

1 **Areally extensive surface bedrock exposures on Mars: Many are clastic** 2 **rocks, not lavas**

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15 **Key points:**

16 1. Many bedrock plains are likely composed of mechanically weak rocks

17 2. Potential origins include lithified detrital sediments, pyroclastics, or impact-generated
18 materials

19 3. High thermal inertia may indicate relatively friable rocks, due to ease of comminution
20 product removal and exposure of lithified surface
21

22 **Abstract**

23 Areally extensive exposures of intact olivine/pyroxene-enriched rock, as well as
24 feldspar-enriched rock, are found in isolated locations throughout the Martian highlands.
25 The petrogenetic origin(s) of these rock units are not well understood, but some previous
26 studies favored an effusive volcanic origin partly on the basis of distinctive composition and
27 relatively high thermal inertia. Here we show that the regolith development, crater
28 retention, and morphological characteristics for many of these “bedrock plains” are not
29 consistent with competent lavas, and reinterpret the high thermal inertia orbital signatures
30 to represent friable materials that are more easily kept free of comminution products
31 through aeolian activity. Candidate origins include pyroclastic rocks, impact-generated
32 materials, or detrital sedimentary rocks. Olivine/pyroxene enrichments in bedrock plains
33 relative to surrounding materials could have potentially formed through deflation and
34 preferential removal of plagioclase.

35 **1. Introduction**

36 The Martian cratered highlands host numerous surface exposures of intact rock,
37 identified by morphologies that indicate lithified materials (scarp-forming, wind-eroded

38 surfaces) and by their high thermal inertia values relative to average surfaces on Mars (e.g.
39 Edwards et al., 2009). Thermal inertia (TI) is defined as $(k\rho c)^{1/2}$, where k is the bulk thermal
40 conductivity, ρ is the bulk density, and c is the specific heat of the material (Kieffer et al.,
41 1977). On Mars, TI is strongly controlled by particle size, porosity and compaction. Typical
42 Martian surfaces, which are dominated by dust- to sand-sized particles, exhibit TI values
43 between 28 and 355 $\text{J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$ (Putzig et al., 2005); bedrock, compacted and/or coarse
44 particulate surfaces exhibit higher TI values.

45 TI values are typically modeled from nighttime surface temperature measurements,
46 for example from the Mars Global Surveyor Thermal Emission Spectrometer (TES) (~ 5
47 km/pixel, Christensen et al., 2001) or Mars Odyssey Thermal Emission Imaging System
48 (THEMIS) (100 m/pixel, Christensen et al., 2004a). Using THEMIS, bedrock can be spatially
49 resolved. Given the uncertainties in partial sediment cover, as well as in atmospheric dust
50 opacity at the time of data acquisition, most studies have not used a strict THEMIS TI
51 threshold to define bedrock. Rather, bedrock is usually identified using a combination of TES
52 $\text{TI} > 350 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$, a relatively higher THEMIS nighttime radiance, and morphological
53 expressions of lithified material (e.g. eroded surfaces, scarps) (e.g. Rogers and Nazarian,
54 2013). We do note that with one exception, the exposures discussed in this work have at
55 least a portion of the bedrock exposure with THEMIS TI values $> 493 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$, from the
56 THEMIS global TI mosaic (Christensen et al., 2013). Bedrock exposures have been identified
57 in various geologic contexts, including flat plains, crater/canyon walls, and canyon floors
58 (Edwards et al., 2009). Our focus here is on the dozens of areally extensive, flat exposures,
59 hereafter referred to as “bedrock plains” (**Figure 1**). We review key details about these
60 surfaces below.

61 Bedrock plains are most commonly found in topographic lows of intercrater surfaces
62 of heavily cratered terrain or as graben- or crater floor-filling materials, and can exceed $\sim 10^4$
63 sq km in area (Rogers et al., 2009; Edwards et al., 2009; Rogers and Fergason, 2011; Rogers
64 and Nazarian, 2013). Portions of the bedrock plains are overlain by relatively lower-TI
65 materials, and are surrounded by lower-TI surfaces previously interpreted as megaregolith,
66 crater ejecta, and/or pyroclastic materials (Rogers et al., 2009; Bandfield et al., 2013) (e.g.
67 **Figure 1b-c**). At the decameter scale, the intercrater and crater-filling bedrock plains are
68 fractured, lack evidence for fine-scale layering, and range from flat/smooth to rugged, with

69 apparent topographic relief. The formation ages of many of the intercrater bedrock plains
70 are likely between Middle and Late Noachian and late Noachian/Early Hesperian, based on
71 stratigraphic relationships for some units (e.g. **Figure 2**) as well as good spatial
72 correspondence with the “Late Noachian highlands” unit mapped by Tanaka et al. (2014).
73 But some units are too small (order of $<1,000 \text{ km}^2$) to demonstrate Noachian formation ages
74 and/or have no dateable cross-cutting units, and thus could be younger than Early
75 Hesperian.

76 Though intercrater plains and crater floors are the most common contexts for
77 bedrock plains, other bedrock plains include the sulfate-bearing “etched unit” of Terra
78 Meridiani (Hynek et al., 2002; Arvidson et al., 2005), the floor materials of Nili Patera caldera
79 (Christensen et al., 2005), the intermontane regions of Libya Montes (Christensen et al.,
80 2004b), and a fractured, banded plateau in the Nili Fossae region (Hamilton and Christensen,
81 2005). The Nili Fossae bedrock plain is of particular interest because it is near two of the
82 proposed landing sites for the Mars 2020 rover: Jezero crater (Goudge et al., 2015) and
83 Northeast Syrtis (Bramble et al., 2017).

84 With the exception of Terra Meridiani, bedrock plains are typically enriched in olivine
85 and/or pyroxene compared to surrounding low-TI surfaces, determined through analyses of
86 infrared spectra (Rogers et al., 2009; Rogers and Ferguson, 2011; Loizeau et al., 2012; Ody et
87 al., 2012; Edwards et al., 2014). Spectral evidence of secondary minerals in
88 olivine/pyroxene-enriched bedrock plains is thus far undiscovered, with the exception of the
89 Nili Fossae olivine-bearing unit, which is altered in places (Ehlmann et al., 2008). Some
90 bedrock plains contain feldspathic rocks (Wray et al., 2013; Carter et al., 2013; Rogers and
91 Nekvasil, 2015), with variable alteration (Wray et al., 2013). Last, it should be noted that the
92 observable characteristics of bedrock plains differ significantly from the Hellas basin rim
93 intercrater plains sedimentary units described by Salese et al. (2016), which show
94 subhorizontal bedding, alteration minerals, no olivine enrichment, and relatively low TI
95 values.

96 The formational mechanism(s) of bedrock plains are not well understood. Previous
97 studies of the intercrater surfaces and crater/graben-filling high-TI units favored an effusive
98 volcanic origin for these materials, primarily based on: the distinctive compositions
99 compared to surroundings (Rogers et al., 2009; Ody et al., 2012; Rogers and Nazarian, 2013;

100 Edwards et al., 2014), the mare-like appearance at THEMIS resolution (smooth plains with
101 wrinkle ridges), and relatively high TI values ($>1200 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$) in a few locations (Rogers
102 et al., 2009; Rogers and Nazarian, 2013; Edwards et al., 2014). Detrital sedimentary origins
103 were less favored due to the lack of olivine-bearing source regions for the olivine sediments
104 (e.g. Rogers and Nazarian, 2013); although some spectrally-undistinctive crater filling
105 bedrock materials were interpreted as sedimentary (McDowell and Hamilton, 2007).
106 Olivine-enriched bedrock plains in Nili Fossae, Libya Montes and Nili Patera have been
107 interpreted as effusive volcanics (Christensen et al., 2004b; Hamilton and Christensen, 2005;
108 Tornabene et al., 2008) or alternatively for Nili Fossae and Libya Montes, impact melts
109 (Mustard et al., 2007).

110 In this work, we re-examine intercrater and crater/graben-filling bedrock plains with
111 a focus on high-resolution morphologies and crater retention, as well as by comparing the
112 thermophysical characteristics of these distinctive units with known volcanic plains in
113 Hesperia Planum and Syrtis Major. Our observations suggest that many bedrock plains are
114 relatively friable materials, consistent with clastic rocks, rather than lavas.

115 **2. Observations**

116 **2.1 Bedrock plains lack thick regolith cover, unlike Hesperian volcanic plains**

117 Hesperian volcanic plains are extensive, flat-lying units thought to have formed
118 during high effusion rate, fissure-fed eruptions (Greeley and Spudis, 1981). Crater/graben
119 wall exposures and ejecta from small diameter craters found in Hesperian volcanic plains
120 exhibit high TI values and/or visibly blocky materials, suggesting mechanically strong
121 materials at depth, consistent with lavas (Bandfield et al., 2013; Warner et al., 2017).
122 Hesperia Planum and Syrtis Major serve as useful volcanic plains reference surfaces because
123 they are at equatorial latitudes (thus less affected by Amazonian periglacial reworking), are
124 not dust mantled, and in some areas directly contact older, bedrock plains exposures.

125 Hesperia Planum and Syrtis Major typically exhibit TES TI values ($170\text{-}310 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$)
126 consistent with fine to coarse sand or a mixture of dust and coarser materials; plains
127 bedrock exposures within these two regions are rare (**Figure 1**). In Hesperia Planum, typical
128 morphologies observed in high resolution imagery are bedforms or smooth, featureless
129 surfaces, suggesting a surficial layer dominated by unconsolidated sediment. These
130 observations indicate that the Hesperian lavas are covered with a regolith, similar in grain

131 size and unit thickness (~1 to 10 m) to Hesperian volcanic units in Gusev crater and Elysium
132 Planitia (Golombek et al., 2006a; 2017; Warner et al., 2017).

133 In the absence of surface processes that remove (e.g. eolian or fluvial resurfacing) or
134 prevent regolith from forming (e.g. burial), a thick regolith is expected for
135 Noachian/Hesperian-age bedrock exposed to repeated impact events (Hartmann et al.,
136 2001). However, a thick regolith is not present on the bedrock plains, as indicated by the
137 relatively higher TI values and morphological indicators of exposed rock (**Section 1**). Why is
138 this? Increased strength or duration of erosional processes (e.g. wind, fluvial) on bedrock
139 plains could potentially explain the minimal regolith cover. However, preferential erosional
140 strength/duration cannot explain all of the bedrock occurrences because there are locations
141 where regolith-covered Hesperian lavas directly contact or are in close spatial and
142 topographic proximity to bedrock plains. For example, at the southern margin of Syrtis Major
143 Planum, bedrock plains are directly subjacent to the Hesperian lavas. A striking difference in
144 TI and sediment cover is observed across this boundary (**Figure 2a-b**), indicating a direct
145 relationship between regolith cover and its underlying source unit. It is unlikely that long-
146 term landscape modification by eolian processes would preserve such a well-defined contact
147 if the two units were similarly resistant to erosion and exposed to the same surface
148 processes. Even if denudation of the landscape occurred prior to emplacement of the
149 Hesperian lavas, a meters-thick regolith should have developed on the denuded surface after
150 that event(s). This, however, is not observed. A sharp difference in TI is also found at the
151 eastern margin of Hesperia Planum, where low-TI, regolith-covered Hesperian lavas contact
152 bedrock plains (**Figure 1c**). This example is discussed further in **Section 3.2**.

153 We hypothesize that the difference in regolith cover across the bedrock plains-
154 Hesperian volcanics contact is related to differences in material properties, where the
155 bedrock plains represent mechanically weak materials relative to Hesperian lavas.
156 Comminution products from mechanically weak materials would be expected to include few
157 blocks and a larger proportion of fine-to-medium sand-sized particles (e.g. Malin and Edgett,
158 2000; Golombek et al., 2006b, 2010) that are easily moved by wind (Greeley et al., 1980). In
159 contrast, regolith developed from mechanically strong materials would include a larger
160 proportion of blocks and coarser particulate material, as observed at other lava plains
161 localities where surface processes are limited to impact gardening and eolian modification

162 (Golombek et al. 2006; Warner et al. 2017). Over time, this would lead to buildup of a regolith
163 dominated by unconsolidated materials that were not mobilized by wind (e.g. coarse sand
164 and larger, as well as subsequently trapped dust) on competent surfaces (e.g. Golombek et
165 al., 2006a, 2017), whereas mechanically weak materials (perhaps consisting of weakly
166 consolidated, dominantly fine-to-medium sand-sized clasts) would experience constant
167 deflation, exposing a lithified surface.

168 This hypothesis of material properties controlling regolith thickness is founded in
169 similar observations of known clastic rocks elsewhere on Mars. For example, Noachian
170 sulfate-bearing sandstones in Meridiani Planum exhibit an Amazonian exposure age, which
171 was attributed to relatively rapid resurfacing from eolian scour of highly erodible rocks
172 (Golombek et al., 2006b, 2014). The Amazonian exposure age and comparatively high
173 erosion rates of finely layered units in Valles Marineris and Arabia Terra have also been
174 attributed to their friable nature (Malin and Edgett, 2000; Grindrod and Warner, 2014). Last,
175 the Columbia Hills of Gusev crater are dominated by clastic rocks, and exhibit less regolith
176 cover than the adjacent (and younger) Hesperian plains (Grant et al., 2006). The Gusev
177 example is described in more detail below, building on the findings of Grant et al. (2006) but
178 with a focus on the orbital TI signatures.

179 In Gusev crater, Mars Exploration Rover observations show that regions of the
180 Columbia Hills exhibit less regolith cover compared to the superjacent Hesperian plains
181 (Grant et al., 2006). This is consistent with TI measurements from orbit, where the Columbia
182 Hills exhibit relatively high THEMIS TI ($\sim 350\text{--}500 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$), compared to the Hesperian
183 basaltic unit ($\sim 180\text{--}210 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$) (**Figure 2c**). In contrast, *in-situ* TI measurements from
184 individual blocks of each of these units show that the Hesperian basalts exhibit TI values of
185 $\sim 1200 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$, which is higher than TI values from rocks of the Columbia Hills (~ 600
186 $\text{J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$) (Fergason et al., 2006). The areally dominant Columbia Hills rock type causing
187 the high TI signature from orbit is likely the “Algonquin”/“Comanche” class, interpreted as
188 variably altered olivine-bearing basaltic tephra (Ruff et al., 2014). To the southwest of the
189 Columbia Hills, materials with similar texture and TI to the Algonquin/Comanche rocks
190 outcrop through windows in the Hesperian basaltic unit (**Figure 2c**), and may be lateral
191 extensions of the Algonquin class (Ruff et al., 2014). The Algonquin class rocks would likely
192 be mechanically weak compared to Gusev plains lavas (e.g. Thomson et al., 2013). The

193 gradient in TI between the Hesperian basaltic unit and the stratigraphically lower unit is
194 sharp; the extent of the regolith cover closely corresponds with the margins of the Hesperian
195 units (**Figure 2c**), suggesting control by the material properties of both units.

196 **2.2 Morphological observations and crater retention**

197 Bedrock plains commonly exhibit parallel to sub-parallel striations that resemble
198 yardangs or wind-eroded morphologies, suggestive of friable materials (**Figure 3a-b**). On
199 Earth, yardangs are typically observed in friable units and are rarely observed in crystalline
200 rocks such as basalts, but examples do exist (Inbar and Risso, 2001). Nonetheless, other
201 morphological indicators of friability, such as outcrops with smooth or scalloped textures,
202 are present in some bedrock plains (**Figure 3c-d**).

203 Lavas are competent, high shear strength materials that retain small craters through
204 increased resistance to comminution, erosion, and diffusive slope processes. In contrast,
205 friable materials do not retain small craters as well, and are subject to faster erosion rates
206 than armored, rocky basaltic surfaces by 1 to 3 orders of magnitude (Golombek and Bridges,
207 2000; Golombek et al., 2006a,b; 2014; Sweeney et al., 2016). We investigated cumulative
208 crater frequency as a function of crater diameter (>200 m) for nine bedrock units, and
209 compared these densities with those from nearby lower-TI surfaces of roughly equal area
210 (**Text S1**). Low-TI surfaces were chosen from the same or younger global
211 chronostratigraphic unit (defined by Tanaka et al., 2014) as the bedrock surface. Low-TI
212 surfaces from Early, Middle or Late Noachian highlands chronostratigraphic units (eNh,
213 mNh, lNh) were presumed to consist of regolith derived from ancient basaltic crust, and were
214 chosen as close in elevation as possible to the bedrock units (<300 m) to reduce possible
215 influence of differences in slope modifying processes and wind activity over the two surfaces
216 (**Table S1**).

217 The bedrock plains show between 18 and 78% lower crater density than adjacent
218 low-TI surfaces (**Table S2, Figure 3e**), and in addition, small craters commonly appear less
219 well-preserved on the bedrock (**Figure 3a-b, Figure S10**). Furthermore, except for regions
220 2 and 3, the crater frequency curves for the bedrock units exhibit shallower slopes than the
221 curves for the low-TI units, particularly for diameter bins below ~500-700 m (varies by
222 region). This is consistent with stronger resurfacing on the bedrock units and easier removal
223 of craters at or below ~500-700 m diameter (note that we ignore possible crater scaling

224 effects with target properties, van der Bogert, 2017; **Text S1**). We caution that differences in
225 crater populations across these two surface types could also arise from spatial differences in
226 wind strength; this could be tested with mesoscale atmospheric modeling. However, in
227 general, these observations suggest that bedrock plains do not preserve small craters as well
228 as adjacent low-TI surfaces.

229 **3. Discussion**

230 **3.1 Does exposed rock always indicate friable rock?**

231 Impact comminution should result in at least a meter of regolith for Late Amazonian
232 surfaces, increasing to tens to hundreds of meters thickness for Hesperian and Noachian
233 surfaces (Hartmann et al., 2001). Though nearly regolith-free ancient surfaces could arise
234 from weak mechanical strength and friability, combined with erosion (e.g. Malin and Edgett,
235 2000; Golombek et al., 2014) (**Section 2**), exposures of Noachian/Hesperian *competent*
236 bedrock could occur through other scenarios. For example, Noachian surfaces that were
237 rapidly buried would have been protected from impact comminution as long as the burial
238 cover was present; later exhumation would result in exposed rock (Hartmann et al., 2001).
239 Exhumation would have had to occur in the late Amazonian, otherwise a ~meters-thick
240 regolith would have subsequently developed (Hartmann et al., 2001). One potential example
241 of where effusive volcanic bedrock may have been protected through burial is at the eastern
242 margin of Hesperia Planum (**Figure 1c**), where there is minimal difference in small crater
243 preservation between the bedrock plains and the Hesperian lavas (**Table S2**).

244 In some cases, fluvial erosion and later wind activity could have helped to expose
245 competent bedrock. Olivine-enriched bedrock in Ares Vallis (Rogers et al., 2005), which
246 exhibits high retention of small diameter craters (**Figure S11**), may be an example of this
247 scenario. Though this bedrock could have been exposed as early as the first outflow event
248 (likely Hesperian), and thus subjected to impacts for significant duration, episodic flooding
249 events in Ares Vallis may have continued through the Early Amazonian (Warner et al., 2009).
250 These later events could have removed any regolith that had formed up until that point.
251 Then, strong katabatic winds, funneled by the canyon, may have continued to keep those
252 olivine-bearing surfaces free of comminution products; continuance of low sediment cover
253 in the mid-to-late Amazonian would have been aided by a reduced cratering rate (and thus
254 reduced sediment production rate) relative to the Noachian/Hesperian.

255 In summary, exposed rock does not necessarily indicate friable rock. But other
256 morphological indicators, such as crater retention and morphology, can be used to assess
257 friability (e.g. Malin and Edgett, 2000). Our observations suggest that bedrock plains may
258 commonly consist of friable rock. If correct, this has implications for using TI to interpret
259 rock mechanical strength from orbit. For example, previous studies have interpreted the
260 relatively high TI values (500 to $>1200 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$) of intercrater and crater floor materials
261 to represent materials of high mechanical strength (e.g. Rogers et al., 2009; Bandfield et al.,
262 2013; Edwards et al., 2014); however, relatively high-TI surfaces could maintain their high
263 values through high erodibility, and thus could instead indicate low mechanical strength
264 relative to Hesperian lavas. Furthermore, these TI values are within the range of those
265 measured from friable volcanoclastic or sedimentary rocks in the Columbia Hills ($\sim 620 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$, up to $\sim 1100 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$; Ferguson et al., 2006), Meridiani Planum (likely 400-1100
266 $\text{J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$ inferred from Rock Abrasion Tool grind energies, Golombek et al., 2008), and
267 Gale crater (370–540, up to $\sim 700 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$ for mudstones/sandstones, Hamilton et al.,
268 2014; Vasavada et al., 2017).

270 **3.2 Potential origin(s) of bedrock plains and causes of olivine enrichment**

271 The evidence for friability suggests that effusive volcanic origins are unlikely for
272 many bedrock plains units, in contrast with interpretations from previous studies (Rogers et
273 al., 2009; Rogers and Ferguson. 2011; Bandfield et al., 2013; Rogers and Nazarian, 2013;
274 Edwards et al., 2014). Given the compositional and morphological variability observed
275 among these dozens of units there is no reason to assume that they all have a single origin.
276 However, origin models must satisfactorily explain the typically distinctive compositions
277 (e.g. olivine enrichments compared to surrounding low-TI surfaces), and the concentration
278 in topographic lows. Candidate petrogenetic processes that deposited these clastic units
279 include explosive volcanism, impact-related processes, and detrital sedimentation.

280 Localized, explosive volcanism could have produced some of the olivine-bearing
281 and/or feldspathic bedrock plains units. Though evidence for draping relationships are
282 absent for the intercrater and crater-filling bedrock plains (Rogers and Nazarian, 2013),
283 vents located within the topographically lower parts of these basins could have produced
284 locally-deposited tephras.

285 Impact-related deposition might explain some of these deposits. Large, basin-scale
286 impacts (e.g. Isidis, Argyre, Hellas), could have produced olivine-bearing clastic rocks in the
287 form of suevites, and also potentially as condensates from silicate vapor created during the
288 impact (Toon et al., 2010). These silicate condensate materials could range from
289 porous/unconsolidated to strongly welded, depending on the thickness of the deposits and
290 temperature at time of deposition (Toon et al., 2010; Palumbo and Head, 2017). Indeed,
291 previous authors have suggested that the Nili Fossae and Isidis olivine-bearing bedrock
292 plains units may represent the silicate condensate (Palumbo and Head, 2017) or impact melt
293 (Mustard et al., 2007) from the Isidis basin impact. For bedrock plains units elsewhere in the
294 highlands, the formation ages would need to be better constrained in order to support or
295 rule out the role of basin scale impacts.

296 Sediment transport and deposition has been discussed as a likely basin-filling
297 mechanism, particularly for the flat-floored, degraded craters that are so common in the
298 highlands (e.g., Craddock et al., 1997; Forsberg-Taylor et al., 2004; McDowell and Hamilton,
299 2007; Irwin et al., 2013). If sedimentary, the lack of spectral evidence for cementing minerals
300 would suggest that they are volumetrically minor and/or spectrally bland in the infrared
301 wavelength ranges used for spectral analysis (e.g. iron oxides). An explanation for the olivine
302 enrichments through a detrital sedimentary process, however, requires more extensive
303 discussion. Mineral enrichments can be produced through hydrodynamic sorting during
304 sediment transport (e.g. Fedo et al., 2016), but it is unclear whether such sorting could occur
305 so uniformly over the 10^2 - 10^3 km scales associated with olivine-bearing plains bedrock.
306 More likely, olivine enrichment occurred after deposition, through deflation. As the bedrock
307 plains are comminuted and eroded, denser and/or coarser olivine-bearing clasts or particles
308 could have remained behind as lag deposits, possibly forming the patches of sediment that
309 can be observed in depressions/hollows in high resolution imagery (e.g. **Figure 3**), and
310 dominating the infrared spectral signatures measured from orbit. Though we introduce
311 deflation here to describe a possible means of olivine enrichment from olivine-poor
312 sedimentary rocks, this process could have produced olivine enrichment from any of the
313 clastic rock types described above.

314 **3.3 Implications for Mars 2020 landing sites**

315 An extensive olivine-bearing unit is observed around the perimeter of the Isidis
316 impact basin that includes the Nili Fossae region. Proposed origins for this unit include lavas
317 (Hamilton and Christensen, 2005; Tornabene et al., 2008), impact melts (Mustard et al.,
318 2007), or silicate condensate from vaporized crust following the Isidis basin impact
319 (Palumbo and Head, 2017). This unit exhibits TI values between $\sim 400\text{-}700 \text{ J}\cdot\text{m}^{-2}\text{K}^{-1}\text{s}^{-0.5}$,
320 consistent with clastic rock or pervasively-fractured crystalline rock (Edwards and Ehlmann,
321 2015). Like other examples presented in this work, this Noachian-aged unit exhibits higher
322 TI than the Hesperian-aged Syrtis Major lavas that overlie the unit (**Figure 2d**). Though
323 portions of the olivine-bearing unit exhibit evidence for aqueous alteration (Ehlmann and
324 Mustard, 2012) that may have weakened the olivine-bearing bedrock, significant portions
325 appear unaltered and exhibit similar THEMIS TI values to the altered regions (Edwards and
326 Ehlmann, 2015). Unaltered regions are found within 300 km distance and within a few
327 hundred meters elevation from the Syrtis lavas, suggesting that preferential wind activity
328 cannot explain the differences in TI and regolith cover between unaltered olivine bedrock
329 and Syrtis lavas. We thus suggest that the Nili Fossae olivine-bearing unit is friable compared
330 to Syrtis lavas, and argue that this favors pyroclastic, detrital sedimentary, or impact-related
331 origins (e.g. Palumbo and Head, 2017).

332 Olivine-bearing light-toned bedrock plains are present beneath a younger, crater-
333 retaining unit in Jezero crater. There, the geologic context would support a sedimentary
334 origin, as suggested by Goudge et al., (2015). Finally, olivine-bearing plains bedrock is
335 observed to the southwest of the Columbia Hills, and could represent tephras (**Section 2.1**;
336 Ruff et al., 2014) but also could have other origins such as those described in **Section 3.2**. No
337 matter which landing site is selected, detailed petrographic analysis of olivine-bearing
338 bedrock plains with the Mars 2020 payload, and in terrestrial laboratories upon sample
339 return, will likely provide insight into the formational mechanism(s) of these distinctive and
340 widespread units.

341 **5. Conclusions**

342 The lack of regolith cover compared to known volcanic surfaces, as well as poor
343 retention of small craters, suggest that many bedrock plains are not composed of
344 mechanically strong materials, such as lavas. Rather, many of these olivine-bearing and
345 feldspathic units likely represent clastic rocks. The high TI values of bedrock plains relative

346 to average Martian surfaces likely reflects weak material properties. Friable materials would
347 break down into fine particulate materials that are more easily moved by wind. In contrast,
348 lavas comminute into blocky, coarse materials that are not easily eroded, resulting in buildup
349 of thick regolith. Thus from orbit, TI differences between adjacent geologic units could, in
350 some cases, appear inverted relative to their underlying differences in mechanical strength
351 and TI. Candidate origins include pyroclastic, impact related materials and/or sedimentary
352 rocks. We suggest that the observed olivine enrichments may have developed over time,
353 through deflation, preferential removal of plagioclase in the finer-particulate fraction, and
354 accumulation of olivine-bearing sediments in patchy lag deposits.

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531

532 **Figure Captions**

533 **Figure 1. (a).** Map of TES-derived TI within study region indicates that bedrock plains are
534 rare in the Hesperian volcanic plains of Hesperia Planum and Syrtis Major Planum and
535 common in the intercrater plains and crater floors of the highlands (white arrows indicate
536 examples). Other locations discussed in the text are labeled. **(b-c)** Example bedrock plains in
537 eastern Noachis Terra (regions 2 and 3, **Table S1**) and Terra Cimmeria (region 5, **Table S1**).

538 **Figure 2.** Differences in regolith cover are apparent between Hesperian volcanic plains and
539 subjacent, older bedrock in Syrtis Major **(a-b)**, Gusev crater **(c)**, and northeast Syrtis **(d)**. **(a)**
540 THEMIS TI shows distinct thermophysical stratigraphy with bedrock plains underlying
541 lower-TI lavas. **(b)** Portion of HiRISE image ESP_036579_1795. Dark-toned Syrtis lavas
542 overlie a light-toned, basaltic fractured unit that is similar in appearance to other bedrock
543 plains. Though the light-toned unit is included in the “Hesperian volcanics” (eHv) unit of
544 Tanaka et al. (2014), it is stratigraphically older, and may represent Noachian bedrock. The
545 dark-toned unit ridges protrude through swaths of sediment in topographic lows; bedforms
546 are absent, possibly indicating a dominance of coarse-particulate materials that are not
547 easily moved by wind. The light-toned higher-TI unit contains sparsely distributed
548 bedforms. **(c)** In Gusev crater, Hesperian lavas exhibit low TI values from orbit, whereas
549 subjacent, dominantly clastic materials in the Columbia Hills exhibit higher TI values. *As
550 suggested by Ruff et al. (2014). **(d)** Northeast of Syrtis Major, Hesperian lavas overlie
551 variably altered olivine bearing rocks (Ehlmann and Mustard, 2012). The younger lavas
552 exhibit lower TI values and thicker regolith.

553 **Figure 3.** Morphological and crater density distribution examples suggesting some of the
554 bedrock plains consist of friable materials. **(a-b)** CTX (~6 m/pixel) image mosaics of bedrock
555 surfaces near **(a)** 215.10°E, 37.99°S and **(b)** 142.50°E, 19.80°S showing oriented, linear
556 ridges that we interpret as yardangs. Note degraded, scalloped rims of small craters on
557 bedrock. Color indicates relative TI (red=high; blue=low). **(c)** Portion of HiRISE image
558 PSP_009339_1585 (25 cm/pixel), located in Peta crater near 350.88°E, 21.00°S. Bedrock unit
559 exhibits smooth or scalloped textures in some locations. **(d)** Portion of HiRISE image
560 ESP_047522_1555 (50 cm/pixel), located near 125.72°E, 24.38°S. Bedrock unit exhibits
561 scalloped textures in some locations (arrows). **(e)** Crater density for bedrock and low-TI
562 surfaces in region 9; locations shown on THEMIS day IR **(f)** and night IR **(g)** radiance. The
563 supplementary material contains additional crater density maps and morphological
564 examples.





