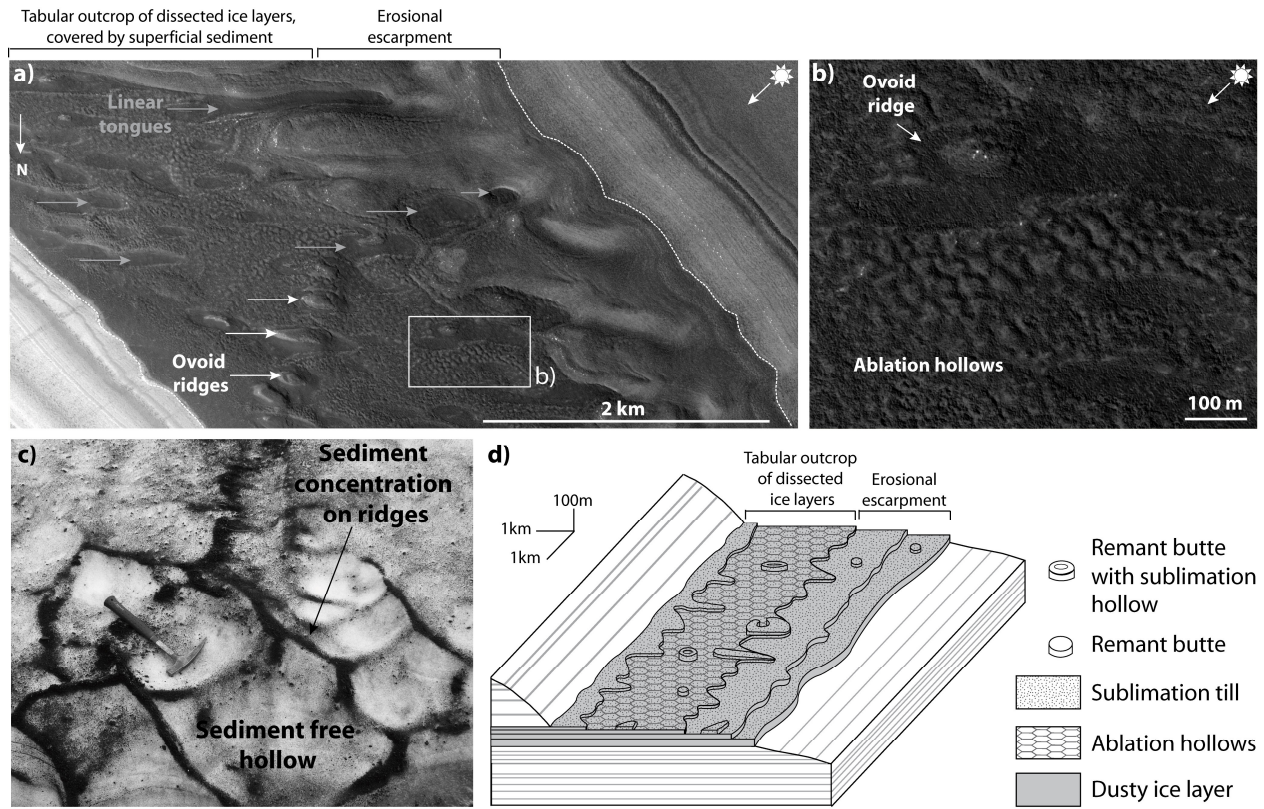
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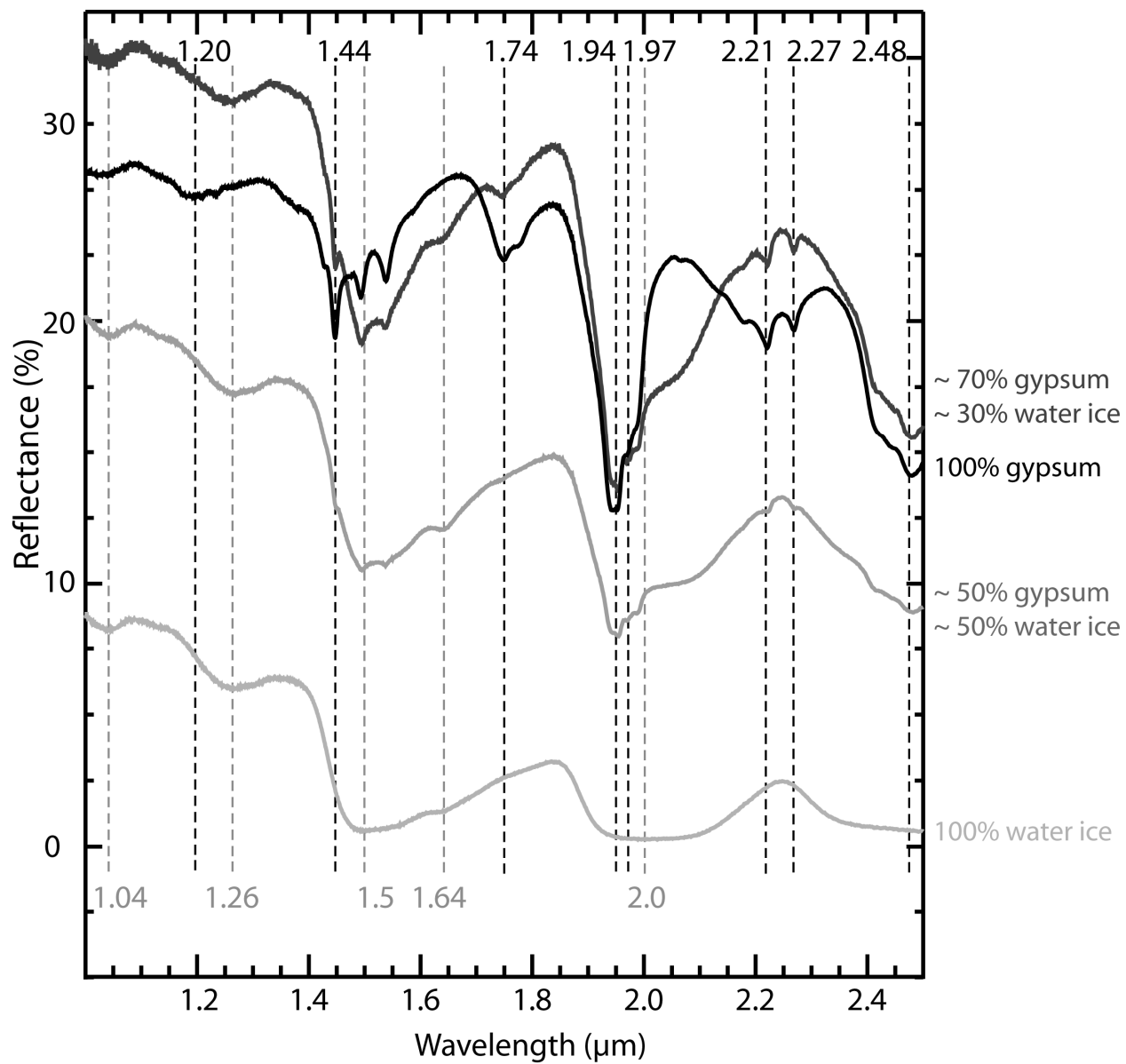
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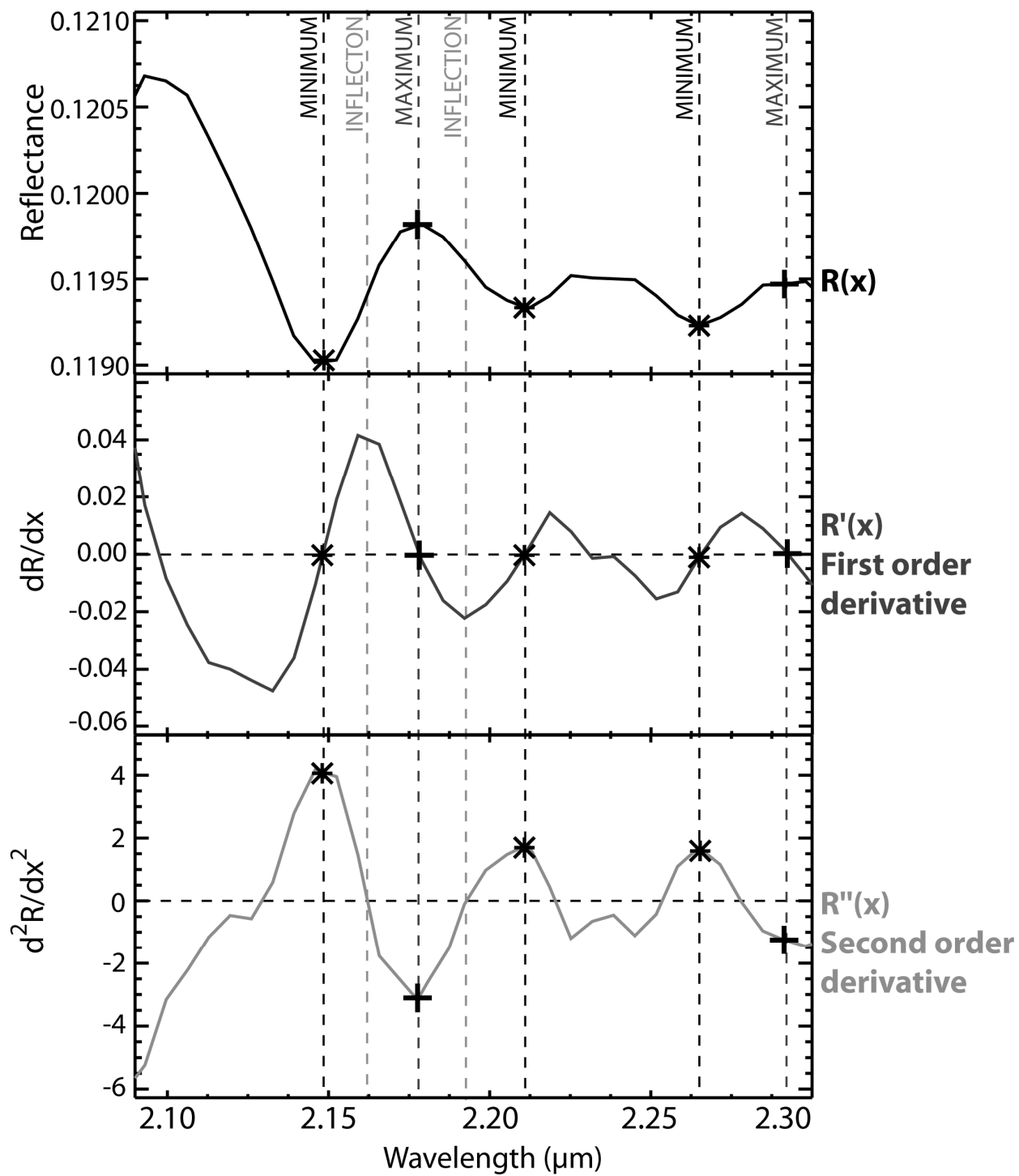
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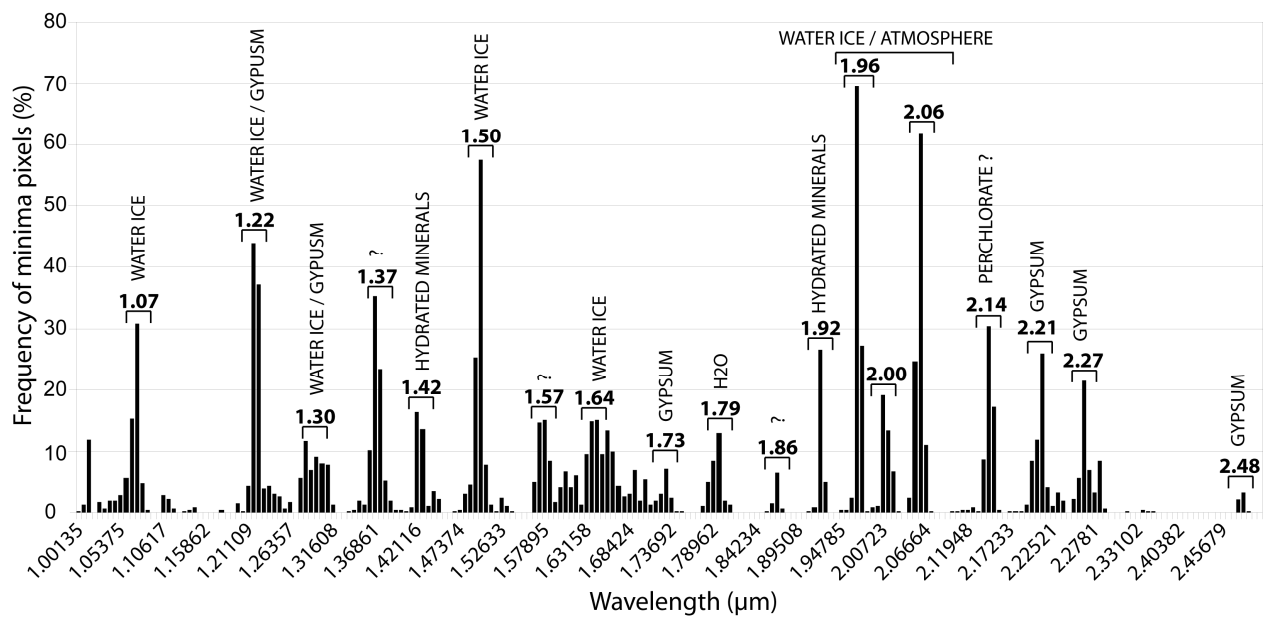


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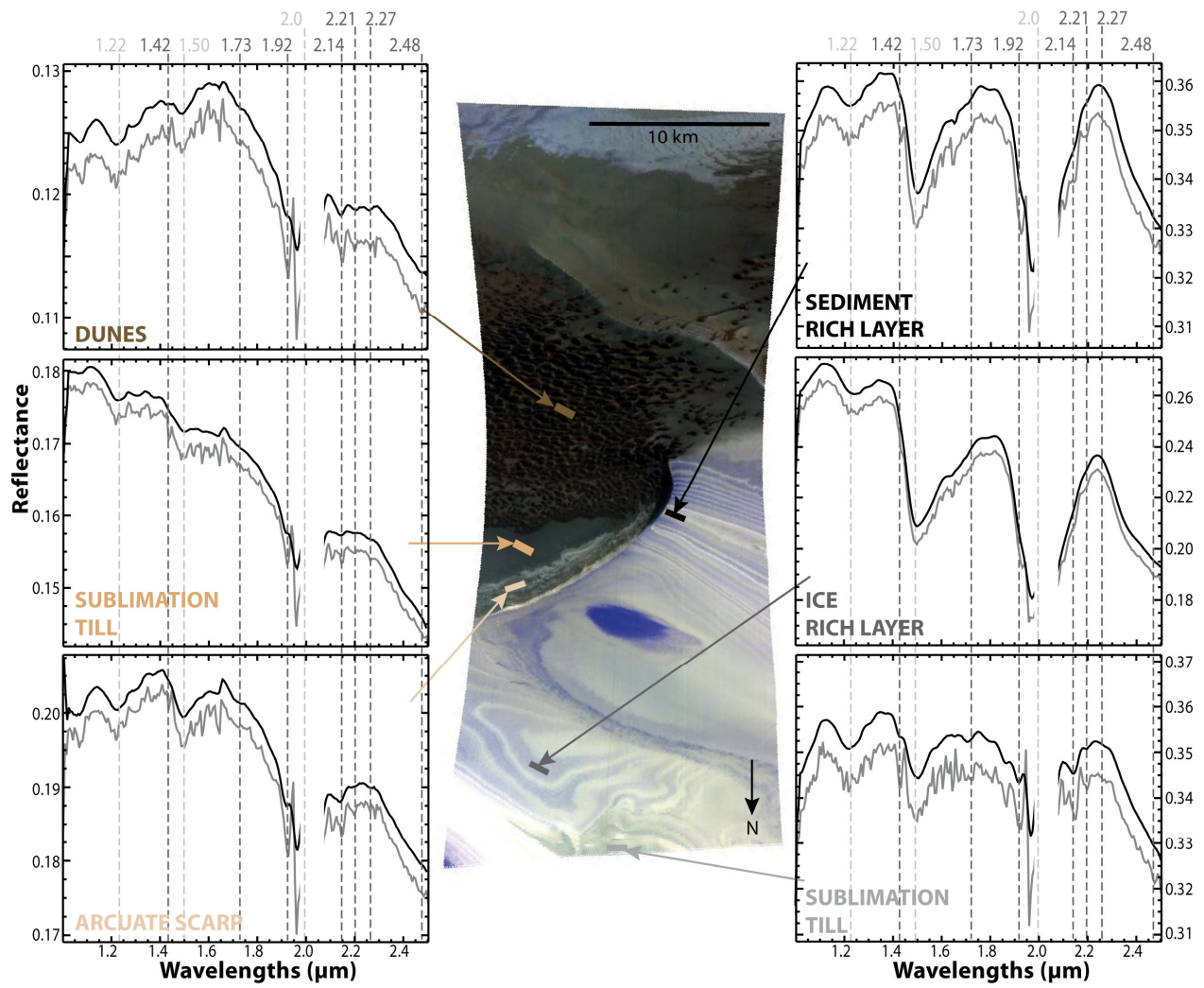


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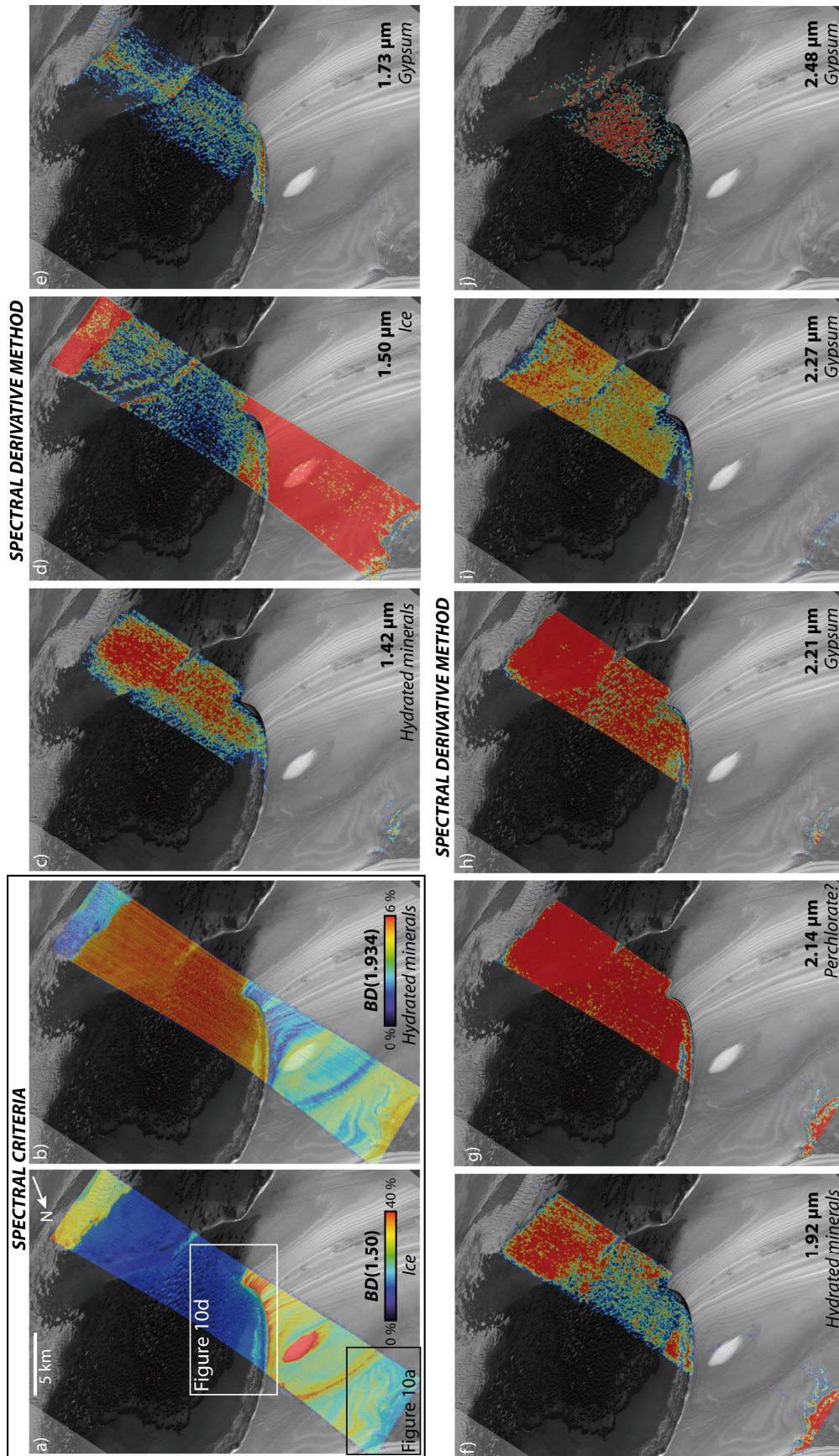
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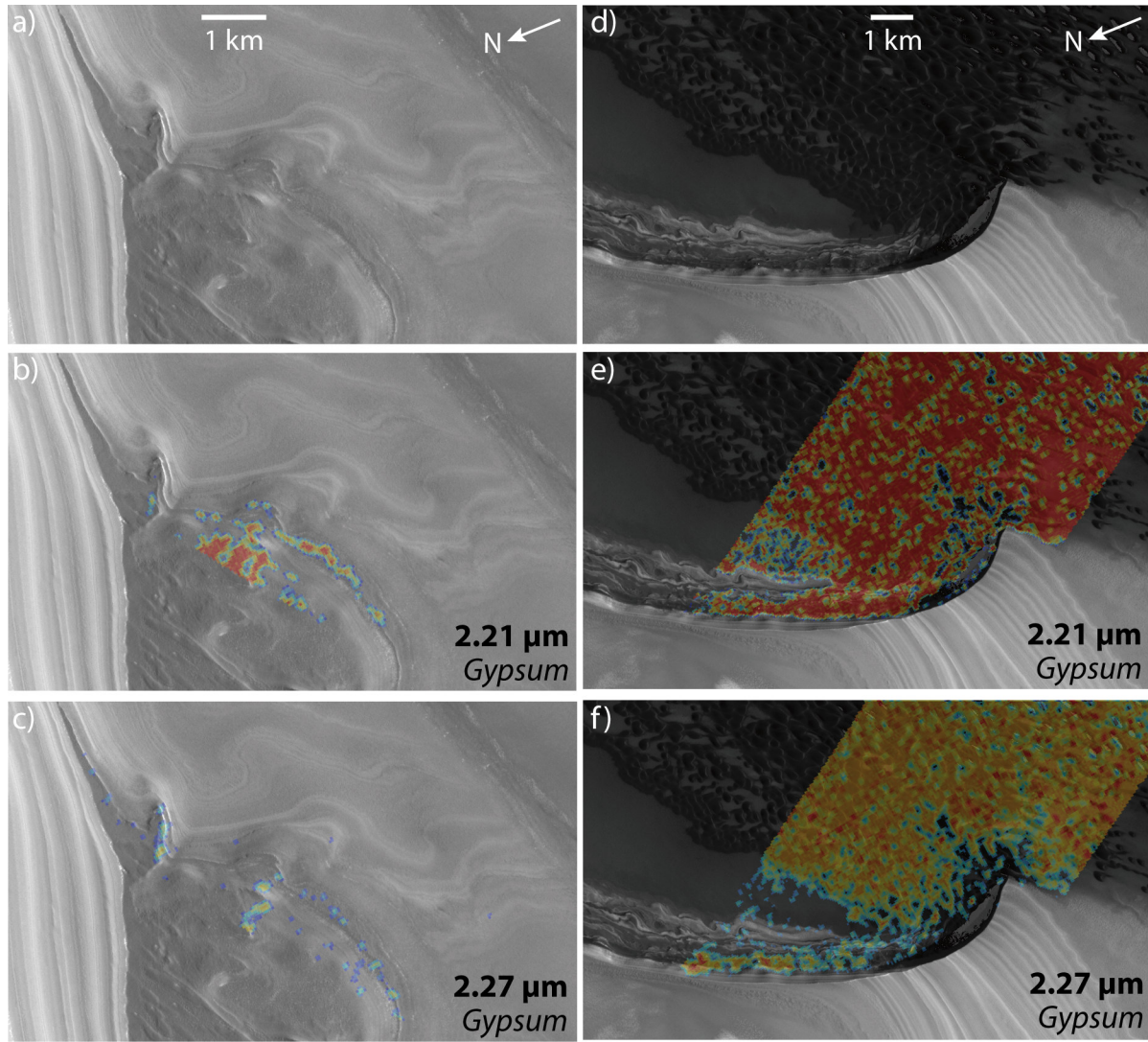
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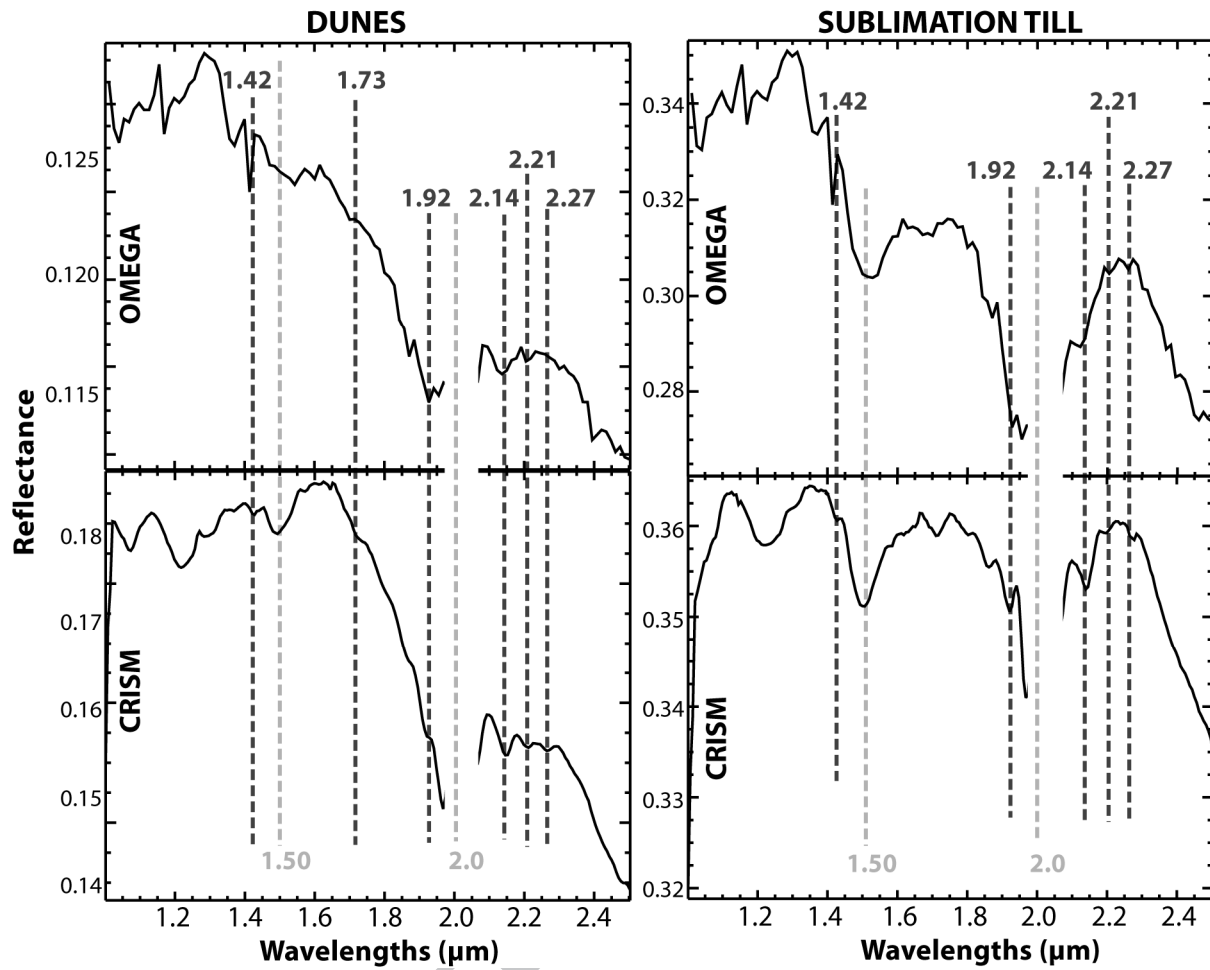
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