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► **To cite this version:**

S. Alwmark, B. Horgan, A. Udry, A. Bechtold, S. Fagents, et al.. Diverse Lava Flow Morphologies in the Stratigraphy of the Jezero Crater Floor. *Journal of Geophysical Research. Planets*, In press, The MARS Perseverance Rover Jezero Crater Floor Campaign, pp.e2022JE007446. 10.1029/2022JE007446 . hal-03978278

HAL Id: hal-03978278

<https://ut3-toulouseinp.hal.science/hal-03978278>

Submitted on 8 Feb 2023

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Diverse Lava Flow Morphologies in the Stratigraphy of the Jezero Crater Floor

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Key Points:

- We investigated Artuby and Rochette member rocks of the Máaz formation in Jezero crater using *Perseverance's* Mastcam-Z and SuperCam data.
- Complex knobby, foliated, vesicular, and layered lithologies are most consistent with lava flows originating through multiple eruptions.
- The Máaz formation in Jezero crater could be unrelated to the regional Circum–Isidis capping unit.

34 Abstract

35 We present a combined geomorphologic, multispectral, and geochemical analysis of crater floor
36 rocks in Jezero crater based on data obtained by the Mastcam-Z and SuperCam instruments
37 onboard the NASA Mars 2020 *Perseverance* rover. The combined data from this analysis
38 together with the results of a comparative study with geologic sites on Earth allows us to
39 interpret the origins of rocks exposed along the Artuby ridge, a ~900 m long scarp of lower Máaz
40 formation rocks. The ridge exposes rocks belonging to two morphologically distinct members,
41 Artuby and Rochette, that both have basaltic composition and are spectrally indistinguishable in
42 our analysis. Artuby rocks consist of morphologically distinct units that alternate over the ridge,
43 bulbous, hummocky, layers with varying thicknesses that in places appear to have flowed over
44 underlying strata, and sub-planar thinner laterally continuous layers with variable friability. The
45 Rochette member has a massive appearance with pronounced pitting and sub-horizontal partings.
46 Our findings are most consistent with a primary igneous emplacement as lava flows, through
47 multiple eruptions, and we propose that the thin layers result either from preferential weathering,
48 interbedded ash/tephra layers, 'a'ā clinker layers, or aeolian deposition. Our analyses provide
49 essential geologic context for the Máaz formation samples that will be returned to Earth and
50 highlight the diversity and complexity of geologic processes on Mars not visible from orbit.

51

52 Plain Language Summary

53 Characterization of the landing site for the Mars 2020 *Perseverance* rover mission yields insight
54 into early solar system processes and provides essential context for Mars Sample Return. Here
55 we have investigated crater floor rocks in Jezero crater that are exposed along a scarp called
56 Artuby ridge with the Mastcam-Z and SuperCam instruments onboard the *Perseverance* rover.
57 The Artuby ridge displays a characteristically layered set of rocks with a basaltic composition
58 that are spectrally and chemically indistinguishable in our investigation. We compare our
59 observations from Jezero with well-understood geologic deposits on Earth, from Hawai'i and
60 New Mexico. We find that terrestrial lava flows can have complex interiors that replicate many
61 features that we see in Mastcam-Z images of the Artuby ridge, and thus, that the series of rocks
62 exposed along the Artuby ridge are dominated by lava flows originating through multiple
63 eruptions. There are a number of layers and textures of the rocks in our investigation that may
64 not have originated as lava flows, that instead may be products of weathering, interbedding of
65 lava and volcanic ash/tephra, or wind-borne sediment deposition. Our results highlight the
66 diversity of geologic units on Mars not visible from orbit.

67

68 1 Introduction

69 The NASA Mars 2020 *Perseverance* rover landed in Jezero crater on February 18th, 2021
70 (Fig. 1a,b). Jezero crater is ~45 km in diameter and located in the Nili Fossae region, on the
71 northwestern margin of the ~1200 km-diameter Isidis impact basin. The formation age of Jezero
72 crater can be bracketed by the formation of the Isidis basin (3.96 ± 0.01 Ga; Werner, 2008; or
73 4.05–4.20 Ga; Marchi et al., 2021) and emplacement of the crater rim-draping regional olivine-
74 and carbonate-bearing unit (3.82 ± 0.07 Ga; Goudge et al., 2015; Mandon et al., 2020). This
75 places its formation in the Noachian period, which was a warmer and wetter period of Mars
76 history (e.g., Irwin et al., 2005; Kite, 2019; Palumbo et al., 2020; Salese et al., 2020). The crater
77 once hosted an open-basin lake fed by a valley network bringing in sediments that built the

78 prominent western delta (e.g., Fassett & Head, 2005, 2008a,b; Goudge et al., 2015, 2017;
79 Mangold et al., 2020, 2021; Schon et al., 2012). *Perseverance* is tasked with characterizing the
80 geology of the landing site, evaluating the astrobiological potential of recorded geologic
81 environments, and selecting, collecting, and documenting a set of in situ samples (Farley et al.,
82 2020).

83 The shallow depth profile of Jezero crater indicates substantial crater-fill with an
84 estimated thickness of ~1 km (Ehlmann et al., 2008; Garvin et al., 2003; Schon et al., 2012).
85 Orbital mapping of the exposed geologic units on the Jezero crater floor prior to landing (Stack
86 et al., 2020) highlights three major crater filling materials: Crater Floor Fractured Rough (Cf-fr)
87 and Crater Floor Fractured 1 and 2 (Cf-f-1, Cf-f-2). Studies previous to the Stack et al. (2020)
88 mapping effort did not distinguish Cf-f-1 and Cf-f-2 as two separate units, due to their similar
89 morphological and spectral signatures. The separation in Stack et al. (2020) is based on subtle
90 differences in surface texture between the two units, and on elevation, where Cf-f-2 crops out at
91 higher elevation inside the crater. During the crater floor campaign of the Mars 2020 mission
92 *Perseverance* traversed over Cf-fr and Cf-f-1 (see Sun et al., 2022; Sun et al., this issue). Cf-f-1
93 is stratigraphically lower than Cf-fr (Farley et al., 2022; Sun et al., 2022; Sun et al., this issue),
94 and has a distinct olivine-like signature in near-infrared (NIR) reflectance spectra, whereas the
95 younger Cf-fr displays NIR reflectance-spectra more indicative of pyroxene (Horgan et al.,
96 2020). In situ analysis by *Perseverance* confirms that Cf-fr, which has informally been named
97 the Máaz formation, is dominated by augite and plagioclase, and Cf-f-1, the Séítah formation,
98 consists of olivine, augite, and plagioclase (Bell et al., 2022; Farley et al., 2022; Liu et al., 2022;
99 Núñez et al., 2022; Schmidt et al., 2022; Udry et al., this issue; Udry et al., 2022). Preflight
100 analysis based on orbital data (Brown et al., 2020; Goudge et al., 2012, 2015; Kremer et al.,
101 2019; Mandon et al., 2020) suggests that Cf-f-1 (and Cf-f-2) is genetically related to the regional
102 olivine-bearing unit, either as simply a part of that unit, or as material from the regional unit
103 eroded, transported, and redeposited within Jezero. Based on in situ data from *Perseverance*,
104 Farley et al. (2022), Liu et al. (2022), and Wiens et al. (2022) interpret the Séítah formation as a
105 slowly cooled olivine cumulate emplaced as a thick lava flow, lava lake, or intrusion, which is
106 inconsistent with fluvial reworking and is difficult to reconcile with Séítah being simply a part of
107 the regional unit, as that would imply a huge volume of regionally spread, slowly-cooling,
108 ultramafic melt.

109 Characterizing the crater floor rocks in Jezero crater is essential for constraining
110 processes that acted during early Martian history and for documenting context for Mars sample
111 return. During sols 169–200 of the *Perseverance* mission, the rover traversed east-to-west along
112 the contact between the underlying Séítah (Cf-f1) and the overlying Máaz (Cf-fr) formations,
113 where an exposed scarp along the contact was named the *Artuby ridge*. During sols 285–351 the
114 rover traversed back west-to-east along the same contact. In this contribution, we present
115 detailed descriptions of Máaz formation rocks exposed along the Artuby ridge, and discuss the
116 genesis of rocks described in this setting.

117 Decades of studies of Mars with orbital instruments, landers, and rovers have revealed a
118 diverse geologic history, that in many aspects is Earth-like (e.g., Carr, 2018). Given the wealth of
119 complex and interconnected geologic processes recorded, such as fluvial and glacial activity,
120 impact processes, igneous processes, and crustal tectonic activity, comparison with known, or
121 more well-understood, terrestrial geologic processes guides our understanding of geologic
122 processes on Mars. Terrestrial analog sites are thus widely accepted as a key source of
123 information in understanding processes that have shaped and modified the geologic history of

124 Mars (e.g., Farr, 2004; see also papers in Chapman, 2007). In this work, we use relevant sites on
125 Earth as a reference to interpret data obtained with *Perseverance*. Here we answer the questions:
126 What features of the Artuby ridge can we resolve in Mast Camera Zoom (Mastcam-Z) images?
127 How do these features and textures compare to Earth analogs? Can we use our observations to
128 constrain the origin of these rocks?
129
130

131 **2 The Mááz formation**

132 The Mááz formation dominates the crater floor in Jezero crater. It has lobate margins that
133 appear to embay topographically higher regions. Several authors have hypothesized that the
134 Mááz formation (Cf-fr unit) in Jezero and a unit mapped outside of the crater, the Nili plains 2
135 unit (Sun & Stack, 2020a,b) or mafic capping unit (Bramble et al., 2017), have a common history
136 (see also Hundal et al., 2022; Sun & Stack, 2020a). Prior to landing, a series of publications
137 hypothesized that the Mááz formation originated as a lava flow (e.g., Goudge et al., 2012, 2015;
138 Schon et al., 2012), or that it had a fluviolacustrine or aeolian origin (Holm-Alwmark et al.,
139 2021; Kah et al., 2020; Shahrzad et al., 2019; Stack et al., 2020; see also discussion in Horgan et
140 al., 2020 and Sun & Stack 2020b). However, surveys in Jezero crater have shown that the crater
141 floor rocks investigated up close with the SuperCam, Planetary Instrument for X-ray
142 Lithochemistry (PIXL), Scanning Habitable Environments with Raman & Luminescence for
143 Organics & Chemicals (SHERLOC), and Wide Angle Topographic Sensor for Operations and
144 eNginneering (WATSON) instruments are igneous, based on the holocrystalline interlocking
145 texture, mineralogy, and bulk composition (Farley et al., 2022; Liu et al., 2022; Sun et al., 2022;
146 Schmidt et al., 2022; Udry et al., 2022). As the unit is crater retaining, it holds great potential as
147 an anchor for calibrating crater chronology if its age can be determined with returned samples
148 (e.g., Herd et al., 2021; Simon et al., this issue; Simon et al., 2022).

149 The Mááz formation also has the potential to constrain processes such as fluvial activity
150 inside Jezero crater. Various publications have proposed cases for the relative timing of
151 deposition of the dominant geomorphologic units in the *Perseverance* vicinity (i.e., the western
152 delta, Cf-fr, and Cf-f1), with scenarios presented for Cf-fr deposition before delta formation
153 (Holm-Alwmark et al., 2021; Horgan et al., 2020; Ruff, 2017; Stack et al., 2020; Sun & Stack,
154 2020a), after delta formation (Goudge et al., 2015; Schon et al., 2012; Stack et al., 2020), and
155 interfingering delta/crater floor rocks (e.g., Horgan et al., 2022; Stack et al., 2020).

156 The Mááz formation is morphologically diverse, consisting of blocky, massive, pitted,
157 and layered rocks, and has been subdivided (descending stratigraphically) into the Ch'ał,
158 Naat'áanii, Roubion, Rochette, and Artuby members (Crumpler et al., this issue; Horgan et al.,
159 this issue; Sun et al., 2022; Sun et al., this issue). The Ch'ał member is seemingly massive, forms
160 blocky terrains (Horgan et al., 2022; Sun et al., 2022), and was traversed near the landing site.
161 The blocky terrain forms the rough surface texture of parts of the crater floor located east of the
162 Octavia E. Butler (OEB) landing site in Jezero (Fig. 1c). Flat-lying polygonally fractured rocks
163 of the Naat'áanii and Roubion members were traversed on the way from OEB to the Artuby
164 ridge. The other members were first observed at or near the Artuby ridge (see below). Roubion
165 rocks were interpreted by Farley et al. (2022) as the lowest stratigraphical exposure of the Mááz
166 formation, overlain by the Artuby member near the Artuby ridge and by the Rochette member
167 just south of OEB. Later interpretations by Horgan et al. (this issue) interpret Roubion to be

168 intermediate stratigraphically between lower (the Artuby and Rochette members) and upper (the
169 Naat'áanii and Ch'ał members) Máaz, at least near the OEB landing site (Fig. 1d).

170 The upper Máaz formation rocks of Ch'ał and Naat'áanii exhibit Mastcam-Z spectra consistent
171 with a mixture of low-Ca pyroxene, hematite, and Fe-bearing feldspar, whereas the lower Máaz
172 formation members exhibit spectra characterized by broad absorption bands centered beyond
173 1000 nm, consistent with a mixture of high-Ca pyroxene and Fe-bearing feldspar (Horgan et al.,
174 2022; Rice et al., 2022a). Exposures in the Guillaumes (Roubion member) and Bellegarde
175 (Rochette member) abrasion patches analyzed with PIXL and SuperCam are consistent with a
176 mineralogy dominated by laths of plagioclase and interlocking pyroxene (Schmidt et al., 2022;
177 Udry et al., this issue). Farley et al. (2022) and Schmidt et al. (2022) classified Guillaumes and
178 Bellegarde as micro-gabbros emplaced as lava flows. The geochemical trend observed across the
179 diverse Máaz rocks shows an overall increase in silica and alkali elements upsection (Wiens et
180 al., 2022). The lower Máaz member targets have mean abundances (standard deviation in
181 parentheses) of SiO₂ of 45.4 (2.42) wt. %, Al₂O₃ of 6.4 (1.1) wt. %, FeO_T of 21.2 (3.0) wt. %,
182 and MgO 4.6 (1.0) wt. %, and upper Máaz targets presented by Udry et al. (this issue) as fine-
183 and coarse-grained have mean abundances of SiO₂ of 51.8 (2.6) wt. %, Al₂O₃ of 8.9 (1.3) wt. %,
184 FeO_T of 17.9 (3.8) wt. %, and MgO 2.3 (0.6) wt. % (Udry et al., this issue).

185 The lower Máaz formation rocks exposed along the Artuby ridge (the Rochette member) were
186 the second sample target of the Mars 2020 mission (Simon et al., this issue; Simon et al., 2022;
187 Sun et al., 2022;), and the first successfully cored set of samples.

188

189 **3 Methods**

190 The Mastcam-Z instrument on the *Perseverance* rover is comprised of a pair of variable
191 focal length multispectral charge-coupled device cameras to document the rover's surroundings
192 (Bell et al., 2021; Bell et al., 2022). The cameras have a lowest zoom setting with a focal length
193 of 26 mm and a highest zoom setting with a focal length of 110 mm. Between these two lowest
194 and highest zoom settings the pixel scale varies between ~540 μm and ~148 μm at 2 m distance,
195 and ~27 cm and ~6.7 cm at 1 km distance. The cameras have a filter set that allows acquisition of
196 11-point narrow-band spectra between 442–1022 nm (Bell et al., 2021). Mastcam-Z relies on
197 radiometric calibration targets (Kinch et al., 2020) for reflectance-calibration of images,
198 following a pipeline that is based on pre-flight testing and in-flight validation (Hayes et al.,
199 2021). The calibration targets are routinely imaged together with commanded multispectral
200 image sequences, and both these activities are typically executed within ± 90 minutes of local
201 noon.

202 Spectra are extracted from multispectral images by selecting specific regions of interest
203 in images from which reflectance factor (R*) values are averaged (Bell et al., 2021). After this,
204 values from each camera are scaled by a single scaling factor per camera to make values from
205 the two cameras agree at their overlapping wavelength at 800 nm. The scaling factor for each
206 camera is defined to set the value at 800 nm to the average between the independently-derived
207 values for the two cameras. Error bars shown indicate the standard deviation of pixels within
208 each region of interest. Variations in spectra derived from multispectral images taken by the
209 Mastcam-Z cameras can be caused by compositional and/or mineralogical characteristics of
210 investigated materials related to the presence of certain Fe²⁺-bearing silicates, Fe³⁺-bearing
211 oxides or oxyhydroxides, and OH⁻ or H₂O-bearing alteration minerals, but also from changes in

212 texture, surface relief, illumination and viewing conditions, or the presence of materials covering
213 surfaces such as dust or coatings (Bell et al., 2021, 2022; Rice et al., 2022a).

214 SuperCam is also situated on the *Perseverance* rover's mast, and is used for imaging and
215 for measuring rock chemistry and mineralogy (Maurice et al., 2021; Wiens et al., 2021). In this
216 study, we report results of rock chemistry measurements using Laser Induced Breakdown
217 Spectroscopy (LIBS) and imaging of rock textures using the Remote Micro Imager (RMI). The
218 LIBS instrument collects plasma light induced by a pulsed 1064 nm laser that is analyzed
219 between 245 and 853 nm. Elements are identified by different emission lines in the LIBS
220 spectrum. To accompany the LIBS data obtained on target surfaces, SuperCam has 23 calibration
221 targets that are routinely imaged (Cousin et al., 2022). To characterize the chemistry of targets
222 reported below we followed the work-flow defined in Udry et al. (this issue), which means we
223 used the major-element oxide composition (MOC) calibration (Anderson et al., 2021) that builds
224 on a suite of 1198 laboratory spectra on a total of 334 reference samples. The first five LIBS
225 shots are removed from calculations because we are interested in the rock composition, not the
226 composition of surface dust. The RMI uses red, green, and blue (RGB) filters to deliver visible
227 color images with a field of view of 19 mrad and an angular optical resolution of 80 μ rad
228 (Maurice et al., 2021; Wiens et al., 2021).

229 Layer thicknesses across the Artuby outcrop were measured using The Planetary
230 Robotics 3D Viewer (PRo3D) software, which is an interactive 3D visualization tool that allows
231 users to work with high-resolution 3D reconstructions of the rover's surroundings (Barnes et al.,
232 2018).

233

234 **4 The Artuby ridge**

235 The Artuby ridge is ~900 m long, constitutes an apparent "boundary" between the Máaz-
236 and Séítah formations, and exposes 2–3 m of SW dipping lower Máaz formation stratigraphy.
237 The outcrop extends from the Artuby East section, with its southeastern-most exposure at the
238 Séítah southeast "thumb", and extends northwest (Figs. 1,2). The best exposures of strata imaged
239 by *Perseverance* are the ~120 m long Artuby East section, which includes "type locality"
240 Artuby_116 (Figs. 2a,3a), and the ~300 m long Artuby West section (Figs. 2a,4). We first
241 obtained Mastcam-Z images of the Artuby ridge traversing south from the OEB landing. In these
242 images the Artuby ridge protrudes as a cliff characterized by interbedded recessive and resistant
243 layers (Fig 2b-d). En route to Séítah, *Perseverance* first investigated the Mure outcrop on sol 168
244 (Fig. 2a), before encountering the Artuby member to the northwest of Mure.

245 At the base of Artuby ridge lies the inferred contact, or transition, between the Séítah and
246 Máaz formations (Farley et al., 2022). The contact itself is obscured by regolith and/or slump
247 material from Artuby ridge and has not been imaged with *Perseverance*.

248

249 **4.1 The Artuby member**

250 The Artuby member rocks overlie the Séítah formation at the Artuby ridge and are
251 characterized by distinct, yet morphologically variable, layering (Fig. 3, 4). The individual
252 layers/laminations are on the order of mm to dm thick, and appear distinct because of differences
253 in resistance to weathering/erosion. These characteristics result in resistant layers that protrude
254 from the ridge, interbedded with more recessive materials (Fig. 3a-g, Fig. 5a). Different portions
255 of the Artuby_116 outcrop, where the Artuby member is ~2.5 m thick, are characterized by mean

256 layer thicknesses that vary distinctly between outcrop sections (Fig. 3c). In the outcrop section
257 that immediately underlies the Rochette member, i.e., the highest stratigraphic portions of the
258 Artuby member, the mean layer thickness is 7–11 mm (variation caused by measurements on
259 mosaics and corresponding 3D meshes from two different rover positions). In the section that
260 underlies this, which is characterized by the presence of more knobby layers, the mean layer
261 thickness is 64–74 mm (Figs. 3, 5; Table S2; Fig. S4).

262 The resistant layers of the Artuby member are often irregular, knobby, in appearance,
263 and display rounded weathering surfaces. As the knobs weather, they expose subtle layering
264 (Fig. 3b). This layering is also evident at exposures where the knobbliness is strongly weathered,
265 and the outcrop appears more recessive (Fig. 3b), for example at the Rimplas workspace (Fig.
266 3h). In some places, pits in the more massive materials appear to cause partings, resulting in a
267 layered, or laminated appearance (Fig. 3b). Accompanying the knobby layers in Artuby are
268 thinner, planar layers (Fig. 3d). These are in some instances highly recessive compared to
269 underlying and overlying strata (Fig. 5c), and in some instances more erosionally resistant than
270 the more massive units. At the Grasse outcrop, we observe thin, erosionally resistant layers that
271 frame the more massive, knobby layers (Fig. 3d, 5d). Generally, over the Artuby outcrops, the
272 lateral appearance of the strata is highly variable, which makes it difficult to follow one layer
273 over tens of meters of distance (Fig. 3). We have observed instances where the knobby layers of
274 Artuby appear to have slumped, or flowed, down over underlying strata (Fig. 3f-g, 4c).

275 Throughout the Artuby outcrops the Artuby member layers have a coarse, granular
276 surface texture (Fig. 5c,d), and sometimes a reddish or purple coating or varnish (see Garczynski
277 et al., this issue; Garczynski et al., 2022). The abraded surface of the Artuby member rock
278 (Montpezat; Fig. S2) at the Rimplas workspace displays a fine-grained rock with interlocking
279 crystal texture. Variation between the natural surface and the abraded surface texture is thus
280 likely an effect of surface weathering.

281 Elemental chemistry obtained by SuperCam LIBS of six Artuby member targets (see
282 Table S1 for a full list of analyzed targets) is consistent with a mineralogy that is dominated by
283 plagioclase and a combination of augite, pigeonite, and possible clino-ferrosilite (Udry et al., this
284 issue; Udry et al., 2022), with minor Fe-Ti-oxides, and an overall basaltic major element
285 chemical composition (see also Wiens et al., 2022). Udry et al. (this issue) reports that SuperCam
286 has not detected olivine in any rock targets of the Artuby member (or any other Mááz formation
287 targets). The chemical composition throughout the entire Artuby ridge section is fairly
288 homogenous, with a mean SiO₂ of 44.1 wt.%, and a mean Mg# [molar 100×(MgO/(MgO+FeO))]
289 of 37 for the SuperCam targets we included in our analysis (i.e., the 13 targets marked in pink in
290 Fig. 6; see also Udry et al., this issue). The stratigraphically lowest targets of the Artuby member
291 are those at the Rimplas workspace (Figs. 2c, 3h), and the composition is consistent with the rest
292 of Artuby.

293

294 4.2 The Rochette member

295 At Artuby ridge, the Artuby member rocks are overlain by the competent caprock
296 member Rochette. The boundary between the two is often not clear, so distinguishing between
297 the two members where only limited sections of rock are exposed is not always straightforward.
298 The Rochette member is 30–50 cm thick and has less granular, smoother surface texture than the
299 Artuby member, thus appearing more competent, or “massive”. The unit displays internal

300 structure in the form of (sub)horizontal layering and is pitted (Figs. 4, S1). Layers are generally
301 approximately 4–5 cm thick and planar when viewed in the Artuby ridge section. Sub-
302 planar/contorted layers are apparent on top of the ridge in the Citadelle workspace area (Fig. S1).
303 The unit is relatively resistant to erosion, compared to some of the layers of the underlying
304 Artuby member rocks, resulting in the cliff-like appearance of the ridge seen in views from
305 Séítah towards Artuby (Fig. 2d). Compared to the Artuby member, where the more massive
306 layers produce rounded weathering faces, weathering of Rochette produces angular, blocky
307 boulders. The Rochette member extends from Artuby ridge south-southeast, forming flat-lying
308 fractured surfaces similar in appearance to the pavers observed near the OEB landing site, that
309 belong to the Naat'áanii member of the Máaz formation. The Rochette member appears to also
310 form the margin of Máaz west of the landing site, but the Artuby member is not observed near
311 the landing site in either outcrop or radar soundings (Crumpler et al., this issue; Horgan et al.,
312 this issue).

313 A paired set of samples, the Montdenier and Montagnac cores, and companion abrasion
314 patch Bellegarde, were obtained in the Citadelle area, sols 181–199 (Fig. S2) on a small boulder
315 (~40 cm across) called Rochette (Fig. S1). This area exposes only the Artuby ridge caprock, the
316 Rochette member, which on top of the ridge displays parallel fluting and prominent mm- to cm-
317 scale pits. Given the ~10° tilt to the south of strata of the Artuby ridge, the Citadelle workspace
318 rocks, including the sampled Rochette boulder, are slightly higher in stratigraphy than the
319 caprocks seen in section along the Artuby ridge. Morphologically (and chemically, Fig. 6 and
320 below), rocks seen on top of the Artuby ridge appear largely consistent with those in the
321 Citadelle workspace. The primary differences are the more pronounced vugginess of rocks in the
322 Citadelle workspace compared to the Rochette member along the Artuby ridge, which may not
323 be a primary difference in the characteristics of the rocks, but a result of better images and/or
324 wind abrasion on the top of the ridge.

325 The Bellegarde abrasion patch exposes a fine-grained igneous textured rock with
326 Mastcam-Z spectral signatures consistent with, for brown areas, a mixture of low-Ca pyroxene
327 and hematite, and, for gray areas, high-Ca pyroxene (Horgan et al., 2022; Horgan et al., this
328 issue). These occur together with plagioclase and minor Fe-Ti-oxides (Schmidt et al., 2022; Udry
329 et al., this issue). Plagioclase and pyroxene crystals are 0.2–0.5 mm across (Schmidt et al., 2022;
330 Simon et al., this issue), which is consistent with an aphanitic texture. Chemically, the
331 Bellegarde abrasion patch is placed in the basalt compositional field (Simon et al., this issue;
332 Udry et al., this issue).

333

334 4.3 Spectral properties and endmembers of Artuby ridge strata

335 We define the following spectral endmembers of Artuby ridge: dust, purple coatings,
336 Artuby member thin layers (i.e., granular, (sub-)planar layers; top of Fig. 3e), Artuby member
337 thick layers (generally knobblier layers of Artuby; Fig. 3f) plus Rochette member, and local
338 regolith (Fig. 7). Based on the features of targeted areas in Mastcam-Z images (such as presence
339 of coatings, thick dust, analyses of borehole tailings, and “clean” bedrock), we have added
340 circles on Fig. 7a to show how the spectrally distinguishable endmembers plot (see Table S1 for
341 a complete list of targets included in analysis). The thick layers of the Artuby member are largely
342 spectrally (and chemically; Fig. 6; Udry et al., this issue; Udry et al., 2022) indistinguishable
343 from the Rochette member on natural surfaces. These both exhibit Mastcam-Z spectra that vary
344 from broad absorption centered beyond 900 nm, consistent with the presence of clinopyroxene,

345 to flat profiles >600 nm (Fig. 7b; see also Horgan et al., 2022; Rice et al., 2022a). No significant
346 spectral indication of olivine has been detected in these rocks (or any other endmembers of the
347 Artuby ridge) with Mastcam-Z, consistent with observations made by SuperCam LIBS (Udry et
348 al., this issue; Udry et al., 2022). The abraded patches, Bellegarde (Rochette member) and
349 Montpezat (Artuby member), are broadly similar spectrally, with one notable exception—it is
350 possible to extract pyroxene spectra from the Bellegarde data, but not from Montpezat. Horgan et
351 al. (2022) attributes this to differences in grain size between the two members.

352 The thin layers in Artuby exhibit Mastcam-Z spectra with a strong red slope <750 nm and
353 a weak ~900 nm band (Fig. 7b). We also identify a 1030 nm downturn in the spectra (Rice et al.,
354 2022a). The absorption band is consistent with a similar pyroxene mineralogy as the thick layers,
355 but the strong red slope at short wavelengths suggests more ferric iron, and the downturn 1030
356 nm sometimes indicates hydration. Thus, the thin layers are likely of similar mineralogy as the
357 thick layers but perhaps with some additional alteration and/or oxidation.

358 Both coatings and dust on rock surfaces can be resolved from bedrock in NIR spectra
359 based on their strong ferric signatures (Fig. 7). Dust is spectrally characterized by flat spectra
360 >750 nm and a strong red slope <750 nm, consistent with nanophase ferric oxides (Fig. 7b;
361 Morris et al., 1993). Coatings exhibit Mastcam-Z spectra with a weak ~900 nm band, moderate
362 red slope <750 nm, and strong 525 nm band (Fig. 7b; see Garczynski et al., this issue;
363 Garczynski et al., 2022). The local regolith observed near the Artuby ridge is spectrally
364 characterized by a broad strong band centered >900 nm, with a strong red slope <750 nm (Fig.
365 7b).

366
367

368 **5 Discussion**

369 The varied nature of the outcrop of Máaz rocks exposed along Artuby ridge tells a story
370 of geologic evolution, be it evolution in the paleoenvironment such as a transition from subaerial
371 igneous deposition to aqueous clastic deposition, or magmatic evolution with long-lived igneous
372 activity with transitions in magma/eruption characteristics. Analysis of rock texture and
373 morphology are crucial tools for determination of rock origin on Mars. Additionally, while we
374 have many natural surface analyses of layers in Artuby ridge, we have only two abrasion patches
375 where the abraded rock surface has been investigated. While the two abraded patches are likely
376 basaltic lavas (Farley et al., 2022; Liu et al., 2022; Schmidt et al., 2022; Sun et al., 2022), the
377 limited number is not representative of the variety of textures and morphologies observed across
378 the Artuby ridge. Thus, further analysis is required to learn about the nature of the rocks that
379 constitute the strata.

380

381 **5.1 Artuby Ridge characteristics**

382 The Artuby ridge displays a rich diversity in texture and morphology of rocks that are
383 basaltic in composition (Udry et al., this issue). The outcrop is characterized by the presence of
384 relatively erosionally resistant, knobbly, layers visible in Mastcam-Z images. The knobbly layers
385 occur together with planar layers that have varied resistance to erosion, all appearing
386 rough/granular in surface texture. The uppermost portion of the Artuby member exhibits finer-
387 scale layering/lamination. The overlying Rochette member weathers differently, producing
388 sharp-edged resistant blocks rather than the knobbly, rounded, faces of the more resistant layers

389 in the Artuby member. The morphological differences between the Artuby and Rochette
390 members are not accompanied by clear differences in the spectral data or chemical data derived
391 from analysis by SuperCam LIBS (Figs. 6, 7). In fact, throughout the entire Máaz formation, the
392 mineral assemblage is uniform (plagioclase, augite, pigeonite, and possibly clino-ferrosilite), and
393 minor variations in abundances and grain size over members/units lead to the different spectral
394 shapes (Horgan et al., this issue). It is possible that some of the spectral differences between
395 Artuby/Rochette and upper Máaz are caused by more high-Ca pyroxene. On Earth, olivine is an
396 important constituent of basaltic rocks (Basaltic Volcanism Study Project, 1981), yet, as we have
397 reported, neither Mastcam-Z nor SuperCam have detected olivine in Artuby ridge rocks. This
398 does not mean that there is no olivine in these rocks, but rather that there is no significant
399 amounts of coarse-grained olivine that can be resolved by these instruments. In fact, PIXL
400 detected fayalitic olivine with secondary serpentine in the Guillaumes (Roubion member) and
401 Bellegarde (Rochette member) abrasion patches (Schmidt et al., 2022), despite the rocks'
402 normative normative mineralogy not including olivine. The lack of olivine, according to the
403 Mastcam-Z and SuperCam investigation, is consistent with the interpretation that the rocks are
404 Fe-rich basalts (Schmidt et al., 2022; Wiens et al., 2022) but we do not exclude the possibility
405 that the rocks contain olivine, either in small amounts, of small grain size, or perhaps in altered
406 form.

407 The thicker, resistant layers and knobs in Artuby spectrally resemble Rochette (Fig. 7),
408 although the Artuby targets are often characterized by an overall redder color. This could be due
409 to higher presence of ferric iron in the Artuby member than the Rochette member (e.g., Rice et
410 al., 2022a). Because of difficulties in isolating spectra from the thinnest and most recessive
411 layers in Artuby due to mantling by dust/regolith and the presence of coatings, we are not able to
412 conclude whether the distinct spectral profile of some of these layers is in fact related to
413 compositional/mineralogical differences, or whether the surface is causing the differences. What
414 we can say, however, is that the Artuby and Rochette members are both distinct from the Bastide
415 member of Séítah based on the absence of an olivine signature and the lower MgO content
416 (Núñez et al., 2022; Udry et al., this issue; Udry et al., 2022; Wiens et al., 2022). Targets within
417 the Séítah formation (excluding the Content member) have a mean of 23.5 wt.% MgO, compared
418 to 2.3–4.6 wt.% MgO for Máaz formation targets (Udry et al., this issue; compare also results
419 shown here in Fig. 6). On the other hand, the similarities in chemistry between the Content
420 member, which overlies Bastide within the Séítah region (Fig. 1d), and Máaz formation rocks
421 (specifically upper Máaz) are so pronounced that Udry et al. (this issue) suggest the two belong
422 to the same petrogenetic sequence. The Content member is characterized by vesicular textures
423 that Udry et al. (this issue) interpret to have formed by effusive volcanism, providing further
424 separation from the rest of Séítah, since the other Séítah formation members are interpreted to
425 represent an olivine cumulate (Farley et al., 2022; Liu et al., 2022). The targets within the
426 Content member show very similar major element compositions to Máaz rocks (Udry et al., this
427 issue), and the mineralogy is also similar in Content and Máaz, particularly for pyroxene
428 compositions (Udry et al., this issue), although the Content member is somewhat more alkaline.
429 The exact stratigraphic relation between the Content member and Máaz is however not clear
430 from field observations with *Perseverance* (Fig. 1).

431 In turn, Rochette can be separated spectrally from the stratigraphically higher Máaz
432 rocks, where Rochette rocks exhibit less absorption at short wavelengths, and overall bluer
433 slopes (see also Horgan et al., 2022; Rice et al., 2022a). Compositionally, both the Artuby and
434 Rochette members are more primitive than upper Máaz (Wiens et al., 2022; Udry et al., this

435 issue). Differences in chemistry of the regolith further highlights the variances in rock
436 composition between on one side the Artuby and Rochette members, and on the other the
437 stratigraphically higher Máaz rocks exposed near the OEB landing. We see that the regolith near
438 Artuby, which is likely sourced from weathering of local bedrock, is spectrally distinct from
439 typical regolith at the OEB landing (Fig. S3; see also Cardarelli et al., 2022; Vaughan et al., this
440 issue).

441

442 5.2 Extrusive igneous emplacement

443 Earth analog sites are essential complements to robotic missions such as Mars 2020 to
444 develop understanding of the geologic evolution of Mars (e.g., Farr, 2004; Osinski et al., 2006).
445 In this contribution, we do not aim to describe any site on Earth as identical to what we see along
446 the Artuby ridge, and do not attempt to present new Mars analog sites on Earth. Instead, we want
447 to use the varied morphological characteristics of diverse well-described geologic deposits on
448 Earth to understand similar characteristics on Mars, and to test hypotheses of martian geologic
449 processes.

450 Earth offers us a wide range of analogs for planetary lava flow features. Even restricting
451 our attention to mafic flows, we can observe a range of lava emplacement styles that depend
452 upon effusion rate and the complexities of lava rheology. Changes in the nature of the eruption
453 and evolution of the source magma over the period that the volcanics are being emplaced, as well
454 as change with distance from the source, result in diverse lava flow morphologies (e.g., Self et
455 al., 2021; Sheth, 2006; Voigt et al., 2021) that vary both laterally and vertically over a
456 stratigraphic sequence (Fig. 8). Deposits of alternating pyroclastics and lava flows are common
457 on Earth, resulting in even more complex morphological variations over a stratigraphic sequence
458 than deposits formed by a single volcanic depositional mechanism (Fig. 8b). These types of
459 alternating deposits in the same stratigraphic section, due to their intrinsic characteristics, have
460 different resistance to weathering, resulting in recessive and resistant materials in the same
461 section (Fig. 8b). We find that these features of terrestrial volcanic deposits have many
462 commonalities with the strata at Artuby ridge, where we observe complex, varied morphologies
463 that include repetition of sequences of units that share characteristics.

464 On Earth, flood basalt lobe thicknesses are generally >5 m (Self et al., 2021), with thinner
465 lobes associated with basaltic shield volcanism, such as on Hawai‘i. The Artuby member is ~ 2 –
466 2.5 m thick at the Artuby ridge (Figs. 3,5), and composed of a series of layers with varying
467 thicknesses between <1 to ~ 20 cm. The Rochette member appears ~ 2 m thick in radar soundings
468 along the ridge (Horgan et al., this issue). Similarly, nearer to the OEB landing site The Radar
469 Imager for Mars Subsurface Experiment (RIMFAX) profiles shows a ~ 2 m thick unit with strong
470 upper surface reflectors interpreted as the Rochette member (the Artuby member is not apparent
471 in this area; Horgan et al., this issue).

472 For low-viscosity, basaltic compositions, low lava effusion rates (less than ~ 5 – 10 m³ s⁻¹
473 for Hawaiian tholeiite) typically lead to emplacement of compound, inflated pāhoehoe flow
474 fields (Rowland and Walker, 1990), in which dominant feeder pathways (i.e., lava tubes or more
475 laterally-extensive lava sheets) supply lava to the flow front beneath a solidified, insulating
476 surface crust. On shallow slopes, flows spread laterally, and continued injection under, and
477 thickening of, the lava crust leads to vertical uplift (“inflation”) of the flow surface (Hon et al.,
478 1994). Cooling and stalling at the flow front, or blockages within tubes/sheets, lead to breakouts
479 of new flow lobes at quasi-random locations within the flow field. Thus, the flow field is

480 constructed from numerous overlapping, interfingering toes and lobes emplaced in a piecemeal
481 manner in space and time. Initial flow lobes are dm to m-scale, but can thicken and inflate to
482 several m over time (Hon et al., 1994). Cross sections through successions of pāhoehoe lobes
483 exhibit highly variable morphologies on a range of scales, with simple lobes and toes showing a
484 variety of features: lack of lateral continuity or uniformity, multiple overlapping lens-shaped or
485 bulbous lobes, vertical or concentric cooling-related lamination within lobes, and smooth, ropey,
486 or platy surface textures. At the Artuby scarp, lack of lateral continuity and undulating partings
487 or contacts in some portions of the Rochette and Artuby members is consistent with multiple
488 overlapping lobes, typically observed on dm- to m-scales (Fig. 3f-g, 4c, compare also knobbly,
489 granular parts of Artuby in Fig. 3b with flow foliation in Fig. 8f). Bulbous morphologies
490 (“knobbliness”) similar to those we have observed in the Artuby member are also observed in
491 terrestrial lava sections, either as a weathering expression or as a primary feature of pāhoehoe
492 toes (Figs. 8b,f, 9a).

493 Higher effusion rates (greater than $\sim 10 \text{ m}^3 \text{ s}^{-1}$ in Hawai‘i; Rowland & Walker, 1990)
494 typically produce channel-fed ‘a‘ā flows ranging up to the order of 10 m thick, characterized by
495 rough clinkery surface and basal layers, with dense flow cores. These produce relatively simple
496 lobate planforms, in contrast to the more complex compound form of pāhoehoe flow fields. In
497 cross-section, the clinker layers of ‘a‘ā flows can preferentially weather out to form recessive
498 layers. At the Artuby scarp, the capping Rochette member appears more laterally uniform than
499 the Artuby layers, and the blocky pattern of jointing and fracturing is consistent with thicker
500 terrestrial flows (Fig. 9b).

501 Vesicularity within successions of pāhoehoe lobes can be highly variable, given the range
502 of morphologies and textures observed in terrestrial pāhoehoe examples (Fig. 9a–d). Vertical or
503 concentric patterns of variability in vesicularity can be induced by the interplay of cooling (i.e.,
504 thermal and hence rheological gradients) and shearing within a flow lobe. Larger ($> \text{m}$ scale)
505 inflated flow sections typically exhibit vesicularity patterns in which a dense, low vesicularity
506 core is sandwiched between upper and lower chilled zones. The upper portion of lava flows are
507 frequently characterized by vertical gradations in vesicle size in which smaller, densely packed
508 vesicles occur near the upper section and larger, but fewer, vesicles occur lower in the section,
509 reflecting buoyant bubble rise and coalescence as the overlying crust continues to thicken (Fig.
510 8a). The upper portion of the Rochette member exhibits examples of vesicle size increasing with
511 height, as might be the case beneath a chilled, more finely-vesicular surface crust (such as in Fig.
512 8a). Alignments of coarse vesicles (vesicle “trains”) can be produced by trapping of buoyant
513 bubbles beneath cooler, viscoelastic lava or solidified crust. The absolute vesicularity or gas
514 volume per unit lava volume is controlled by the ideal gas law (Cashman & Kauahikaua, 1997)
515 whereas vesicle sizes are controlled by coalescence according to the time available for growth
516 before the downward and upward-moving solidification fronts set in. Similar patterns of
517 vesicularity are observed to be suppressed in typical ‘a‘ā lava flows in which the interiors are
518 relatively massive and vesicles are sparse (Figs. 8b, d). Weathering of especially vesicular zones
519 can leave the dense portions protruding beyond the recessive weathered vesicular zones.

520 Development of paleosols or deposition of pyroclastic fall layers deposited between
521 flows can produce prominent partings, as can internal shearing and continued injection of lava
522 beneath cooling crusts, particularly in basal sections of lava flows. Complex layering or foliation
523 can develop in this manner within lava flows over a range of viscosities in both pāhoehoe and
524 ‘a‘ā sections from shearing strains during the late stages of flow (Fig. 8c–f). Finer laminations
525 (such as we observe in Artuby scarp strata) within individual lobes may result from such

526 shearing and lead to vertical gradients or contrasts in rheology, vesicularity, and/or crystallinity
527 of the lava. The recessive layers that we have observed can be produced by preferential
528 weathering of more vesicular lava; differential aeolian erosion of the fine (cm-scale) vertical
529 variability in vesicularity could potentially lead to the fine-scale laminations observed in the
530 Rochette member (Fig. 4b-d).

531 Given that Jezero crater once hosted a lake, it is important to consider textures of
532 subaqueously emplaced lavas. Lavas emplaced in water produce a range of morphologies,
533 including pillows, lobes, and tubes (Skilling, 2002). Quench shattering of rapidly cooled lava
534 skin and explosive magma–water interactions can occur to produce abundant glassy clastic
535 material and accompanying granular textures. We do not observe features such as obvious pillow
536 lavas or pillow breccias within the Artuby member. The mix of intact and fragmented lava
537 morphologies with tephra and hyaloclastite can produce complex, rather disorderly sequences
538 (White et al., 2015), but may include layered beds of clastic material.

539 Pyroclastic density current (PDC) deposits (Dufek et al., 2015) might exhibit a wide
540 range of facies, depending on the PDC formation mechanism, current volume, particle
541 concentration (dense vs. dilute), and evolution with distance from source. Deposits might exhibit
542 massive to planar- or cross-bedded interior structures, complex interior stratigraphy, poor
543 sorting, reverse grading, and welding. PDC deposits by nature tend to thicken in topographic
544 lows, and thin over topographic highs. Fall deposits, on the other hand, tend to mantle
545 topography more evenly, are better sorted than PDC deposits, and, depending on the relative
546 steadiness of the eruptive source, may produce massive to planar-bedded deposits and grading.
547 These are expected to contain significant amounts of glass (e.g., Houghton & Carey, 2015). The
548 granular/clastic texture that we observe in at least some of the layers in the Artuby member (Fig.
549 5c) could be consistent with being a product of explosive volcanism, but we see no definitively
550 diagnostic features of either pyroclastic fall or PDC deposits. Fall deposits could be consistent
551 with the laterally continuous, planar layers of the Artuby member, which also tend to be more
552 recessive. Furthermore, both fall deposits and hyaloclastites are typically associated with high
553 abundances of glass. Glass has not been detected with SuperCam, although glass presence was
554 inferred by Udry et al. (this issue) based on the presence of flow-banded textures, conchoidal
555 fractures, and shiny luster observed on some rock targets (although wind polishing could also
556 result in glassy-like appearance). Although the thinner planar layers were difficult to isolate in
557 multispectral images, the slight spectral differences observed could be caused by alteration,
558 which would be expected of more glassy layers.

559

560 5.3 Aqueous deposition

561 When first observing the fine-scale foliation, including sub cm-scale, in Artuby ridge in
562 Mastcam-Z images, a sedimentary origin came to mind since fine-scale laminations are
563 characteristic of deposition of clastic sediments in aqueous environments (see also Sun et al., this
564 issue). Such laminations can typically be tied to a clastic sedimentary origin based on the
565 presence of other accompanying morphological characteristics, such as cross-laminations. We
566 have observed no unequivocally cross-laminated layers in the Artuby ridge. We have also not
567 observed any other evidence of clastic sedimentary deposition, such as grain rounding, or
568 presence of cement in the abrasion patches in the Rochette or Artuby members (see also
569 discussion in Farley et al., 2022).

570 The well-developed state of the western delta in Jezero crater indicates that clastic
571 sediment input was significant, which also means that lacustrine sediments would have been
572 deposited in the lake. The depth of the Jezero crater lake is estimated to have been approximately
573 40 m based on the bottomset-foreset relationship in the western delta (Mangold et al., 2021). The
574 lacustrine sediments were deposited on top of the Mááz formation, if the delta does in fact post-
575 date these rocks (e.g., Holm-Alwmark et al., 2021; Stack et al., 2020). This means that the lack
576 of observed sedimentary rocks in Jezero crater thus far (the Séítah formation is also interpreted
577 as igneous rock; e.g., Farley et al., 2022; Liu et al., 2022; Núñez et al., 2022; Wiens et al., 2022)
578 opens up the question—where are the lake sediments?

579 Thus, erosion must have removed the entire lacustrine sedimentary record in Jezero
580 crater, at least from the most recent lake episode, that formed the western delta (see discussion in
581 Quantin-Nataf et al., 2021, this issue). The only exception is isolated mesas of deltaic/lacustrine
582 deposits (e.g., Holm-Alwmark et al., 2021; Mangold et al., 2021). If parts of the lacustrine phase
583 in Jezero crater that is tied to formation of the western delta pre-dates deposition of (parts of) the
584 Mááz formation, then lacustrine sedimentary rocks can be preserved either in the Mááz formation
585 rock sequence, or in a stratigraphically lower position. Lacustrine phases in Jezero crater that
586 pre-date the phase of formation of the western delta are supported by the presence of the much
587 less well-preserved northern fan with a possibly separate inflow channel (e.g., Jodhpurkar &
588 Bell, 2021). No direct evidence of clastic sediment deposition was observed in situ in the lower
589 Mááz formation by us, or by Horgan et al. (2022; this issue) and Udry et al. (this issue) did not
590 observe any such lithologies in the upper Mááz members either. However, RIMFAX soundings
591 show that it is possible that Rochette is overlain by a localized ~1 m thick low-density
592 sedimentary unit at least in places along the western margin of Mááz in the landing site area
593 (Horgan et al., this issue), meaning that preserved clastic sediments could be present both
594 stratigraphically lower than lavas of the Mááz formation, and in higher stratigraphic positions
595 than Artuby/Rochette within the Mááz formation.

596

597 5.4 Intrusive igneous emplacement

598 On Earth, olivine cumulates are part of a petrogenetic sequence of rocks that also usually
599 contain overlying ferrogabbros, or pyroxene-rich cumulates (e.g., Namur et al., 2015). Farley et
600 al. (2022) and Liu et al. (2022) propose that the Séítah formation is an olivine cumulate, and that
601 the lower Mááz rocks may be petrogenetically related (see also Sun et al., this issue; Wiens et al.,
602 2022). Thus, evaluating the potential origin of Artuby ridge strata as part of the same
603 petrogenetic sequence is necessary. We evaluate the potential of the foliation observed in lower
604 Mááz to have been formed in an igneous intrusion first. Layered structures and banding of
605 intrusive igneous bodies on Earth can be highly variable (Davies et al., 1980; Hirschmann, 1992;
606 Hoover, 1989; McBirney, 1996; Namur et al., 2010, 2015), such as in the Skaergaard intrusion,
607 and in the Sept Iles layered intrusion, where individual recognizable layers range from ~10 cm to
608 > 1 m. Thinner, cm-scale, beds occur in for example the Stillwater Complex (McCallum, 1996),
609 and sub-cm scale beds, are known for example from the La Cordadera gabbro intrusion in Chile
610 (Namur et al., 2015). These would be a closer analog to the scale of bedding/lamination that we
611 observe in Artuby ridge. However, we are not aware of any terrestrial layered igneous intrusion
612 that display such flow-like structures as observed in the Artuby ridge, or form lobate margins
613 such as observed within the Mááz formation rocks exposed at the Artuby scarp. We also note
614 that the pitted texture of the Rochette member may be difficult to reconcile with a petrogenetic

615 relationship with the cumulate rock in Séítah. The mineralogy of Rochette/Artuby is also
616 inconsistent with an origin from the same magma source as Séítah, as the pyroxene suites are of
617 very different compositions (Udry et al., this issue).
618

619 5.5 Summary of the origin of Artuby ridge strata

620 The knobbly layers and the Rochette member of the Artuby ridge share many
621 morphological characteristics with lava flows on Earth, such as the complex lobe/intraflow
622 textures and presence of vesicles (compare Figs. 3f,g with 8b and 9 a,b, and d). We also note
623 similarities in variable friability that in our Hawai'i example are due to the presence of a
624 somewhat weathered 'a'ā clinker. The textures and morphologies of the Artuby and Rochette
625 members at Artuby ridge are thus consistent with interior structures of terrestrial mafic lava
626 flows having a range of emplacement styles and weathering expressions (Fig. 5). Our conclusion
627 is further supported by the spectral character and chemical composition, as well as textures
628 observed in the abraded patches in both Artuby and Rochette, which lack evidence for typical
629 clastic sedimentary rocks such as roundedness of grains and presence of cement.

630 Taken together, the morphology and compositional data of the more massive layers in
631 Artuby (i.e., the knobbly layers and Rochette member), provide convincing evidence that these
632 units are effusive in origin. Along the Artuby ridge, the Artuby and Rochette members range up
633 to 2–2.5 and 2 m thick, respectively. While we can infer the thickness of both members from
634 RIMFAX data (Horgan et al., this issue) without seeing the lower portion of Artuby due to
635 regolith cover, these are likely underestimates of the true thickness of the members because the
636 upper portion of Rochette may have been eroded (Quantin-Nataf et al., 2021, this issue).
637 Likewise, the Artuby member may have been eroded before emplacement of Rochette, and may
638 include erosional surfaces. In addition, each member may be composed of multiple, thinner flow
639 lobes. Typical basaltic shield flows are on the order of dm to a few m; flood basalts can be
640 substantially thicker. Nevertheless, the small thicknesses of Artuby and Rochette might be
641 suggestive of thin, low-viscosity flow units and/or the distal lobes of compound flow fields.

642 The rough, somewhat granular surface texture of Artuby member rocks could be the
643 result of the breakdown of finely vesicular lava and resembles the foliated interior shown in Fig.
644 8f. The finely-layered, planar material that appears to be intercalated or to surround the bulbous
645 Artuby member morphologies (Fig. 3d) is difficult to explain by an effusive origin. These seem
646 to be broadly consistent in composition to the rest of the Artuby and Rochette member targets,
647 which agrees with a scenario where these layers are derived from the same source as the more
648 massive layers, albeit via a different mechanism (e.g., tephra deposits). They could also be
649 weathered 'a'ā clinker layers, or even hyaloclastites produced in the presence of external water.
650 It is also possible that the Artuby ridge contains layers of material deposited via aeolian
651 processes during periods of time when volcanic products were not emplaced.

652 It is likely that the Artuby ridge section exposes material that experienced weathering for
653 various durations, and development of regolith could give rise to the variable morphologies and
654 resistance to weathering that we observe. This is also consistent with the similarities in chemistry
655 and spectral properties across the rock members. More pronounced weathering of some sections
656 of knobbly materials could also give rise to more layered than knobbly appearance, since the
657 knobbly material appears to have an internal foliation (Figs. 3b, h).
658

659 5.6 Implications for the Séítah/Máaz relationship

660 The Artuby ridge exposes the stratigraphically lowest Máaz formation rocks, the Artuby
661 and Rochette members. These members overlie the Séítah formation, which occurs outside of the
662 lobate margins of Máaz (Fig. 1b; see also Hamran et al., 2022). However, with the contact
663 obscured, a genetic relationship has not been tested directly during the Mars 2020 crater floor
664 exploration campaign.

665 Orbital mapping along the Máaz–Séítah contact farther to the south in Jezero crater
666 revealed discrete layers in the basal 2–6 m of the Máaz formation (exposed both outside and
667 inside of the lobate margins), which change in overall thickness and in the number of observed
668 layers (Alwmark et al., 2021). Alwmark et al. (2021) suggested that these observations indicate
669 that layered strata were likely deposited over existing topographic relief associated with an
670 erosional unconformity between the units. Based on our results, it is possible that these variations
671 in both thickness and in the number of observed layers from orbit could result (at least partly)
672 from lobe geometry rather than an erosional unconformity between the formations. In situ
673 analysis with RIMFAX on *Perseverance* has shown that layers at the Séítah and Máaz contact
674 are tilted at the same angle (Farley et al., 2022). This was used to argue in Farley et al. (2022)
675 that both formations were tilted at the same time, potentially indicating that the formations were
676 deposited relatively close in time.

677 While Máaz and Séítah share some morphological characteristics, such as interlayering of
678 thin beds and lateral transitions from recessive to resistant behavior, they are chemically and
679 mineralogically distinct (Rice et al., 2022a; Sun et al., 2022; Udry et al., this issue; Udry et al.,
680 2022; Wiens et al., 2022). The one exception is the Content member, representing the highest
681 stratigraphic targets analyzed within the Séítah region, which is distinct from the rest of Séítah
682 and has both textural and chemical/mineralogical similarities with the Máaz formation (Udry et
683 al., this issue). Additionally, whereas Máaz has clear boundaries that define the extent of the
684 formation, Séítah does not, although the exact geographic limits and extents of the different
685 Máaz members is somewhat unclear at this time.

686 Olivine cumulates have been described in settings of layered mafic intrusions, thick lava
687 flows/ponds, shallow sills/laccoliths, or in impact melt sheets (e.g., Brown et al., 2004). In
688 terrestrial examples olivine cumulates are known to occur stratigraphically below pyroxene-rich
689 igneous rocks, as part of the same sequence. In these rocks the overall chemical trend of the
690 minerals increase in Fe, Na, and K upwards in stratigraphy. The chemical composition of the
691 overlying Máaz formation rocks may thus at first hand appear to be the natural follow-on of that
692 same petrogenetic sequence. However, our conclusion that Artuby ridge rocks represent lavas
693 with potential intercalated tephra deposits and weathering products, is not consistent with a
694 scenario where lower Máaz is part of the same petrogenetic sequence as Séítah. Analysis of
695 SuperCam data further supports our observations that it is unlikely that the Artuby ridge rocks
696 represent the same petrographic sequence as Séítah. Udry et al. (this issue) exclude the same
697 parental magma for the Séítah and Máaz formations based on the chemical composition and
698 mineralogy of the rocks. An extended period of erosion would thus be necessary to remove the
699 upper portion of the Séítah cumulate (the portion poor in olivine) since the Artuby ridge rocks
700 are not compatible with a scenario where those rocks represent that upper cumulate portion,
701 indicating that some time passed between emplacement of the Séítah and Máaz formations (see
702 also discussion in Horgan et al., this issue).

703

704 5.7 Implications for regional geology and Mars sample return

705 The Máaz formation has been proposed to be related to the capping unit mapped outside
706 of Jezero crater, the Circum-Isidis regional capping unit (e.g., Bramble et al., 2017; Hundal et al.,
707 2022). The age of this unit is likely somewhere between 3.6 and 3.96 Ga, constrained by the
708 formation of the Isidis basin (Werner, 2008) and Syrtis Major lavas (Mustard et al., 2009). A
709 lava flow is not a viable explanation as an emplacement mechanism for the whole capping unit
710 because it spans too large a range in elevation (see also discussion in Sun & Stack 2020a,b), and
711 because exposures of the unit tend to drape topography (Hundal et al., 2022). Hundal et al.
712 (2022) favors a pyroclastic depositional scenario for the capping unit because it thinly drapes
713 topography, including depressions, it is regionally uniform, has m-scale bands of variable
714 tonality, and additionally because of the stratigraphic and geomorphic similarities to the regional
715 olivine-rich unit that is proposed to be an ashfall deposit (Kremer et al., 2019; Mandon et al.,
716 2020). The capping unit is always found overlying the olivine-rich unit, suggesting a shared
717 origin (Hundal et al., 2022).

718 A pyroclastic depositional scenario for the regional unit is, however, inconsistent with the
719 scenario proposed here for lower Máaz, where the majority of the Máaz formation in the Artuby
720 ridge is lava, and also with upper Máaz, which is also interpreted to be lava (e.g., Crumpler et al.,
721 this issue; Horgan et al., 2022; Horgan et al., this issue). Our results thus indicate that either
722 Máaz is not related to the regional capping unit, or the unit mapped as the regional capping unit
723 itself consists of a number of individual geologic deposits with similar orbital morphological and
724 spectral properties, that are not necessarily related. The Rochette member has also been mapped
725 along the borders of Séítah further north of the Artuby ridge (Crumpler et al., this issue), was
726 emplaced as lava flow(s) embaying Séítah, and is thus likely the unit (perhaps together with the
727 Artuby member) that is forming the lobate margins of the “Jezero crater dark-toned floor”
728 throughout Jezero crater (Horgan et al., this issue). The upper Máaz formation, on the other hand,
729 seems to have a more limited spatial occurrence in Jezero crater (Horgan et al., this issue), and
730 was not identified as separate from the unit causing the lobate margins in earlier orbital
731 investigations (e.g., Goudge et al., 2015; Hundal et al., 2022; Stack et al., 2020). Because of the
732 presence of potential erosional surfaces between lower and upper Máaz, and within both upper
733 Máaz (Horgan et al., this issue) and lower Máaz, there may be significant time periods separating
734 emplacement of lower Máaz and upper Máaz, despite the similar chemistries. This could mean
735 that individual exposures of the regional capping unit, despite similar chemistry (i.e., spectral
736 signature), may also have been deposited with significant time gaps.

737 Alternatively, the individual geologic deposits that constitute the regional capping unit
738 may be related but were formed by different emplacement processes, for example, both effusive
739 and explosive phases from a given source. Or, in a different scenario, the capping unit could be
740 comprised of deposits from multiple, compositionally similar sources with a range of eruption
741 styles. Some of the regional outcropping units (including the Artuby formation) may have been
742 emplaced as lava flows during a major regional volcanic episode, followed by emplacement of
743 other units as pyroclastic fall, such as is indicated in the stratigraphy of Artuby ridge (Fig. 5).
744 Then, more lava flows, including the Rochette member, were emplaced. Further comparison with
745 the regional capping unit (and comparison between olivine-bearing units within and outside
746 Jezero crater) will be possible as *Perseverance* continues to traverse up and beyond the western
747 delta. Until then, the issue of the relationship between the Máaz formation and the regional
748 capping unit remains an issue to be resolved.

749 A global shift in the character of Mars volcanism has been inferred to have happened in
750 the Late Noachian/Early Hesperian (Bandfield et al., 2013; Baratoux et al., 2013; Mustard et al.,
751 2005; Poulet et al., 2009; Robbins et al., 2011), both in terms of eruptive character (from
752 generally explosive to generally effusive) and in dominant pyroxene mineralogy of the volcanic
753 products (from low-Ca to high-Ca). Although the underlying cause for this is poorly understood,
754 the observations made by *Perseverance* of the Jezero crater floor have important implications for
755 the evolution and character of the martian mantle. We show that the nature of martian volcanic
756 activity is complex and variable, at a resolution not discernable with orbital data, and that
757 effusive volcanic products dominate the Artuby and Rochette members of the Máaz formation in
758 Jezero crater, interbedded with products that are potentially compatible with explosive
759 volcanism.

760 Thin flow lobes, such as those exposed in the Artuby ridge, differ from flood basalts on
761 Earth, and may indicate relatively small eruptive fluxes (Self et al., 2021). Thin flow lobes could
762 also be an indication of being far from the source. Horgan et al. (2020) hypothesized that the
763 source for lava flows in the crater could be fissures and dikes in the crater floor that are covered
764 by the unit, or the conical and cratered edifice located on the southeastern rim of Jezero crater
765 (Fig. 1). Ravanis et al. (2022) noted the presence of enigmatic conical features located in the
766 marginal unit adjacent to the crater rim inside the crater, and Horgan et al. (this issue) reported
767 on a ~200 m high isolated edifice located just inside the northern crater rim of Jezero. The
768 presence of fine-scale layers near the top of the Artuby member could indicate a transition to
769 more explosive eruptions, followed by a hiatus of deposition and then formation of the Rochette
770 member as a lava flow. The superposition and similar chemistries of the two members argue in
771 favor of the Artuby and Rochette members being derived from a single reservoir, yet their
772 distinctly different morphologies—with Artuby having a rough, granular, texture that appears
773 almost clastic, and Rochette being more massive with vesicles—indicate some change in the
774 eruptive characteristics, perhaps through replenishment of the magma reservoir.

775 Mars Sample Return will provide a set of samples from the Jezero crater floor that when
776 analyzed on Earth will yield unprecedented insight into martian magmatic evolution. However,
777 these are limited in number compared to the complex set of layers and morphologies that we
778 observe in the Artuby ridge, and alone cannot provide the whole picture of geologic evolution.
779 Thus, establishing the geologic context is extremely important for both the interpretation of
780 samples and for increasing our understanding of martian geologic evolution.

781

782 **6 Conclusions**

783 We have presented a geomorphologic analysis of Mastcam-Z images and multispectral
784 information alongside chemical analysis by SuperCam compared with analog sites on Earth that
785 we used to interpret the origins of rocks found along the Artuby Ridge, a ~900 m scarp on the
786 floor of Jezero crater. There are two members of the Máaz formation exposed at the Artuby
787 ridge: the Artuby member, which consists of morphologically distinct units, and the Rochette
788 member, which overlies the Artuby member. Our analysis of the varied morphological
789 expression of the rocks exposed shows that they are consistent with a primary igneous deposition
790 as lava flows. This is due to the primary mineralogy, the basaltic major element composition, and
791 the presence of interlocking grains in abrasion patches, vesicles and vesicle trains, jointed
792 crystals, potential flow textures, variations between more massive, knobby appearance, and
793 layers of alternating thicknesses. There remain a number of textures which are difficult to

794 explain with an origin as purely lava flows, namely the presence of laterally continuous,
795 sometimes recessive, thin layers of alternating friability that “frame” more bulbous portions of
796 the Artuby member. We suggest that these thin layers may be the result of preferential
797 weathering of shear zones of high vesicularity layers within the lava, the interbedding of lava and
798 ash/tephra, weathered ‘a’ā clinker layers, or aeolian deposition, or alternatively, that they may be
799 hyaloclastite developed in the presence of external water, although we do not see compelling
800 evidence of emplacement in water.

801 Our results show that either (1) the Máaz formation and the regional mafic capping unit
802 are unrelated, or (2) that the two are related, but were formed via different emplacement
803 mechanisms, to explain both the observations at Artuby ridge and the broad-scale characteristics
804 of the regional unit (e.g., its draping morphology).

805 Our analyses provide geologic context for the Máaz formation samples that, upon their
806 arrival to Earth, will be analyzed with high-end laboratory equipment available at that time. Our
807 results highlight the diversity of geologic units on Mars not visible from orbit.

808

809 **Acknowledgments**

810 The authors would like to thank the Mars 2020 science and engineering teams for their efforts in
811 running the mission and in collecting data, and for discussion. A special thanks is extended to
812 Mastcam-Z Payload Downlink Leads for regenerating mosaics for this contribution. Professional
813 editorial handling by L. Montési and thorough reviews by Christopher Hamilton and one
814 anonymous reviewer is gratefully acknowledged. S. Alwmark was supported by an International
815 Postdoc grant from the Swedish Research Council (grant # 2017-06388). A. Udry was supported
816 by a NASA grant (80NSSC21K0330). The participation of O. Gasnault in SuperCam on
817 *Perseverance* was supported by CNES. D. Flannery acknowledges funding by the Australian
818 Research Council. K. B. Kinch was supported by the Carlsberg Foundation grant CF19-0023. E.
819 Cloutis acknowledges funding from the Canadian Space Agency and the Natural Sciences and
820 Engineering Research Council of Canada. B. Horgan, J. Núñez, E. Ravanis, and S. Fagents were
821 funded by NASA's Mars 2020 Project via a subcontract from the California Institute of
822 Technology/Jet Propulsion Laboratory to Arizona State University (subcontract 1511125).

823

824 **Open Research**

825 The data in this publication are from the Mastcam-Z and SuperCam instruments onboard the
826 NASA Mars 2020 *Perseverance* rover. The SuperCam data are from LIBS and RMI. The data
827 are available through the Planetary Data System Imaging Node ([https://pds-](https://pds-imaging.jpl.nasa.gov/portal/mars2020_mission.html)
828 [imaging.jpl.nasa.gov/portal/mars2020_mission.html](https://pds-imaging.jpl.nasa.gov/portal/mars2020_mission.html)), and GeoSciences Node ([https://pds-](https://pds-geosciences.wustl.edu/missions/mars2020/)
829 [geosciences.wustl.edu/missions/mars2020/](https://pds-geosciences.wustl.edu/missions/mars2020/)). For Mastcam-Z data, see Bell & Maki (2021). The
830 Mastcam-Z multispectral database from sols 0–380 is published in Rice et al. (2022b). For
831 SuperCam data, see Wiens & Maurice (2021).

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1231

1232 **Figure Captions**

1233 **Figure 1. a)** Color-coded elevation (in m) overview image of Jezero crater based on the High
1234 Resolution Imaging Science Experiment (HiRISE) digital elevation model and corresponding
1235 orthoimages (modified after Holm-Alwmark et al., 2021). **b)** Enlarged view of area of interest
1236 with Máaz and Séítah formation rocks and rover traverse in black. White dashed line indicates
1237 approximate division between the Máaz and Séítah formations. Basemap is HiRISE color image
1238 (Ferguson et al., 2020). **c)** View of upper Máaz formation Ch'ał member rocks overlaying the
1239 Naat'áanii member pavers near the Octavia E. Butler landing site. This enhanced color Mastcam-
1240 Z mosaic was obtained on sol 369 at a focal length of 34 mm. Credit: NASA/JPL-
1241 Caltech/ASU/MSSS/K. Powell. Numbers along borders of mosaics are azimuth and elevation
1242 coordinates (in relation to rover position when images were obtained). **d)** Stratigraphic profile

1243 and members of the Séítah and Máaz formations (based on Crumpler et al., this issue; Horgan et
 1244 al., this issue; Sun et al., this issue).

1245
 1246 **Figure 2. a)** Magnified view of the Artuby ridge with outcrops, workspaces, and samples
 1247 relevant to this study indicated by numbers. For full list of targets analyzed see Table S1.
 1248 Basemap is HiRISE IRB color (Ferguson et al., 2020) from CAMP (e.g., Stack et al., 2020). **b)**
 1249 Mastcam-Z sol 173 enhanced color mosaic of Artuby ridge at a focal length of 110 mm. Credit:
 1250 NASA/JPL-Caltech/ASU/MSSS/C. Rojas. **c)** Mastcam-Z sol 174 enhanced color mosaic of
 1251 Artuby ridge with the Artuby_116 outcrop at a focal length of 63 mm. Credit: NASA/JPL-
 1252 Caltech/ASU/MSSS/L. Mehall. **d)** Mastcam-Z sol 305 enhanced color mosaic of the Séítah-
 1253 Máaz transition from Lombards to Artuby ridge at a focal length of 34 mm. Credit: NASA/JPL-
 1254 Caltech/ASU/MSSS/K. Powell.

1255
 1256 **Figure 3.** Enlarged views of the Artuby_116 and Rimplas outcrops at the Artuby ridge. **a)**
 1257 Mastcam-Z left eye sol 175 image of the Artuby_116 outcrop at a focal length of 63 mm. The
 1258 whole outcrop section in this view is approximately 3 m in height. **b)** Enlarged view of knobbly
 1259 layers in the Artuby_116 outcrop. **c)** Mastcam-Z sol 342 enhanced color mosaic at a focal length
 1260 of 110 mm that shows the Artuby_116 outcrop and the Rimplas workspace to the left. Marked on
 1261 this image with numbers 1 and 2 are the two sections where we have measured layer thicknesses
 1262 (see main text). Credit: NASA/JPL-Caltech/ASU/MSSS/L. Mehall. **d)** Mastcam-Z sol 177
 1263 enhanced color mosaic of the Grasse outcrop at Artuby_116, taken at a focal length of 110 mm.
 1264 Credit: NASA/JPL-Caltech/ASU/MSSS/K. Powell. **e)** Mastcam-Z left eye image obtained on sol
 1265 342 at a focal length of 110 mm showing both bulbous morphologies and quasi-planar layers at
 1266 the Artuby_116 outcrop. **f)** Mastcam-Z sol 175 enhanced color mosaic of bulbous layers at
 1267 Artuby ridge (Artuby_116 outcrop) at a focal length of 110 mm. Credit: NASA/JPL-
 1268 Caltech/ASU/MSSS/L. Mehall. **g)** Enlarged view of bulbous morphologies of layers at the
 1269 Artuby ridge. Arrows point to places where layers appear to have flowed over underlying strata,
 1270 and where individual lobes pinch out. Mastcam-Z right eye image obtained on sol 175 at a focal
 1271 length of 110 mm. **h)** Mastcam-Z sol 343 enhanced color mosaic of the Rimplas workspace at
 1272 Artuby ridge, taken at a focal length of 110 mm. Note that the knobbly layers have internal
 1273 foliation which is not as evident on knobs that are less weathered (right side of image). Credit:
 1274 NASA/JPL-Caltech/ASU/MSSS/K. Powell.

1275
 1276 **Figure 4. a)** Mastcam-Z sol 292 enhanced color mosaic of part of the Artuby West ridge at a
 1277 focal length of 34 mm. Credit: NASA/JPL-Caltech/ASU/MSSS/L. Mehall. **b)** Mastcam-Z right
 1278 eye image of layers on Artuby West ridge obtained on sol 292 at a focal length of 110 mm. **c)**
 1279 Cropped section of Mastcam-Z left eye image obtained on sol 288 of layers on Artuby West
 1280 ridge at a focal length of 110 mm. Note what could be overlapping flow lobes (arrow) on the
 1281 right side of the regolith-covered crevasse. Or alternatively, that the top layer appears to have
 1282 slumped, or flowed, over underlying strata. **d)** Mastcam-Z left eye image obtained on sol 288 of
 1283 layers on Artuby West ridge at a focal length of 110 mm.

1284
 1285 **Figure 5.** Schematic stratigraphy of the Artuby ridge and examples of the various morphologies
 1286 and surface textures of Artuby ridge targets, with hypotheses for formation of strata summarized.
 1287 **a)** Schematic and simplified stratigraphic column of the Artuby_116 outcrop at Artuby ridge.
 1288 Letters refer to where RMIs targets in b, c, and d correspond to in stratigraphy. Sections with

1289 numbers 1 and 2 correspond roughly to where layer thickness have been measured (see also Fig.
 1290 3c). **b)** SuperCam RMI of target Amirat obtained on sol 290. **c)** SuperCam RMI of target
 1291 Majastres obtained on sol 294. **d)** SuperCam RMI of target Grasse obtained on sol 177.

1292

1293 **Figure 6.** Magnesium number and major element compositions ($\text{SiO}_2 + \text{Al}_2\text{O}_3$) of SuperCam
 1294 targets on the Artuby ridge (light gray circles), and targets located off the ridge (see Fig. 2a) that
 1295 are, or may be, in Séítah stratigraphy (dark gray circles).

1296

1297 **Figure 7.** Summary of Mastcam-Z multispectral data from Artuby ridge. **a)** Plot of all high-
 1298 quality spectra extracted from Artuby ridge, comparing spectral slopes at long and short
 1299 wavelengths with 525 nm band depth. All three parameters are high for ferric oxides, but vary
 1300 for ubiquitous purple coatings and exposed rock. Squares indicate Artuby member observations
 1301 and circles indicate Rochette member observations, which follow similar patterns. **b)**
 1302 Representative Mastcam-Z spectra, displaying the similarity between Artuby member knobs,
 1303 resistant Rochette member cap rocks, and local regolith.

1304

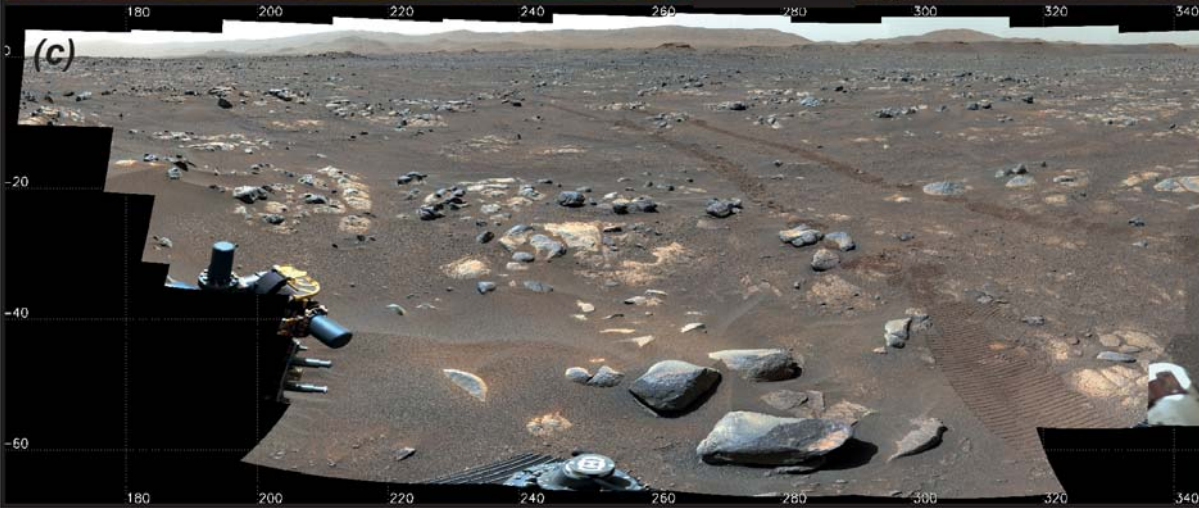
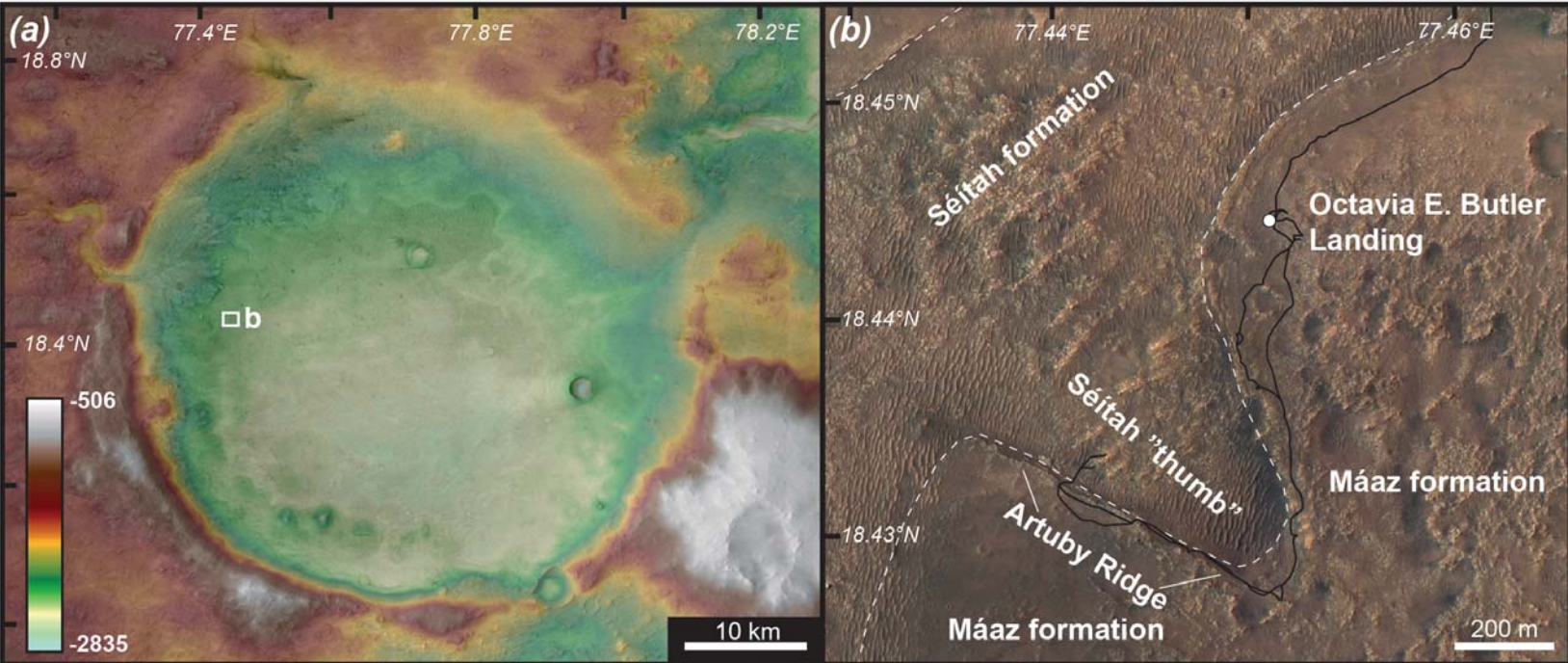
1305 **Figure 8.** Examples of complex layers, foliations, and fractures in the interior of lava flows over
 1306 a range of morphologies, compositions, and viscosities. **a)** Typical inflated pāhoehoe lava section
 1307 with massive interior and vesicular upper and lower sections. Note sub-orthogonal joints in upper
 1308 section. Laguna lava flow, New Mexico. **b)** Typical ‘a‘ā lava section with foliated basal zone and
 1309 chaotic interior fractures. Sub-alkaline basalt, Cerros del Rio volcanic field, west of Santa Fe,
 1310 New Mexico. **c)** Shearing foliations within the massive interior of a trachybasalt lava section.
 1311 Mt. Taylor volcano, New Mexico. **d)** Basal structure of an ‘a‘ā trachybasalt lava flow, note the
 1312 relatively vesicle-free and massive interior underlain by a scoriaceous basal breccia. The basal
 1313 scoria results from combined over-riding of brittle chill zone fragments, basal initial melt
 1314 vesicularity, and local pyroclastic fallout. Mt. Taylor volcanic field, New Mexico. **e)** Inflated
 1315 tholeiitic pāhoehoe flow section with interior late-stage tube-like ellipsoidal zone. Note
 1316 orthogonal jointing in upper section. McCartys lava flow, New Mexico. **f)** Basal shear foliations
 1317 at the base of a thick massive, columnar zone within a porphyritic trachyandesite lava flow. Mt.
 1318 Taylor volcano, New Mexico.

1319

1320 **Figure 9.** Photographs of volcanic terrestrial analog sites compared to morphologies of the
 1321 Artuby ridge. **a)** Bulbous textures in weathered/exposed stack of pāhoehoe lobes on Oahu. **b)**
 1322 Blocky lava outcrop atop laterally discontinuous pāhoehoe lobes on Oahu. **c)** Road cut through a
 1323 stack of flows on Oahu. **d)** Magnified view of the same flow stack as in a), highlighting the
 1324 variable friability of the layers and the complex lobe/intraflow textures with relatively fine-scale
 1325 foliation.

1326

Figure 1.



(d) Stratigraphic column showing the relationship between the Mááz formation (Mááz fm) and the Séítah formation (Séítah fm).

Mááz fm	Séítah fm
Ch'at	Content
Naat'áanii	?
Roubion	
Rochette	
Artuby	
	Issole
	Bastide

Figure 2.

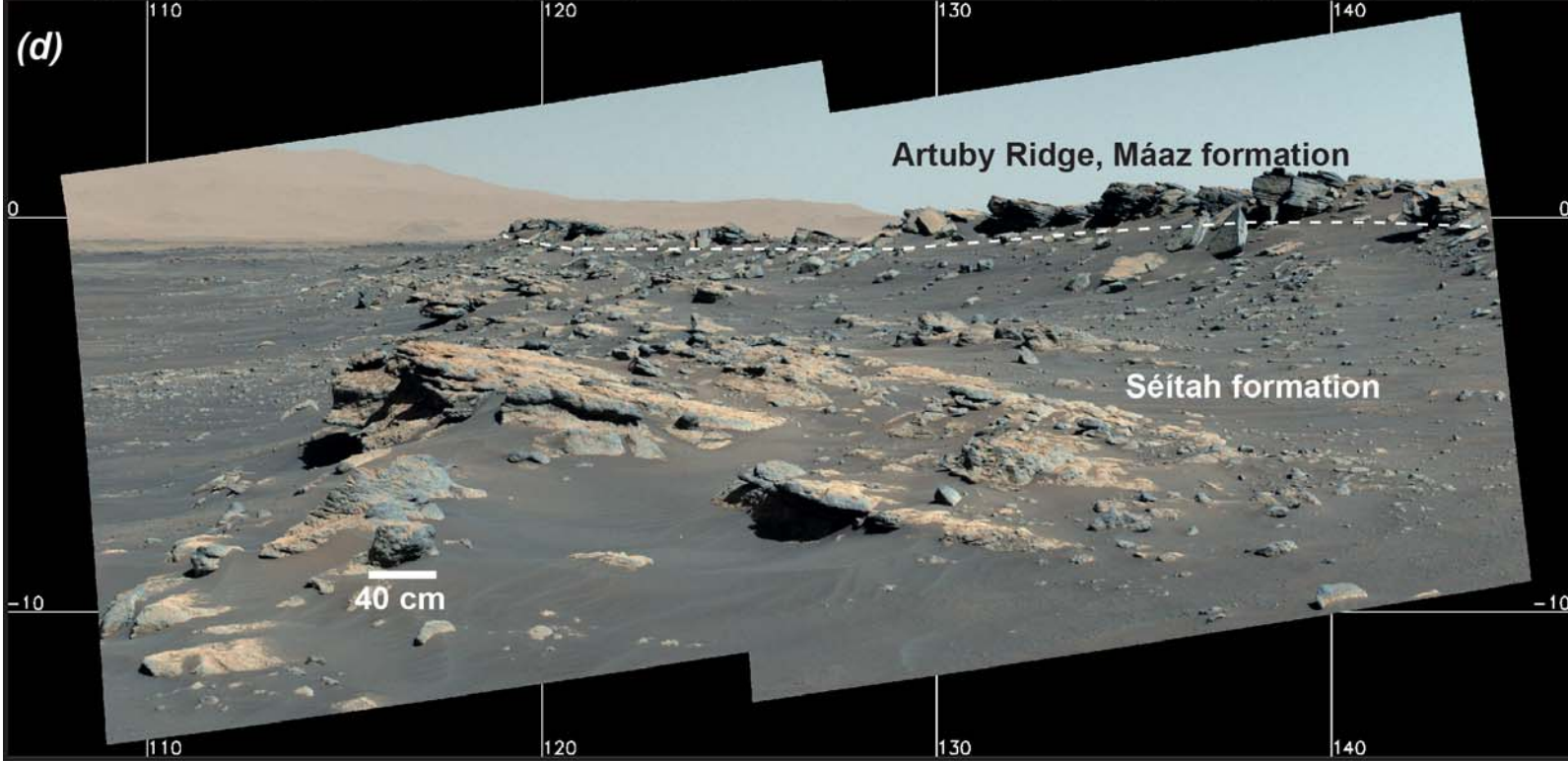
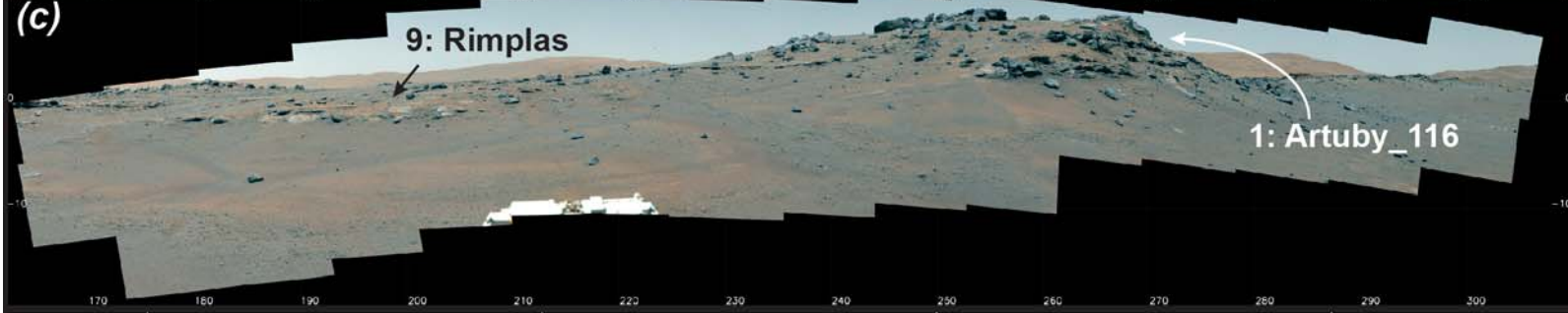
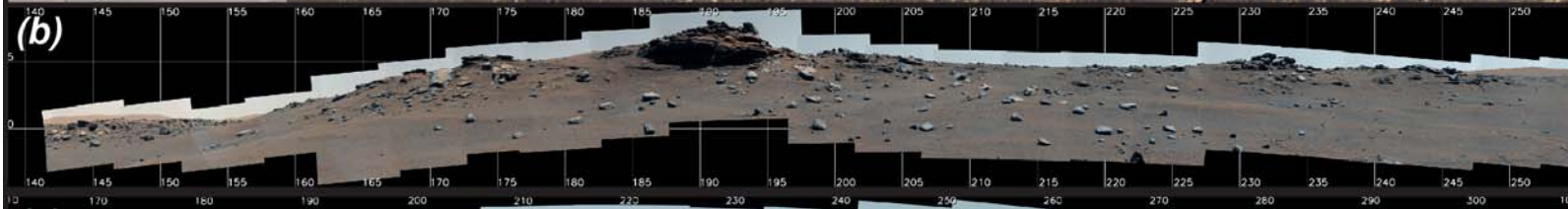
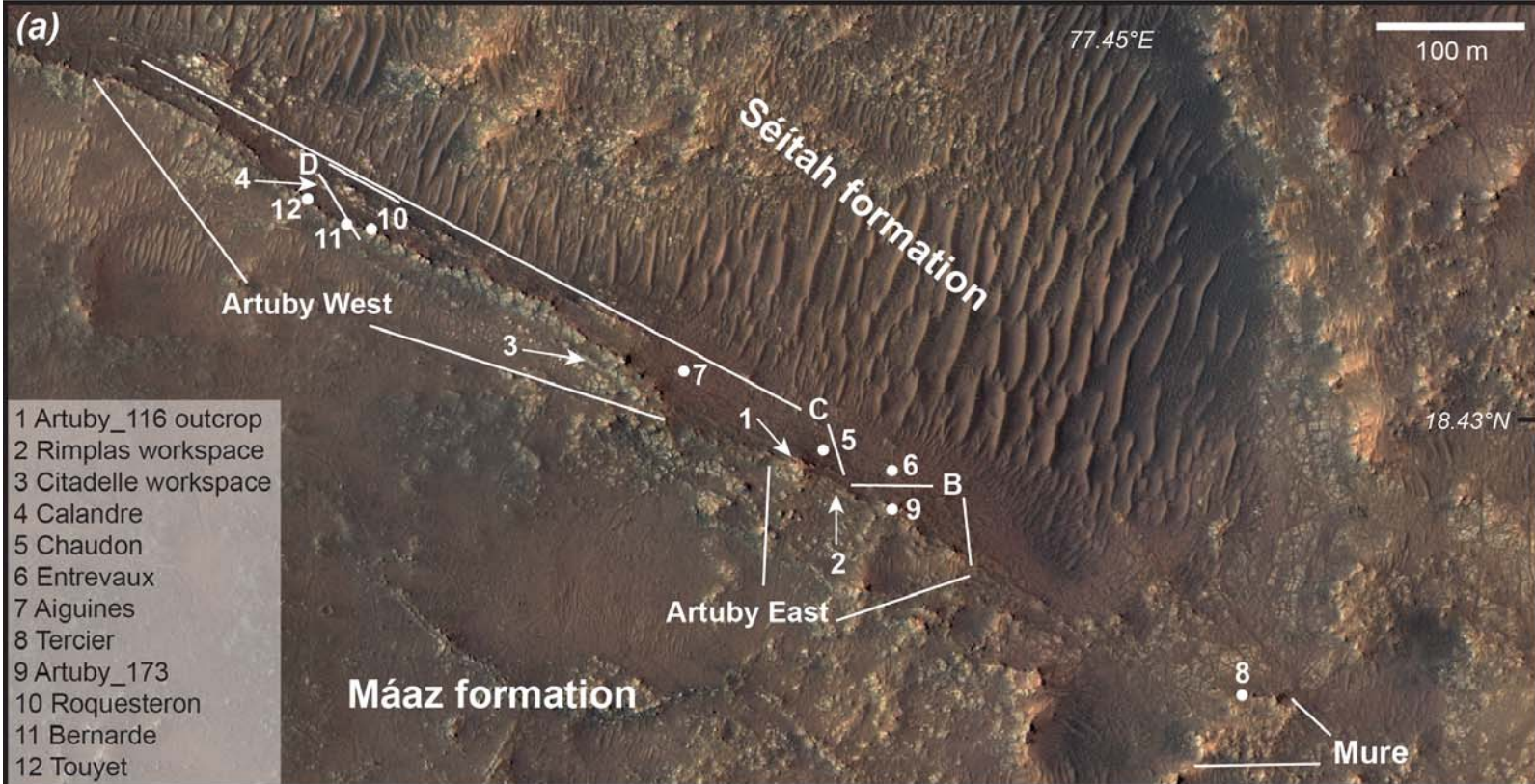


Figure 3.

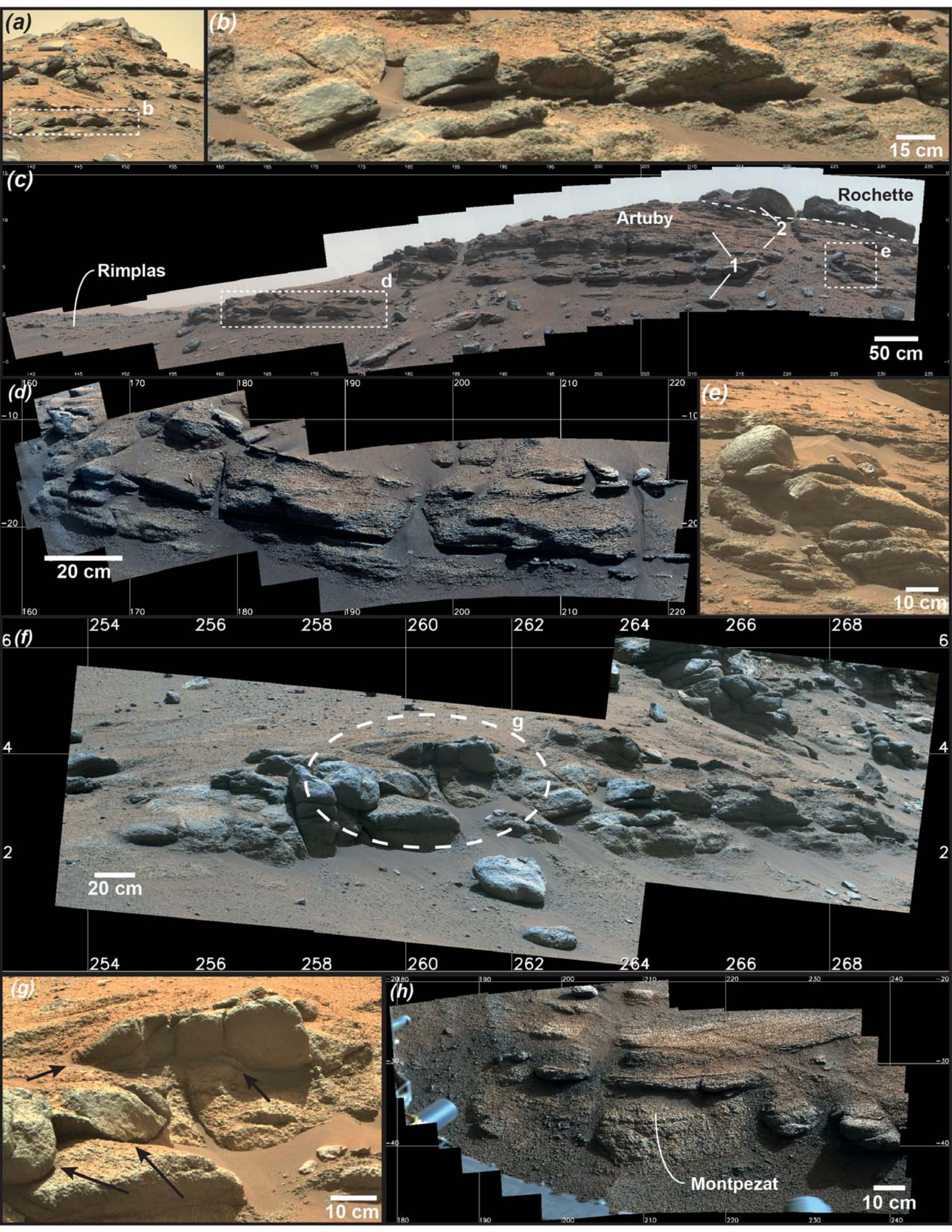


Figure 4.

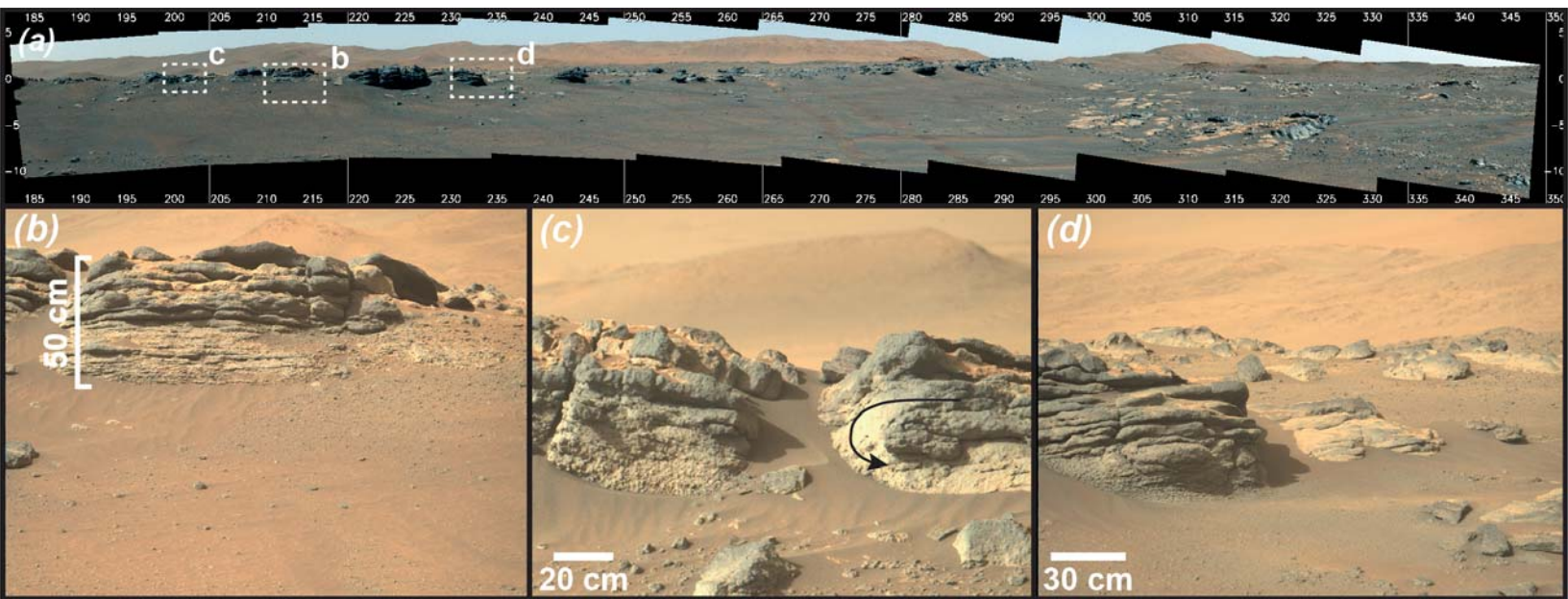


Figure 5.

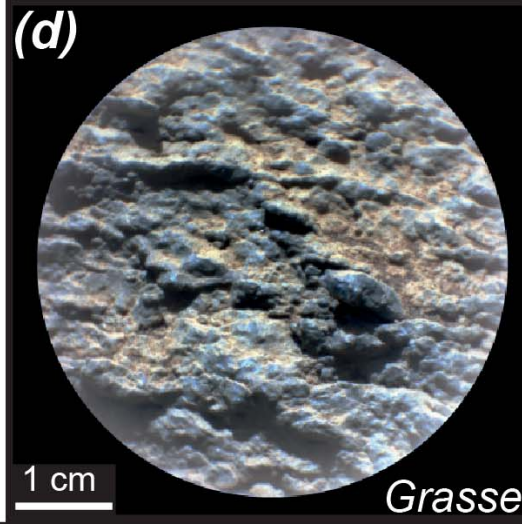
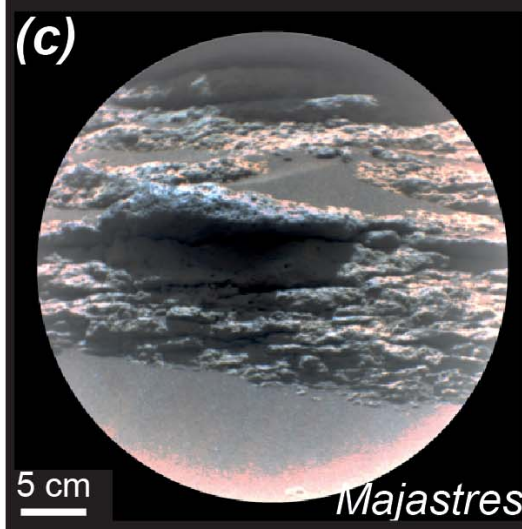
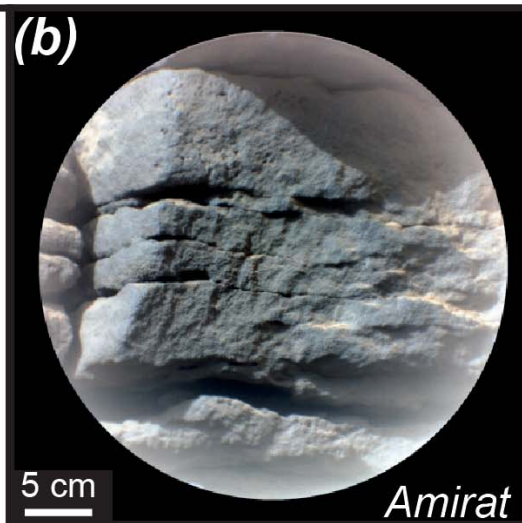
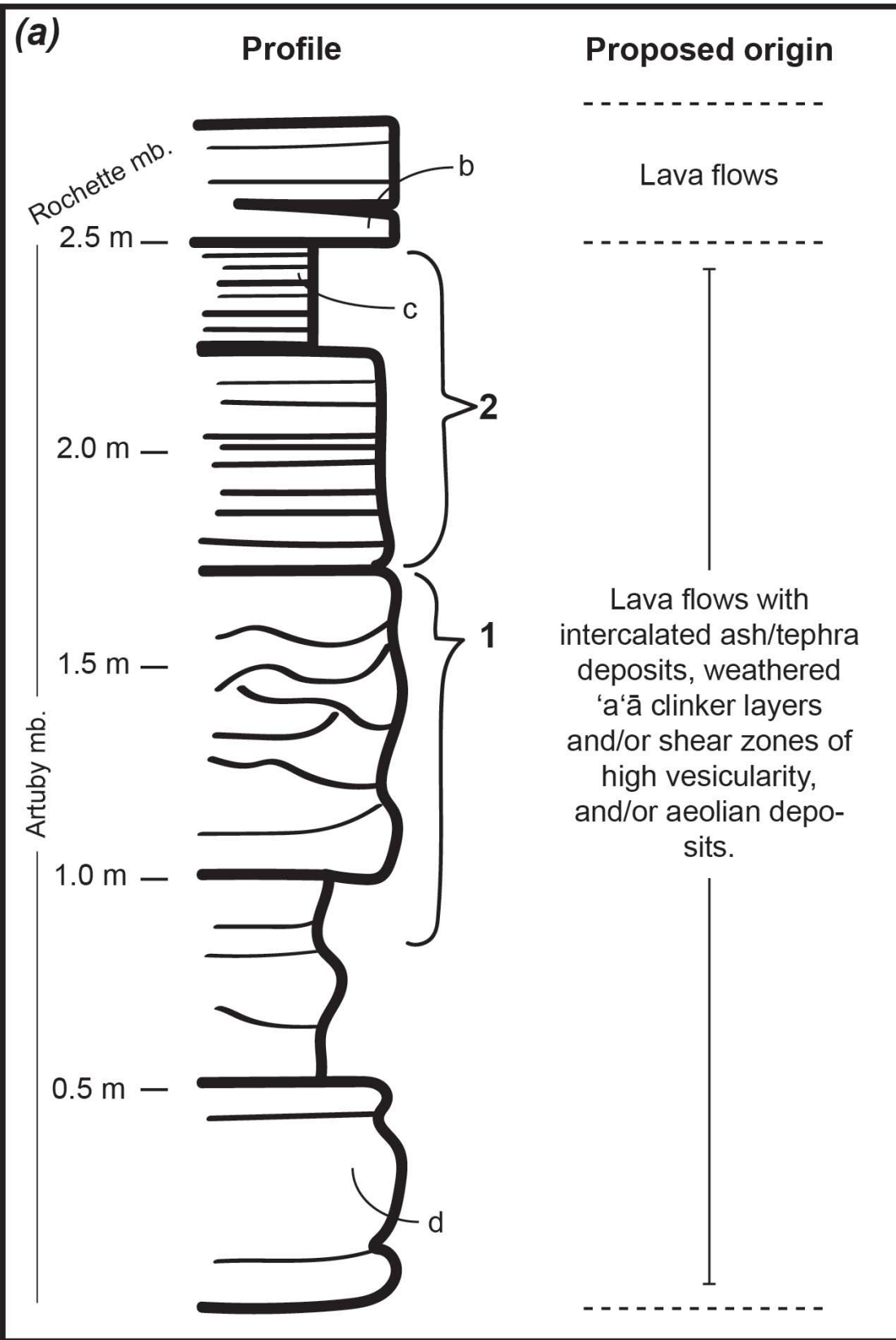
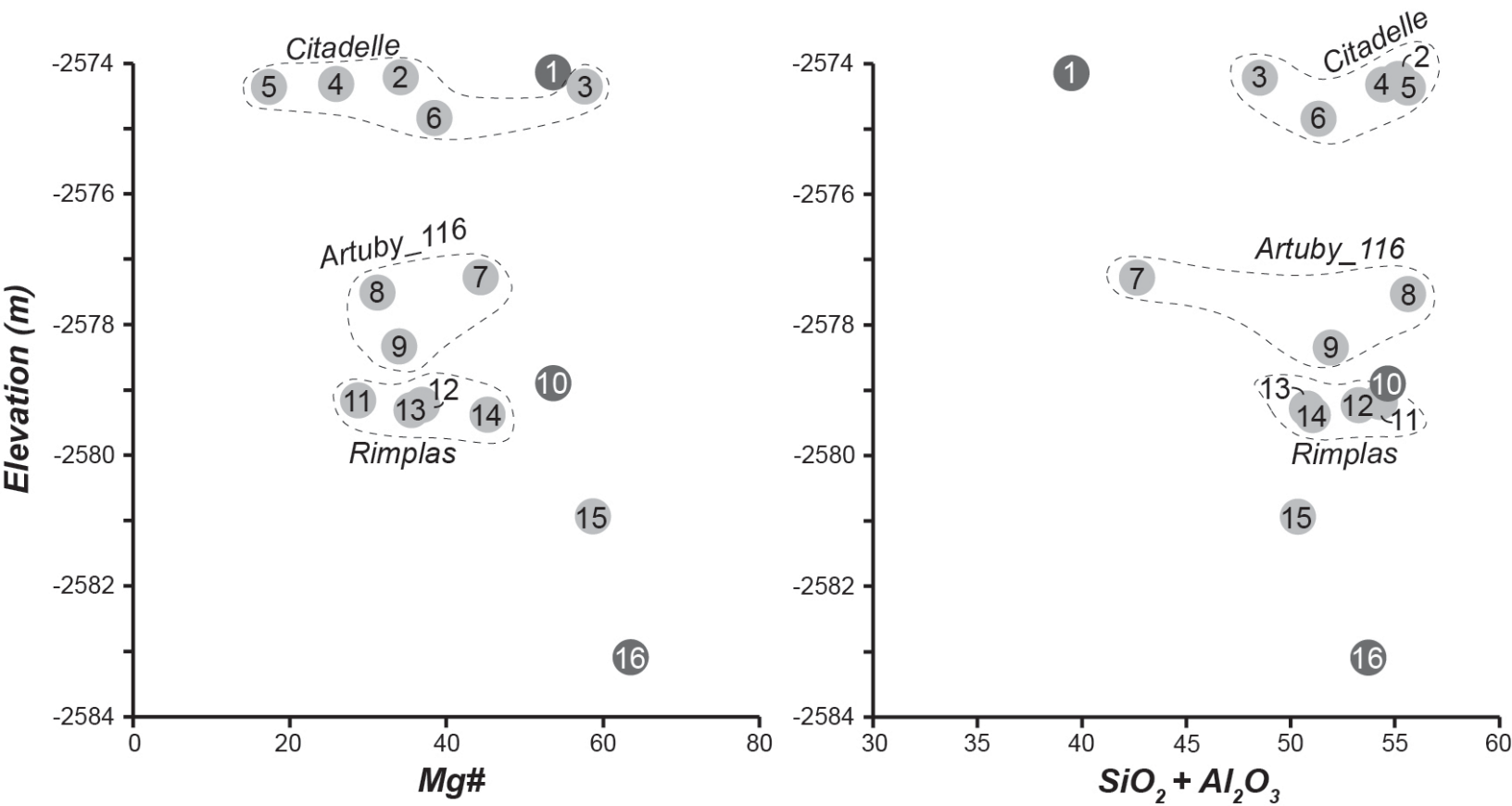


Figure 6.



- (1) Calandre, (2) Sauzeries_Haute, (3) Grandes_Tours_du_Lac, (4) Estoublon, (5) Manior, (6) Rigaud, (7) Vacluse, (8) Auribeau, (9) Grasse, (10) Aiguines, (11) Blache, (12) Vergons, (13) Montpezat, (14) Galabre, (15) Chaudon, (16) Entrevaux

Figure 7.

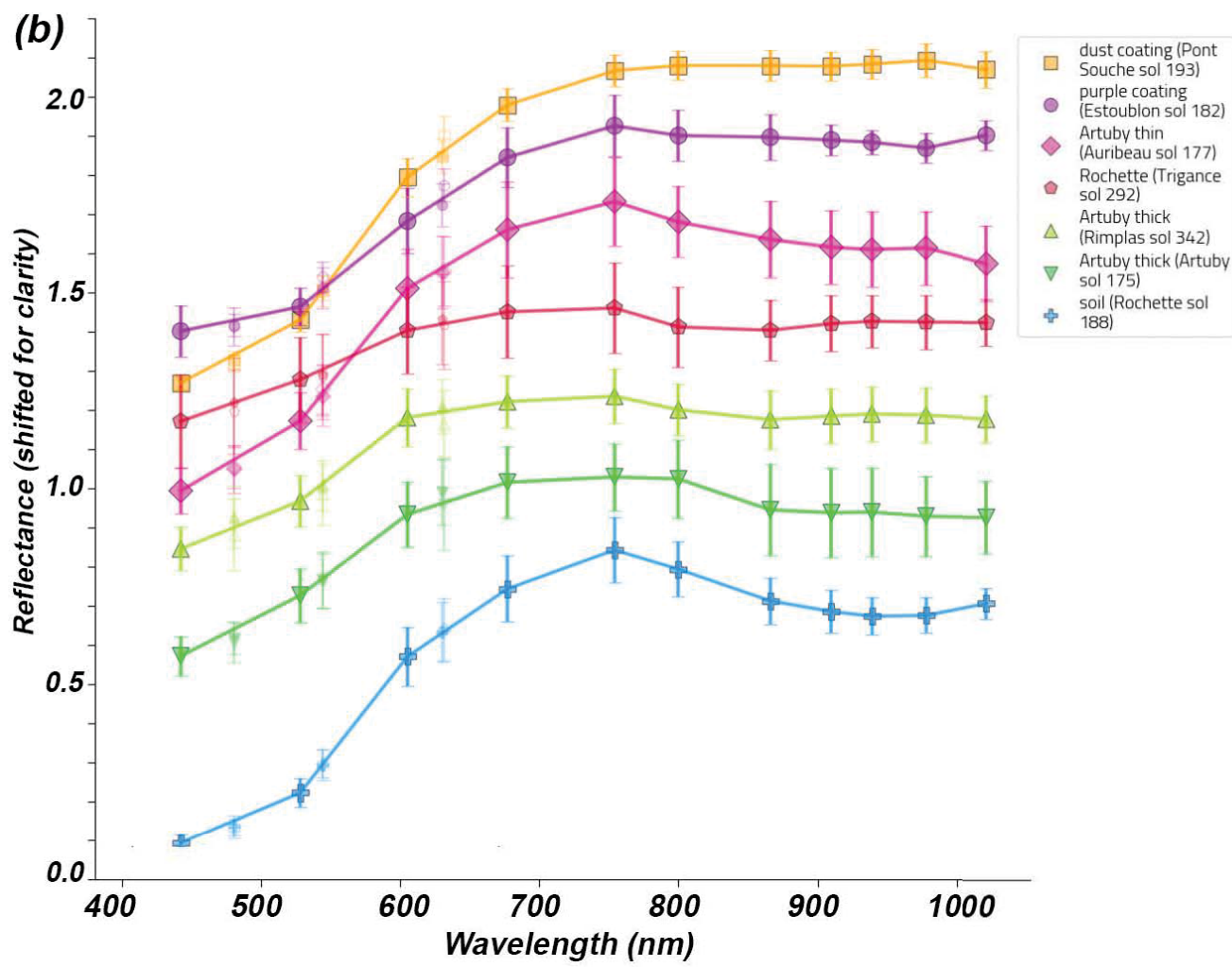
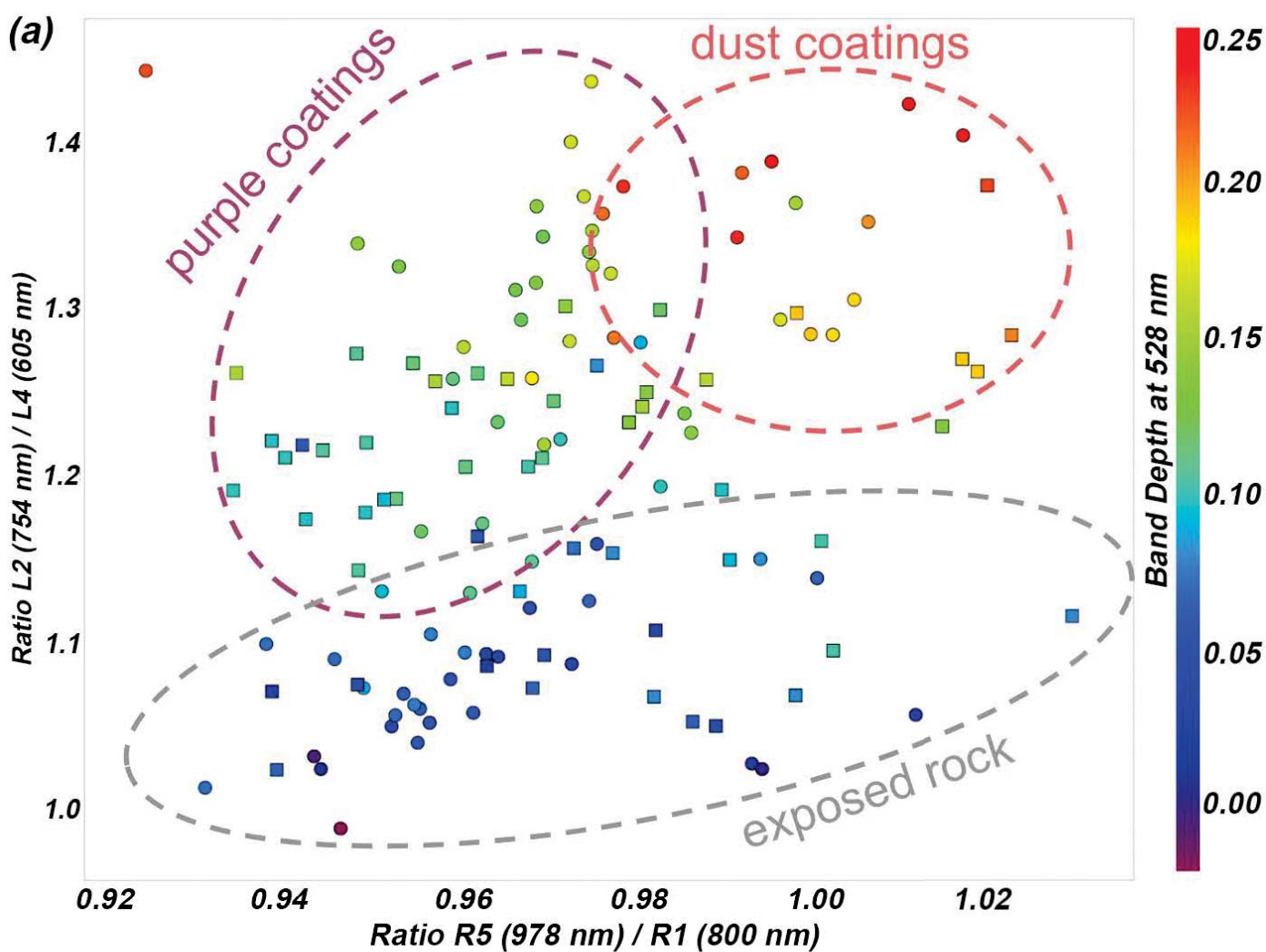


Figure 8.

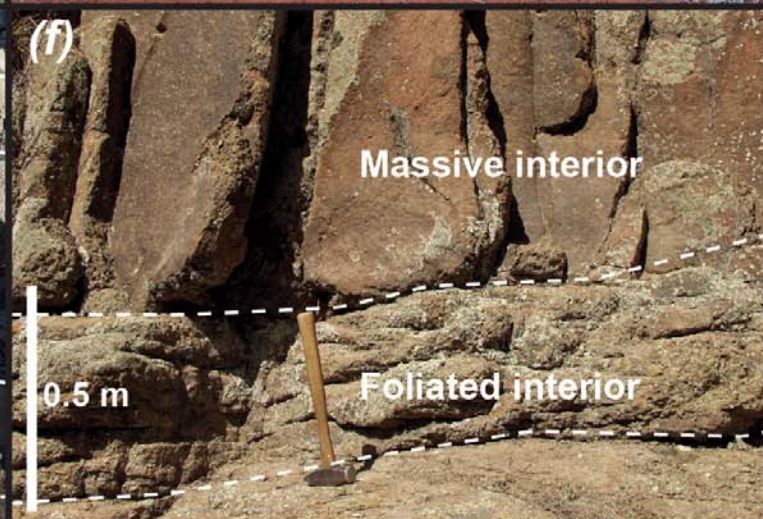
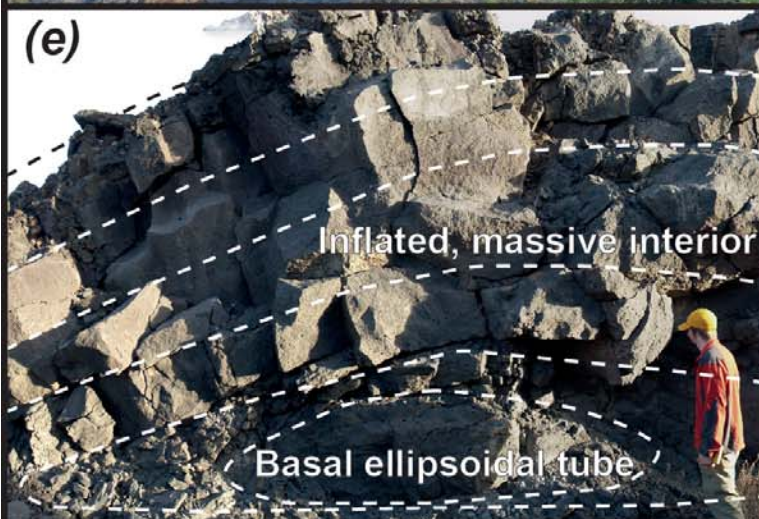
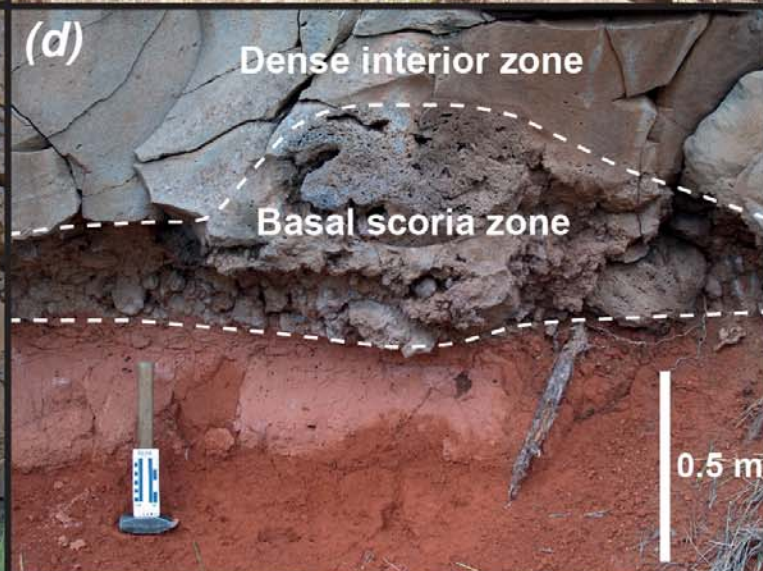
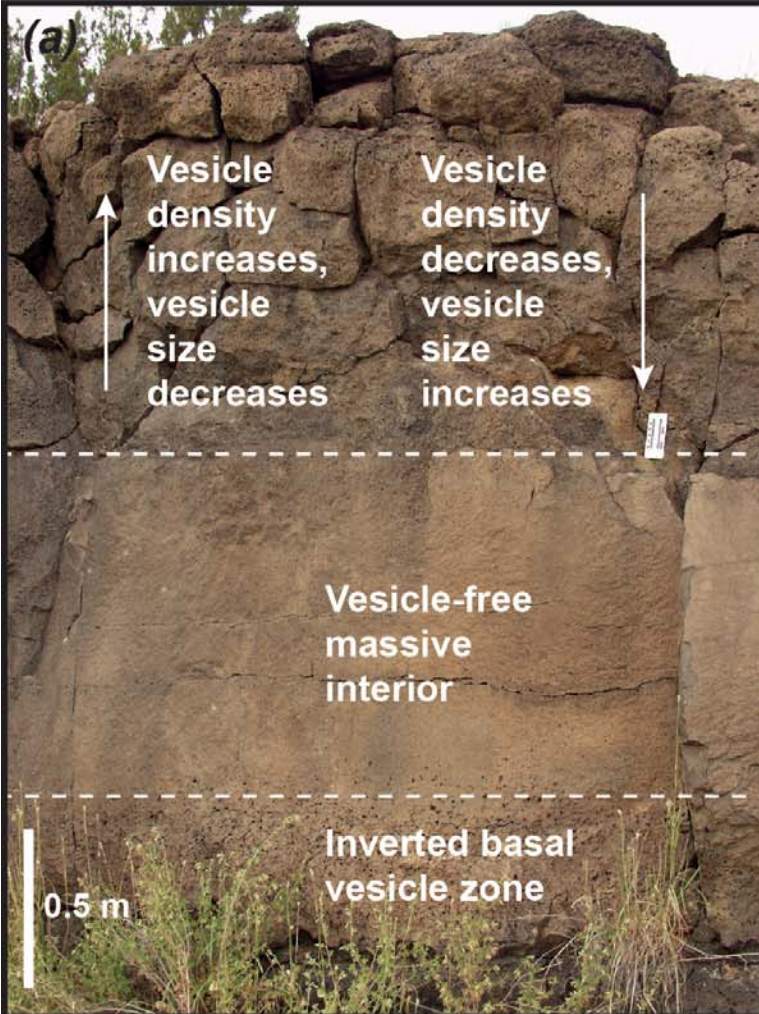


Figure 9.

