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**Title: Modern Mars' geomorphological activity, driven  
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1 **Title: Modern Mars' geomorphological activity, driven by wind, frost, and gravity**

2  
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31

32  
33 **ABSTRACT**

34 Extensive evidence of landform-scale martian geomorphic changes has been acquired in the  
35 last decade, and the number and range of examples of surface activity have increased as more  
36 high-resolution imagery has been acquired. Within the present-day Mars climate, wind and  
37 frost/ice are the dominant drivers, resulting in large avalanches of material down icy, rocky, or  
38 sandy slopes; sediment transport leading to many scales of aeolian bedforms and erosion; pits of

39 various forms and patterned ground; and substrate material carved out from under subliming ice  
40 slabs. Due to the ability to collect correlated observations of surface activity and new landforms  
41 with relevant environmental conditions with spacecraft on or around Mars, studies of martian  
42 geomorphologic activity are uniquely positioned to directly test surface-atmosphere interaction  
43 and landform formation/evolution models outside of Earth. In this paper, we outline currently  
44 observed and interpreted surface activity occurring within the modern Mars environment, and tie  
45 this activity to wind, seasonal surface CO<sub>2</sub> frost/ice, sublimation of subsurface water ice, and/or  
46 gravity drivers. Open questions regarding these processes are outlined, and then measurements  
47 needed for answering these questions are identified. In the final sections, we discuss how many  
48 of these martian processes and landforms may provide useful analogs for conditions and  
49 processes active on other planetary surfaces, with an emphasis on those that stretch the bounds of  
50 terrestrial-based models or that lack terrestrial analogs. In these ways, modern Mars presents a  
51 natural and powerful comparative planetology base case for studies of Solar System surface  
52 processes, beyond or instead of Earth.

53

#### 54 KEY WORDS

55 Geomorphological activity; Mars; Comparative Planetology; Aeolian; Sublimation; Mass  
56 wasting

57

#### 58 HIGHLIGHTS

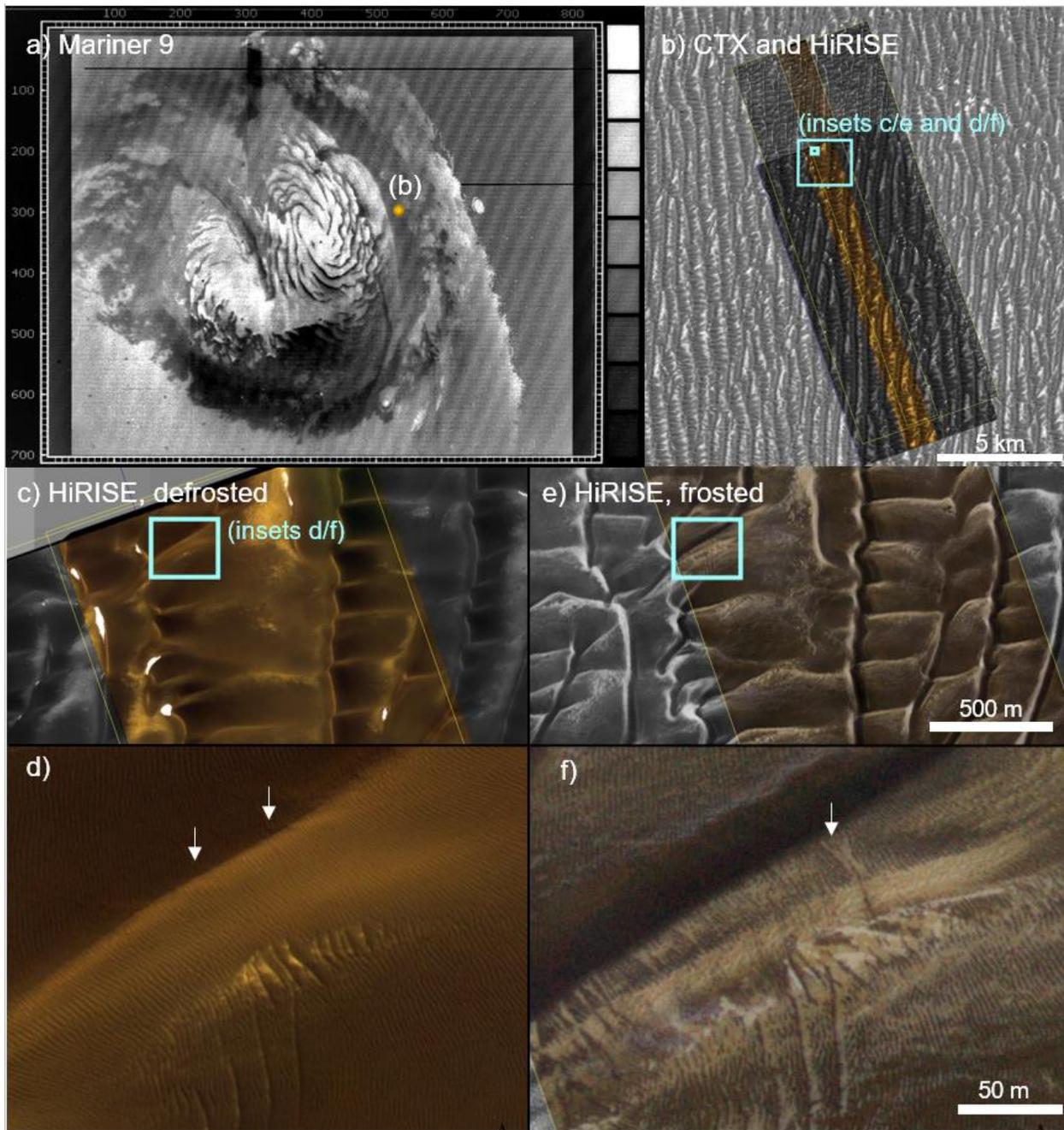
- 59
- 60 • Mars' surface is actively shaped in the present due to wind, frost/ice, and gravity.
  - 61 • Overlapping, high-resolution images from orbit are key for detection of activity.
  - 62 • In situ and orbital data are needed to fully characterize the active Mars processes.
  - Mars studies provide critical information about activity beyond that seen on Earth.

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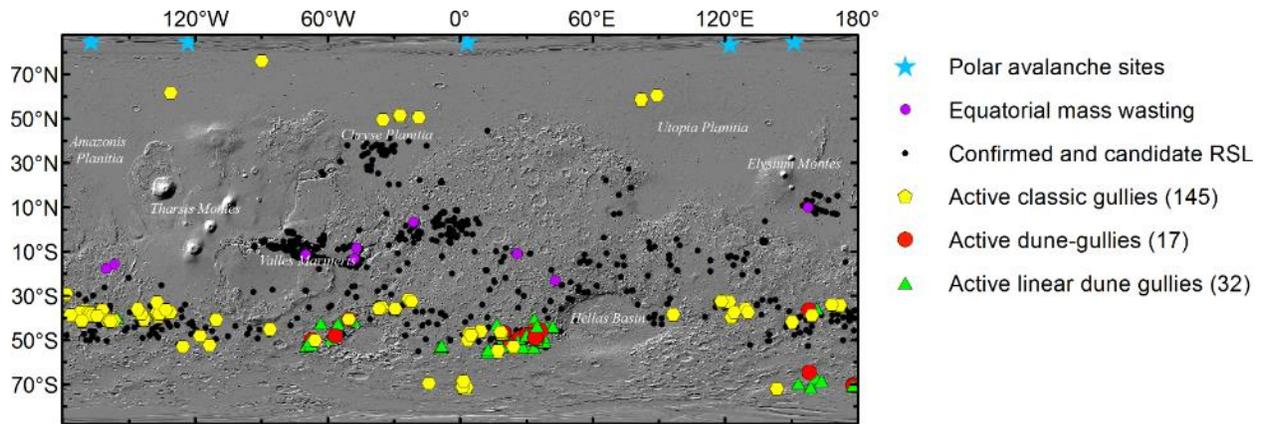
116 **1 Introduction**

117 Numerous studies since the early Mars missions have documented evidence of surface  
118 activity on Mars (e.g., Figure 1), but it was only with the advent of high-resolution and repeat  
119 imaging over multiple martian years that the full scope of present-day martian surface activity,  
120 including topographic changes, has been appreciated. Direct observations of geomorphological  
121 activity, over all areas of Mars (Figure 2), has enabled testing of hypotheses about the driving  
122 conditions and processes causing observed changes. In particular, many of the types of  
123 landforms hypothesized to be recently formed have been shown to form or be modified in the  
124 present day, with observations and models tying activity rates and timing to frost(s), wind, and  
125 gravity (i.e., movement down slopes).  
126



127  
 128  
 129 Figure 1. Example images of early and recent images of martian polar dunes, and the details that  
 130 become apparent in zoomed-in images. (a) Mariner 9 image of the north polar cap and polar erg,  
 131 acquired 1972-10-12 (= Mars Year (MY) 10  $L_s$  95°, see §1.2 for date nomenclature), ~3 km/px  
 132 (MTVS 4297-47). (b) Basemap is from CTX images (~5 m/px) and shown are 2 pairs of  
 133 monochromatic and false color HiRISE images (50 cm/px): ESP\_027012\_2610, acquired 2012-  
 134 05-01 (MY 31  $L_s$  104°) and ESP\_058950\_2610, acquired 2019-02-22 (MY 34  $L_s$  345°). At  $L_s$   
 135 104° (early summer), the seasonal frost layer is subliming, with only a few small patches of ice  
 136 remaining (i.e., the bright spots). The large ripples are clearly visible on the dunes (§2.1.3), along  
 137 with a few new dune furrows (arrows; §3.3.2). At  $L_s$  354° (end of winter), the surface is  
 completely covered in CO<sub>2</sub> frost, so is more uniform in color and brighter. Some underlying

138 ripples are visible beneath the frost, as is a new dune alcove (arrow; §3.2.2). The false color  
139 scheme is based on an automatic contrast enhancement algorithm with further manual tweaks to  
140 increase visibility of small features. The dunes are dark in color because they are made of basalt  
141 and the interdune substrate is lighter, with frost/ice brightest. North is up and illumination is  
142 from the left in all HiRISE images. Scale bars are approximate as images are not orthorectified.  
143



144 Figure 2. Map showing observations of activity for features where global surveys have been  
145 completed; note that this map is incomplete due to the patchy spatial coverage of repeat imaging:  
146 monitoring locations for polar avalanches (Becerra et al., 2020) and presumed to occur at all  
147 steep scarps, equatorial mass-wasting (M.F. Thomas et al., 2020), RSL (Stillman et al., 2020),  
148 classic gullies (Dundas et al., 2019a), dune gullies (Dundas et al., 2019a; Diniega et al., 2010),  
149 and linear dune gullies (Dundas et al., 2019a; Pasquon et al., 2016).  
150

151  
152 Studies of present-day geomorphological activity, especially if tied to correlated observations  
153 of activity and the relevant environment, are uniquely positioned to directly test surface-  
154 atmosphere interaction and landform formation/evolution models. Such studies are of great  
155 importance for understanding Mars' environmental and geologic history because landforms can  
156 serve as proxy records of specific processes and environmental conditions, such as surface  
157 thermo/mechanical properties, grain size(s), and wind velocities and variability. Studies of  
158 present-day activity on Mars are also uniquely enabling for studies of processes active on other  
159 planetary bodies because these either provide a matchless detailed planetary data point outside of  
160 Earth's gravity, atmosphere, and other conditions for comparison to terrestrial studies and  
161 derived models, or provide a detailed look into a process that has no terrestrial analog.

162 This review is on martian landforms that can be robustly connected to specific surface  
163 environmental conditions and processes. We focus on martian surface activity that (1) is  
164 observed or hypothesized to be happening in the present climate (albeit, in some cases,  
165 potentially at very slow, not yet directly observable rate), and (2) creates a specific and

166 interpretable change to the martian rocky or icy surface's shape that can be detected for >1  
167 martian year.

168 Throughout this paper, we discuss the “modern Mars” environment, and in particular on the  
169 present martian climate, which has been observed at high frequency and resolution over the last  
170 few decades via spacecraft. However, modern Mars also includes the “recent” climate—a term  
171 generally used to refer to the time since the last major obliquity excursion (around 500 kyr,  
172 *Laskar et al.*, 2004), as this is the time period for the most recent significant sculpting of the  
173 currently observable landscape. This time period is important as it is the only period where direct  
174 characterization of the environment, based on present-day measurements, can be paired with  
175 specific surface changes and thus hypothesized landform formation and evolution models can be  
176 robustly tested with observations. Additionally, the climate conditions during the present and  
177 recent past are thought to be representative of the martian climate over the last few billion years.  
178 Throughout this period, called the Amazonian, Mars is thought to have been dry and cool, with  
179 very low surface pressures and little liquid water. (Recent studies and our present understanding  
180 about the Amazonian climate are summarized in *Diniaga and Smith*, 2020.) Thus, while this  
181 review focuses on activity that has been observed in or hypothesized to be occurring in the  
182 present day, what we learn about surface-altering processes and driving environmental  
183 conditions is likely to be relevant through a few billion years of Mars' geological and  
184 climatological history—and so interpretations of even relict landscapes should take into account  
185 the presently observed surface-altering processes.

186 In this review, we will outline observed and interpreted surface activity occurring within the  
187 modern Mars environment. This activity, when tied to specific environmental drivers, has been  
188 shown to be primarily caused by wind- and frost-related processes, although in many cases the  
189 exact mechanism driving the geomorphological change has not yet been determined. The frost or  
190 ice involved in present-day landform evolution is of two broad classes: the atmosphere-sourced  
191 CO<sub>2</sub> and H<sub>2</sub>O frost that accumulates each winter on the martian surface, and the previously  
192 buried/preserved ice deposits beneath the martian surface or within the polar cap that now are  
193 undergoing long-term loss (although for the cap, short-term loss occurs in some areas but it is  
194 unclear if there is total net long-term loss or gain). Three sections describe landforms with  
195 formation mechanisms associated with wind (i.e., aeolian features) and these two classes of frost.  
196 Gravity also plays a role as many of these landforms involve material moving downslope (i.e.,

197 mass-wasting features). In a separate section, we describe a few landforms where the initiation of  
198 or additional environmental control on such downslope movement has not yet been determined.

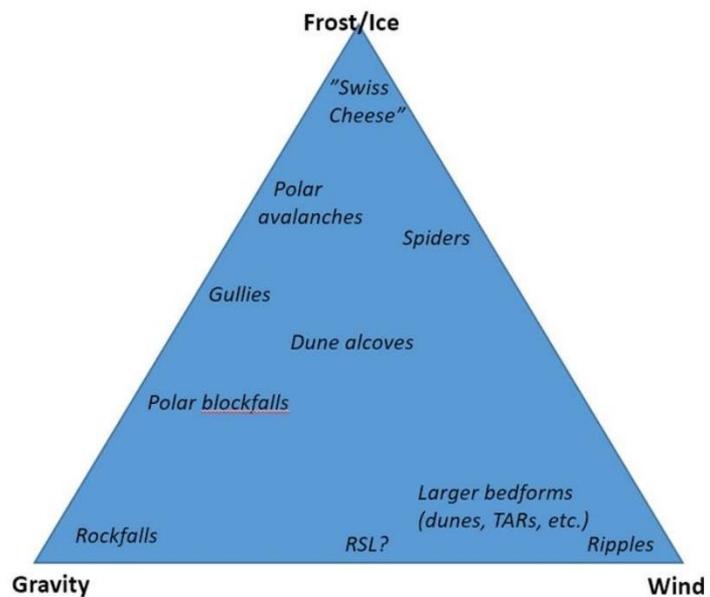
199 In each of these sections, after outlining what is known, we outline open questions about the  
200 exact processes and environmental thresholds/controls. We also summarize current big questions  
201 about the martian surface and atmosphere environment in the present and through the  
202 Amazonian that could be addressed through continued study of these specific surface changes  
203 and landforms. Additionally, we identify the measurements needed to answer these questions.  
204 Finally, we discuss how many of these environmental conditions and processes may provide  
205 useful analogs for conditions and processes active on other planetary surfaces, with an emphasis  
206 on those that lack terrestrial analogs and for which Mars is a more natural comparative  
207 planetology base case.

208

### 209 **1.1 Why focus on wind and sublimation as drivers for surface activity?**

210 Some of the earliest Mars investigations via Earth-based telescopes or spacecraft observed  
211 Mars' atmosphere and seasonal frost (e.g., *Johnson*, 1965; *Lowell*, 1895). Orbital observations  
212 have enabled tracking of seasonal frost caps (§3.1) and movement of dust and sand (§2), as  
213 reflected in bedform movement and regional albedo changes. In situ indications of wind and at  
214 least trace amounts of seasonal frost (H<sub>2</sub>O and/or CO<sub>2</sub>) have been observed by all Mars landers  
215 that have survived through a martian winter (missions listed in Table 2, references listed in §3.1).  
216 The northernmost lander (*Phoenix*) even had one of its solar panels crushed due to accumulation  
217 of a thick layer of frozen CO<sub>2</sub>  
218 ([https://www.nasa.gov/mission\\_pages/phoenix/news/phx20100524.html](https://www.nasa.gov/mission_pages/phoenix/news/phx20100524.html)). Evidence that both  
219 wind and seasonal frost affect landform evolution has built up over the last two decades as the  
220 martian surface environment and morphology, and changes in the surface morphology, have  
221 been observed and characterized globally at sub-landform-scales and in places at sub-meter-scale  
222 resolution (§2–3, 5). As shown in Figure 3, many of the examples of observed present-day  
223 surface changes appear to be explained through some combination of wind, annual frost/ice  
224 formation or sublimation, and gravity (i.e., mass wasting). Close study of these  
225 landforms/surface changes along with concurrent measurement of their environment enables  
226 testing and refinement of quantitative models of the underlying processes, under Mars  
227 conditions.

228 In addition, there are martian landforms that have not yet been directly observed to form and  
229 change, but which are interpreted to be forming in the present climate due to long-term (i.e.,  
230 multi-annual) sublimation or  
231 modification of surface or  
232 subsurface water ice reservoirs (§4).  
233 Such landforms are also important to  
234 study, again with concurrent detailed  
235 measurement of their present  
236 environments, because they provide  
237 a bridge to a recent past climate  
238 when that ice was deposited.



239 **Figure 3.** A ternary diagram  
240 illustrating the proposed relative  
241 controls by frost/ice, wind, and  
242 gravity on many of the landforms discussed in this review (§2–3, 5). (Not included here are the  
243 landforms created through long-term subsurface ice processes, §4, and landforms not discussed  
244 within this paper.)  
245

246  
247 Beyond wind and frost, a few other known or hypothesized present-day surface processes are  
248 widespread and can move large amounts of material over the martian surface. However, for  
249 reasons described here, these processes and landforms are not discussed further within this paper.

- 250 • Rocky landforms and textures that appear similar to terrestrial features formed through wind  
251 erosion, such as ventifacts (*Laity and Bridges, 2009*) and yardangs (*Liu et al., 2020; Ward,*  
252 *1979*) have been identified on Mars. Such erosion is likely occurring in the present martian  
253 climate but would be occurring at very slow rates; terrestrial sand abrasion occurs at tens to  
254 thousands of microns per year (discussed in *Laity and Bridges, 2009*) and bulk Mars surface  
255 aeolian erosion rates are at the low end of that range (§2.2.3). Thus, we are not yet able to  
256 draw quantitative connections to specific environmental conditions, including roughly when  
257 these environmental conditions existed, and constrained modeling of the formation process  
258 is difficult. Hence, we do not discuss such landforms in this review.
- 259 • Impact cratering is also actively changing the shape of the martian surface in the present  
260 climate (e.g., *Daubar et al., 2013; 2019*), but the dominant controls for that process are

261 characteristics of the impactor and the impacted surface structure, not the environment at the  
262 time of impact. Thus, impactor-related processes are not a focus of this review.

263 • We will not discuss processes that generally change the appearance of the surface by  
264 moving around only a surficial layer, such as insolation-driven dust lifting or dust devils  
265 (*Balme and Greeley, 2006*) or thin slope streaks (*Chuang et al., 2007*), although slope  
266 streaks do occasionally transport greater thicknesses (*Dundas, 2020b*). Such processes do  
267 not yield a significant change in the shape of the landscape and/or a clear geomorphic  
268 change retained for >1 martian year, and so are considered beyond the scope of this study.  
269 One exception is recurring slope lineae (RSL, discussed in §5.1), which is a landform of  
270 recent high interest.

271 Other drivers, such as volcanism or liquid water, have been proposed to explain observed  
272 geomorphologies and, in a few cases, observed surface activity. Some studies have suggested  
273 that these other drivers may be important and influential for shaping martian geology during the  
274 Amazonian and into the present. However, as outlined above, we focus this review on  
275 geomorphic processes known or generally thought to be active in the present day, and processes  
276 where existing observations of the martian environment are (so far) at least qualitatively  
277 consistent with the models we describe. For example, we note that this review does not include  
278 discussion of liquid water-driven geomorphic activity because, although many studies have  
279 proposed recent or present-day water-driven activity to explain observed geomorphologies (e.g.,  
280 *Chevrier and Rivera-Valentin, 2012; Malin et al., 2006*), no studies have yet been able to explain  
281 a water source that is consistent with all environmental observations or with behaviors/timing of  
282 activity that is consistent with observations of changes. In addition to liquid water-driven  
283 processes, volcanism, tectonics, glacial flow, rainfall, and biological activity will not be  
284 discussed within this review.

## 285 286 **1.2 Sources of seminal data**

287 The primary data that have led to studies of present-day surface geomorphological changes  
288 have been high-resolution visible images that allow for identification of smaller surface features  
289 and changes. For the latter, a key enabler were visible images of the same site, repeated over  
290 time—between the high resolution of these images (down to 0.25 cm/px by the Mars  
291 Reconnaissance Orbiter (MRO) High Resolution Imaging Science Experiment (HiRISE)) and a  
292 longer temporal baseline (currently at seven martian years for the highest-resolution images;

293 longer for comparisons to coarser resolution data: Table 1), many more examples of surface  
294 changes have been identified. An example of such repeat images is shown in Figure 4;  
295 comparisons of such images need to consider different resolutions and illumination conditions  
296 (i.e., time of day).

297 In discussions of the timing of observed activity, we use the common Mars Year (MY – note  
298 the difference from million years = Myr) and solar longitude ( $L_s$ ) nomenclature. Enumeration of  
299 Mars Years and seasons is described in detail by *Piqueux et al. (2015a)*, but a brief description is  
300 as follows:

- 301 • A Mars year is nearly twice as long as an Earth year (~687 Earth days).
- 302 • The solar longitude denotes the position of Mars in its orbit, running from  $L_s 0^\circ$  to  $360^\circ$ .  
303 (Due to Mars' orbital eccentricity, a degree of  $L_s$  spans 1.5-2.2 martian days or 'sols'.)
- 304 • A Mars Year starts at  $L_s 0^\circ$  = northern spring equinox, and proceeds to  $L_s 90^\circ$  = northern  
305 summer solstice,  $L_s 180^\circ$  = northern autumnal equinox, and  $L_s 270^\circ$  = northern winter  
306 solstice.
- 307 • MY 1,  $L_s 0^\circ$  started on April 11, 1955.

308 After a surface change has been clearly identified, science investigations of that present-day  
309 activity generally aim to identify the driving environmental conditions and relevant processes.  
310 With visible imagery, a temporal survey can be done over a sequence of overlapping images to  
311 determine when the change occurs, or a spatial survey can be used to constrain where this  
312 landform exists (as well as where it doesn't exist). With other observational datasets,  
313 environmental information can be gathered—for example:

- 314 • spectral data can yield constraints for the surface composition;
- 315 • atmospheric observations and modeling can yield information about wind patterns and  
316 surface pressure variations; and
- 317 • topographic data, measurements of shadows, or photoclinometry analysis can yield  
318 estimates of heights and slopes.

319 Such environmental information can be gathered from orbit or in situ. Table 2 contains a listing  
320 of the Mars rovers and landers often referenced in studies of present-day activity and  
321 surface/atmosphere environmental conditions.

322

323 Table 1. Primary instruments used to acquire orbital visible imagery used in studies of present-day surface activity, in reverse  
 324 chronological order of start of operations. To definitively measure a feature or surface change in an image, at least three pixels are  
 325 generally needed. The date of last contact is used to denote the end of operations.

<b>Instrument</b>	<b>Spacecraft</b>	<b>Period of Operation</b>	<b>Nadir Pixel Scale</b>	<b>Field of View</b>	<b>Global Coverage (as of July 2020)</b>
Colour and Stereo Surface Imaging System (CaSSIS: <i>N. Thomas et al., 2017</i> )	ESA's ExoMars Trace Gas Orbiter (TGO)	2016-11-22 to present	~5 m/px	>8 km-wide swath in 3-colors or >6 km-wide swath in 4 colors; typical length of 50 km	Total area of images is 2.3%, but <1.6% after removing overlap
High Resolution Imaging Science Experiment (HiRISE: <i>McEwen et al., 2007</i> )	NASA's Mars Reconnaissance Orbiter (MRO)	2006-03-24 to present	~0.3 m/px for most images (from 300 km altitude)	6 km-wide swath in grayscale, with nested 1.2 km wide swath in 3-colors; typical length of 10 km	Total area of images is 3.6%, but <2.5% after removing overlap
Context (CTX) Camera ( <i>Malin et al., 2007</i> )	(MRO)	2006-04-13 to present	~6 m/px (from 300 km altitude)	30 km-wide swath in grayscale; typical length of 90 km	~100%, and a global mosaic has been created ( <i>Dickson et al., 2018</i> )
High Resolution Stereo Camera (HRSC: <i>Neukum et al., 2004a</i> )	ESA's Mars Express (MEx)	2004-01-14 to present	~10 m/px for nadir channel (from 250 km altitude)	53 km-wide swath in 4-colors with length at least 300 km for regular images	~75% with resolution 10-20 m/px; 100% with resolution >100 m/px ( <i>Gwinner et al., 2019</i> )
Mars Orbital Camera (MOC: <i>Malin et al., 1992</i> )	NASA's Mars Global Surveyor (MGS)	1997-09-15 to 2006-11-02	1.4–12 m/px for narrow angle; 225–7500 m/px for wide angle (from 378 km altitude)	3 km-wide swath in grayscale for narrow angle; 115 km-wide swath in 2-colors for wide angle; typical length of 30 km	0.5% at better than 3 m/px; 5.45% at better than 12 m/px ( <i>Malin et al., 2010</i> ); 100% for wide angle

326

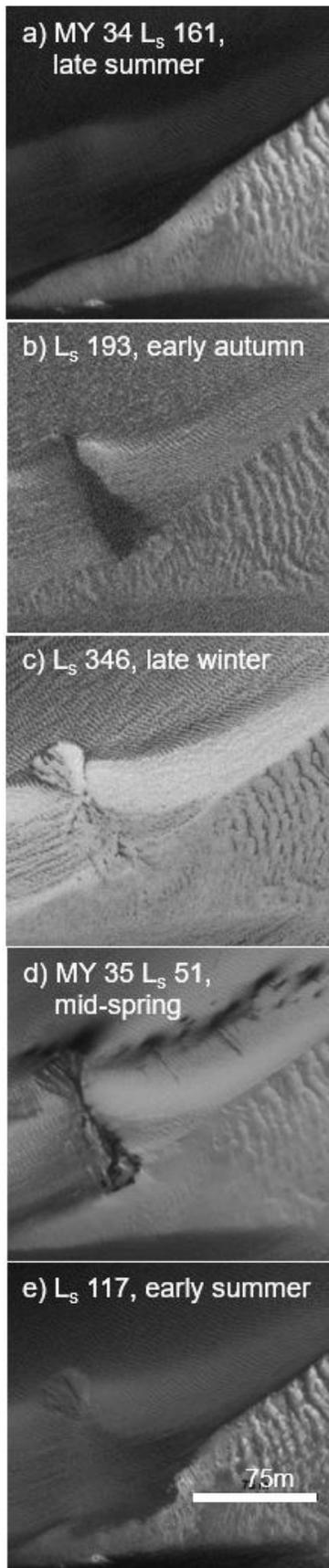


Figure 4. An example of how repeat imagery enables identification of activity, with timing constraints. In this chronological sequence of images, the polar dune slope (Tleilax dune field, 83.5°N, 118.5°E) becomes covered with frost and a new dune alcove forms (*b*) during L<sub>s</sub> 161–193° (and likely during L<sub>s</sub> 180–189°, intervening images are shown in SOM1). The alcove has formed by L<sub>s</sub> 346° (*c*) and is clearly present under the seasonal frost layer. Sublimation begins in spring, with spots appearing preferentially along the dune brink and within the alcove and apron (*d*). Sublimation completes L<sub>s</sub> 51–117°. HiRISE images are (*a*) ESP\_054971\_2635, (*b*) ESP\_055683\_2635, (*c*) ESP\_058967\_2635, (*d*) ESP\_060734\_2635, and (*e*) ESP\_062633\_2635. A scale bar is shown in the last image (75 m), but absolute distances are approximate as images are not orthorectified. North is up and illumination is from the left.

343 Table 2. Successful Mars rovers and landers are often referenced in studies of present-day  
 344 activity and surface/atmosphere environmental conditions, as these provide critical in situ data.  
 345 These are listed here, in reverse chronological order of start of operations. For missions that have  
 346 ended, the date of last contact is given to denote the end of operations.

<b>Mission/Spacecraft</b>	<b>Landing site</b>	<b>Period of Operation</b>
NASA's <i>InSight</i> lander	4.5°N, 135.9°E Elysium Planitia	2018-11-26 to present
NASA's Mars Science Laboratory (MSL) rover, <i>Curiosity</i>	4.6°S, 137.4°E Gale crater	2012-08-06 to present
NASA's <i>Phoenix</i> lander	68.2°N, 234.3°E Vastitas Borealis	2008-05-05 to 2008-11-02
NASA's Mars Exploration Rover (MER)-B, <i>Opportunity</i>	1.9°S, 354.5°E Meridiani Planum	2004-01-25 to 2018-06-10
NASA's Mars Exploration Rover (MER)-A, <i>Spirit</i>	14.6°S, 175.5°E Gusev crater	2004-01-04 to 2010-03-22
NASA's Mars Pathfinder <i>Sojourner</i> rover	19.1°N, 326.7°E Ares Vallis	1997-07-04 to 1997-09-27
NASA's <i>Viking 2</i> lander	47.64°N, 134.3°E Utopia Planitia	1976-09-03 to 1980-04-12
NASA's <i>Viking 1</i> lander	22.27°N, 312.1°E Chryse Planitia	1976-07-20 to 1982-11-11

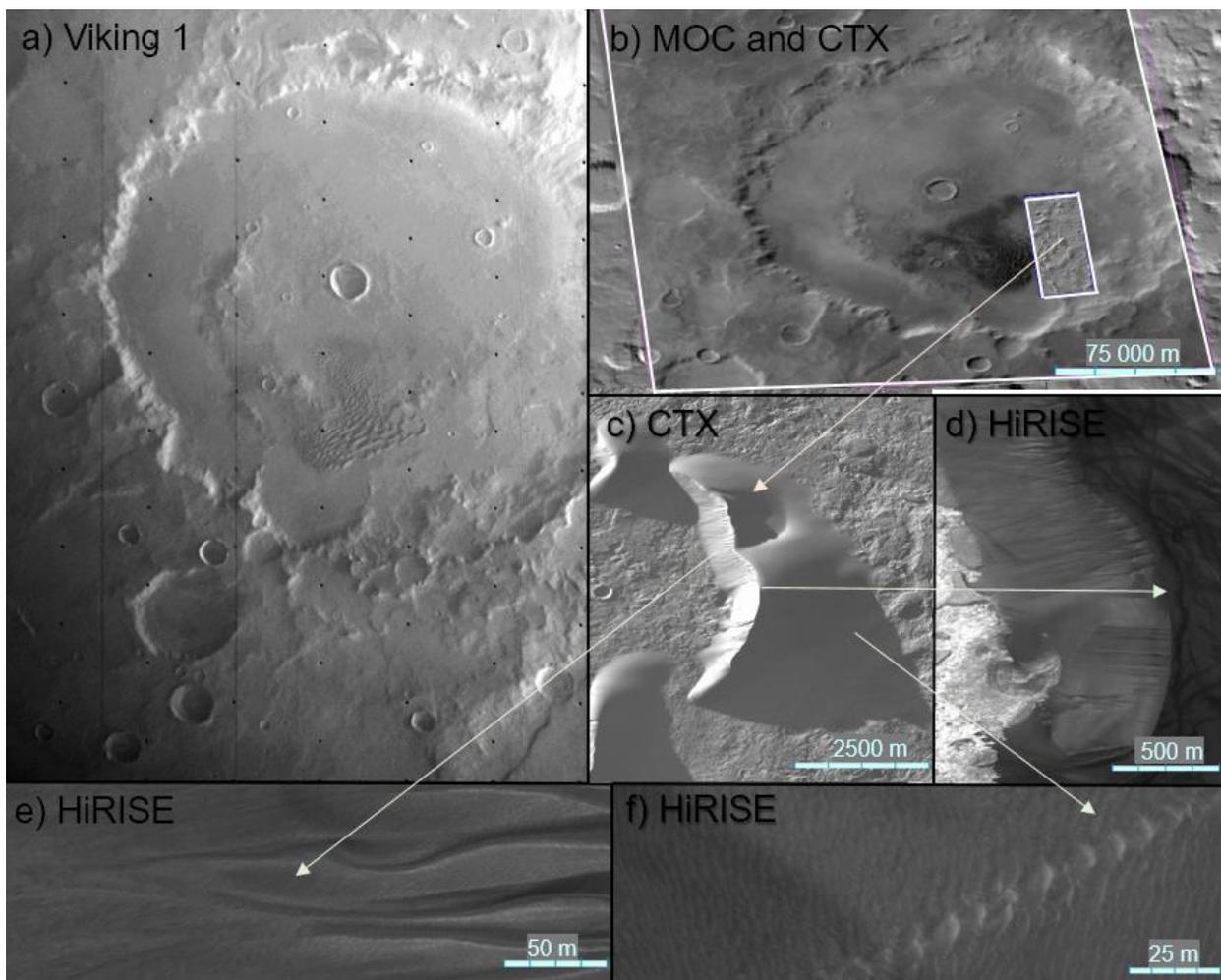
347  
 348 **2 Wind-formed landforms**

349 Although the martian atmosphere is very thin compared to the Earth's (~0.1–1% surface  
 350 pressure; e.g., *Banfield et al.*, 2020; *Harri et al.*, 2014; *Hess et al.*, 1976; *Taylor et al.*, 2010;  
 351 *Withers and Smith*, 2006), wind-driven sediment transport has been observed (e.g., *Baker et al.*,  
 352 2018a; 2018b; *Bridges et al.*, 2012a; 2012b; 2017) and aeolian bedforms analogous to terrestrial  
 353 features have been found over all scales. These include decimeter- and decameter-scale  
 354 windblown ripples (e.g., *Lapôtre et al.*, 2018; *Silvestro et al.*, 2010), migrating barchan dunes  
 355 (e.g., *Chojnacki et al.*, 2015; 2018; 2019), and megadunes (e.g., Figure 5; *Silvestro et al.*, 2012).  
 356 In many cases, higher-resolution images show that multiple scales of bedforms are superimposed  
 357 over each other (Figures 5, 6), reflecting the dominant local wind conditions over different  
 358 spatial and temporal scales and potentially different types of surface-atmosphere interaction  
 359 dynamic regimes (e.g., see §2.2). Studies of these bedforms have yielded much insight into  
 360 atmospheric characteristics, wind directions and speeds (and variability), surface grain  
 361 characteristics and availability, and sediment fluxes (as summarized for a range of planetary  
 362 bodies in *Diniaga et al.*, 2017). Environmental and geologic history, as inferred from these  
 363 bedforms and the grains that compose them, are the focus of much of this section. Additionally,

364 we briefly describe some of the fundamental models used to relate atmospheric and grain  
365 characteristics to sediment fluxes (§2.2).

366 Sediment flux rates can also be estimated based on the appearance of denuded surfaces and  
367 degraded crater forms, where the amount of removed sediment and absolute age of the present  
368 surface can be estimated (§2.2.3). Wind-driven erosion can also be inferred from features such as  
369 yardangs and ventifacts which are common on the martian surface; however, rates of erosion or  
370 sediment flux are difficult to estimate from studies of these features, so we do not discuss them  
371 in this review.

372



373  
374 Figure 5. Images of Kaiser crater (47.4°S, 18.8°E) and its dunes, acquired over the last 50 years.  
375 (a) This Viking 1 image (094A42) was acquired 1976-09-22 (MY 12 L<sub>s</sub> 126°) and has a  
376 resolution of 259 m/px. The identification of a dune field in a nearby crater in a Mariner 9 image  
377 was the first indication that sufficient sediment transport for bedform development was occurring  
378 in the thin martian atmosphere (Sagan *et al.*, 1972 -- see Figure 11, image ID MTVS 4264:16).  
379 However, >20 years later, the MOC camera on the Mars Global Surveyor orbiter reimaged

380 Kaiser crater (the large image in (b): MOC M0101026, was acquired MY 24 L<sub>s</sub> 137° or 1999-05-  
381 11). While some MOC images had resolution ~5 m/px, the image shown in (b) is of resolution  
382 233 m/px – but in all cases, no signs of dune shape change or migration were identified. It wasn't  
383 until MRO arrived that clear signs of present-day activity were identified: (b/c) this CTX image  
384 (P21\_009193\_1329\_XI\_47S339W, acquired MY 29 L<sub>s</sub> 97° or 2008-07-12) with resolution <4  
385 m/px showed mass-wasting features in the barchan megadune. HiRISE images of that same dune  
386 yielded clear information about surface activity, including: dust devil tracks (the dark curvy lines  
387 along the upwind/right slope of the dune in (d)), large-scale mass-wasting in the dune gullies  
388 along the southern side of the downwind slope of the dune in (d) (first reported in *Diniega et al.*,  
389 2010), (e) small-scale downslope sediment avalanching is evident in surface roughness changes  
390 over the ripple patterns, and (f) several scales of ripples migrating up the stoss dune slope. The  
391 HiRISE image shown in (d-f) is ESP\_058972\_1330\_RED (acquired 2019-02-24 or MY 34 L<sub>s</sub>  
392 346°, resolution 0.25 m/px). The arrows extending from (b) to (c) and from (c) to (d-f) are to  
393 show the locations of the zoom-in views. The megadune that is the main landform within (c,d) is  
394 ~750 m tall—the largest known barchan dune in the Solar System. In all images, north is up. In  
395 the Viking image illumination is from the bottom right; in the MOC image illumination is from  
396 the upper left; in the CTX and HiRISE images illumination is from the upper left. Contrast has  
397 been tweaked to bring out details for each view.  
398

## 399 **2.1 Depositional and Erosional Aeolian Landscapes: Materials and Landforms**

### 400 **2.1.1 Wind-transported sediment grain properties**

401 In situ observations of bedforms and transported grains by MER showed that a variety of  
402 grain sizes, from dust (<62 µm diameter) to sand (up to 2 mm) to granules (> 2 mm) (*Greeley et*  
403 *al.*, 2006; *Sullivan et al.*, 2008), are transported via present-day aeolian processes. *Curiosity*  
404 visited the Bagnold dune field and confirmed that martian dune sand was unimodally distributed  
405 in the very-fine-to-fine sand range, with median grain size of 100–150 µm (*Ehlmann et al.*, 2017;  
406 *Ewing et al.*, 2017; *Lapôtre et al.*, 2016; *Sullivan and Kok*, 2017; *Weitz et al.*, 2018). This  
407 countered early hypotheses, based on coarse resolution (2–30 km/px) orbital thermal  
408 measurements, that martian dune sand was composed of coarse sand grains with diameter ~500  
409 µm (*Edgett and Christensen*, 1991; *Pelkey and Jakosky*, 2002); *Edwards et al.* (2018) suggested  
410 that some of these previous estimates may have overestimated grain size due to subpixel mixing  
411 of sandy and rocky materials. Much of the dune sands actively transported on Mars today are  
412 likely to be similar in size to the grains found in the Bagnold dunes because this size is thought  
413 to be most easily mobilized by winds (§2.2.1).

414 Coarser grains have been observed by *Curiosity* along the crest of large ripples along the  
415 trailing edge of the Bagnold dune field (median ~ 350 µm; *Weitz et al.*, 2018), where coarse  
416 grains are expected to concentrate based on analogy with terrestrial dune fields (*Ewing et al.*,  
417 2017; *Lapôtre et al.*, 2016), as well as along the crests of isolated ripples (e.g., *Day and Kocurek*,

418 2016), in a sand shadow (*Minitti et al.*, 2013), and in ripple fields outside of the Bagnold dunes  
419 (median ~300–500  $\mu\text{m}$  with  $<\sim 1\%$  grains  $>1$  mm; *Weitz et al.*, 2018). Some of these coarser-  
420 grained ripples were covered in dust (*Lapôtre et al.*, 2018; *Weitz et al.*, 2018), similar to  
421 observations by the *Spirit* rover at El Dorado (*Sullivan et al.*, 2008). On bedrock surfaces,  
422 coarser sediments (~1–3 mm) appear to be mobilized by aeolian processes (*Baker et al.*, 2018a).

423 Other observations of contiguous dune fields near their putative sand sources show  
424 decreasing thermal inertia downwind, suggesting some grain size variability (*Chojnacki et al.*,  
425 2014). Alternatively, in situ investigations have shown surface crusts composed of dust,  
426 including on bedforms, that have been suggested to form from slow chemical weathering and/or  
427 salt formation as duricrusts (*Ewing et al.*, 2017; *McSween et al.*, 2004; *Moore et al.*, 1999;  
428 *Sullivan et al.*, 2008). This apparent induration of dusty bedform surfaces may play a role in the  
429 occurrences of lithified dune fields found with largely intact morphologies (*Chojnacki et al.*,  
430 2020; *Edgett and Malin*, 2000; *Milliken et al.*, 2014).

431 In situ observations of grains generally find them to be subangular with high circularity (e.g.,  
432 *Ehlmann et al.*, 2017; *Weitz et al.*, 2018), which is consistent with the observed properties of  
433 many smaller grains in desert aeolian systems on Earth (e.g., *Goudie and Watson*, 1981). Slip  
434 faces of dunes and large ripples at Gale crater have Earth-like  $\sim 30^\circ$  angle of repose, with a few  
435 steeper outliers, possibly indicating local cohesion but otherwise largely loose sand (*Atwood-*  
436 *Stone and McEwen*, 2013; *Ewing et al.*, 2017). Their results are consistent with MER in situ  
437 observations of grains that suggest low-to-no cohesion in non-dusty aeolian materials (*Sullivan et*  
438 *al.*, 2008).

### 439 440 **2.1.2 Wind-transported sediment composition**

441 One important control in aeolian processes is the source of wind-transported sediment  
442 because the source region influences the availability of sediment and physical properties of the  
443 grains. (Here we discuss primarily the ‘latest’ erosion-source of the sediment, which is not  
444 necessarily the original source because grains presently eroding out of a crater or icy wall may be  
445 exhumed from sedimentary deposits formed during past aeolian transport (e.g., *Chojnacki et al.*,  
446 2014b; *Fenton*, 2005; *Tirsch et al.*, 2011).) In some areas the sources of sediment are easy to  
447 identify in visible images, such as in the north polar erg where sand is clearly seen to be eroding  
448 from within the north polar basal unit (*Byrne and Murray*, 2002; *Massé et al.*, 2012; *Tsoar et al.*,  
449 1979; SOM 2). In other areas, it is likely that aeolian sands have been transported long distances

450 (up to hundreds of kilometers) and may then be mixed with several sources within sediment  
451 “sinks,” such as topographic lows (*Dorn and Day, 2020*).

452 However, in most cases, visible imagery is insufficient to definitively locate sediment  
453 sources for specific observed bedforms. In conjunction with visible imagery, orbital and in situ  
454 compositional data can be used to attempt to constrain the source regions of saltating sand. Such  
455 studies consistently find that martian dune sand is primarily basaltic (pyroxene-rich and olivine-  
456 bearing sands), consistent with the bulk composition of the martian surface, with some broad  
457 concentrations of gypsum and other sulfates primarily within the north polar erg (e.g., *Achilles et*  
458 *al., 2017; Chojnacki et al., 2014b; Ehlmann et al., 2017; Fenton et al., 2019; Gendrin et al.,*  
459 *2005; Johnson et al., 2017; 2018; Rampe et al., 2018; Rogers and Aharonson, 2008; Sullivan et*  
460 *al., 2008*). This broad similarity in martian dune sand composition makes it difficult to link  
461 aeolian sediments to their potential source, although a few studies have attempted to do this.  
462 Based on orbital data, candidate sources for intracrater fields were identified in nearby mafic  
463 layers outcropping in crater or valley walls (e.g., *Chojnacki et al., 2014b; Fenton, 2005; Lapôtre*  
464 *et al., 2017; Stockstill-Cahill et al., 2008; Tirsch et al., 2011*) based on the presence of a few  
465 minor phases. (Although, in the study by *Fenton (2005)*, the grain composition was also traced to  
466 other, widespread rocky units in the region, suggesting that sand-bearing layers may have first  
467 accumulated through region-wide deposition(s), with numerous local exposures being now  
468 exhumed and recycled.) However, both orbital and in situ data have also shown compositional  
469 variation within a dune field (e.g., *Chojnacki et al., 2014a; Lapôtre et al., 2017; Pan and Rogers,*  
470 *2017; Seelos et al., 2014*), suggesting that winds may sort grains by mineralogy, potentially due  
471 to correlations between grain composition and phenocryst size, grain density, and shape  
472 (*Baratoux et al., 2011; Fedo et al., 2015; Lapôtre et al., 2017; Mangold et al., 2011*). For  
473 example, observations from the *Spirit* rover found that mafic minerals were concentrated in the  
474 coarse-grained targets in Gusev crater (e.g., *Morris et al., 2006; Ming et al., 2008; Sullivan et al.,*  
475 *2008*), and observations from *Curiosity* found Mg, Fe, Ni, and Mn to be enriched in coarser  
476 samples (*O’Connell-Cooper et al., 2017; 2018*) and more crystalline and amorphous ferric  
477 materials in finer-grained targets (*Johnson et al., 2017; 2018*). Such sorting would affect bulk  
478 mineral composition measurements and cause significant variation from that of the parent rock  
479 after aeolian transport.

480 Grain composition can also yield information about the general history of sediment on a  
481 planetary body and provide clues for a grain's original source region. For example, both orbital  
482 and in situ measurements of the dune sand within Gale crater showed that a small (<10%)  
483 fraction of sand was composed of X-ray amorphous materials, indicating the presence of  
484 weathered silicates and nanophase Fe oxides and sulfates (*Achilles et al.*, 2017; *Ehlmann et al.*,  
485 2017; *Lane and Christensen*, 2013; *Rampe et al.*, 2018). Such materials suggest the grains were  
486 weathered through contact with water (*Ehlmann et al.*, 2017). However, without clear knowledge  
487 of the source region of the grains, it is difficult to tie their history to the geologic history of a  
488 specific site. Additionally, this amorphous component may come from martian dust, which is  
489 well mixed globally, reflecting regular global circulation of fine particles (*Berger et al.*, 2015;  
490 *Lasue et al.*, 2018). In situ measurements of martian airfall dust by multiple rovers have shown  
491 that it is very consistent over the martian surface and is reflective of the global Mars soil unit and  
492 its general basaltic crust, but is elevated in S and Cl relative to martian rocks and sand (*Berger et al.*  
493 *et al.*, 2015; *Ehlmann et al.*, 2017; *Lasue et al.*, 2018; *Yen et al.*, 2005).

### 494 495 **2.1.3 Bedforms: Types and Morphologies**

496 From orbiter images, aeolian bedforms of meter-wavelength ripples (i.e., just visible at the  
497 highest image resolution: Table 1) to kilometers-scale megadunes have been mapped and  
498 measured around the globe (e.g., *Bridges et al.*, 2007; *Brothers and Kocurek*, 2018; *Hayward et al.*  
499 *et al.*, 2007; 2014). Rovers have driven through a few centimeters-high to meter-high ripples (e.g.,  
500 *Curiosity* drove through Dingo gap (*Arvidson et al.*, 2017)) and around the meters-high dunes in  
501 Bagnold dune field (*Bridges and Ehlmann*, 2017; *Lapôtre and Rampe*, 2018). In this section, and  
502 again when discussing bedform migration (§2.2.2), we discuss five classes of bedforms (their  
503 names are underlined) because these are presently proposed to reflect different regimes of  
504 aeolian bedform dynamics; however, questions remain about how distinct these bedforms may  
505 be.

506 Sand ripples with decimeter wavelength have been observed by all Mars rovers (Table 2), but  
507 are below the image resolution limit of orbital data. (Larger ripples observed by these rovers  
508 were recognizable in orbital images, see below.) The *Spirit* rover observed dark decimeter-scale  
509 ripples within the El Dorado ripple field (*Sullivan et al.*, 2008), which were some of the first  
510 documented to migrate (*Sullivan et al.*, 2008), demonstrating that these were active aeolian  
511 bedforms (discussed further in §2.2.2). Within Gale crater, similar ripples with wavelengths of

512 ~5–12 cm, straight crests, and subdued sub-centimeter topography were observed in fine sand  
513 (*Ewing et al.*, 2017; *Lapôtre et al.*, 2016; 2018). On Earth, ripples of similar size and  
514 morphology (i.e., with straight crests and relatively subdued profiles) are called impact ripples  
515 and are created through grain splash (*Bagnold*, 1941; *Rubin*, 2012; *Sharp*, 1963; *Werner et al.*,  
516 1986; *Wilson*, 1972). By analogy, decimeter-scale ripples on Mars have been interpreted as  
517 impact ripples—an interpretation also consistent with numerical (*Yizhaq et al.*, 2014) and  
518 theoretical modeling (*Andreotti et al.*, 2006; *Duran Vinent et al.*, 2019), predicting impact ripples  
519 should have decimeter-scale wavelengths on Mars.

520 Larger (i.e., meter- to decameter-wavelength) ripples were originally grouped together within  
521 the polygenetic class of “mega-ripples,” similar to how aeolian ripples on Earth were originally  
522 designated by scale and grain size population as smaller unimodal impact ripples and larger  
523 bimodal “mega-ripples” (*Bagnold*, 1941; *Sharp*, 1963). However, the martian megaripple class  
524 has since been divided into three groups based on observed activity, morphology, albedo, and  
525 consistency of grain size within the features. Dark meter-scale ripples are visible in orbital  
526 imagery of dune fields and sand sheets, which shows them to be ubiquitous and to migrate over  
527 seasonal timescales (§2.2.2) (*Bridges et al.*, 2012a; SOM 3). In addition to being larger than  
528 martian impact ripples, these features differ in morphology. For example, their crestline  
529 geometry and orientation is highly variable, with dark meter-scale ripples on the gentle stoss of  
530 dunes tending to be transverse-to-oblique and highly sinuous, whereas those on steeper slopes  
531 have straight linear crests (*Ewing et al.*, 2017; *Lapôtre et al.*, 2016; 2018). Their downwind  
532 profiles also vary, from transverse large ripples with asymmetric profiles, gentle stoss slopes, and  
533 near-angle of repose lee faces (*Ewing et al.*, 2017; *Lapôtre et al.*, 2016; 2018; *Sullivan et al.*,  
534 2008) to longitudinal large ripples with symmetric profiles (*Lapôtre et al.*, 2018). Furthermore,  
535 grainfall and grainflow deposits are observed on the lee of transverse large ripples (*Ewing et al.*,  
536 2017; *Lapôtre et al.*, 2016; 2018), and decimeter-scale impact ripples form concurrently and  
537 migrate on the stoss of large ripples (*Ewing et al.*, 2017; *Lapôtre et al.*, 2016; 2018; *Sullivan et*  
538 *al.*, 2008). These bedform wavelengths were found to correlate negatively with elevation on the  
539 planet (*Lapôtre et al.*, 2016; *Lorenz et al.*, 2014) (discussed again in §7.3).

540 Two additional classes of meter-scale (and larger) ripples are observed on Mars with coarser  
541 grains inferred to occur along the crest and limited observed activity: coarse-grained ripples and  
542 transverse aeolian ridges. As with terrestrial mega-ripples, the coarser fraction along the crest has

543 implications for both the morphology and activity of these bedforms. As coarser elements  
544 accumulate near the crests, mega-ripple dimensions (spacing and heights) gradually increase  
545 (*Andreotti et al.*, 2002). Mega-ripples on Earth migrate and respond to changes in winds  
546 relatively slowly as typical wind stresses are below the threshold to initiate and sustain surface  
547 creep of coarser sand (*Bagnold*, 1941; *Lämmel et al.*, 2018). Critically, mega-ripples may need  
548 ample saltating sand driven by a formative, preferentially uni-directional wind regime to migrate  
549 (e.g., during infrequent storms).

550 Meter-scale, bimodal coarse-grained ripples (descriptive term employed here without  
551 implication of specific modes of transport) were identified during MER traverses at Gusev crater  
552 (*Sullivan et al.*, 2008) and Meridiani Planum (*Jerolmack et al.*, 2006; *Sullivan et al.*, 2005). For  
553 example, with the active decimeter ripples, ~3 m wavelength and ~30 cm tall dark ripples were  
554 observed in the El Dorado ripple field. This location contained both fine- and coarse-grained  
555 ripples, and both appeared to be static based on the grain size distribution and lack of sediment  
556 mobility, except for dust removal (*Sullivan et al.*, 2008). Coarse-grained ripples have also been  
557 observed by the *Curiosity* rover in Gale crater along the trailing edge of Bagnold dune field and  
558 outside of the active dune field in isolated sand sheets (Figure 6), with variable crest grain sizes  
559 and amount of dust cover (*Lapôtre et al.*, 2018; *Weitz et al.*, 2018). More recently martian  
560 “mega-ripples” were interpreted using orbital data due to their greater dimensions (5–20 m  
561 spacing, ~1–5 m tall) and brighter crests than typical dark decameter ripples, where the latter was  
562 inferred as a coarser grain size component (*Silvestro et al.*, 2020). These intermediate-scale  
563 bedforms are typically trailing the stoss side of or flanking dunes, dominantly transverse in  
564 morphology, and some were recently reported to be migrating (*Chojnacki et al.*, 2019; *Silvestro*  
565 *et al.*, 2020; SOM 4).

566 Larger, martian bedforms (10–200 m wavelength, 1–14 m tall) termed transverse aeolian  
567 ridges (TARs) were first noted and debated following their discovery in early high-resolution  
568 image data (*Bourke et al.*, 2003). TARs tend to have longer, more widely distributed  
569 wavelengths and are brighter than the large ripples (*Lapôtre et al.*, 2016), and tend to have more  
570 symmetric profiles than most bedforms (*Zimelman*, 2010). These enigmatic bedforms are  
571 concentrated in the martian tropics, appearing in isolated or expansive fields across plains, within  
572 craters or canyons, or in association with large dark dunes (*Balme et al.*, 2008; *Berman et al.*,  
573 2011; 2018; *Bourke et al.*, 2003; *Geissler*, 2014; *Geissler and Wilgus*, 2017; *Hugenholtz et al.*,

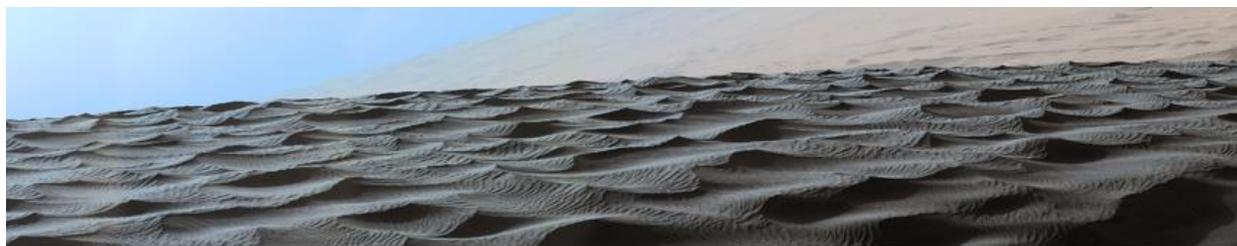
574 2017; *Wilson and Zimbelman*, 2004; *Zimbelman*, 2010; 2019). TARs are generally thought to  
575 form from surface creep of coarse-grained particles (e.g., *Bourke et al.*, 2003; *Hugenholtz et al.*,  
576 2017; *Zimbelman*, 2010) or the deposition, induration, and erosion of dominantly dust-sized  
577 particles (*Geissler*, 2014). Although initially without a good terrestrial analog, moderate-scale  
578 aeolian bedforms (2–250 m wavelength, 1–4 m tall) were recently identified in deserts of Iran  
579 and Libya (*Foroutan and Zimbelman*, 2016; *Foroutan et al.*, 2019). It was also recently proposed  
580 that *Curiosity* traversed a TAR in Gale crater (*Zimbelman and Foroutan*, 2020).

581 As on the Earth, the largest aeolian bedform class are sand dunes. These features were seen  
582 in some of the earliest imagery (*Greeley et al.*, 1992; *Masursky*, 1973; *Sagan et al.*, 1972; 1973)  
583 and have been mapped globally (*Hayward et al.*, 2007; 2010; 2012; 2014; *Fenton*, 2020). The  
584 most extensive coverage of dune sand occurs within the northern circum-polar basins as nearly  
585 continuous sand seas (e.g., Olympia Undae) (*Hayward et al.*, 2014; *Lancaster and Greeley*,  
586 1990). Impact craters are the most wide-spread locale for dune fields because these serve as a  
587 natural sediment sink (*Dorn and Day*, 2020; *Greeley et al.*, 1992; *Hayward et al.*, 2007; 2014;  
588 *Roback et al.*, 2020). Other common settings for dune fields are topographic depressions such as  
589 troughs, valleys, and chaotic terrain, including the great structural rift system of Valles Marineris  
590 (*Chojnacki et al.*, 2014a). Less commonly, extra-crater plains may host dispersed clusters of  
591 dunes (*Chojnacki et al.*, 2018; *Fenton*, 2005; *Hayward et al.*, 2007).

592 Specific dune morphologies could be properly classified following the advent of high-  
593 resolution image data (*Malin et al.*, 1992; 2007; *McEwen et al.*, 2007) and are sorted using  
594 classic terrestrial classifications as defined by *McKee* (1979) (SOM 5). The vast majority of  
595 martian dune morphologies occur as crescent-shaped dunes (i.e., barchan, barchanoid) where  
596 horns overall point in the downwind direction (e.g., Figures 4, 5), although the occurrence of  
597 asymmetric barchans and linear dunes growing through a fingering instability (*Courrech du Pont*  
598 *et al.*, 2014) has been recognized on Mars (e.g., *Ewing et al.*, 2017; *Silvestro et al.*, 2016). Other  
599 not-uncommon dune types include linear, transverse (e.g., Figure 1), star, sand sheet, and dome  
600 dunes (*Davis et al.*, 2020; *Fenton et al.*, 2013; *Hayward et al.*, 2007). Overlapping dunes  
601 (compound) and/or combinations of dune morphologies (complex) are also very commonly  
602 observed in large dune fields or ergs (*Brothers and Kocurek*, 2018; *Chojnacki et al.*, 2014a;  
603 *Fenton et al.*, 2013). Less common classes of topographically related dunes may be found on  
604 crater or canyon walls as falling or climbing dunes (*Bourke et al.*, 2004; *Chojnacki et al.*, 2010).

605 Additional occurrences of dunes possessing unusual morphologies that were not readily  
606 classified using terrestrial types were also found (*Hayward et al.*, 2007). For example, “bullseye”  
607 dune fields, based on their concentric ring patterns, only occur in high-southern latitude craters  
608 and are unreported on Earth (*Fenton and Hayward*, 2010; *Hayward et al.*, 2014). More broadly,  
609 sand dune morphology of the high southern latitudes (poleward of 50°S) show well-rounded  
610 crests and lee-sides below the angle of repose, likely due to limited aeolian activity and the  
611 prominence of ground ice (*Banks et al.*, 2018; *Fenton and Hayward*, 2010). Ultimately, these  
612 different dune morphologies form in response to the numerous extraneous environmental factors  
613 of Mars (e.g., wind direction and variability, transport capacity, sand supply, topography,  
614 seasonal frost/ice) (e.g., *Courrech du Pont et al.*, 2014; *Ewing and Kocurek*, 2010; *Gao et al.*,  
615 2015; *Kocurek and Lancaster*, 1999, *Rubin and Hunter*, 1987). The only ground observations of  
616 martian dunes to-date come from *Curiosity*’s investigation of the Bagnold dune field, where  
617 barchans migrate along the field’s trailing edge, transitioning into barchanoidal ridges and into  
618 linear oblique dunes further south towards Aeolis Mons (informally known as Mount Sharp)  
619 (*Bridges and Ehlmann*, 2017; *Lapôtre and Rampe*, 2018).

620



621  
622 [Figure 6. MSL Mastcam image showing two sizes of ripples over the stoss slope of Namib dune,](#)  
623 [Bagnold dune field. The large ripples have ~2-m wavelength, and the smaller ones have ~10-cm](#)  
624 [wavelength. Image is NASA photojournal PIA20755, acquired 2015-12-13 on \*Curiosity\*’s 1192<sup>nd</sup>](#)  
625 [sol.](#)

626

## 627 **2.2 Aeolian Transport, Fluxes, and Erosion Rates**

628 Knowledge of the minimum wind speed capable of inducing aeolian transport is central in  
629 predicting bedform migration rates, resurfacing rates, and dust emissions in ancient and  
630 contemporary martian climates (e.g., *Bagnold*, 1941; *Greeley and Iversen*, 1985; *Kok et al.*,  
631 2012; *Sullivan and Kok*, 2017). Winds below the threshold, or minimum wind speed for motion,  
632 are not sufficient to mobilize material; thus, determining the minimum wind speed required to  
633 initiate motion on the surface of Mars can unlock clues regarding Mars’ past climate and weather  
634 phenomena. For example, aeolian sedimentary strata reveal the sizes of grains transported under

635 past climates and directional changes in transport. Such strata are found throughout Mars’  
636 landscape giving us hard evidence for how the wind has interacted with the surface, especially  
637 when having speeds greater or equal to the threshold for grain motion (e.g., *Banham et al.*, 2018;  
638 *Chojnacki et al.*, 2020; *Day et al.*, 2019; *Grotzinger et al.*, 2005; *Milliken et al.*, 2014). By  
639 understanding how the threshold of wind-driven grain motion has changed over time as the  
640 climate shifted, we can begin mapping aeolian processes throughout Mars’ history using the  
641 process-based evidence solidified in martian sedimentary strata. We can also use these thresholds  
642 to predict contemporary activity on Mars—in particular to forecast surface dust emission rates,  
643 which is critical for landed robotic and human exploration.

### 644 **2.2.1 Thresholds of motion and transport hysteresis**

645 The fluid threshold for wind-blown sand is the minimum shear velocity required to initiate  
646 grain movement by the force of the wind alone and was developed to predict dust emission and  
647 landform change in sandy environments on Earth (*Bagnold*, 1936; 1937). The Shields-type  
648 function is central to most modern threshold equations for Mars and uses shear velocity, a height  
649 independent variable that represents the momentum transfer from the boundary layer to the  
650 surface,  $u_* = \sqrt{\frac{\tau}{\rho}}$  (in m/s; where  $\tau$  is stress in Pa,  $\rho$  is density in kg/m<sup>3</sup>), to predict the onset of  
651 motion:

$$652 \quad u_{*t} = A \sqrt{\left(\frac{\rho_s - \rho}{\rho}\right) g d} \quad (\text{Eqn 1})$$

653 where  $u_{*t}$  is the threshold shear velocity (m/s),  $\rho_s$  and  $\rho$  are sediment and fluid densities (kg/m<sup>3</sup>),  
654  $g$  is gravitational acceleration (m/s<sup>2</sup>),  $d$  is grain size (m), and  $A$  is an empirically derived  
655 constant (equal to square root of the Shields criterion) that includes a dependence on particle  
656 Reynolds number at threshold conditions,  $Re_{pt} = \frac{u_* d}{\nu}$ , where  $\nu = \frac{\mu}{\rho}$  is the kinematic viscosity of  
657 the winds (m<sup>2</sup>/s), with  $\mu$  as their dynamic viscosity (Pa·s). The first threshold models for Mars  
658 resolved estimates of the  $A$  parameter based on wind tunnel experiments in the Planetary Aeolian  
659 Laboratory’s MARTIAN SURFACE WIND TUNNEL (MARSWIT) at NASA’s Ames Research Center  
660 (*Greeley et al.*, 1976; 1980; *Iversen and White*, 1982): the threshold was reached when  
661 “...saltation (along the entire wind tunnel) test bed was initiated (following *Bagnold* (1941))”  
662 (*Greeley et al.*, 1976, p. 418). Observations of  $u_{*t}$  were used to back out detailed models for  
663 estimating the  $A$  parameter using  $Re_{pt}$ , resulting in three conditional models:  
664

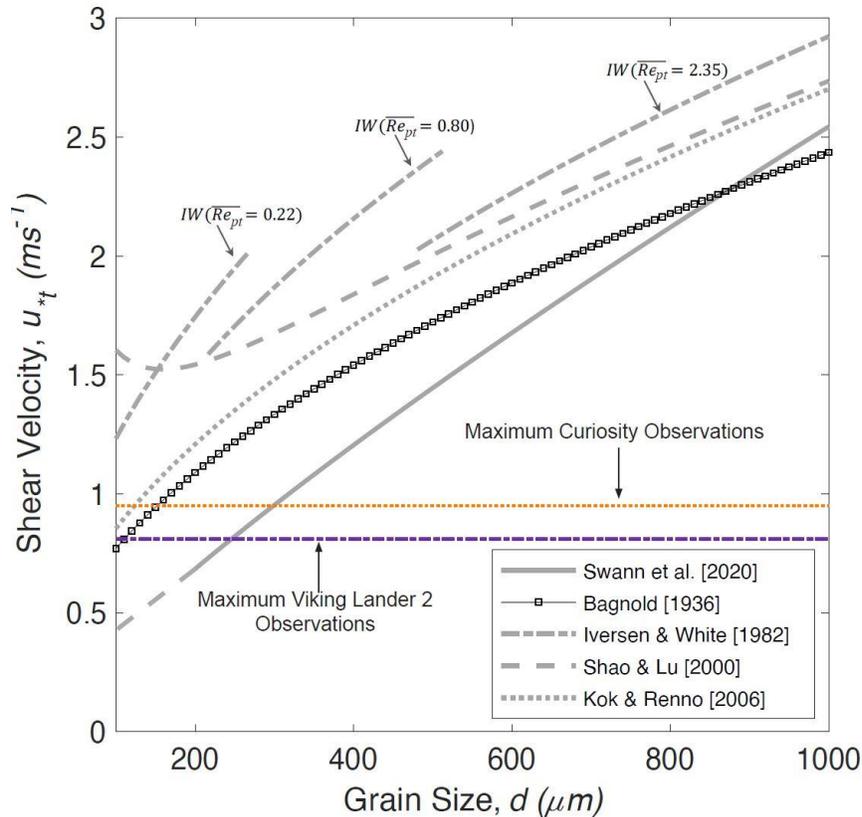
665 
$$A = 0.2 \sqrt{\frac{\left(1 + \frac{0.006}{\rho_s g d^{2.5}}\right)}{1 + 2.5 Re_{pt}}} \text{ for } 0.03 \leq Re_{pt} \leq 0.3 \quad (\text{Eqn 2})$$

666 
$$A = 0.129 \sqrt{\frac{\left(1 + \frac{0.006}{\rho_s g d^{2.5}}\right)}{1.928 Re_{pt}^{0.092} - 1}} \text{ for } 0.3 \leq Re_{pt} \leq 10 \quad (\text{Eqn 3})$$

667 
$$A = 0.120 \sqrt{\left(1 + \frac{0.006}{\rho_s g d^{2.5}}\right)} \{1 - 0.0858 \exp[-0.0671(Re_{pt} - 10)]\} \text{ for } Re_{pt} \geq$$
  
 668 10 (Eqn 4).

669 These original models predicted the minimum shear velocity required to mobilize sand as well as  
 670 the optimum grain size for windblown transport over a range of atmospheric densities (Figure 7).  
 671 Yet, these equations predict threshold winds speeds higher than those modelled or measured at  
 672 the surface of on Mars (e.g., *Gomez-Elvira et al.*, 2014; *Lorenz*, 1996; *Newman et al.*, 2017) and  
 673 leading to a discrepancy between lower than threshold martian wind speeds and active sediment  
 674 transport observed from orbital imagery and landers on Mars.

675



676  
 677 Figure 7. Plot showing different Martian threshold models for sediment transport, along with  
 678 estimated shear velocities from *Curiosity* and Viking Lander 2. ‘IW’ refers to the Iversen and  
 679 White (1982) model, estimated using specified values of  $Re_{pt}$ .  
 680

681 Two reasons for this discrepancy are (1) the experimental criterion used to define the  
 682 threshold and (2) the absence of a complete dimensional transformation of their empirical data  
 683 (Swann *et al.*, 2020). Defining the threshold as the onset of continuous motion over the test bed,  
 684 a common practice on Earth, disregards intermittent sporadic motion that occurs at slower shear  
 685 velocities. In particular, this definition disregards the ability for a small burst of sand grains to  
 686 induce equilibrium transport downwind through impact cascades (Bauer *et al.*, 2009; Sullivan  
 687 and Kok, 2017). Through a set of numerical experiments, Sullivan and Kok (2017) determined  
 688 that, on Mars, cascading saltation can lead to continuous saltation but over distances much longer  
 689 than available in laboratories. This finding is highly significant for martian aeolian processes.  
 690 High-frequency turbulent fluctuations that momentarily exceed the threshold for motion can  
 691 induce transport of a small patch of grains that, downwind, can become equilibrium transport.  
 692 The concept is hinged on a lower, impact threshold,  $u_{*it}$ . The impact threshold occurs at slower  
 693 shear velocities because the momentum transferred to particles at rest is a function of the wind  
 694 and the impact of saltating grains. Thus, the momentum from the wind does not need to be as

695 great in order to sustain motion because the impact of saltating grains dislodges particles at rest.  
696 On Earth, the impact threshold is approximately 80% of the fluid threshold, but on Mars it is  
697 predicted to be as low as 10–20% of the fluid threshold due to the much lower atmospheric  
698 density (*Kok, 2010*). Thus, once particles are mobilized, wind speed has to drop significantly in  
699 order for particle motion to cease.

700 In light of these findings, new experimental observations were recently conducted in the  
701 MARSWIT to resolve the threshold at the onset of cascading saltation of sand-sized particles  
702 (*Burr et al., 2020: 150–1000 μm; Swann et al., 2020: 200–800 μm*). Incrementally increasing the  
703 speed over a bed of particles at rest, these studies dimensionally transformed wind tunnel  
704 observations from a set of vertically stacked pitot tubes to calculate shear velocities  
705 corresponding to discontinuous, sporadic motion; here, we report primarily on the results from  
706 *Swann et al. (2020)*. These shear velocities were used to resolve Bagnold’s *A* parameter for  
707 cascading motion from Equation (1):

$$708 A_{Fluid} = 0.0502 D_*^{0.3157} \quad (\text{Eqn 5})$$

$$709 A_{General} = 0.0646 D_*^{0.2426} \quad (\text{Eqn 6})$$

710 where

$$711 D_* = d \left( \frac{\rho(\rho_s - \rho)}{\mu^2} \right)^{\frac{1}{3}} \quad (\text{Eqn 7}).$$

712 The new model predicts threshold shear velocities that are slower than previous models by a  
713 factor of 1.6 to 2.5. In their model, for a surface with an average grain size of 200 μm, the  
714 minimum shear velocity required to initiate cascading motion ranges from 0.63 to 0.81 m/s at  
715 atmospheric densities between 0.013 to 0.025 kg/m<sup>3</sup>, reconciling theory with measured wind  
716 speeds (Figure 7). However, their model is only valid for particles ranging from 200 to 800 μm,  
717 excluding values for finer particles where interparticle cohesion increases the threshold for  
718 motion (*Bagnold, 1937; Iversen and White, 1982; Shao and Lu, 2000*). The transition from  
719 cohesion-dominated to gravity-dominated threshold is represented by a marked upturn, or  
720 inflection, in threshold curves where forces required to initiate motion increase due to an  
721 increase in interparticle attractive forces between finer particles (Figure 7). Predicting the  
722 inflection point in the threshold curve determines the optimum grain size (i.e. the easiest particles  
723 to move by the force of the wind) and represents the most commonly mobilized particles. Early  
724 workers estimated that this inflection point should lie between 100 and 200 μm (*Bagnold, 1937;*

725 *Iversen and White, 1982; Shao and Lu, 2000*); this prediction is consistent with *Curiosity's*  
726 observations of well sorted, unimodally distributed 100–150  $\mu\text{m}$  sand in the active Bagnold dune  
727 field (e.g., *Weitz et al., 2020*).

728

### 729 **2.2.2 Bedform migration and evolution**

730 Bedforms, from small impact ripples up through mature dunes, have been observed to  
731 migrate in a range of locations on Mars. These migration rates and the scale of the bedforms  
732 indicate variable sediment flux rates, which are typically an order of magnitude lower than  
733 terrestrial rates (*Bridges et al., 2012b; Chojnacki et al., 2019*).

734 Small ripples with decimeter wavelength have been observed to migrate short distances over  
735 a few sols around Mars rovers during windy seasons (i.e., southern summer (*Ayoub et al., 2014;*  
736 *Baker et al., 2018b*)). For example, poorly sorted <300- $\mu\text{m}$  sand at El Dorado were observed to  
737 migrate about 2 cm over 5 sols (*Sullivan et al., 2008*), and small ripples in fine sand were  
738 observed to migrate by up to 2.8 cm/sol in sand patches at Gale crater (*Baker et al., 2018b*).  
739 Assuming activity during half of the martian year, extrapolated migration rates range from 10 cm  
740 to 10 m per martian year (*Baker et al., 2018b*).

741 Migration of dark meter-scale ripples, ubiquitous in association with dark dunes (*Bridges et*  
742 *al., 2007*), has been observed in high-resolution repeat orbital images (e.g., SOM 3). In these  
743 images, ripple displacements can be measured manually or in aggregate for larger areas using the  
744 Co-registration of Optically Sensed Images and Correlation (COSI-Corr) methodology (*Bridges*  
745 *et al., 2013; Leprince et al., 2007*). The first unambiguous meter-scale modification of ripples  
746 and dune edges was documented in Nili Patera (*Silvestro et al., 2010*), where superposed  
747 decimeter-tall ripples (*Ewing et al., 2017; Lapôtre et al., 2018*) may migrate up to several meters  
748 per year, but average  $\sim 0.5$  m/yr from larger sampling (*Ayoub et al., 2014; Bridges et al., 2012a;*  
749 *Chojnacki et al., 2018; Preston and Chojnacki, 2019; Runyon et al., 2017; Silvestro et al., 2013*).  
750 Ripples are swiftest mid-way up a dune's stoss slope through the dune crest:  $\sim 5$ x faster than  
751 ripples at the base of the stoss or in the lee or flanks areas (*Bridges et al., 2012a; Preston and*  
752 *Chojnacki, 2019; Roback et al., 2019; Runyon et al., 2017*). In general a linear relationship  
753 between ripple migration rate and ripple elevation on the dune has been demonstrated (*Bridges et*  
754 *al., 2012a; Runyon et al., 2017*), likely due to streamline compression from dune topography as  
755 winds are pushed upslope. Isolated ripple patches not associated with a dune field have the  
756 lowest migration rates; such rates are detected using image pairs spanning two or more martian

757 years. These measurements reflect sand flux rates between  $0.1\text{--}2.3\text{ m}^3\text{ m}^{-1}\text{ yr}^{-1}$ , which are  
758 typically several factors less than the saltation rate suggested by the migration rate of  
759 neighboring dunes (Ayoub *et al.*, 2014; Bridges *et al.*, 2012b; Roback *et al.*, 2019; Runyon *et al.*,  
760 2017; Silvestro *et al.*, 2013). Saltation rates also appear higher during the northern hemisphere  
761 autumn/winter, which is also when driving winds are likely greatest (Ayoub *et al.*, 2014; Roback  
762 *et al.*, 2019).

763 The first clear observation of bedform change from orbital data was the gradual  
764 disappearance of two small ( $\sim 1000\text{ m}^2$ ) north polar dome dunes and  $\sim 85\%$  deflation of a third  
765 over a five-year time span (1999–2004) in MOC images (Bourke *et al.*, 2008). Since then,  
766 several studies have used various combinations of HiRISE pairs and topography to estimate  
767 migration rates and sand fluxes for dunes (Ayoub *et al.*, 2014; Bridges *et al.*, 2012a; 2012b;  
768 Cardinale *et al.*, 2020; Chojnacki *et al.*, 2015; 2017; 2018; Hansen *et al.*, 2011; Runyon *et al.*,  
769 2017; Silvestro *et al.*, 2013; Figure 8; SOM 2–4, 6). Reported average migration rates are  
770 consistently  $\sim 0.5\text{ m/yr}$  ( $\pm 0.4\text{ m/yr}$ ,  $1\sigma$ ) for dunes that are  $\sim 2\text{--}120\text{-m}$  tall (average height  $19\pm 14$   
771 m) (Chojnacki *et al.*, 2019). These reported rates are typically for barchan or barchanoid dune  
772 morphologies in uni-directional wind regimes (e.g., SOM 6), but include some instances of  
773 linear, dome, and falling dunes. Average crest flux measurements for dune fields ranged between  
774  $1\text{--}18\text{ m}^3\text{ m}^{-1}\text{ yr}^{-1}$  (average across  $q_{\text{crest}} = 7.8\pm 6.4$  ( $1\sigma$ )  $\text{m}^3\text{ m}^{-1}\text{ yr}^{-1}$ ), where the maximum flux for an  
775 individual dune was  $35\text{ m}^3\text{ m}^{-1}\text{ yr}^{-1}$  (Chojnacki *et al.*, 2019). These rates and fluxes are relatively  
776 variable in terms of geography and timing. For example, the highest sand fluxes documented to  
777 date appear to concentrate in three regions: Syrtis Major, Hellespontus Montes, and the north  
778 polar erg (Chojnacki *et al.*, 2019). Poleward of  $45^\circ\text{ S}$ , dunes sites show limited bedform  
779 mobility, and southward of  $57^\circ\text{ S}$  only ripple migration has been detected (i.e., no bulk dune  
780 movement) (Banks *et al.*, 2018). Dunes surrounding the north polar layered deposits and residual  
781 cap display the greatest migration rates and fluxes:  $\sim 50\%$  greater than on average for Mars ( $11.4$   
782 vs.  $7.8\text{ m}^3\text{ m}^{-1}\text{ yr}^{-1}$ ) (Chojnacki *et al.*, 2019). These higher values are found in the polar regions  
783 despite the limited sediment state caused by autumn/winter  $\text{CO}_2/\text{H}_2\text{O}$  ice accumulation that  
784 reduces surface interactions with the wind (Diniaga *et al.*, 2019a; Hansen *et al.*, 2011; 2015).

785

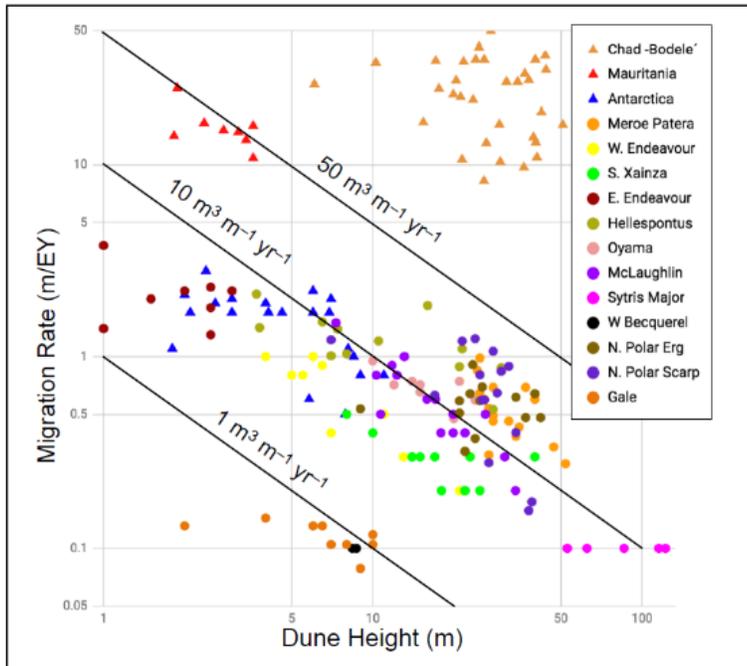


Figure 8. Log-log plot of dune migration rates vs. dune heights for select martian and terrestrial dune fields. Dune rates are averages per site using the longest-baseline HiRISE orthoimages available. Migration rate confidence intervals are typically  $\pm 0.1$ – $0.3$  m/yr and account for orthoimage registration offsets and human error in manual measurements. The "terrestrial" estimates are from dune field data collected in Chad, Mauritania, and Antarctica (Bourke *et al.*, 2009; Ould Ahmedou *et al.*, 2007; Vermeesch and Drake, 2008). Mars dune flux estimates are from earlier work by Chojnacki *et al.* (2017; 2018; 2019).

805        Aside from many ripples within the southern dune fields, many smaller martian bedforms  
 806        show no sign of present-day migration; such features may also have superposed craters, debris,  
 807        and fracturing that indicate a long-term lack of migration and renewal. In particular, with a few  
 808        newly identified exceptions (Silvestro *et al.*, 2020), TARs appear to be inactive based on  
 809        morphology and context (Berman *et al.*, 2018). For example, crater age dating indicates certain  
 810        TAR fields in Schiaparelli crater have been inactive for the last  $\sim 100$  kyr to  $\sim 2$  Myr, suggesting  
 811        that they are relict deposits (Berman *et al.*, 2018). Numerous authors investigating dark ripples or  
 812        dunes via comparison of HiRISE image pairs have reported on the lack of apparent motion for  
 813        nearby TARs. However, efforts just may not have used sufficiently long temporal baselines; for  
 814        example, these investigations typically used images spanning 2–3 martian years for a survey of  
 815        low sand flux regions (e.g., Valles Marineris, Meridiani) or dune migration (Banks *et al.*, 2015;  
 816        Berman *et al.*, 2018; Bridges *et al.*, 2012a; Chojnacki *et al.*, 2014a; 2017; Geissler *et al.*, 2012).  
 817        Using longer baseline images ( $>4$  martian years) and targeting known high flux dunes within  
 818        McLaughlin crater, several bright-toned TAR-like bedforms showed unambiguous crest  
 819        displacements (Silvestro *et al.*, 2020). It may be that certain TAR populations within high flux  
 820        sand corridors are subjected to enough repeated saltation to dislodge their presumably coarser-  
 821        grained crest areas. Preliminary results suggest mega-ripple and TARs that are migrating today

822 are doing so with rates and fluxes an order of magnitude lower than those estimated for adjacent  
823 sand dunes (*Silvestro et al.*, 2020; SOM 4).

### 824 825 **2.2.3 Erosion Rates**

826 We focus here on bulk surface erosion rates that are likely to be primarily driven by aeolian  
827 erosion (versus mass wasting, which is discussed in §3, 5), predominantly via sand abrasion  
828 (*Laity and Bridges*, 2009). Bulk surface erosion rates have generally been estimated based on the  
829 existence, age, and geomorphology of various crater populations along with geologic setting. For  
830 example, locations in Gusev crater showed in situ and orbital estimates of  $10^{-3}$ – $10^{-5}$  m/Myr (note  
831 that m/Myr is equivalent to  $\mu\text{m}/\text{yr}$ ) (*Golombek et al.*, 2006) and rates of  $10^{-2}$ – $10^{-3}$  m/Myr were  
832 estimated for Elysium Planitia based on crater depth degradation and rim erosion (*Sweeney et al.*,  
833 2018). Based on deviations in small-crater counts from expected isochrones, the crater  
834 obliteration rate for light-toned sedimentary rocks suggests an average erosion rate of  $10^{-1}$   
835 m/Myr (*Kite and Mayer*, 2017). Younger terrain in Meridiani Planum yielded higher rates 1–10  
836 m/Myr (*Golombek et al.*, 2014), which may be more similar to wind-driven scarp retreat in Gale  
837 crater, as suggested by radiogenic and cosmogenic dating of exposed sediments within Aeolis  
838 Mons (*Farley et al.*, 2014). (As noted in those studies, terrestrial continental denudation rates for  
839 arid regions are still a few (2–5) orders of magnitude higher.)

840 To quantitatively connect surface abrasion rates to aeolian sand flux rates, the total sand flux  
841 (i.e., saltation plus reptation) is needed. This can be estimated from the dune crest fluxes and  
842 making some assumptions about the mass loss from impacting sand on the target material. For  
843 basalt sand grains hitting basaltic rocks at the impact threshold for Mars, this value of abrasion  
844 susceptibility is  $\sim 2 \times 10^{-6}$ , based on laboratory measurements (*Greeley et al.*, 1982) and  
845 accounting for the energetics of martian saltation and reptation (*Bridges et al.*, 2012b). Taking  
846 the estimated saltation and reptation trajectories for Mars of 0.1–0.5 m (*Kok*, 2010) and interdune  
847 sand fluxes, abrasion rates for a range of sloping surfaces (i.e., flat ground to a vertical rock face)  
848 can be approximated (a detailed methodology for doing this is explained in *Bridges et al.*  
849 (2012b)). Abrasion rates for several sites have been reported (e.g., Nili Patera, Gale crater,  
850 Mawrth Vallis, Jezero crater) and range 0.01–1.3 m/Myr for flat ground and 0.3–47 m/Myr for  
851 vertical rock faces (*Bridges et al.*, 2012b; *Chojnacki et al.*, 2018; *Farley et al.*, 2014).

852

### 853 **2.3 Open questions for martian aeolian landforms and sediment history**

854 Major questions remain open about the age, sources, and amounts of dust and sand on Mars.  
855 The few areas where dune sand is traced back to a source involve eroding crater walls or polar  
856 layered deposits, where sand appears to be recycled from sandstone or an ancient erg,  
857 respectively (e.g., *Chojnacki et al.*, 2014b; *Tirsch et al.*, 2011). On Earth, most sand grains form  
858 from chemical and physical erosion of quartz down to a stable grain size (*Krinsley and Smalley*,  
859 1972); Mars instead is predominantly basaltic (*Ehlmann et al.*, 2017; *Greeley and Iverson*, 1985;  
860 *Minitti et al.*, 2013; *Yen et al.*, 2005). Models predict that sand-sized grains could be created  
861 through explosive volcanic processes (*Edgett and Lancaster*, 1993; *Wilson and Head*, 1994), but  
862 the most recent volcanism occurred 2–10 Mya (e.g., *Neukum et al.*, 2004b). Others have  
863 proposed that sand grains may form by fragmentation driven by impact and aeolian processes  
864 (*Golombek et al.*, 2018; *McGlynn et al.*, 2011). Some have proposed that the general generation  
865 and flux of granular material on Mars has declined over time, with the impact, volcanic and  
866 chemical weathering processes on an ancient, wet Mars generating the majority of sediment  
867 (*Grotzinger and Milliken*, 2012; *McLennan et al.*, 2019). However, it is not currently known if  
868 most martian sand has been recycled or if a significant amount is actively forming in the present  
869 climate.

870 Similar questions can be asked about dust. The global dust budget and surface reservoir  
871 distribution, as well as the dust lofting rate, present important controls on climate models. It is  
872 important to understand not only the present state, but also how dust availability and distribution  
873 may have changed through climate cycles (i.e., thousands to millions of years) and climate  
874 epochs (i.e., to billions of years).

875 Observations of a few dune fields suggest that sand is size and compositionally sorted (e.g.,  
876 *Chojnacki et al.*, 2014b; *Lapôtre et al.*, 2017; *Pan and Rogers*, 2017; *Seelos et al.*, 2014) as it  
877 progresses through a transport pathway and aeolian bedforms. Such observations present an  
878 interesting feedback question, as grain size can influence evolution/mobility of the bedforms and  
879 further grain transport. This also suggests that additional complexity may be needed in models  
880 connecting landform morphology to formation history.

881 Although hypotheses for the growth-limiting mechanism of meter-scale ripples are  
882 converging towards an aerodynamic process (e.g., *Duran Vinent et al.*, 2019; *Lapôtre et al.*,  
883 2016; 2021; *Sullivan et al.*, 2020) the nature of their inception mechanism is still being debated

884 (*Duran Vinent et al.*, 2019; *Sullivan et al.*, 2020). Questions about present-day activity rates (if  
885 nonzero) and formative history of such features, and why this diversity of bedforms is found,  
886 remain an open area of study. While these questions are about the evolution of landforms, such  
887 models are built from sediment flux and saltation layer models, which in turn depend on models  
888 of how individual grains are moved along the surface (e.g., the fractional contributions of  
889 saltation versus reptation to a wind-driven sand flux)—discussed more in §2.4.

890

## 891 **2.4 Open questions for the physics of aeolian processes**

892 Regarding the fundamental physics of aeolian grain transport, terrestrial and laboratory  
893 studies form the basis of the majority of information known about the influence of different  
894 parameters (summarized in *Pahtz et al.*, 2020). In application of these models towards the  
895 martian environment, current threshold models predict minimum wind speeds that align with  
896 observed wind speeds on Mars (*Burr et al.*, 2020; *Kok et al.*, 2012; *Swann et al.*, 2020).  
897 However, a number of uncertainties in the application of thresholds to natural boundary layers  
898 acting over spatially heterogeneous surfaces and bedforms on Mars remain, including: (1) the use  
899 of idealized surface conditions for threshold model derivation, (2) the difficulty in obtaining  
900 necessary parameters such as grain size, shape and density on Mars, and (3) potential errors in  
901 estimating shear velocity from single-height wind speed observations.

902 Empirical coefficients in Martian threshold models are derived for idealized surface  
903 conditions, saltating particles moving over flat beds of cohesionless grains with uniform size  
904 distributions. These do not represent the more complex surfaces and bedforms found on Mars,  
905 e.g., stoss slopes of dunes, mixed grain size surfaces and bedforms, and coarse-lag deposits.  
906 Surfaces and bedforms with mixed grain size distributions, sediment consolidation levels, or  
907 coarse lag deposits can act to increase the minimum wind speed required to initiate saltation or  
908 become active by a different mode of transport (e.g., saltation vs. rolling particles or reptation).

909 Uncertainty in threshold predictions also arises from the difficulty in determining grain size,  
910 shape, and density comprising aeolian bedforms on Mars. In situ observations from landers and  
911 rovers have been successful at determining these characteristics. For example, *Curiosity*'s Mars  
912 Hand Lens Imager (MAHLI) determined particle sizes and shapes within ripples throughout Gale  
913 crater (*Weitz et al.*, 2018). However, these observations are geographically limited and remote  
914 sensing techniques do not have the resolution required to determine grain size and density that  
915 are required to predict the threshold for motion.

916 Finally, there is uncertainty in estimating shear stress, or shear velocity, on the surface of  
917 Mars. Shear velocity, a surrogate for bed shear stress, can be estimated from single-height wind  
918 speed observations using either the covariance of 2D or 3D velocity components or von  
919 Karman's Law of the Wall. However, local thermal convection at the surface on Mars induces a  
920 dynamically unstable boundary layer (*Fenton and Michaels, 2010*). The instability in the  
921 boundary layer, represented by deviations from typical logarithmically distributed velocity  
922 fluctuations, is difficult to predict. At present, wind speeds on Mars are observed at a single  
923 height (typically ~1.5 m above the surface) and sampled at low frequencies. Thus, we have yet to  
924 measure how the boundary layer responds to variations in local convection, or estimate the error  
925 associated in low-frequency sampling that can alias shear velocity calculations, in particular  
926 when using the covariance derivation. In situ measurements of vertical velocity gradients within  
927 unstable boundary layers at the surface of Mars are necessary to reduce error in shear velocity  
928 estimation.

929 Unfortunately, testing different sediment transport processes further is not possible with  
930 existing rover payloads or from orbit. Additionally, it is difficult to mimic martian conditions,  
931 especially over sufficient distances to allow full formation of the saltation layer, within present  
932 terrestrial laboratories. As will be discussed in §6, in situ investigations are needed to acquire the  
933 high-frequency, high-resolution measurements that can correlate driving environmental  
934 conditions (such as wind velocities, including gusts, and surface pressure) with the sediment  
935 movement.

936

### 937 **3 Seasonal Frost/Ice-formed Landforms**

938 The martian atmosphere is ~95% CO<sub>2</sub> and contains trace amounts of water vapor (e.g., on the  
939 order of a few tens of precipitable microns). Under typical present-day martian surface  
940 conditions, CO<sub>2</sub> and H<sub>2</sub>O condenses near ~145 K and ~198 K, respectively (*Ingersoll, 1970*;  
941 *James et al., 1992*). Frost condensation temperatures are reached at virtually all latitudes  
942 (*Piqueux et al., 2016*), although, as on Earth, the exact duration of the period when the  
943 environment is sufficiently cold for frost or ice to accumulate (e.g., seconds, to seasons, to  
944 astronomical cycles) depends on latitude and local surface and subsurface conditions (e.g., grain  
945 size and composition, subsurface water ice content and depth) that influence the local thermal  
946 inertia, shadowing due to topography, and atmospheric conditions such as dust opacity (*Putzig  
947 and Mellon, 2007*).

948 In this section, we describe the frost and ice types that currently form on the martian surface  
949 (§3.1). Sublimation of this diurnal (i.e., only overnight) or seasonal frost/ice is highly energetic  
950 and is thought to cause erosion by inducing and enhancing mass wasting (§3.2) or by  
951 digging/scouring out material from the surface directly under a subliming ice slab (§3.3).

952

### 953 **3.1 Currently formed surface frost/ice types on Mars**

954 Present-day CO<sub>2</sub> and H<sub>2</sub>O deposition can be in the form of diurnal frosts, seasonal frosts, or  
955 snowfall. As the amount of precipitable water is so limited in the tenuous martian atmosphere,  
956 water frost/ice condensation will depend on the local partial pressure of water vapor. In contrast,  
957 CO<sub>2</sub> ice requires significantly lower temperatures to condense out of the atmosphere, but it is  
958 more abundant than water and thus is not limited by diffusion through the lower atmosphere.  
959 Tens to hundreds of micrometer thick diurnal CO<sub>2</sub> frost layers form overnight over a significant  
960 fraction of the planet (*Piqueux et al.*, 2016). During current martian winters, as much as a third  
961 of atmospheric CO<sub>2</sub> can be deposited onto the surface (*James et al.*, 1992; *Leighton and Murray*,  
962 1966) dramatically redistributing CO<sub>2</sub> and decreasing surface pressures. Accumulated decimeters  
963 or thicker depth layers of seasonal CO<sub>2</sub> frost will sinter, forming polycrystalline CO<sub>2</sub> slab ice(s)  
964 (*Matsuo and Heki*, 2009) with optical and thermal properties very different from terrestrial water  
965 frost and ice. In particular, CO<sub>2</sub> ice is transparent to visible wavelengths but opaque to thermal  
966 infrared (*Matsuo and Heki*, 2009). As the surface warms moving towards spring, the  
967 accumulated CO<sub>2</sub> and H<sub>2</sub>O frost/ice will sublime, but not uniformly. Visible solar radiation can  
968 penetrate the CO<sub>2</sub> ice layer and, via a process known as the solid state greenhouse effect (*Matson*  
969 *and Brown*, 1989) because it is analogous to the greenhouse effect in planetary atmospheres but  
970 happens in a transparent solid body instead of a gaseous atmosphere, lead to the springtime  
971 insolation-induced basal sublimation of the translucent, impermeable slab ice (*Kieffer et al.*,  
972 2006). Defrosting marks will appear first, readily visible in high-resolution images, such as  
973 sublimation spots, fans, and dark linear ‘flow’ features (*Gardin et al.*, 2010; *Kaufmann and*  
974 *Hagermann*, 2017; *Kieffer*, 2007; *Malin and Edgett*, 2001; *Pilorget et al.*, 2011; 2013) and  
975 polygonal fracturing of the ice slab (*Piqueux and Christensen*, 2008; *Portyankina et al.*, 2012).  
976 In general, sublimation can be very energetic and is thought to be a key driver for the formation  
977 of many landforms (as described below). Seasonal frosts and snowfall events have some  
978 interannual variability in terms of location and duration, which have begun to be documented in

979 a systematic manner as more complete records of the present-day climate and weather are  
980 acquired (*Calvin et al.*, 2015; *Hayne et al.*, 2016; *Piqueux et al.*, 2015b; *Widmer et al.*, 2020).

981 To date, the majority of present-day surface activity connected to surface frost/ice has been  
982 hypothesized to be controlled primarily by the deposition and/or sublimation of seasonal CO<sub>2</sub>  
983 frost/ice. The seasonal frost cap begins to form early in the martian fall, reaches maximal extent  
984 (i.e., equatorward reach) at the end of the fall, and sublimates between the end of the winter and  
985 into the spring (*Piqueux et al.*, 2015b). CO<sub>2</sub> snowfall is observed to contribute to the seasonal  
986 frost accumulation (*Gary-Bicas et al.*, 2020; *Hayne et al.*, 2012; 2014). Seasonal ice sheets reach  
987 up to ~2 m in thickness near the poles (*D.E. Smith et al.*, 2001) and fractured ice layers and  
988 detached ice blocks have been observed in the mid-latitudes (e.g., *Dundas et al.*, 2012). From  
989 orbital observations, patchy seasonal surface deposits of CO<sub>2</sub> frost have been observed as far  
990 equatorward as ~42° N (*Widmer et al.*, 2020) and 33° S (*Schorghofer and Edgett*, 2006;  
991 *Vincendon et al.*, 2010a). In the north, which is the hemisphere with more water in its polar cap  
992 (*Ojha et al.*, 2019) and atmosphere (*M.D. Smith*, 2002), a ring of water ice is annually observed  
993 equatorward of the CO<sub>2</sub> seasonal frost cap (*Appéré et al.*, 2011; *Langevin et al.*, 2005; 2007;  
994 *Wagstaff et al.*, 2008). H<sub>2</sub>O frost has been detected from orbit as far equatorward as 32° N and  
995 13° S (*Vincendon et al.*, 2010b), while in situ observations suggest H<sub>2</sub>O frost at 48° N with the  
996 *Viking 2* lander (*Hart and Jakosky*, 1986; *Svitek and Murray*, 1990; *Wall*, 1981), at 2° S with the  
997 *MER Opportunity* (*Landis*, 2007), and at 5° S with *Curiosity* at Gale crater (*Martinez et al.*,  
998 2017).

999 Although not yet well characterized through observations or models, it is likely that H<sub>2</sub>O and  
1000 CO<sub>2</sub> frost/ices do not form and evolve independently of each other and that their interplay, and  
1001 interaction with incorporated atmospheric dust, constitutes an additional control on  
1002 geomorphological activity that is not yet well understood. For example, CO<sub>2</sub> ice can serve as a  
1003 sink for water vapor (*Houben*, 1997; *Houben et al.*, 1997), and H<sub>2</sub>O deposits affect basal  
1004 sublimation of CO<sub>2</sub> (*Titus et al.*, 2020). In addition to influencing accumulation and sublimation  
1005 timing and rates, mechanical interactions between different types of frost/ice may create another  
1006 control on some geomorphological activity. For example, differences in grain sizes between a  
1007 surface condensed frost layer and snowfall may enhance wintertime mass-wasting activity  
1008 (*Hansen et al.*, 2018; §3.2.2).

1009 Over recent Mars history, Mars' obliquity shifts have also affected the spatial distribution  
1010 and stability of accumulated seasonal frost/ice. Past multi-annual (up to tens of thousands of  
1011 years) accumulation of CO<sub>2</sub> ice has formed up to ~1 km thick units within the polar regions  
1012 (*Phillips et al.*, 2011); similarly, over long periods of time, fluxes of water through the  
1013 atmosphere can result in the formation of large reservoirs at the poles (*Bierson et al.*, 2016;  
1014 *Buhler et al.*, 2020; *Manning et al.*, 2019), as well as within middle and equatorial latitudes  
1015 (*Jakosky et al.*, 2005; *Mellon et al.*, 2004; *Mellon and Jakosky*, 1993; 1995; *Mellon et al.*, 1997).  
1016 As the orbital parameters change, these water ice reservoirs can become unstable. Landforms  
1017 created through present-day or recent sublimation of such ice reservoirs are discussed in §4, and  
1018 study of these units, coupled with studies of present-day frost/ice driven surface activity, is  
1019 necessary to extend models to past climatic periods and interpret relict landforms (§7.3).  
1020 (However, discussion of the past formation and preservation of such perennial ices is outside the  
1021 scope of this review.)

1022

## 1023 **3.2 Seasonal sublimation triggered mass-wasting landforms**

### 1024 **3.2.1 Gullies**

1025 Based on morphological similarity to terrestrial gullies (i.e., comprising alcove, channel, and  
1026 apron features), martian gullies were initially hypothesized to be formed through liquid water  
1027 flow, perhaps through groundwater seepage (*Malin and Edgett*, 2000), but a source for the water  
1028 was not apparent. Early MOC observations showed signs of defrosting activity in south polar  
1029 gullies (*Bridges et al.*, 2001; *Hoffman*, 2002) but did not document any significant changes to the  
1030 frost-free surface. *Malin et al.* (2006) provided the first detailed description of contemporary  
1031 gully activity, reporting two new digitate light-toned deposits in southern-hemisphere craters.  
1032 Both deposits were associated with poorly developed gullies and were relatively superficial.  
1033 *Malin et al.* (2006) suggested that these flows indicated discharge of shallow groundwater, but  
1034 *Pelletier et al.* (2008) modeled one of the deposits in detail and found that it could be explained  
1035 by dry granular flow. *Kolb et al.* (2010) carried out similar modeling of additional light-toned  
1036 deposits without constrained formation ages and found that they too could be reproduced by dry  
1037 flows.

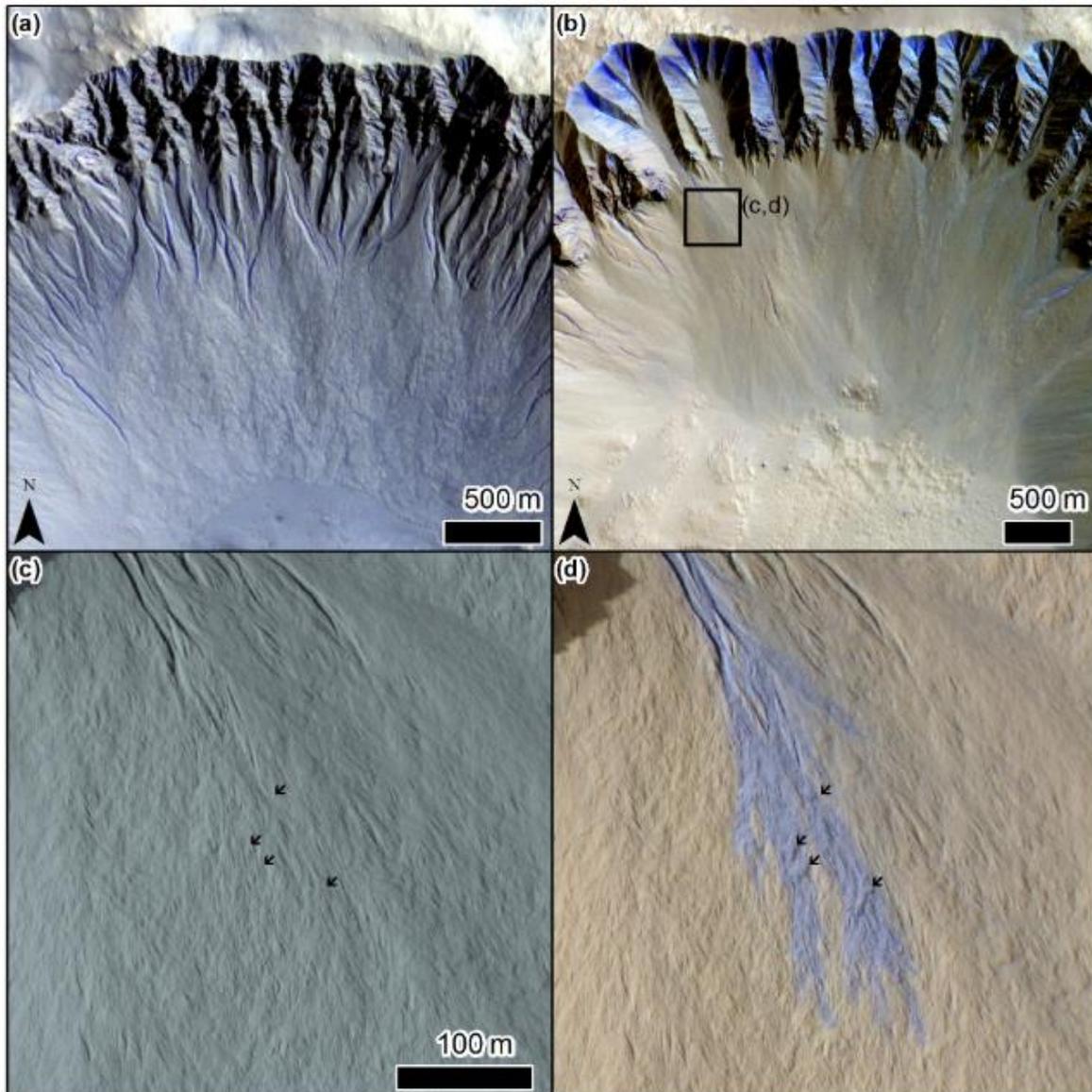
1038 Subsequent detections of more active flows in gullies along both crater walls and dune slopes  
1039 (e.g., Figures 4, 9) have led to better constraints on the processes causing activity. *Harrison et al.*  
1040 (2009), *Diniega et al.* (2010), and *Dundas et al.* (2010) all examined active gullies and reported

1041 weak seasonal constraints favoring cold-season activity for gullies on both sand dunes and other  
1042 surfaces. *Harrison et al.* (2009) favored seasonal occurrence of liquid water based on  
1043 geomorphological similarities to terrestrial debris flows, while *Diniega et al.* (2010) and *Dundas*  
1044 *et al.* (2010) proposed that winter CO<sub>2</sub> frost was driving activity in some fashion. The latter  
1045 option was strongly supported when *Dundas et al.* (2012) reported active flows with much  
1046 tighter timing constraints that correlated well with observed CO<sub>2</sub> frost, including observations of  
1047 creeping flows slowly advancing down frosted channels over a period of weeks, as well as one-  
1048 off events producing larger morphologic changes. Expanded observations with more locations  
1049 are consistent with these behaviors (*Dundas et al.*, 2015a; 2019a), as is a detailed study of gullies  
1050 in a pit near the south pole (*Raack et al.*, 2015; 2020). Morphological changes in gullies can be  
1051 extensive (*Dundas et al.*, 2012; 2015a; 2019a).

1052 Gullies located on sand dunes with classic alcove-channel-apron morphology appear to be  
1053 more active and have even larger morphological changes, possibly because of the loose substrate  
1054 (*Diniega et al.*, 2010; *Dundas et al.*, 2012; 2015a; 2019a). These features have been found on  
1055 sand dunes through the southern mid-latitudes, many with extensive annual activity (*Dundas et*  
1056 *al.*, 2019a). *Pasquon et al.* (2019a; 2019b) documented several styles of CO<sub>2</sub>-frost driven activity  
1057 that drove changes in channel sinuosity, noting an initial alcove-collapse stage followed by  
1058 transport into the lower parts of the gullies.

1059 The details of the processes by which CO<sub>2</sub> frost causes gully activity on rocky or sandy  
1060 slopes are not yet well understood. Starting shortly after the discovery of gullies, several frost-  
1061 driven processes were proposed. *Hoffman* (2002) suggested gas-lubricated flows initially  
1062 triggered by basal sublimation of translucent CO<sub>2</sub> frost and further mobilized by additional  
1063 sublimation during transport. *Ishii and Sasaki* (2004) proposed that avalanches of CO<sub>2</sub> frost  
1064 could occur, while *Hugenholtz* (2008) suggested that frosted granular flow could operate in  
1065 gullies. In the latter process, coatings of frost help to lubricate flow. *Cedillo-Flores et al.* (2011)  
1066 showed that sublimating CO<sub>2</sub> frost could effectively fluidize overlying granular material but did  
1067 not provide a mechanism for how such frost would be emplaced under regolith. *Pilorget and*  
1068 *Forget* (2016) demonstrated that basal sublimation would be effective at generating gas eruptions  
1069 in some gullies, and also that CO<sub>2</sub> ice could condense in the regolith pore space under some  
1070 conditions, but did not explain the formation of channel morphologies (that study specifically  
1071 focused on new channels forming amongst linear gullies (§3.3.1)). *Dundas et al.* (2019a) carried

1072 out calculations that showed the available energy budget was sufficient to generate gas from  
1073 entrained CO<sub>2</sub> frost during the flow, providing fluidization, and *de Haas et al.* (2019) provided a  
1074 more detailed description of the relevant physics.  
1075



1076  
1077 Figure 9. (a) Gullies in Galap crater in CaSSIS image MY34\_005744\_220\_1. (b) Gullies in Gasa crater  
1078 in CaSSIS image MY34\_005684\_218\_1, with black box showing the location of detailed panels (c, d). (c)  
1079 Image of a gully fan prior to a depositional event, HiRISE image ESP\_012024\_1440. (d) Image of a gully  
1080 fan after a depositional event, HiRISE image ESP\_020661\_1440. Arrows point to lobate deposits not  
1081 visible in previous image. Incidence angles for (c) and (d) are 57.4 and 57.9°, respectively, so these  
1082 changes are not an illumination effect; see *Dundas et al.* (2012) for an animation that provides a blink  
1083 comparison. Scale is the same as in (c).  
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### 1085 **3.2.2 Dune alcoves**

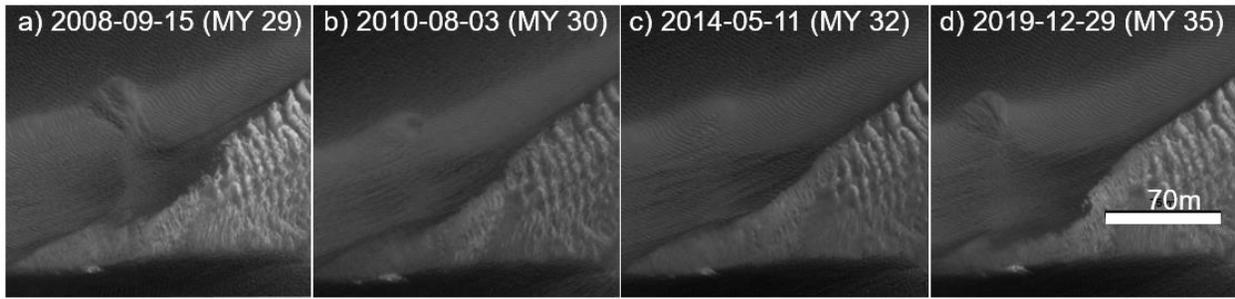
1086 New 5–40 m wide erosional alcoves are actively forming on dune lee-side slopes in the north  
1087 polar erg, often connected to a depositional apron (Figure 10); these features are called “dune  
1088 alcoves” rather than “dune gullies” (§3.2.1), as they generally lack a channel. Newly formed  
1089 alcove-apron features in the north polar erg were first reported on by *Hansen et al.* (2011), who  
1090 noted that these features formed annually and were found on ~40% of the dune slopes. In that  
1091 study, the alcove-aprons were correlated with springtime sublimation activities, such as the  
1092 appearance of dark spots and flows (*Gardin et al.* (2010); this timing and morphological  
1093 similarity to dune gullies in the southern mid-latitudes led to the hypothesis that these features  
1094 were formed through a seasonal-frost driven process— and specifically that springtime  
1095 sublimation was leading to the alcove formation activity (*Hansen et al.*, 2011). Later studies  
1096 demonstrated that new alcoves were visible under winter frosts (*Horgan and Bell*, 2012), but  
1097 were likely forming after the first autumnal frosts, thus moving the timing of alcove formation  
1098 activity to early autumn (*Diniega et al.*, 2019a; *Hansen et al.*, 2015). This led to a new  
1099 hypothesis that sublimation of diurnal frosts or interactions between early autumnal surface  
1100 frosts and snowfalls may initiate this mass wasting (*Diniega et al.*, 2019a; *Hansen et al.*, 2018).

1101 Due to this difference in timing of activity as well as dune alcoves not being reactivated in  
1102 subsequent Mars years (i.e., once the alcove forms, it fills in due to aeolian sand transport  
1103 (Figure 10), but does not widen or lengthen during subsequent winters, as many gullies are  
1104 observed to do), these features appear to be different from gullies, including dune gullies  
1105 (§3.2.1). The morphologies are also different, as few dune alcoves have a channel connecting  
1106 them to their depositional aprons (potential exceptions discussed in *Grigsby and Diniega*  
1107 (2018)).

1108 Subsequent studies have documented that dune alcoves are found within some mid-latitude  
1109 dune fields (*Diniega et al.*, 2019b). Although a much lower number of overlapping images  
1110 means timing of formation of these mid-latitude dune alcoves cannot yet be well constrained,  
1111 these features are all found in dune fields that experience seasonal frost and snowfalls, thus  
1112 remaining consistent with the polar erg-based hypotheses. However, improved constraints on the  
1113 environments where dune alcoves form, versus environments where they do not form, remain  
1114 under study and a formation mechanism model has not yet been developed.

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Figure 10. Example dune alcoves forming on a polar dune slope (Teilax dune field, 83.5°N, 118.5°E). The slope fills in due to aeolian sand transport (*c,d*) and then a new nearby dune alcove forms (*d*). This is the same location as shown in Figure 4; here, all images were acquired at about the same mid-summer period ( $L_s$  127–129°), so illumination conditions are consistent. HiRISE images are (*a*) PSP\_010019\_2635, (*b*) ESP\_018839\_2635, (*c*) ESP\_036510\_2635, and (*d*) ESP\_062923\_2635. A scale bar is shown in the last image, but absolute distances are approximate as images are not orthorectified. The bright ripples in the interdune region appear immobile over this timescale and can be used to determine relative location. North is up and illumination is from the left.

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### 3.3 Basal sublimation formed landforms

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#### 3.3.1 Linear gullies

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Martian “linear gullies” were first identified within Russell crater (*Mangold et al.*, 2003) and these decameters to kilometers-long, meters-wide troughs have since been found on a range of sandy slopes (e.g., *Dundas et al.*, 2012; *Pasquon et al.*, 2016; Figure 11). Based on their resemblance to terrestrial rills, these features were originally likened to terrestrial debris flows and proposed to be formed by surface water flow due to meltwater following a period of high obliquity (e.g., *Jouannic et al.*, 2012; *Mangold et al.*, 2003; 2010; *Miyamoto et al.*, 2004) or present-day atmospheric condensates (*Vincendon et al.*, 2010b). However, their terminal morphology (i.e., lack of debris aprons and instead ending abruptly or with pits) and the observation that some linear gully troughs and pits were forming in the present day (*Dundas et al.*, 2012; *Reiss et al.*, 2010) did not support this model.

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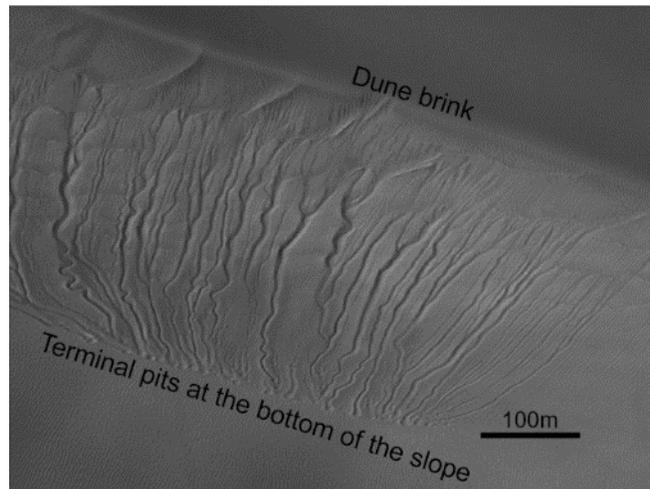
1145

1146

*Diniaga et al.* (2013) proposed a dry model, with the idea that these features may be similar to boulder-tracks—although lacking a boulder. According to this model, the trough forms due to a block of CO<sub>2</sub> ice rolling or sliding downslope, carving out its path on a sandy surface. Such ice forms within the seasonal frost layer that is observed to be deposited across these mid-latitude slopes each martian winter. As the frost on the dune slopes sublimates, ice remains cold trapped in shaded dune alcoves at the top of the dunes, but eventually may become dislodged, falling onto relatively warm, exposed dark sand. Sublimation at the base of this dry ice block lifts the block

1147 slightly from the sandy surface (in a manner similar to the Leidenfrost effect), allowing it to  
1148 freely roll or slide down the sandy slope, unencumbered by friction and carving out a trough.  
1149 Upon stopping, the dry ice block would continue to sublime in situ, digging out a pit that then  
1150 would be the remaining record after the block disappears. Field experiments with dry ice slabs  
1151 slid down terrestrial desert dune slopes (*Bourke et al.*, 2016a; 2016b; *Diniega et al.*, 2013) and  
1152 laboratory experiments that examine interactions between sublimating dry ice blocks and a  
1153 granular substrate (*McKeown et al.*, 2017) have shown that it is feasible for the “hovercrafting  
1154 dry ice block” model to broadly produce many of the observed linear gully morphologies. CO<sub>2</sub>  
1155 ice blocks, up to ~3-m diameter, have also been observed to form and migrate downslope within  
1156 martian features (e.g., *Dundas et al.*, 2012), and further modeling of interactions between CO<sub>2</sub>  
1157 ice and sediment support development of this process in the present martian climate (e.g.,  
1158 *Pilorget and Forget*, 2016). However, a refined model of block transport and quantitative  
1159 understanding of how linear gully morphological characteristics, such as width and sinuosity,  
1160 relates to formation history has not yet been developed, limiting interpretation of these features.  
1161 In particular, it is not yet known if the ~10 m-wide troughs seen in Russell crater formed under a  
1162 past climate when significantly larger blocks of CO<sub>2</sub> may have formed, or if these have widened  
1163 (albeit slowly) in the present climate due to ice blocks sliding down over many martian winters,  
1164 with blocks similar to those forming new ~meter-wide troughs (*Dundas et al.*, 2012; *Jouannic et*  
1165 *al.*, 2019; *Reiss et al.*, 2010).

1166  
1167 **Figure 11.** Example of a linear gully cluster  
1168 on a climbing dune slope (50.2°S, 292.1°E),  
1169 along the inside of the rim of an unnamed  
1170 crater. Note the range of sinuosities, trough  
1171 widths (1-10 m including levees), and pits  
1172 (2-5 m) even in this one cluster. Image ID:  
1173 HiRISE ESP\_030624\_1295, north is up and  
1174 illumination is from the left.



### 1175 **3.3.2 Araneiforms**

1176 Araneiforms (also known as “spiders”)  
1177  
1178 are unique surface features that have no Earth analogs. Located primarily on the south polar  
1179 layered deposits and surroundings (*Piqueux et al.*, 2003; *Schwamb et al.*, 2018), these features  
1180 are characterized by dendritic, tortuous troughs several meters wide and deep which extend from  
1181 a central pit and range from <50 m to 1 km in diameter (Figure 12d). Their specific morphology

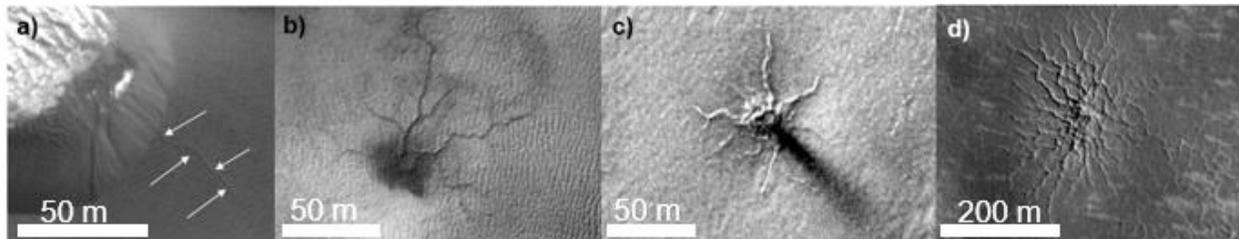
1182 types range from 'fat' to 'starburst' (*Hansen et al.*, 2010) and these sub-types tend to cluster non-  
1183 randomly (*Hao et al.*, 2020).

1184 These features are widely accepted to form via basal sublimation of CO<sub>2</sub> slab ice due to the  
1185 solid state greenhouse effect (*Matson and Brown*, 1989) (see §3.1 for a longer description). As a  
1186 consequence of this phenomenon, informally called the “Kieffer model,” sublimation at the base  
1187 of the CO<sub>2</sub> ice leads to a buildup of gas pressure beneath the ice overburden. Eventually this gas  
1188 pressure exceeds the strength of the ice, causing it to crack or rupture at a weak spot. Pressurized  
1189 gas rushes towards the vent, emerging as a plume and depositing the entrained material as fans  
1190 and spots (Figure 12c). The escaping gas entrains particulates from the substrate, gradually  
1191 eroding troughs (*Kieffer*, 2007; *Kieffer et al.*, 2006; *Piqueux et al.*, 2003).

1192 Repeated venting episodes are believed to build the full extent of araneiforms over thousands  
1193 of martian years (*Hansen et al.*, 2010; *Piqueux and Christensen*, 2008; *Portyankina et al.*, 2010;  
1194 *N. Thomas et al.*, 2010). Most araneiform terrains show annual repeating sublimation activity: in  
1195 spring, dark fans and blotches drape over troughs of araneiforms indicating CO<sub>2</sub> jet activity. As  
1196 the ice layer continues to sublime, the bright frost is removed and so the dark deposits fade or  
1197 even completely disappear, and the cycle repeats again in the next spring. However, despite  
1198 continuous monitoring of araneiform terrains by high-resolution remote sensing, no detection of  
1199 changes in the topography of large, well-developed araneiforms has been reported over the last 6  
1200 martian years. This leads to the question about whether the large araneiforms are currently  
1201 evolving with a slow erosive process by the sub-ice CO<sub>2</sub> gas flow that modifies the substrate at a  
1202 rate below current detection limits, or are dormant remnants of some past climate. It also  
1203 highlights uncertainty in existing estimates of araneiform ages: *Piqueux and Christensen* (2003)  
1204 estimated that they are at least 10<sup>4</sup> martian years old based on an erosion-rate estimate of ~1  
1205 m<sup>3</sup>/yr, but *Portyankina et al.* (2017) observed erosion rates of ~8 m<sup>3</sup>/yr.

1206 Away from old(er) araneiform terrains, newly forming dendritic troughs have been recently  
1207 detected with HiRISE (*Portyankina et al.*, 2017). These troughs form in the vicinity of sand  
1208 dunes and have been observed to grow interannually (Figure 12b). Dendritic troughs are  
1209 proposed to represent the early stages of araneiform formation (*Portyankina et al.*, 2017). Even  
1210 smaller-scale (~tens of meters long) dendritic features known as sand furrows annually scour  
1211 northern hemisphere dune slopes, but these are erased in summer (*Bourke and Cranford*, 2011;  
1212 *Diniega et al.*, 2019a; Figure 12a). While laboratory experiments have replicated dendritic

1213 patterns on granular substrate via CO<sub>2</sub> sublimation (*Mc Keown et al.*, 2021), the factors that  
1214 distinguish the apparent disparity in activity, scale, latitudinal distribution and morphology  
1215 between sand furrows, dendritic troughs and araneiforms have not yet been delineated.  
1216



1217  
1218 Figure 12. Examples of dendritic troughs, increasing towards right in network-complexity. The  
1219 first 2 show annually active features: (a) dune furrows (extending towards bottom right corner  
1220 from the dune brink, between arrows) and (b) the dendritic features described in *Portyankina et al.*  
1221 *et al.* (2017). The furrows disappear before the following year, while the dendritic trough has  
1222 grown through multiple martian years. The two on the right (c, d) are araneiforms, which have  
1223 not yet been observed to change. HiRISE images are from (a) *Diniega et al.* [2018] and (b-d)  
1224 *Portyankina et al.* [2017]: ESP\_017895\_2650, ESP\_011842\_0980, ESP\_023600\_1095, and  
1225 ESP\_032009\_0985, respectively. (NASA/JPL/UA).  
1226

### 1227 3.4 Open questions for seasonal frost/ice and related landforms

1228 For many of the landforms discussed above, observations for the timing and locations of  
1229 activity implicate some form of seasonal frost/ice as a driver, and it is generally thought that the  
1230 energy generated through frost/ice sublimation are a key control. However, models of frost  
1231 formation and sublimation have not yet been quantitatively connected to sediment fluxes or  
1232 erosion rates, and thus it is not known exactly what form(s) and amount of frost/ice may be  
1233 needed to induce landform creation, evolution, or modification. Terrestrial analog studies are  
1234 often used to provide a starting model (with an analog chosen based on similar geomorphology),  
1235 but recent studies demonstrating that CO<sub>2</sub> frost/ice is a major geomorphic agent suggests that  
1236 there are limits in how far Earth-based (often liquid water controlled) models can be applied. The  
1237 sublimation-dominant dynamics of CO<sub>2</sub> frost and ice, and even H<sub>2</sub>O when exposed under Mars  
1238 pressure conditions (e.g., *Herny et al.*, 2019; *Massé et al.*, 2016; *Raack et al.*, 2017), have no  
1239 terrestrial analog.

1240 Generation of landform evolution models are hampered by lack of knowledge about the  
1241 behavior and properties of the martian seasonal frost/ice layer, including how this layer evolves  
1242 through the winter. Passively sensing orbital instruments generally cannot observe during the  
1243 period of interest due to polar night; additionally, as the surface first reaches CO<sub>2</sub> condensation

1244 temperatures in the autumn, an atmospheric haze (i.e., the polar hood) obscures visible images.  
1245 No in situ measurements of the seasonal CO<sub>2</sub> frost layer have yet been collected due to technical  
1246 challenges in having a spacecraft survive through the winter (*ICE-SAG*, 2019). Laboratory  
1247 experiments have begun to look at CO<sub>2</sub> frost/ice formation (e.g., *Portyankina et al.*, 2019), as  
1248 well as at how both H<sub>2</sub>O and CO<sub>2</sub> sublimation may interact with granular materials (e.g.,  
1249 *Chinnery et al.*, 2018; *Herny et al.*, 2019; *Kaufmann and Hagermann*, 2017; *Massé et al.*, 2016;  
1250 *Mc Keown et al.*, 2017; 2021; *Pommerol et al.*, 2019; *Portyankina et al.*, 2019; *Raack et al.*,  
1251 2017; *Sylvest et al.*, 2016; 2019; *Yoldi et al.*, 2021). By necessity due to present lab capabilities,  
1252 such experiments are small-scale and simplified in terms of the variables incorporated.  
1253 Eventually such experiments will also need to consider the interactions between varying amounts  
1254 of different types of frost (e.g., fine-grained CO<sub>2</sub> snowfall over or under a layer of surface frost,  
1255 or how H<sub>2</sub>O and CO<sub>2</sub> surface frost may interlayer and affect optical and mechanical properties of  
1256 the full frost layer). Models are also needed to scale laboratory results to natural martian  
1257 conditions and to extrapolate to past environmental conditions, as well as observation of present  
1258 Mars surface and atmospheric conditions to constrain and refine such models.

1259 Finally, martian landforms with similar morphologies are often studied in aggregate, with  
1260 active examples treated as analogs for similar-appearing landforms not (yet) observed to be  
1261 active. In such studies, a common question is if much larger and/or more complex features that  
1262 have not yet been observed to be active are active at very slow rates or if they are instead records  
1263 of a past, more intense frost environment. Older records may evolve, possibly via a process  
1264 continuing through shifting climate conditions, or via process(es) different from that involved in  
1265 initial formation. Additionally, just as in application of comparative geomorphology between the  
1266 Earth and Mars, it is possible that similar appearing morphologies on Mars may form through  
1267 different processes (i.e., the principle of equifinality).

1268

#### 1269 **4 Long-term sublimation of ices**

1270 In the present martian climate, water ice is found in the polar caps (§4.1) and in the  
1271 subsurface (§4.2–4). For all of this ice, connection to the atmosphere allows sublimation of the  
1272 ice and specific geomorphologies have been tied to this volatile transport (§4.1, 4.3–4). If the  
1273 rate of sublimation can be determined from the observed activity or geomorphology, the absolute  
1274 ages and stability of the ice deposits can be estimated—yielding environmental constraints on  
1275 recent past climates, including variations in ice formation and stability under different obliquities

1276 (see §4.2.1). Alternatively, if the ages and sizes of past ice reservoirs can be estimated, this can  
1277 yield bounds on the rate at which water ice is subliming and escaping through the regolith.

1278 Interpreting the geological and climatological history reflected in subsurface water ice  
1279 deposits requires comparison between two types of analysis: model predictions of water ice  
1280 stability (as a function of depth, latitude, and subsurface thermophysical properties: §4.2.1) and  
1281 observational evidence of where water ice is or was present (§4.2.2, 4.3–4). When these two  
1282 lines of investigation are consistent, this provides credence to the applied models and  
1283 environmental parameters. When these two lines of investigation differ, this leads to either  
1284 focused questions about the models and/or assumed environmental parameters or constraints on  
1285 the age of geomorphic features and preservation mechanisms (i.e., if ice may have been present  
1286 in the past, but no longer exists).

1287

## 1288 **4.1 Polar surface landforms**

### 1289 **4.1.1 South Polar Residual Cap**

1290 The South Polar Residual Cap (SPRC) is a <10-m-thick layer of CO<sub>2</sub> ice overlying a layer of  
1291 H<sub>2</sub>O ice (*Bibring et al.*, 2004; *Byrne and Ingersoll*, 2003b; *Titus et al.*, 2003) that covers  $7.9 \times$   
1292  $10^9$  m<sup>2</sup> (*P.C. Thomas et al.*, 2016) offset slightly west from the south pole, potentially due to  
1293 broad-scale topography modulation of south polar circulation (*Colaprete et al.*, 2005). The CO<sub>2</sub>  
1294 ice is incised into discrete mesas by ubiquitous sublimation pits that annually enlarge in diameter  
1295 by meters per year (dubbed “swiss cheese terrain”: *Byrne and Ingersoll*, 2003b; *Malin et al.*,  
1296 2001; *P.C. Thomas et al.*, 2005) and manifests in a spectacular variety of planform shapes  
1297 (Figure 13; *P.C. Thomas et al.*, 2016). This annual pit enlargement led to the initial hypothesis  
1298 that the SPRC is only a few hundred years old, perhaps indicating that its existence is evidence  
1299 of recent climate change (*Byrne and Ingersoll*, 2003a; *Malin et al.*, 2001).

1300 Continued cataloguing and documentation of the SPRC’s diverse landforms (*P.C. Thomas et*  
1301 *al.*, 2005; 2009; 2013) culminated in a comprehensive map of its morphology and refined  
1302 estimates of its mass balance (*P.C. Thomas et al.*, 2016), including the revelation that some  
1303 regions show evidence of net local and regional accumulation over the recent past (*Buhler et al.*,  
1304 2017; *P.C. Thomas et al.*, 2016). Importantly, observations indicate that widespread net annual  
1305 vertical accumulation may offset local horizontal pit wall ablation, leading to net mass  
1306 equilibrium with a complete turnover in material every ~100 martian years (*P.C. Thomas et al.*,  
1307 2016). Mass equilibrium is consistent with the observation that Mars’ mean annual pressure is

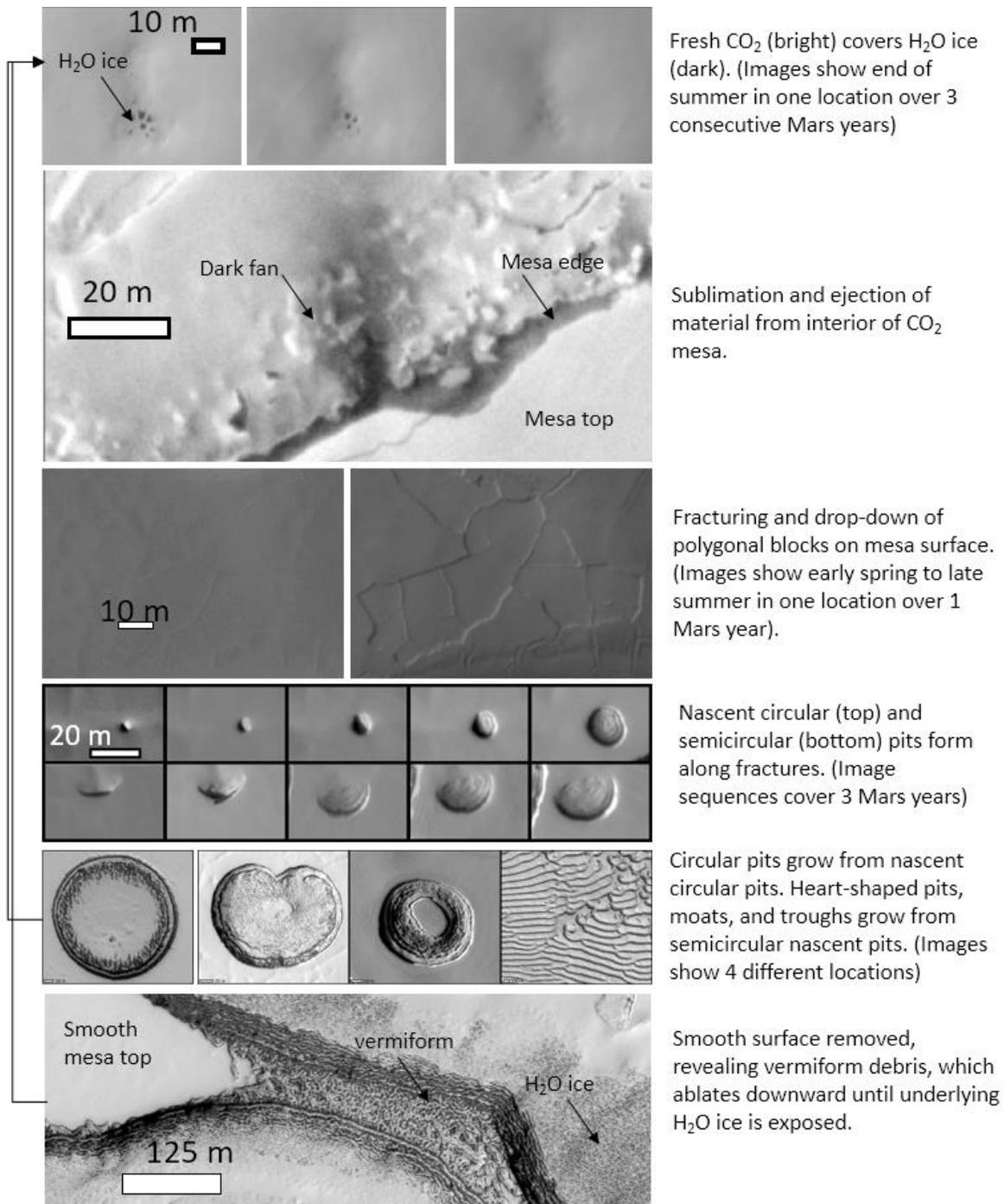
1308 the same to within ~10 Pa between *Viking* lander measurements and the present day (*Haberle et*  
1309 *al.*, 2014) and SPRC landform modeling (*Byrne et al.*, 2015). Further refinement of the SPRC's  
1310 annual mass balance will require observations with higher vertical accuracy and a longer  
1311 baseline (*Buhler et al.*, 2018; *P.C. Thomas et al.*, 2016).

1312 There are four broad categories of SPRC pit morphologies: circular and heart-shaped pits,  
1313 linear troughs, and moats (Figure 13; *P.C. Thomas et al.*, 2016). Thus far, quantitative numerical  
1314 morphological modeling can only produce circular pits (*Byrne et al.*, 2015). However, a  
1315 conceptual model based upon observation indicates that all four main types of morphologies  
1316 develop via the interplay of wintertime accumulation and summertime ablation (*Buhler et al.*,  
1317 2017). Aeolian reworking (*P.C. Thomas et al.*, 2020) and dust storms (*Becerra et al.*, 2015;  
1318 *Buhler et al.*, 2017) may also influence morphologic development and mass balance. Future  
1319 maturation of numerical landform models will be essential for quantifying the mass balance of  
1320 the SPRC under orbital (i.e., polar insolation) conditions different from the modern day.

1321 In the conceptual model of morphologic development (Figure 13; *Buhler et al.*, 2017),  
1322 summertime sunlight causes internal sublimation of the SPRC, leading to the collapse of its  
1323 surface, creating fractures. The roughness caused by fracturing leads to enhanced local  
1324 sublimation, forming nascent pits. Two types of nascent pits form: circular, where a fracture  
1325 widens uniformly at a point, and semicircular, where one side of the fracture falls lower, forming  
1326 a steep scarp and a smooth ramp. The circular pits grow larger and stay circular. The  
1327 semicircular pits grow into either heart-shaped pits or linear troughs with scrolled edges. The  
1328 intersection of growing scarps and slopes can create geometries where moats form.

1329 Ablating pit walls typically leave behind an extended debris ramp of blocky, vermiform  
1330 material with a lower albedo relative to the smooth-topped CO<sub>2</sub> mesas (*P.C. Thomas et al.*,  
1331 2020). In some regions where the surface of the CO<sub>2</sub> ice reaches a critical roughness, the  
1332 morphology degrades into extensive (>1 km diameter) vermiform debris fields that ablate over  
1333 the course of typically tens of years until the underlying H<sub>2</sub>O ice is exposed (*P.C. Thomas et al.*,  
1334 2020). Within a few years, fresh seasonal CO<sub>2</sub> ice survives the summer where the H<sub>2</sub>O ice was  
1335 exposed, restarting the growth of a new, smooth-topped perennial CO<sub>2</sub> ice mesa.

1336



1337  
 1338 Figure 13. Cycle of CO<sub>2</sub> deposition and ablation in the South Polar Residual Cap (SPRC).  
 1339 Seasonal CO<sub>2</sub> deposits on exposed H<sub>2</sub>O ice, survives summer, accumulates year-over-year to  
 1340 become perennial SPRC CO<sub>2</sub>. Sunlight penetrates and heats CO<sub>2</sub> within mesa, causing material  
 1341 loss and fracturing of the mesa surface. Nascent pits form along fractures. Pits that are initially  
 1342 circular in shape grow into larger circular pits. Pits that are initially semicircular in shape  
 1343 develop into heart-shaped pits, moats, or troughs, depending on local sublimation and  
 1344 accumulation conditions. Horizontal pit ablation exposes underlying H<sub>2</sub>O ice. Where the mesa  
 1345 surface is sufficiently damaged, vermiform terrain develops, which ablates downward until H<sub>2</sub>O  
 1346 ice is exposed. Then the cycle repeats. Adapted from *Buhler et al. (2017)*.

1347

#### 1348 **4.1.2 Massive CO<sub>2</sub> Ice Deposit and its capping H<sub>2</sub>O ice layer**

1349 Exposed beneath the SPRC and covering the recently discovered Massive CO<sub>2</sub> Ice Deposit  
1350 (MCID; *Phillips et al.*, 2011; *Putzig et al.*, 2018) is a <20-m-thick layer of H<sub>2</sub>O ice (Figure 14c).  
1351 This H<sub>2</sub>O ice layer hosts 1- to 100-km-scale circular, scalloped, and trough depressions with  
1352 depths up to ~100 m, which likely form due to sublimation and collapse of the underlying MCID  
1353 (Figure 14; *Phillips et al.*, 2011). Viscous flow of the MCID likely also shapes the topography of  
1354 the H<sub>2</sub>O ice layer (*Cross et al.*, 2020; *I.B. Smith et al.*, 2016).

1355 The H<sub>2</sub>O ice layer is heavily fractured, which may derive from volumetric collapse due to  
1356 sublimation of the underlying MCID (Figure 14; *Buhler et al.*, 2020; *Phillips et al.*, 2011) or  
1357 thermal expansion (*Bierson et al.*, 2016). Material exchange between the MCID and the  
1358 atmosphere through this H<sub>2</sub>O ice layer likely keeps the SPRC in mass balance over obliquity  
1359 cycles (*Buhler et al.*, 2020). Based on models, the morphology of the H<sub>2</sub>O ice layer may be  
1360 changing at ~1 mm/yr rates (*Buhler et al.*, 2020; *Jakosky et al.*, 1990) and hold important clues  
1361 to whether this H<sub>2</sub>O ice layer is permeable to CO<sub>2</sub> gas. Because the permeability of the H<sub>2</sub>O ice  
1362 layer is debated (*Manning et al.*, 2019), continued study of the morphology and thermal behavior  
1363 of this H<sub>2</sub>O ice layer is important for understanding the long-term (>10<sup>4</sup> yr) behavior of Mars'  
1364 global atmospheric pressure and climate.

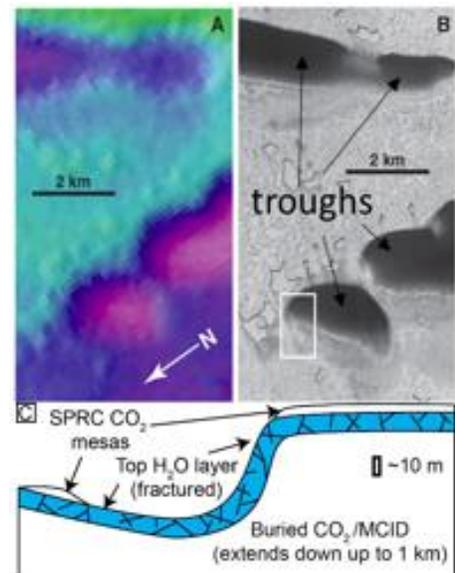
1365

1366 Figure 14. Depressions in the H<sub>2</sub>O ice layer beneath the  
1367 SPRC near 87° S, 268° E. (A) MOLA topography, ~75 m  
1368 elevation range from pink (low) to green (high). (B)  
1369 Context Camera image of the same location. H<sub>2</sub>O ice  
1370 (dark) is exposed in the troughs through windows in SPRC  
1371 CO<sub>2</sub> ice mesas (bright). (C) Schematic cross section of  
1372 SPRC, H<sub>2</sub>O ice layer, and the MCID, illustrating proposed  
1373 layering. A and B modified from *Phillips et al.* (2011).

1374

#### 1375 **4.1.3 North Polar hummocky H<sub>2</sub>O ice surface**

1376 The surface of the perennial North Polar Residual Cap  
1377 (NPRC) is primarily H<sub>2</sub>O ice, as opposed to the CO<sub>2</sub> ice  
1378 deposits in the south (*P.C. Thomas et al.*, 2000). The H<sub>2</sub>O  
1379 ice has a rough, hummocky texture of ~10 m-scale semi-regular depressions and mounds  
1380 (*Nguyen et al.*, 2020; *Parra et al.*, 2017; *Russell et al.*, 2019). Although observations of changes  
1381 within the polar region indicates there are seasonal and interannual periods and locations of both



1382 net deposition and ablation (e.g., *Brown et al.*, 2016; *Calvin et al.*, 2015), other observations  
1383 have been proposed to indicate that the current NPRC surface is underdoing net, long-term  
1384 ablation. For example, large-grained ice dominates the NPRC at the end of northern summer,  
1385 possibly implying exhumation of older, sintered ice (*Langevin et al.*, 2005); however, the ice also  
1386 has a low dust content, indicating a lack of the dust lag that might be expected if the ice were  
1387 ablating (*Langevin et al.*, 2005). Another example is that *Milkovich et al.* (2012) observed that  
1388 the wavelength of the hummocky texture has a positive correlation with elevation and latitude,  
1389 which they interpreted as indicating the formation of the hummocky texture via ablation.  
1390 However, modeling by *Wilcoski and Hayne* (2020) indicates that hummocks would form in both  
1391 ablational and depositional settings and, further, that hummock wavelength correlates primarily  
1392 with age regardless of net ablation or deposition; they find that the typically observed ~10 m  
1393 hummock wavelengths are reached after ~1 kyr. This timescale is consistent with ~1.5 kyr ages  
1394 derived from cratering statistics (*Landis et al.*, 2016). Ablation and deposition of H<sub>2</sub>O ice also  
1395 modify craters on the NPRC, with the current crater population being estimated to have  
1396 accumulated within the last ~20 kyr and ice accumulation rates within craters of ~3–4 mm/yr  
1397 (*Banks et al.*, 2010).

1398 Wind is also likely a driver for evolving geomorphology of icy features on the NPRC  
1399 because it can influence volatile fluxes at the surface, vapor transport, and distribution of both  
1400 CO<sub>2</sub> and H<sub>2</sub>O ice. However, most of these processes driven by both wind and sublimation are not  
1401 directly observable due to very slow rates. Large-scale features like the chasmae/spiral troughs in  
1402 the NPRC are thought to have formed into their present state over millions of years of erosion by  
1403 katabatic winds and asymmetric insolation/sublimation (e.g., *Bramson et al.*, 2019; *Howard*,  
1404 2000; *Smith and Holt*, 2010). Smaller-scale periodicities in the landscape may be related to  
1405 sublimation dynamics of perennial ice layers interacting with the winds over thousands of years  
1406 (e.g., *Bordiec et al.*, 2020; *Herny et al.*, 2014; *Howard*, 2000; *Nguyen et al.*, 2020).

1407 Because the NPRC is the uppermost layer of the NPLD (*Tanaka et al.*, 2005), further study of  
1408 NPRC surface morphology evolution will be important for understanding how layers  
1409 accumulated in the NPLD, as well as how structure of the layers may record aeolian and  
1410 sublimation interactions when that layer of material was exposed on the surface, and how to  
1411 interpret the climate under which those layers formed (*I.B. Smith et al.*, 2020). The layers within  
1412 the NPLD (and in the Southern Polar Layered Deposits (SPLD)) are of high interest for Mars

1413 polar and climate studies because they are thought to be analogous to the layers found within  
1414 terrestrial ice cores and thus record martian climate cycling (*I.B. Smith et al.*, 2020).

## 1415 1416 **4.2 Present/recent subsurface water ice**

### 1417 **4.2.1 Present-day water ice stability**

1418 The fundamental principles controlling subsurface H<sub>2</sub>O ice stability have been understood for  
1419 some time (e.g., *Smoluchowski*, 1968). Given some water vapor content in the atmosphere, ice  
1420 will be deposited in locations that are below the frost point temperature and will sublime at  
1421 locations that are warmer. Integrated over the course of a martian year, ice is stable at locations  
1422 where the average water vapor pressure over ice is less than or equal to that in the atmosphere;  
1423 these locations will also experience net deposition. At equatorial latitudes under current  
1424 conditions, temperatures are too warm for ice to be stable, and ice does not accumulate. At  
1425 middle to high latitudes, peak surface temperatures may still be high, but annual and seasonal  
1426 variations are damped in the subsurface. The equilibrium water vapor pressure is nonlinearly  
1427 dependent on temperature, which allows ice to become stable in the shallow subsurface, at a  
1428 depth that becomes shallower with increasing latitude.

1429 The development of increasingly sophisticated maps of the distribution of stable ice, based  
1430 on the above framework, is summarized by *Mellon et al.* (2004). Recent ice stability maps have  
1431 also been produced by *Chamberlain and Boynton* (2007), *Schorghofer and Aharonson* (2005),  
1432 and *Steele et al.* (2017). Broadly, these all place the present-day stability boundary near 45–60°  
1433 latitude, and show similar longitudinal variations correlated with surface albedo and thermal  
1434 inertia. Differences in ice stability predictions are largely due to assumptions of near-surface  
1435 atmospheric water vapor content and past atmospheric conditions.

1436 Mars' orbit varies over time, leading to H<sub>2</sub>O ice sublimation and deposition as the global ice  
1437 stability field evolves in response. *Dundas et al.* (2014) found that ice newly exposed by fresh  
1438 impact craters persisted for longer than expected, given the currently measured atmospheric  
1439 water vapor column. Their observations indicated that either the lowest-latitude ice has not  
1440 equilibrated with the current climate or that the ice is stabilized by local factors, such as  
1441 enhanced near-surface water vapor concentration. Remnant out-of-equilibrium ice that was  
1442 deposited recently (within ~1 Ma) provides one possible explanation for the low-latitude icy  
1443 craters (*Schorghofer and Forget*, 2012). Similar modeling by *Bramson et al.* (2017) concluded  
1444 that such ice could be considerably older (>10s Ma) than estimated by *Schorghofer and Forget*

1445 (2012) if the ice started sufficiently thick and has also been protected by thick lag deposits built  
1446 up by dust and lithic debris released from the sublimating ice.

1447 The current solutions for Mars' orbital variations (*Laskar et al.*, 2004) indicate that ~4 Ma,  
1448 Mars entered a low obliquity epoch in which mid-latitude ice has generally been less stable and  
1449 polar ice has been more stable. Correspondingly, modeling and observations indicate net  
1450 transport of ice from the middle latitudes to the poles over this timeframe (e.g., *Levrard et al.*,  
1451 2007; *I.B. Smith et al.*, 2016). However, higher frequency (~50 to 100s of kyr) periodicities in  
1452 Mars' orbital parameters (particularly obliquity) likely generate many excursions away from the  
1453 long-term average conditions. Therefore, the spatial distribution of ice deposition and  
1454 sublimation has likely been very dynamic over ~50 to 100 kyr timescales. Understanding this  
1455 dynamic movement of ice is essential to deciphering the evolution of icy terrains. The modeling  
1456 described above shows that the current distribution and stability state of mid-latitude ice could  
1457 place important constraints on Mars' recent past climate. Such model predictions can be coupled  
1458 with observations of landforms (such as sublimation thermokarst, patterned ground, and viscous  
1459 flow features) to improve interpretation of such landforms or to test the climate condition  
1460 assumptions.

1461

#### 1462 **4.2.2 Present-day water ice distribution**

1463 Presently, exposed water ice is only stable on the surface at the poles; it is stable at lower  
1464 latitudes when buried in the subsurface under an insulating, desiccated coating of dust and/or  
1465 regolith. Data from the Mars Odyssey Neutron Spectrometer show the near subsurface (within  
1466 the upper meter) at middle and high latitudes to be hydrogen rich, which has been attributed to  
1467 the presence of water ice (*Boynton et al.*, 2002; *Pathare et al.*, 2018). As discussed in §4.3–4.4,  
1468 numerous geomorphological features, including glacial and viscous flow features, thermal  
1469 contraction polygons, and ice-loss (also referred to as thermokarstic) terrains, suggest a recent  
1470 and/or present-day ice-rich subsurface across most of the mid-latitude plains. Thermal analysis  
1471 indicates widespread water ice at latitudes as low as 35°N/45°S, with high lateral ice depth  
1472 variability—sometimes buried only a few centimeters below sand-like material—and correlated  
1473 with putatively periglacial features (*Piqueux et al.*, 2019).

1474 Due to warm temperatures at the equator, in general, the ice must be buried at greater depth  
1475 for increasingly-equatorward locations (e.g., *Fanale et al.*, 1986; *Leighton and Murray*, 1966;

1476 *Paige*, 1992). For example, the Phoenix lander excavated nearly pure water ice in the upper  
1477 centimeters of the surface at 68°N (*P.H. Smith et al.*, 2009). Recently-discovered scarps near  
1478 ~55° latitude in both the northern and the southern hemispheres expose thick, massive ice that  
1479 appears to extend to within a meter of the surface in high-resolution images (*Dundas et al.*,  
1480 2018). These slopes can be several kilometers long and over 100 m tall. Bare ice at these  
1481 locations likely are actively subliming; *Dundas et al.* (2018) observed boulders falling from one  
1482 scarp and estimated that the sublimation rate was on the order of millimeters per year.  
1483 Additionally, H<sub>2</sub>O ice spectral features at a second scarp weakened over the course of the  
1484 summer, suggesting the gradual accumulation of a thin sublimation lag of dust. This ice loss  
1485 results in ongoing slope retreat and the growth of depressions, which are morphologically  
1486 distinct from the thermokarst features discussed below (§4.3), likely because a bare ice surface is  
1487 maintained in these landforms (*Dundas et al.*, 2018).

1488       Recent (<15 years old) impact craters have exposed and excavated nearly-pure water ice  
1489 (likely >90% ice by volume) within a meter of the surface as close to the equator as 39°N (*Byrne*  
1490 *et al.*, 2009; *Dundas et al.*, 2014). The appearance of these exposures slowly fades over time as  
1491 the exposed ice sublimates (*Dundas and Byrne*, 2010). Some of the ice remains distinctive in color  
1492 for several martian years (*Dundas et al.*, 2014), indicating clean ice (>90% water ice by volume)  
1493 with a low lithic content.

1494       Ice is not expected to be stable near the equator at any depth, though the most equatorward  
1495 boundary of subsurface mid-latitude ice is still an outstanding question and unstable, sublimating  
1496 ice may exist. Recent efforts have focused on integrating numerous datasets and techniques to  
1497 constrain the distribution of mid-latitude ice, especially from the perspective of its utilization as  
1498 an in situ resource for crewed missions, and include studies across swaths of the northern plains  
1499 regions (*Orgel et al.*, 2019; *Ramsdale et al.*, 2019; *Sejourne et al.*, 2019) as well The Mars  
1500 SWIM (Subsurface Water Ice Mapping) Project ([swim.psi.edu](http://swim.psi.edu)). So far, the areas found to be  
1501 most consistent with abundant water ice occur poleward of ~40°N, in the northern plains region  
1502 of Arcadia Planitia, where widespread ground ice was previously inferred in radar sounding and  
1503 geomorphic crater studies (*Bramson et al.*, 2015; *Viola et al.*, 2015), and within an extensive  
1504 network of debris-covered glaciers in the Deuteronilus Mensae region (*Petersen et al.*, 2018).  
1505 This is generally more poleward than the regions focused on by human exploration planners, due  
1506 to other constraints on human access and operations (e.g., *ICE-WG*, 2015).

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### 4.3 Sublimation thermokarst

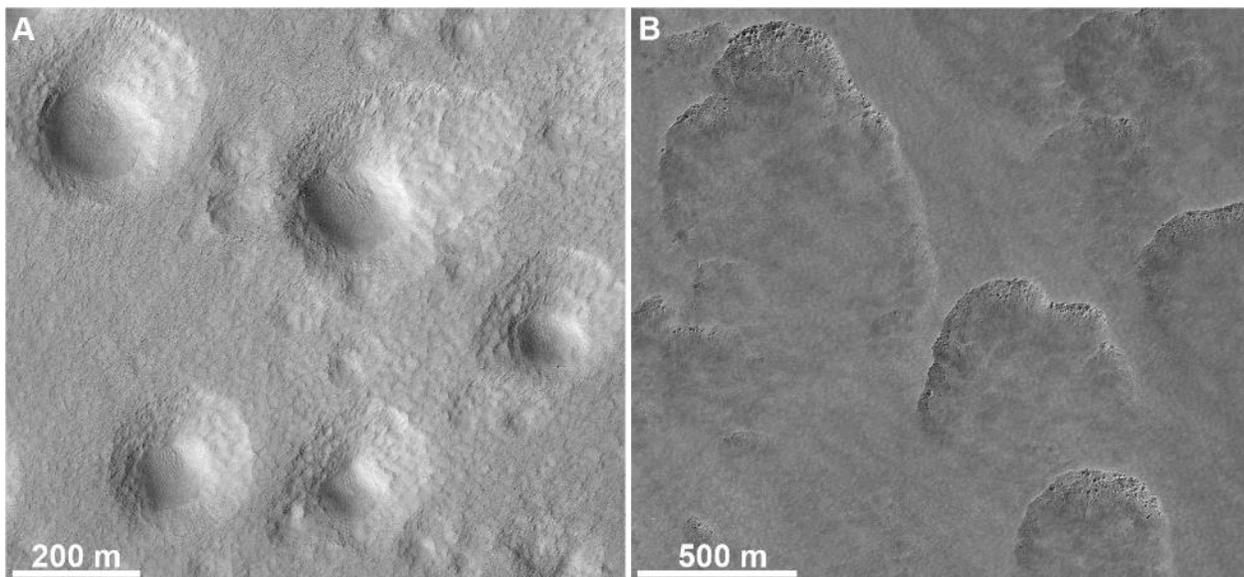
Candidate thermokarstic landscapes were identified in images dating back to the Mariner and Viking missions (*Anderson et al.*, 1973; *Costard and Kargel*, 1995; *Sharp*, 1973). These landforms result from surface collapse following loss of subsurface ice, leading to rimless depressions that are often hundreds of meters in size, and meters to tens of meters deep. Scalloped features (e.g., *Dundas et al.*, 2015b; *Lefort et al.*, 2009; 2010; *Morgenstern et al.*, 2007; *Séjourné et al.*, 2011; 2012; *Soare et al.*, 2007; 2008; 2011; *Ulrich et al.*, 2010; *Zanetti et al.*, 2010) and expanded craters (e.g., *Dundas et al.*, 2015b; *Viola et al.*, 2015; *Viola and McEwen*, 2018) are considered to be some of the most iconic examples of ice loss features (Figure 15). Initial interpretations included formation via melting akin to terrestrial thermokarst and alases (*Soare et al.*, 2007; 2008; 2011), but the present general consensus is that these are formed via sublimation.

Given the importance of temperature on ice stability, evolution of the terrain often proceeds through enhanced retreat of the warmer slopes. Under the present obliquity those are equator-facing. This leads to asymmetric landforms, with scalloped terrains often being elongated in the direction of retreat and exhibiting shallower slopes on the equatorward-facing slopes. It has been debated whether the scalloped depressions form primarily via retreat of the pole-facing slope at high obliquity (*Séjourné et al.*, 2011; *Ulrich et al.*, 2010) or retreat of the equator-facing slope under conditions similar to the present (*Lefort et al.*, 2009; *Morgenstern et al.*, 2007; *Zanetti et al.*, 2010). More complicated landscapes can form through the merging of the features, and the intervening terrain between collapse features is generally thought to retain the ice-rich subsurface unit.

Numerical landscape evolution modeling (*Dundas*, 2017; *Dundas et al.*, 2015b) shows that standard martian ice-stability theory (§4.1) can produce both scalloped-depression and expanded-crater morphologies via sublimation. The model is driven by surface topography and a high subsurface ice content, which produce uneven sublimation and an evolving landform. In the model, retreat of both pole- and equator-facing slopes occur, although the latter appears to be most important. The subsurface ice loss is triggered by some local disturbance (such as an impact, in the case of expanded craters). Modeling of the process suggests that these landforms may take  $10^4$ – $10^5$  years or more to form, though the development of a sublimation lag will eventually help preserve these landforms from additional ice loss (*Dundas et al.*, 2015b).

1539 These landforms likely develop gradually over tens to hundreds of thousands of years  
1540 (*Dundas, 2017; Dundas et al., 2015b*) since the surface debris slows sublimation and protects the  
1541 ice from high peak temperatures. As such, landform evolution is unlikely to be occurring at  
1542 scales that are observable from orbit, unless slow ice loss occasionally triggers larger mass  
1543 wasting events. However, given the likelihood of out-of-equilibrium ice (§4.1), it is likely that at  
1544 least some of these sublimation-thermokarst features are evolving at present. It is even possible  
1545 that this could occur in ice that is generally in equilibrium, since the process will work on any  
1546 slope that is locally out of equilibrium.

1547



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1549 **Figure 15.** (A) Expanded secondary craters on the northern plains. Note funnel shape suggesting  
1550 widening and shallowing of the rim. HiRISE image ESP\_045303\_2320. (B) Scalloped  
1551 depressions south of the Hellas basin (58.1° S, 74.0° E). Note the steep pole-facing slopes.  
1552 HiRISE image ESP\_049581\_1215. In both images, illumination is from the left and north is up.

1553

#### 1554 **4.4 Patterned Ground**

1555

1556 Polygonally patterned ground, containing polygons with a wide large range of diameters  
(meters to tens of kilometers), is one of the most common and, based on superposition, youngest  
1557 landforms on Mars. Polygonally patterned ground is either sorted (with surface patterns defined  
1558 by rock fragments) or unsorted landforms (with surface patterns defined by thermal contraction  
1559 cracks without sediment motion) (*French, 2007*). On Earth, sorted patterned ground is most  
1560 commonly found in fine-grained sediments overlain by coarser rock fragments because these are  
1561 the most susceptible to frost heave and sorting as a function of grain size during freeze-thaw  
1562 cycling (*Kessler and Werner, 2003*). In addition to the polygons, stripes, piles, and other

1563 morphologies found on Earth are also found on Mars, and with transitions between them  
1564 occurring in the same way as on Earth, lending support for the “convection” model that is freeze-  
1565 thaw sorting (e.g., *Gallagher and Balme, 2011; Gallagher et al., 2011; Soare et al., 2016*),  
1566 although such processes can also occur on Earth without freeze-thaw (e.g., *Sletten et al., 2003*).  
1567 The limit of HiRISE resolution (Table 1) necessitates that all clasts used to define sorted  
1568 patterned ground consists of boulder-sized or larger sediments. *Soare et al. (2019)* also used  
1569 locality with potential pingos in such feature identification.

1570 Non-sorted polygonally patterned ground (thermal contraction crack polygons) are  
1571 widespread on Mars and dominate surfaces poleward of ~30-40° latitude (*Levy et al., 2009b*;  
1572 *Mangold, 2005*) where ground ice is abundant (*Boynton et al., 2002*) and where thawed active  
1573 layers (portions of the soil column that seasonally freeze and thaw) have been rare to absent on  
1574 flat-lying surfaces over at least the past ~5 Ma (*Kreslavsky et al., 2008*). Thermal contraction  
1575 crack polygons on Mars are overwhelmingly high-centered features with low bounding troughs,  
1576 although examples of low-centered polygons with elevated shoulders occur in a few locations  
1577 (Figure 16) (*Soare et al., 2014; 2018*). The more common high-centered morphology indicates  
1578 that either excess ground ice has escaped via sublimation along the fracture traces, and/or that  
1579 infilling of fractures by fines is slow compared to subsurface ice loss. Alternatively, the center  
1580 might be deformed upwards by pressures created by aggrading sand wedges (e.g., *Sletten et al.,*  
1581 *2003*).

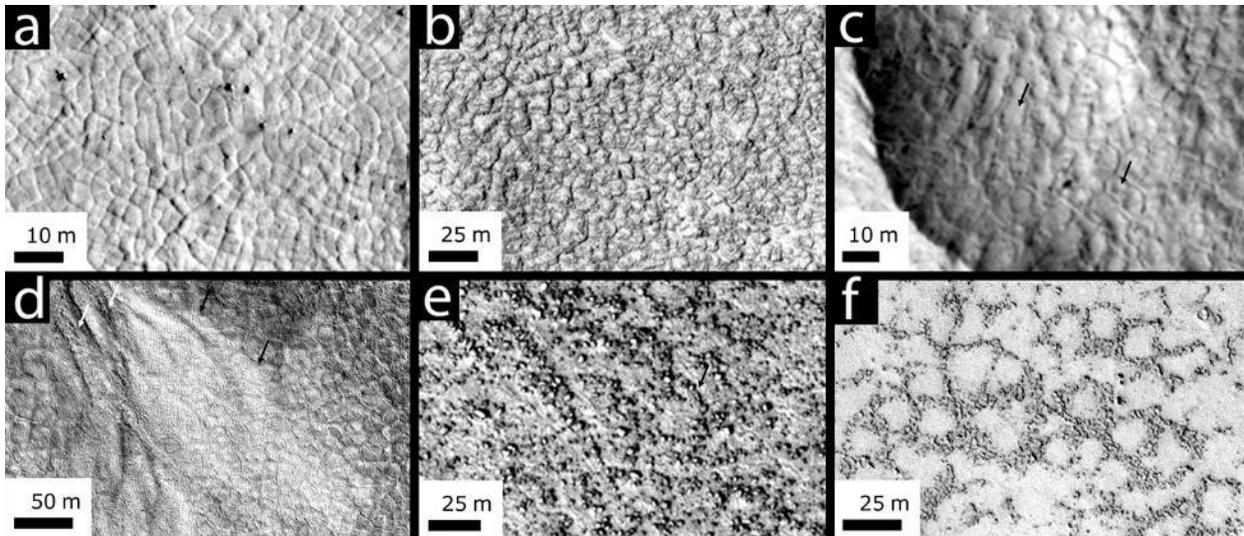
1582 Thermal contraction cracks can form under modern martian climate conditions (*Mellon,*  
1583 *1997*), suggesting that polygons are actively forming and expanding locations where ice-rich  
1584 permafrost and/or buried ice is present (*Mellon et al., 2008*). Continued formation and growth of  
1585 thermal contraction crack polygons is consistent with observations that polygon fracture  
1586 networks crosscut many young deposits on Mars, including gully fans (*Levy et al., 2010*) (Figure  
1587 16). In many locations, ice-rich mantling deposits (e.g., *Head et al., 2003*) are extensively  
1588 fractured, with polygon troughs cross-cutting almost all but the most recent impact craters (*Byrne*  
1589 *et al., 2009; Levy et al., 2010*), suggesting that mantling units may have crater retention ages of  
1590 10-100 kyr.

1591 While there is little debate about the origin and current activity of unsorted patterned ground  
1592 on Mars, the widespread existence and mechanism of formation for sorted patterned ground are  
1593 both topics of considerable debate, particularly as to whether freeze-thaw is necessary for its

1594 formation. Potential sorted patterned ground consists of three main groups: high-latitude  
1595 boulders concentrated in thermal contraction crack troughs (*Levy et al.*, 2009b; *Mellon et al.*,  
1596 2008; *Orloff et al.*, 2011), clasts arranged in boulder halos (*Barrett et al.*, 2017; *Levy et al.*,  
1597 2018), and low-latitude albedo networks (*Balme et al.*, 2009). Boulders may be sorted into  
1598 polygon troughs at high latitude via slumping of over-steepened trough shoulders (*Levy et al.*,  
1599 2010; *Mellon et al.*, 2008), seasonal frost-related locking and sliding mechanisms (*Orloff et al.*,  
1600 2013), or differential inflation of soil profiles at polygon troughs vs. centers (*Levy et al.*, 2018).  
1601 At middle latitudes where boulders are present in rock rings called boulder halos, boulders  
1602 commonly cluster in beaded networks (Figure 16), some of which are confined to polygon  
1603 troughs and some of which are not, leading (*Barrett et al.*, 2017) to interpret these sites as  
1604 possible evidence of freeze-thaw-driven sorting under near-recent climate conditions. Finally,  
1605 equatorial examples of potential clastic networks have been identified near Cerberus Fossae  
1606 (*Balme et al.*, 2009). These occur in the absence of thermal contraction crack polygons and have  
1607 been interpreted as evidence of relict freeze-thaw heaving mechanisms. A lack of meter-scale or  
1608 larger boulders in these deposits (Figure 16) makes these deposits less comparable to the sorted  
1609 clasts observed in the other two examples, and their proximity to volcanic deposits associated  
1610 with Cerberus Fossae outflow raises the possibility that they are features of volcanic origin.

1611 On Earth, sorted patterned ground can form on timescales of years to millennia (*Hallet*,  
1612 2013), while unsorted patterned ground typically matures over millennial to million-year  
1613 timescales (*Levy et al.*, 2006; *Marchant et al.*, 2002; *Sletten et al.*, 2003). Rates of thermal  
1614 contraction crack wedge expansion typically are on the order of millimeters per year, challenging  
1615 efforts to detect change in martian patterned ground. However, it is likely that patterned ground  
1616 formation and evolution—especially thermal contraction crack fracturing and wedge growth are  
1617 occurring on modern Mars and are actively working to resurface middle- and high-latitude  
1618 landscapes.

1619



1620 Figure 16. Polygonally patterned ground on Mars. (a) High-centered, unsorted patterned ground.  
 1621 Portion of HiRISE image PSP\_001474\_2520. (b) High-centered, unsorted patterned ground.  
 1622 Portion of HiRISE image PSP\_003217\_1355. (c) Example of low-center / raised-rim unsorted  
 1623 patterned ground. Arrows point to low-centered polygons. Portion of HiRISE image  
 1624 PSP\_002175\_2210. (d) Bright gully fan material (black arrows) emerging from a gully channel  
 1625 (white arrows) that has been crosscut by underlying thermal contraction cracks. Portion of  
 1626 HiRISE image PSP\_001846\_2390. (e) Candidate sorted patterned ground. Portion of HiRISE  
 1627 image ESP\_017580\_2460. (f) Candidate low-latitude, sorted patterned ground. Portion of  
 1628 HiRISE image PSP\_004072\_1845.  
 1629

1630

#### 1631 4.5 Open questions for long-term sublimation of ice

1632 A primary goal of Mars polar science investigations is to unlock the climate history stored in  
 1633 Mars' polar deposits (*I.B. Smith et al.*, 2020). Improved estimates of the modern annual polar  
 1634 CO<sub>2</sub> and H<sub>2</sub>O ice and nonvolatile mass balance, coupled with a better understanding of the  
 1635 annual reworking of the SPRC and NPRC surface, will be essential to interpreting the climatic  
 1636 conditions encoded in layers of the NPLD and SPLD. Improved observational resolution,  
 1637 cadence, and baseline of observations as well as development of physics-based numerical  
 1638 simulations that are capable of reproducing morphologic observations of polar landforms will  
 1639 both be critical to this endeavor. Additionally, the net annual mass flux of polar material remains  
 1640 poorly constrained. Improved measurement of atmospheric transport of volatiles and dust as well  
 1641 as observation and modeling of surface morphology (including its evolution), seasonal frost layer  
 1642 evolution (§3.4), and thermal cycles of the surfaces of the SPRC and NPRC will greatly enhance  
 1643 our understanding of the current net annual polar mass balance.

1644 Outside of the polar region, a global map of where ice is presently found in the subsurface  
 1645 (and at what depth) along with the structure, volume, and purity of that ice is needed. This is of

1646 particular interest for understanding where ice is currently aggrading or sublimating. Improved  
1647 knowledge of the near-surface water vapor concentration is also needed to understand the  
1648 equilibrium distribution of ice (i.e., where it “should” be). Phoenix measurements suggest a  
1649 stronger atmosphere-regolith interchange in the martian arctic than at lower latitudes (*Fischer et*  
1650 *al.*, 2019) and orbital column measurements are challenged by possible near-surface  
1651 concentration of vapor (*Tamppari and Lemmon, 2020; Tamppari et al., 2010*). Also needed is  
1652 characterization of the surface materials over this ice—especially its depth and bulk  
1653 thermoconductive properties. More refined information about the structure of the lag (and  
1654 variations in its thermoconductive properties) would provide additional constraints for  
1655 development and testing of models related to the preservation of ice (Figure 17). In particular, an  
1656 understanding of the impact of the dust component of lags would contribute to estimates of dust  
1657 accumulation through recent martian history, feeding new constraints into studies of present and  
1658 past climates, how dust can influence ice layer evolution and accumulation, and sediment  
1659 transport pathways and reservoir amounts.

1660 Generally, improved measurements of the thermoconductive, mechanical, and compositional  
1661 properties of regolith where discussed geomorphologies and/or ground ice are found would  
1662 enable improved modeling and laboratory investigations of the processes discussed here.  
1663 Improved process models would in turn enable improved interpretation of these landforms,  
1664 regarding their formative environments and ages. In general, global identification of these  
1665 geomorphological features is complete (or can be completed) down to the decameters-scale  
1666 within global CTX imagery; higher resolution images are needed to map polygons and other  
1667 smaller morphologies.

1668 Addressing these open science questions would also contribute towards high-priority human  
1669 exploration questions regarding in situ resource utilization (ISRU). Water ice deposits within a  
1670 few meters of the surface are of high interest for human and fuel needs, and regolith (including  
1671 sublimation lag deposits) properties would feed into mining and operations designs.

1672

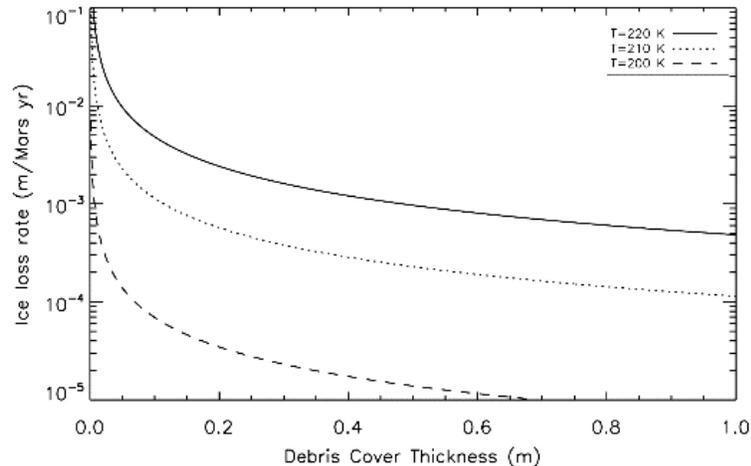


Figure 17. Retreat rate of ice under lags of varying thickness at T=200, 210, and 220 K, assuming background atmospheric water vapor with a frost point of 198.5 K. Calculation follows *Hudson et al. (2007)*.

1686

## 1687 **5 Mass-wasting aided landforms**

1688 In addition to the wind- and frost-driven features discussed above that involve mass wasting  
 1689 (§3.2: gullies, dune alcoves) and gravity-driven transport (§3.3: linear gullies), here we discuss  
 1690 two additional examples of observed downslope movement of materials. For these two cases, a  
 1691 suite of processes may be involved in initiation and enhancement of the transport; gravity is the  
 1692 only well-established driver. Recurring slope lineae (§5.1) were originally proposed to be driven  
 1693 by liquid water, but, as is discussed, the initiation mechanism for the formation of these features  
 1694 is not yet conclusively established and current observations may be more consistent with a dry  
 1695 mass-wasting mechanism. Avalanches and rockfalls (§5.2) both obviously occur due to gravity,  
 1696 with initiation mechanisms potentially related to thermal stresses and, from at least icy slopes,  
 1697 sublimation.

1698

### 1699 **5.1 Recurring Slope Lineae (RSL)**

1700 Recurring slope lineae (RSL; Figure 18; SOM 7) are relatively dark linear markings on steep  
 1701 slopes with low albedos (indicating relatively little coverage by bright dust), typically originating  
 1702 at bedrock outcrops (*McEwen et al., 2011; 2014*). Individual lineae are up to a few meters wide  
 1703 and up to 1.5 km long. The lineae grow incrementally or gradually over several months, usually  
 1704 during the warmest time of year, then fade (and typically disappear) when inactive. RSL recur in  
 1705 multiple martian years (by definition) over the same slopes, but not necessarily every year and  
 1706 not necessarily at the exact same locations. RSL often follow pristine small gullies or channels.  
 1707 Hundreds of individual lineae may be present over a local slope, and thousands are captured in  
 1708 single HiRISE images in some cases. A confirmed site (each HiRISE image sequence is  
 1709 considered a site) is where repeat images show incremental growth and fading, repeated over

1710 multiple martian years. A candidate site has similar-looking features in the same settings and  
1711 seasons as typical RSL, but repeat imaging of the site is insufficient to document growth, fading,  
1712 and recurrence. There were at least 98 confirmed and 650 candidate sites prior to MY34  
1713 (*Stillman, 2018*) (Figure 2).

1714 RSL are common in (1) the southern middle latitudes ( $-60^{\circ}$  to  $-30^{\circ}$  latitude) where they are  
1715 most active in southern summer on generally equator-facing (including east- and west-facing)  
1716 slopes; (2) the equatorial regions where activity is usually coincident with the local slope  
1717 receiving peak insolation; and (3) in Acidalia/Chryse Planitia and other northern middle latitudes  
1718 with activity in northern spring and summer (*McEwen et al., 2011; 2014; Stillman, 2018;*  
1719 *Stillman and Grimm, 2018; Stillman et al., 2014; 2016; 2017*). However, exceptions to these  
1720 timing patterns do occur (*Dundas, 2020a; Ojha et al., 2014*).

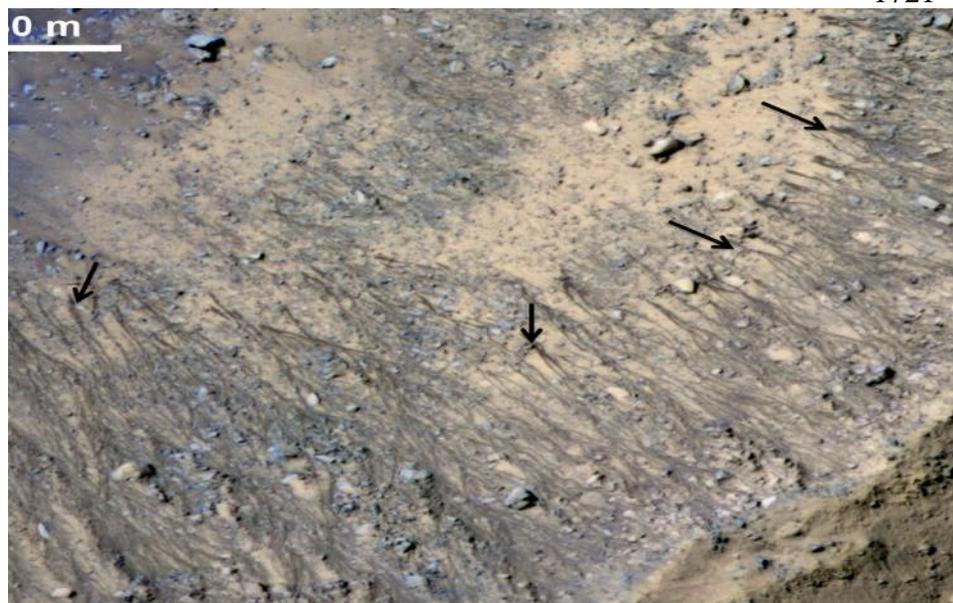


Figure 18. Hundreds of RSL present on the northwest-facing slope of a 935 m diameter simple impact crater, imaged at  $L_s$  313 $^{\circ}$  (southern summer) of MY34, located at 47.3 $^{\circ}$  S, 1.3 $^{\circ}$  E. Black arrows point at the lowermost tips of a few lineae, which begin near the crater rim. The full HiRISE image shows hundreds of

1739 dust devil tracks, indicating fresh dust deposition during the MY34 planet-encircling dust event  
1740 (PEDE) in southern spring. Color composed of infrared, red, and blue-green bandpasses,  
1741 stretched to increase contrast. North is up and illumination is from the northwest (upper left).  
1742 [ESP\\_058208\\_1325](#).

1743  
1744 Many publications have favored wet models for RSL activity (e.g., *Chevrier and Rivera-*  
1745 *Valentin, 2012; Grimm et al., 2014; Huber et al., 2020; Levy, 2012; McEwen et al., 2011; 2014;*  
1746 *Ojha et al., 2013; 2014; 2015; Stillman, 2018; Stillman and Grimm, 2018; Stillman et al., 2014;*  
1747 *2016; 2017; Wang et al., 2019*). The darkening and gradual growth resembles seeping water, and  
1748 the fading could be explained by drying. RSL appearance and temporal behavior are similar to

1749 that of water tracks in Antarctica (*Dickson et al.*, 2013; *Levy*, 2012). The surface temperatures  
1750 corresponding to RSL activity are above the freezing points for salty solutions, which can be as  
1751 low as nearly 200 K (e.g., *Möhlmann and Thomsen*, 2011). However, explaining the source of  
1752 sufficient water for seepage is extremely difficult in the present-day martian environment (e.g.,  
1753 *Dundas et al.*, 2017, and references therein). Evidence for water playing some role in RSL from  
1754 detection of rare hydrated salts (*Ojha et al.*, 2015) now appears to be a data processing artefact  
1755 (*Leask et al.*, 2018; *Vincendon et al.*, 2019). Deep groundwater may persist in Mars and might  
1756 occasionally reach the surface (*Abotalib and Heggy*, 2019; *Stillman et al.*, 2016), but RSL are  
1757 found over a wide range of elevations and settings not consistent with natural groundwater  
1758 discharge, including the tops of isolated peaks and ridges (*Chojnacki et al.*, 2016). Highly  
1759 deliquescent salts are known to exist on Mars and may temporarily trap atmospheric water in  
1760 extremely small quantities, perhaps sufficient to darken the surface (*Heinz et al.*, 2016), but not  
1761 sufficient for seepage down slopes (*Gough et al.*, 2019a; 2019b). Some workers have speculated  
1762 that small quantities of water could trigger granular flows (*Dundas et al.*, 2017; *McEwen*, 2018;  
1763 *Wang et al.*, 2019). Relatively small quantities of boiling water may trigger granular flows  
1764 (*Herny et al.*, 2019; *Massé et al.*, 2016; *Raack et al.*, 2017), but these quantities are far more than  
1765 can be supplied by the martian atmosphere with a typical water column abundance of 10  
1766 precipitable microns (*M.D. Smith*, 2008). Other hypotheses are that mass wasting may occur  
1767 when damp surface materials dehydrate (*Schorghofer et al.*, 2002) or from migration of  
1768 subsurface brines (*Bishop et al.*, 2020). Surface frost (CO<sub>2</sub> and H<sub>2</sub>O) forms in only some RSL  
1769 source regions and will sublime before RSL typically become active (*Schorghofer et al.*, 2019).

1770 Some recent papers have favored dry RSL models. *Edwards and Piqueux* (2016) found that  
1771 the thermal signature of RSL-bearing slopes at Garni crater was consistent with <3% water,  
1772 although *Stillman et al.* (2017) pointed out that none of the thermal observations were  
1773 synchronous with observations of sufficient coverage by lineae to enable thermal detection.  
1774 *Schmidt et al.* (2017) suggested that RSL could operate via granular flows driven by a Knudsen-  
1775 pump gas-flow mechanism enhanced by distinct shadowing. *Dundas et al.* (2017) found that  
1776 RSL terminate on slopes matching the dynamic angle of repose for dry sand. *Tebolt et al.* (2020)  
1777 reported RSL that terminate on lower slopes, but *Dundas* (2020a) noted that some of their  
1778 reported locations do not correspond to RSL. *Stillman et al.* (2020) concluded that the slopes in  
1779 Garni crater were consistent with granular flows within slope errors; *Munaretto et al.* (2020)

1780 reached the same conclusion about RSL in Hale crater. *Schaefer et al.* (2019) reported evidence,  
1781 including relative albedo analysis that RSL in Tivat crater fade similarly to boulder and dust  
1782 devil tracks, potentially due to dust removal from the larger region, and proposed that RSL are  
1783 dry features that mobilize dust. *Vincendon et al.* (2019) also proposed that RSL are due to dust  
1784 removal based on relationships between RSL and aeolian activity. *Dundas* (2020a) proposed that  
1785 RSL are grainflows where sand is seasonally replenished by the uphill migration of ripples, most  
1786 of which are smaller than the 25-30 cm/pixel scale of HiRISE. Following the MY34 planet-  
1787 encircling dust event (PEDE) in 2018, there was a pronounced increase in RSL activity (>5x the  
1788 activity in other years), showing a close connection to recent atmospheric deposition of dust  
1789 (*McEwen et al.*, 2019; 2021). Dust lifting activity (also forming dust devils) may directly cause  
1790 RSL formation on steep slopes, and/or dust storms may correlate with some other factor, such as  
1791 sand transport, that facilitates later RSL activity (*Dundas*, 2020a).

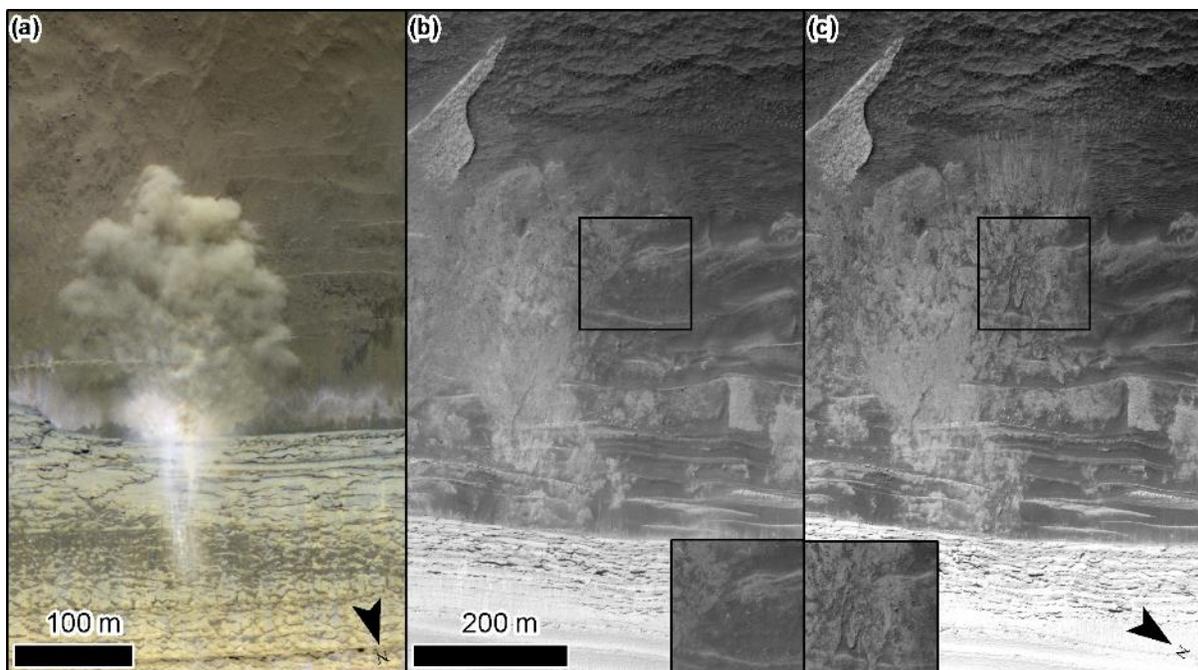
1792

## 1793 **5.2 Avalanches/block falls from rocky and icy slopes**

1794 Both individual fragments and clouds of material have been mobilized down steep slopes and  
1795 cliffs of both rock and ice on Mars. Rocks tumbling down hillslopes leave bounce and roll-marks  
1796 in their wake, whose distribution around impact craters has been used to infer that thermal stress  
1797 is a necessary pre-conditioning factor for their release (*Tesson et al.*, 2020). Other studies have  
1798 used the presence of these rockfall tracks as evidence of seismic activity on Mars (*Roberts et al.*,  
1799 2012; *Brown and Roberts*, 2019; *Senthil Kumar et al.*, 2019). Rocks have also been observed to  
1800 move downslope without leaving a visible track on the surface, in a manner that suggests an  
1801 independent rock-transport process (*Dundas et al.*, 2019b; *Raack et al.*, 2020). Precise timing  
1802 data are often lacking and sometimes contradictory, therefore further work is needed to  
1803 investigate connections between rock breakdown and frost accumulation/sublimation or other  
1804 seasonal effects.

1805 Avalanches and blocks (Figure 19) have been observed to descend from the steep scarps of  
1806 the northern polar ice cap of Mars (*Fanara et al.*, 2020a; 2020b; *Herkenhoff et al.*, 2007; *Russell*  
1807 *et al.*, 2008;). These appear to be two separate categories of mass movement because the  
1808 avalanches are rarely associated with mobilized blocks and spectral evidence suggests they could  
1809 simply be mobilizing the surface seasonal frost deposits (*Pommerol et al.*, 2013). Avalanches  
1810 observable at the time of day of HiRISE observations occur exclusively in early spring (*Russell*  
1811 *et al.*, 2014). Some scarps appear to be more active than the others, hinting at the importance of

1812 some type of localized conditions such as near-surface winds or sun exposure (*Russell et al.*,  
1813 2014). However, it is still an open question regarding whether the origin of these avalanches is  
1814 caused by the thermal stresses in the scarp, sublimation of the seasonal CO<sub>2</sub> deposits, the wind,  
1815 or a yet-identified process (*Becerra et al.*, 2020; *Byrne et al.*, 2017). Conversely, the fallen  
1816 blocks of ice are thought to be detached by thermal stresses on these exposed locations and were  
1817 thought to potentially balance the deformation of the scarps via viscous deformation (*Sori et al.*,  
1818 2016). However, the fallen blocks at one well-studied north polar site account for a minimum  
1819 average scarp retreat rate of ~0.2 m/kyr, which does not balance the published 0.01–1 m/yr  
1820 viscous flow rates, suggesting that either viscous flow rates are lower than modeled or that  
1821 additional processes act to maintain the scarps' steepness (*Fanara et al.*, 2020a). The activity of  
1822 both the blocks and avalanches are important to understand because they give us insight into the  
1823 mass balance of the polar cap and the climate record exposed at these steep scarps.  
1824



1825  
1826 Figure 19: (a) Polar avalanche in enhanced-color HiRISE image ESP\_016228\_2650. (b)  
1827 “Before” HiRISE image ESP\_027750\_2640 and (c) “after” HiRISE image ESP\_036888\_2640  
1828 showing a blockfall with insets showing detail of the blocks and other changes between the  
1829 images. The scalebar and north arrow apply to both (b) and (c) panels.  
1830

### 1831 5.3 Open questions for these mass-wasting aided landforms

1832 In general, with these mass-wasting landforms, we do not yet definitively understand the  
1833 suite of processes involved and thus cannot interpret the observed landforms and activity as

1834 markers of specific environmental conditions. Volatiles, thermal cycling, aeolian processes,  
1835 and/or seismicity are often invoked as drivers for initiation of the downslope transport of  
1836 materials, and determining specifically (and preferably, quantitatively) what causes the activity  
1837 would be useful because then these features can be used as a proxy indicator of their initiation  
1838 mechanism. Additionally, understanding the role of volatiles in rock breakdown and  
1839 mobilization is important for constraining long-term erosion rates on Mars and interpreting the  
1840 degradation state of landforms (e.g., relative dating of fan surfaces in gullies).

1841 In particular, RSL represent the latest in a series of surface features interpreted as evidence for  
1842 flowing water on Mars today, given the clear preference for warmer slopes and the temporal  
1843 behavior mimicking that of seasonal seeps of water on Earth. Although further observations and  
1844 analyses have led to difficulties with every proposed water-driven model, given the planetary  
1845 protection significance of potential water on Mars today, there is large interest in further  
1846 measurements to conclusively determine the formation mechanism for these features (*ICE-SAG*,  
1847 2019; *McEwen*, 2018; *NEX-SAG*, 2015). At present, RSL sites have been classified as potential  
1848 sites where terrestrial microbes might flourish (denoted “unknown special regions”: *Rummel et*  
1849 *al.*, 2014; *Kminek et al.*, 2017) and thus areas that spacecraft must avoid unless they can achieve  
1850 very high levels of sterilization; for example, the presence of RSL was used to rule out candidate  
1851 landing sites for the Mars 2020 rover (*Grant et al.*, 2018). If RSL are dry or only transiently wet  
1852 at very cold temperatures, then this restriction on future Mars exploration could be lifted  
1853 (*McEwen*, 2018), but recent work suggests that putative deliquescent RSL sites could be  
1854 habitable (*Maus et al.*, 2020).

## 1855 **6 Summary of the measurements needed to answer remaining** 1856 **questions**

1857 The general aim of geomorphic studies is to connect quantitatively the observed landforms to  
1858 their formative environmental drivers, via models of the active process(es). For the landforms  
1859 and surface activity we have described, advances generally require additional information about  
1860 the specific environmental drivers for formation and subsequent modification and/or evolution.  
1861 For example, additional information about environmental drivers is needed to progress our  
1862 knowledge of araneiform activity (§3). Both surface properties in araneiform-forming regions as  
1863 well as time-resolved global atmospheric conditions and dynamics (including winds, clouds, and  
1864 dust content) are needed to determine how and where CO<sub>2</sub> ice accumulates and evolves  
1865

1866 seasonally, which is essential information for constraining models of CO<sub>2</sub> basal sublimation and  
1867 thus the scale of eruptive vents and basal erosion in araneiform terrain. Such information,  
1868 combined with observations of the growth of the seasonal ice cap, would also allow for tests of  
1869 predictions of where CO<sub>2</sub> ice is of sufficient thickness and strength for the formation and growth  
1870 of araneiforms—in the present climate or during a recent past climate.

1871 Information needed to address the open questions outlined in previous sections (§2.3, 2.4,  
1872 3.4, 4.5, 5.3) can be gathered through a few complementary study types:

1873 With observational data:

- 1874 1) Mapping where the landforms exist and/or are active, and where they are not
- 1875 2) Geomorphological measurement of the landform and its activity
- 1876 3) Characterization of the timing of activity (e.g., in season, in time of day, in event duration,  
1877 and identification of interannual variation)
- 1878 4) Characterization of the surface (and potentially subsurface) and atmospheric environment  
1879 where and when the activity occurs

1880 With laboratory, terrestrial field analog, and physics modeling studies:

- 1881 5) Identification of possible environmental drivers and investigation of scaling relationships,  
1882 temporal evolution rates, and interactions between materials

1883 Table 3 summarizes which of these areas are most needed for studies of the martian surface  
1884 activities discussed above.

1885 To acquire the observational data related to mapping and timing, continued high-resolution  
1886 orbital imagery is key. The advent of HiRISE-type imaging demonstrated that the martian  
1887 surface is active in the present climate, yielding a paradigm shift from the view that most of the  
1888 interesting martian geologic activity occurred in the ancient past (i.e., during the Noachian and  
1889 Hesperian). Continued repeat imaging of the surface with similar sub-meter resolution and  
1890 illumination will enable identification of yet more surface changes, including those with slower  
1891 activity rates, and potentially tie activity timing to specific seasons. Additionally, increased  
1892 spatial coverage will enhance mapping studies; HiRISE has so far imaged only ~2% of Mars'  
1893 surface. To aid image comparison and identification of geomorphic changes, MRO's current  
1894 orbit is sun-synchronous (i.e., observations recur at specific local solar times of 3 a.m. and 3  
1895 p.m.). The ability to observe the surface at different times of day is also critical because some  
1896 active processes on Mars may only occur during a specific time-of-day. Spacecraft in other orbits

1897 can view different times of day, but with other constraints such as the changing viewing  
1898 conditions making change detection analysis more difficult. For example, a spacecraft in a  
1899 circular orbit with inclination of  $75^\circ$  would drift through all times of day,  $\sim 3x$  per season (*NEX-*  
1900 *SAG*, 2015). Currently, MEx and TGO are both in elliptical orbits and can view the surface  
1901 during different times of day, albeit with visible imagery at lower resolution than HiRISE.  
1902 Alternatively, in situ observations can provide high-resolution and high-frequency observations  
1903 throughout a Mars day, at the location of the sensors. Correlation of observations acquired by  
1904 different spacecraft enables a powerful confluence of high temporal and spatial resolution  
1905 information within regional/global coverage, as well as imagery over a range of wavelengths.

1906 An additional benefit to continuation of global imaging is that interannual variations in  
1907 surface activity can be tracked, yielding another way to constrain environmental drivers. In  
1908 particular, observations of activity before and after the 2018 PEDE have shown that the  
1909 redistribution of dust and related atmospheric effects have increased the frequency of some  
1910 surface changes, such as RSL formation (*McEwen et al.*, 2021), suggesting that these activities  
1911 may involve more dust than was originally hypothesized. The extensive dust activity also altered  
1912 the seasonal frost cap formation/sublimation cycle and related landform activity (e.g., *Calvin and*  
1913 *Seelos*, 2019; *Hansen et al.*, 2020).

1914 To connect the landforms to environmental controls, we also need measurements of the  
1915 environment where these landforms and activity are found. Coupling surface and subsurface  
1916 compositional, thermophysical, and structural measurements with meteorological conditions over  
1917 sites where a specific landform and/or activity is observed allows for a holistic analysis of the  
1918 full system. From orbit, globally distributed (if not with global coverage) compositional and  
1919 thermophysical information has been gleaned from spectral images through the near-infrared to  
1920 thermal wavelengths. However, these datasets are limited to spatial resolutions much coarser  
1921 than the scale of the activity: many are 100 m/pixel or coarser, the best is CRISM with  $\sim 20$   
1922 m/pixel (*Murchie et al.*, 2007). Furthermore, many spectral datasets are only sensitive to surface  
1923 exposures, so a thin layer of dust is enough to obscure the surface materials. In such areas,  
1924 geologic unit mapping can provide some constraints, based on extrapolation from visible  
1925 outcrops, topography, and radar analysis. In situ compositional data, as collected by the Mars  
1926 rovers and landers, allow for much more detailed measurement of surface and near-subsurface

1927 properties. Coupling the in situ data with the global perspective provided through orbital  
 1928 observations has been key to constraining some of the interpolative analysis.

1929 Such analysis will also be important for studies of meteorological conditions, with orbital  
 1930 data providing a look at global circulation and atmospheric features such as clouds. However,  
 1931 existing in situ aeolian and other meteorological data are insufficient to robustly answer surface-  
 1932 atmosphere interaction questions because no dedicated sediment sensors were included in past  
 1933 missions and the meteorological instruments flown were not well accommodated and were not  
 1934 designed to be part of a comprehensive aeolian/meteorological experiment (*ICE-SAG*, 2019;  
 1935 *MEPAG*, 2020). In situ monitoring of surface atmosphere exchanges would provide key new  
 1936 information for constraining volatile and sediment flux models under Mars conditions (Table 4).

1937  
 1938 Table 3. A high-level summary of the types of data currently thought to be needed to advance  
 1939 studies of these features. As hypotheses evolve, definition of the next-needed data would likely  
 1940 change; in all cases more or new data could prompt unexpected new questions or analyses. The  
 1941 numbers/headers for the columns are discussed at the start of this section. Note that columns #1-  
 1942 4 are more focused on spacecraft-acquired observational data, and #5 is more focused on  
 1943 laboratory, terrestrial field analog, and physics modeling studies. Color coding: (Green)  
 1944 Extensive analysis exists or future analysis of existing data types and coverage would be  
 1945 sufficient to assess the broad questions; (Blue) The existing data type(s) are sufficient but  
 1946 increased spatial and/or temporal coverage is needed to address the broad questions, (Purple)  
 1947 New types of data/investigations are needed to address the broad questions.

	1: mapping	2: geomorphology	3: timing	4: environment	5: drivers/ interactions
Decimeter-scale ripples	B	G	B	P	B
Meter-scale ripples	G	B	B	P	P
Decameter-scale ripples	B	B	B	P	P
Dunes	B	G	B	P	B
Gullies	G	G	G	P	P
Dune alcoves	B	G	B	B	P
Linear gullies	B	G	G	B	B
Araneiforms	B	G	G	P	B
SPRC	G	B	B	B	B
H <sub>2</sub> O ice layers within the MCID	G	G	G	B	P
MCID	P	G	P	P	P
Northern cap	G	G	G	P	P

Sublimation thermokarst	G	G	P	P	B
Ice Scarps	G	G	B	P	B
Patterned Ground	G	G	P	B	G
RSL	G	P	B	P	P
Avalanches from rocky slopes	B	B	P	P	B
Avalanches from icy slopes	G	G	B	B	B
Fallen blocks from icy slopes	G	G	G	B	B

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1950  
1951  
1952

Table 4. Some of the Mars in situ aeolian and meteorological concurrently-collected information needed to fill critical gaps in surface-atmosphere interaction and landform formation models (compiled from recent Mars community discussions and reports such as *ICE-SAG*, 2019; *MEPAG*, 2020).

Science Investigation		In situ Measurements
<b>Responses</b> Measure the surface-atmosphere fluxes of sand, dust, volatiles, heat, and momentum.	Volatiles	Absolute concentration in the atmosphere
	Dust	Local surface dust erosion and deposition rates
		Lofted dust flux and grain sizes
	Sand	Surface sand erosion and deposition rates
		Saltation profile: Number, sizes, and velocities of grains in motion, as a function of height
		Reptation flux rates and grain sizes involved
		Creep flux rates and grain sizes involved
	Heat	Temperature profile
		Net downwelling and upwelling radiation
	Momentum	Horizontal wind measurements from at least <u>three</u> heights – to derive surface shear stress; of frequency to determine “average” velocities and the gust velocity distribution (or 3D wind measurements)
<b>Drivers</b> Determine the controls on mechanisms that lead to sediment, volatiles, and heat being moved from the surface	Meteorological Controls (+ winds, above)	Atmospheric temperature and pressure – to derive atmospheric density
		Atmospheric composition (including trace gases/humidity)
		Surface pressure and temperature
		Turbulence (i.e., high-frequency 3D wind measurements)
		Vortices/dust devils: number/frequency, surface shear stress, and amount of dust carried
		Overhead clouds, coverage, characteristics, and altitude
		Atmospheric electric field and electric conductivity
		Grain size distribution on the nearby surface

into the atmosphere and transported.	Surface Controls	Surface grain properties (e.g., angularity, composition, electrostatics)
		Local surface topography, geologic-type (e.g., bedrock exposure, dust cover), and surface roughness elements
		Local surface mechanical and thermoconductive properties (e.g., cohesion, thermal inertia)

1953

1954 **7 Mars as a “natural laboratory” for comparative planetology studies**

1955 As discussed above, Mars is the only body outside of the Earth-Moon system where we have  
 1956 acquired sufficient data (in time/space/type) to observe present-day activity and to investigate  
 1957 processes within a measured environmental and geological “system.” In particular, repeat high-  
 1958 resolution observations have yielded many examples of present-day surface activity and these  
 1959 changes have or could be studied in enough detail to suggest the driving process(es) and/or  
 1960 influential environmental conditions. Furthermore, decades of previous work have yielded ample  
 1961 geologic and atmospheric contextual information and models that will greatly enhance  
 1962 incorporation of the broader geologic and atmospheric/climatological context within process-  
 1963 focused investigations. Timely acquisition of new data needed to fully constrain and calibrate a  
 1964 process model (§6) may also be possible because commercial and international interest in  
 1965 sending spacecraft (and humans) to Mars, along with Mars’ relatively close proximity to Earth,  
 1966 suggest that access opportunities should exist for sending new spacecraft and instruments to  
 1967 Mars over the next decade. This may be especially true for small spacecraft that do not need a  
 1968 dedicated launch vehicle—which may be sufficient for targeted environmental monitoring  
 1969 studies.

1970 Furthermore, Mars’ atmospheric, surface, and planetary conditions are different enough from  
 1971 Earth’s to test and stretch terrestrial study-based models, but similar enough that the terrestrial  
 1972 models are a reasonable starting point. As will be described, Mars’ conditions often have values  
 1973 between those on Earth and other planetary bodies, such as Kuiper Belt Objects (KBOs) or the  
 1974 Moon. Mars’ surface activity in modern times and through the last few billions of years can,  
 1975 thus, serve as a valuable and unique comparative planetology experiment for both (1) processes  
 1976 active on Earth, but under very different conditions (e.g., aeolian processes within a low density  
 1977 atmosphere or with much higher obliquity), and (2) processes not active on Earth but that are  
 1978 active on other planetary bodies (e.g., sublimation-dominated frost dynamics; or records of

1979 active processes without complications from recent extensive fluvial or biological activity) (also  
1980 discussed in *Lapôtre et al.*, 2020).

1981 In the following sections, we describe how Mars serves as a great (perhaps the current best)  
1982 comparative planetology basis for studies of (1) aeolian surface processes and meteorological  
1983 dynamics, (2) sublimation-driven geomorphic dynamics, and (3) planetary bodies with variable-  
1984 density atmospheres. Although not covered in this review, we note that, at Mars, one can also  
1985 study both the above individual phenomena and interactions between them. There are also many  
1986 additional study areas where Mars serves well as a non-Earth comparison point or extraterrestrial  
1987 “laboratory” for testing and refining models, such as impact cratering rates and processes,  
1988 habitability/life evolution, atmospheric dynamics, and polar/ice climate records (e.g., *Lapôtre et*  
1989 *al.*, 2020; *I.B. Smith et al.*, 2020).

1990

## 1991 **7.1 Aeolian surface processes and meteorological dynamics**

1992 Aeolian sand and dust are known to significantly influence landscape evolution and climate  
1993 across the Solar System. In addition to the aeolian landforms and dynamics studied on Earth and  
1994 Mars (§2):

- 1995 • A few dune fields (*Greeley et al.*, 1995; *Weitz et al.*, 1994; Figure 20e), many wind streaks,  
1996 and a few potential yardangs (*Greeley et al.*, 1995) have been identified on Venus, under an  
1997 atmosphere 9× thicker than Earth’s.
- 1998 • On Saturn’s moon Titan, sand produced from photochemical organic aerosols and water ice  
1999 has formed vast dunes and sand seas (*Barnes et al.*, 2015; *Lorenz et al.*, 2006; *Radebaugh*  
2000 *et al.*, 2008; Figure 20c); dust storms may also occur during equinox (*Jackson et al.*, 2020),  
2001 similar to those observed on Mars (*P. Thomas and Gierasch*, 1985).
- 2002 • Even on comets (*N. Thomas et al.*, 2015; Figure 20i) and icy worlds, such as Pluto (*Telfer et*  
2003 *al.*, 2018; Figure 20a), aeolian processes within a transient, rarified atmosphere appear to  
2004 have formed bedforms.

2005 Interpretations of these landforms and processes have generally relied upon models of sediment  
2006 fluxes and transport dynamics (i.e., saltation and reputation rates and profiles) derived primarily  
2007 from terrestrial field and laboratory experiments, with scaling applied based on specific planetary  
2008 conditions. This has led to new tests of bedform evolution models (e.g., *Claudin and Andreotti*,  
2009 2006; *Duran Vinent et al.*, 2019; *Kok*, 2010; *Kok et al.*, 2012; *Parteli and Herrmann*, 2007;

2010 *Sullivan and Kok, 2017; Sullivan et al., 2020; Vaz et al., 2017*) and proposal of a new scaling  
2011 relationship to predict ripple equilibrium wavelength (e.g., *Lapôtre et al., 2016; 2017; 2018;*  
2012 *2021*) after Earth conditions-based model predictions were found to be inconsistent with  
2013 bedforms observed on another planet.

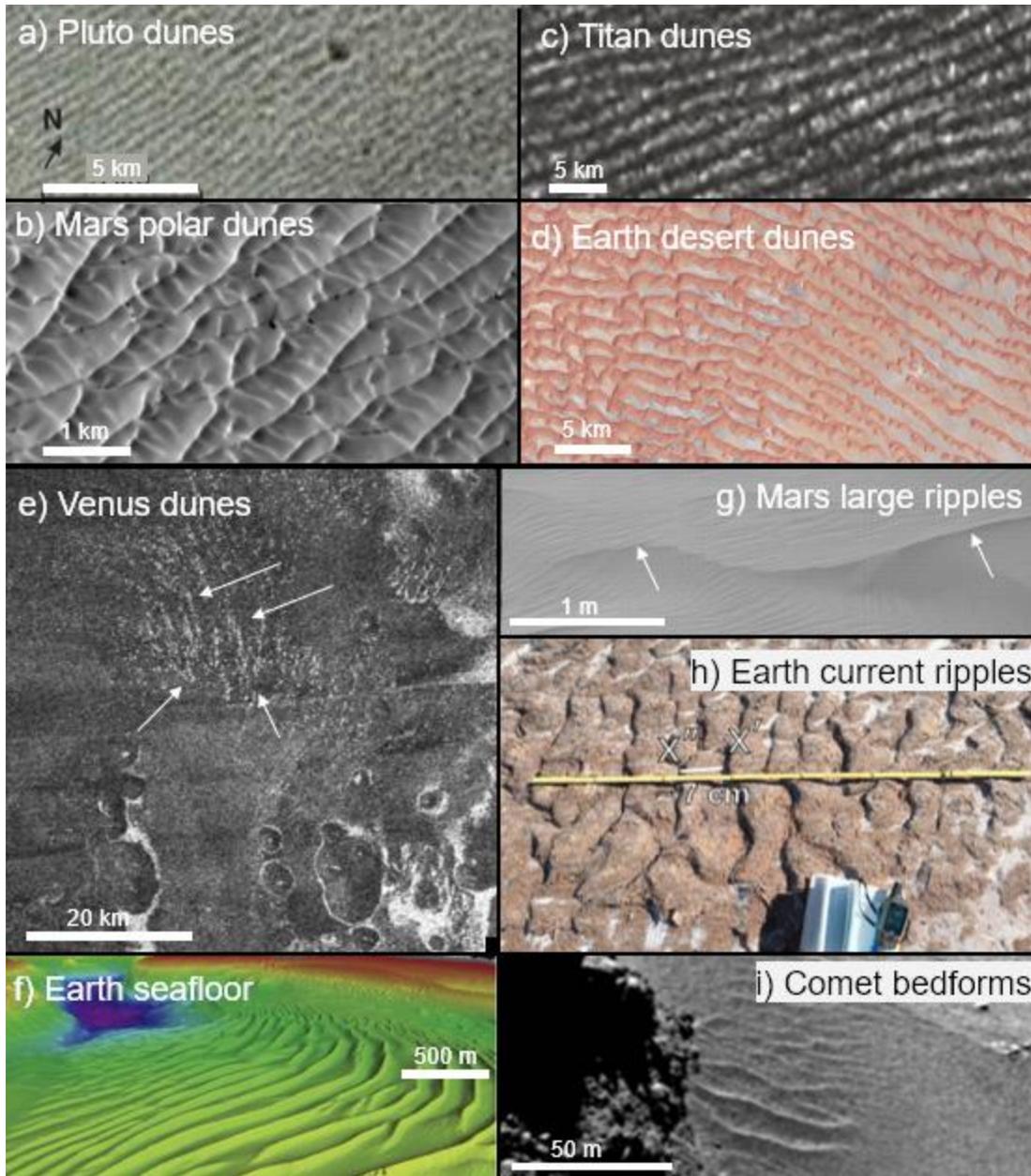
2014 Further study of martian aeolian and other meteorological systems, and how conditions drive  
2015 surface activity, will enable more detailed testing and refinement of surface-atmosphere  
2016 interaction process models. In particular, in situ “field” measurements of martian atmospheric  
2017 boundary layer dynamics driving sand/dust and volatile transport would provide novel  
2018 calibration data for models and wind-tunnel experiments within an environment with a  
2019 substantially lower impact threshold than fluid threshold (*Kok, 2010*). Such “ground truth” is  
2020 needed to advance a cross-planet model to describe sand and dust lofting and transport, including  
2021 helping to discriminate between models such as fluid-dominated (e.g., *Bagnold, 1941; Shao and*  
2022 *Lu, 2000*) or impact-dominated (e.g., *Kok and Renno, 2009; Sullivan and Kok, 2017*) transport or  
2023 coarse grain motion via direct-drag or impact-driven creep (e.g., *Baker et al., 2018a, Silvestro et*  
2024 *al., 2020*).

2025 Another model presently untested under extraterrestrial conditions is the one used to estimate  
2026 turbulent eddy fluxes (which result in the exchange of energy, momentum, and quantities like  
2027 dust, water, and other chemical species between the surface and atmosphere). On Earth, turbulent  
2028 fluxes can be directly calculated from correlated, high frequency measurements of the 3D wind  
2029 components and the quantity of interest; such fluxes are also related to large-scale (and more  
2030 easily measured) quantities such as the vertical gradient of temperature and of the horizontal  
2031 wind (i.e., the wind shear) (*Businger and Yaglom, 1971; Businger et al., 1967; Monin and*  
2032 *Obukhov, 1954*), with ample testing and calibration of the physical model through terrestrial field  
2033 and laboratory studies. These relationships are assumed in planetary studies but have never been  
2034 shown to extend to those environments despite generally being far outside of terrestrial  
2035 conditions (e.g., Mars has an extremely stable nocturnal inversion and unstable afternoon  
2036 convective layer). New in situ martian meteorological measurements would enable validation  
2037 and calibration of this theory within a wider range of atmospheric conditions.

2038 Mars is already used as the comparative planetology basis for some studies, due to its low  
2039 atmospheric density. For example, the threshold curve under low-density gas conditions derived  
2040 for Mars analog conditions in the MARSWIT (*Greeley and Iversen, 1985*) has also been used to

2041 model aeolian-type transport resulting from jetting on comets (*Cheng et al.*, 2013). Dune-like  
2042 patterns on the surface of comet 67P/Churyumov-Gerasimenko (*N. Thomas et al.*, 2015) have  
2043 been suggested to result from thermal winds, although the process involves outgassing from  
2044 cometary jets feeding a rarefied atmosphere rather than from atmospherically driven winds as on  
2045 the surfaces of Mars and Titan (*Jia et al.*, 2017). In imaging data from the New Horizons  
2046 mission, Pluto was shown to have 0.4–1 km-wavelength bedform-like morphologies (*Stern et al.*,  
2047 2015). Proposed to be aeolian dunes (*Telfer et al.*, 2018), a minimum wind shear required for  
2048 saltation was estimated based on work performed in the MARSWIT that separated Reynolds  
2049 number and interparticle cohesion effects (*Iversen and White*, 1982). The important point of  
2050 analogy here is the substantially lower impact threshold than fluid threshold, as discussed in  
2051 §2.2.1.

2052



2053  
 2054 Figure 20. Examples of bedforms on different planetary bodies: (a-d) Planetary features with  
 2055 remarkable geomorphic similarities, leading to hypotheses of aeolian dune fields. (e-j) Examples  
 2056 from the diverse suite of bedforms found on other bodies, which may be more analogous to  
 2057 terrestrial subaqueous bedforms than subaerial ones (proposed terrestrial analogs are shown in (f  
 2058 and h), for Venus and Mars/comet, respectively). Rough scale bars are included, but distances  
 2059 are not exact because images are generally not orthorectified. Images were selected/adapted  
 2060 based on: (a) *Telfer et al.*, 2018: New Horizons color-composite MVIC images, (b) *Diniega et*  
 2061 *al.*, 2017: HiRISE image PSP\_007115\_2600, (c) *Radebaugh et al.*, 2010: Cassini synthetic  
 2062 aperture radar (SAR) image, (d) Landsat image of Rub' al Khali in the Arabian peninsula  
 2063 (<https://earthobservatory.nasa.gov/blogs/earthmatters/2012/11/02/dune-gallery/>), (e) *Diniega et*  
 2064 *al.*, 2017: Magellan SAR image of Menat Undae dune field/NASA Photojournal PIA00483, (f)  
 2065 *Neakrase et al.*, 2017: sand waves in San Francisco Bight (from Fig. 6), (g, h) *Lapôtre et al.*,

2066 2018: MSL Mastcam image acquired in Sol 1221, image of current ripples from dry riverbed in  
2067 Death Valley, CA (from Fig. 2), (i) *Thomas et al.*, 2015: Rosetta image NAC\_2014-09-  
2068 18T00.33.01.377Z\_ID10\_1397549800\_F22.  
2069

## 2070 **7.2 Sublimation-driven geomorphic dynamics**

2071 As discussed above, each martian fall/winter CO<sub>2</sub> frost and ice are deposited in a thick  
2072 seasonal layer and in the winter/spring hemisphere this layer sublimates. Mars has lower gravity  
2073 and a lower pressure and temperature environment than Earth, causing sublimation processes to  
2074 differ from terrestrial analogs, including laboratory analogs. Mars' atmosphere is in vapor  
2075 equilibrium with surface CO<sub>2</sub> frost/ice (*Leighton and Murray*, 1966), thus Mars has an  
2076 environment closer to those of Triton, Pluto, and possibly other Kuiper Belt Objects (*Ingersoll*,  
2077 1990; *Owen et al.*, 1993), rather than terrestrial and laboratory analogs. For example, no large-  
2078 scale sublimation dynamics naturally occur on Earth, and within laboratory studies we cannot  
2079 replicate low gravity over timescales relevant for sublimation.

2080 Sublimation is thought to be the main-driver forming a range of depressions or pits on Mars,  
2081 including SPRC pitting (§4.1.1), icy scarps (§4.3.1), and scalloped depressions (§4.3).  
2082 Sublimation has also been linked to the formation of other (non-active) depressions which have  
2083 not been discussed above, including pitted terrain in and around impact craters (*Boyce et al.*,  
2084 2012; *Tornabene et al.*, 2012), crenulated and labyrinthine pitted surfaces of martian glaciers  
2085 (e.g. *Levy et al.*, 2009a; *Mangold*, 2003), and dissected latitude dependent mantle (*Milliken et al.*,  
2086 2003; *Mustard et al.*, 2001; *Soare et al.*, 2017). These different features are proposed to form due  
2087 to sublimation from volatile reservoirs of different sizes/forms/ages, and vastly different  
2088 sublimation rates. Thus, studies of these features on Mars enable discrimination between  
2089 different proposed models and identification of key geomorphological or environmental  
2090 signatures for separating features that likely formed through slightly different processes, within  
2091 different environments, and/or over very different timescales.

2092 *Mangold* (2011) provides a detailed review of sublimation landforms in the Solar System,  
2093 and here we add new results with the additional perspective of using Mars as an analog. Some  
2094 sublimation-generated depressions on Mars that have already been compared to depressions on  
2095 the surfaces of other bodies in the Solar System include the following:

- 2096 • Hollows on Mercury (e.g., *Blewett et al.*, 2011; *R.J. Thomas et al.*, 2014) (Figure 21b)  
2097 which are thought to be due to volatile-loss or sublimation of sulfur-related compounds,

2098 likely sulfides (e.g., *Bennett et al.*, 2016) and analogies have been drawn to martian swiss  
2099 cheese terrain (§4.1.1; Figure 21a).

2100 • Pitted surfaces have been discovered on asteroids, notably pitted crater floors on Vesta  
2101 (*Denevi et al.*, 2012) and Ceres (*Sizemore et al.*, 2017; 2019) which are remarkably similar  
2102 to pitted terrain found in craters on Mars (*Boyce et al.*, 2012; *Tornabene et al.*, 2012).

2103 Additionally, pitted areas and scarps on outer Solar System moons have been grouped with  
2104 Mars polar features as evidence of sublimation degradation (*Moore et al.*, 1996).

2105 • Pitted terrains on Pluto (Figure 21c) that are thought to represent erosion from sublimation-  
2106 driven winds of the surface of nitrogen ice and possibly methane ice (*Buhler and Ingersoll*,  
2107 2018; *Moore et al.*, 2017). Similar terrains are thought to exist on other Kuiper Belt  
2108 Objects, and the origins of flat-floored pits on Arrokoth remain mysterious (*Schenk et al.*,  
2109 2020). Bladed terrains on Pluto are also thought to derive from sublimation (*Moore et al.*,  
2110 2017).

2111 • Pitted surfaces found on comets (e.g., *Sunshine et al.*, 2016). Recent exploration of 67P  
2112 Churyumov–Gerasimenko by Rosetta maintains that these pits, which span a range of sizes,  
2113 grow primarily via sublimation (e.g., tens to hundreds of meters-wide pits: *Vincent et al.*  
2114 (2015), meters-wide pits: *Birch et al.* (2017)).

2115 On Mars, sublimation is thought to have an important role to play in initiating and/or  
2116 enhancing mass wasting over sandy slopes (§3.2.1-2, 3.3.1, 5.2) and rocky slopes (§3.2.1, 5.2)  
2117 (Figure 21d). Martian mass-wasting features have been proposed to form useful analogies for:

2118 • Gully-like landforms identified on Mercury (*Malliband et al.*, 2019) and Vesta (*Krohn et al.*  
2119 *et al.*, 2014; *Scully et al.*, 2015) (Figure 21e,f), potentially related to sublimation of sulfur-  
2120 compounds and water, respectively.

2121 • Downslope features that have been observed on Helene, one of Saturn's Trojan moons,  
2122 whose formation has been ascribed to sublimation (*Umurhan et al.*, 2016) (Figure 21g).

2123 Furthermore, it is likely that closer inspection of other planetary bodies will reveal further  
2124 examples of sublimation-driven mass-wasting processes.

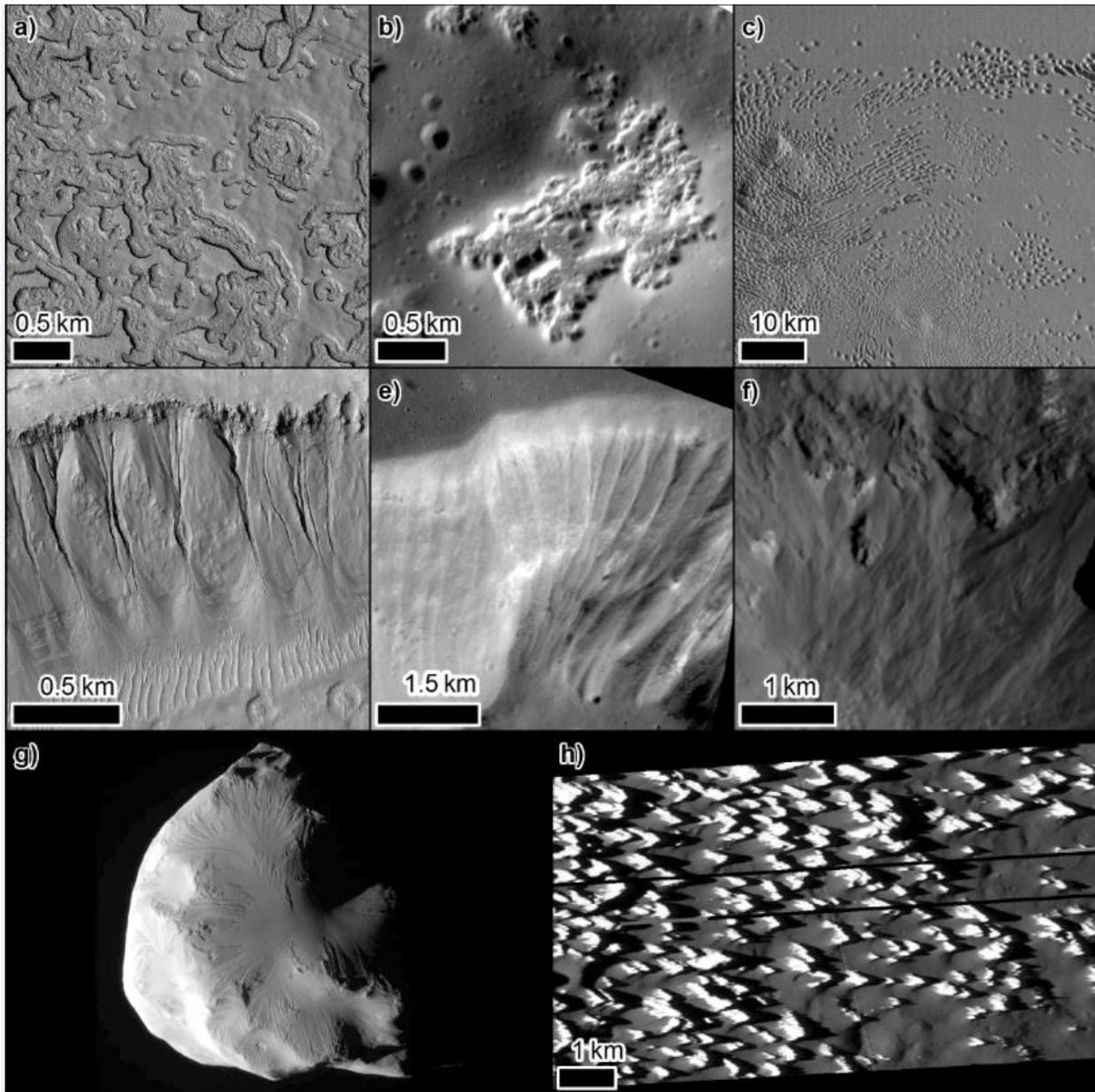
2125 In general, sublimation and related processes drive dust and other contaminants up to the  
2126 atmosphere or, in the case of comets, into their coma. Escaping pressurized gas from sublimation  
2127 at the base of the CO<sub>2</sub> seasonal ice cap on Mars erodes araneiforms (*Hansen et al.*, 2010; *Kieffer*,  
2128 2007; *Piqueux et al.*, 2003; *Portyankina et al.*, 2017), radially organized and /or dendritic

2129 channels (*Mc Keown et al.*, 2017) initially dubbed “spiders,” and deposits material across the  
2130 surface (*N. Thomas et al.*, 2010). These jets of gas, whose origin is via basal sublimation and the  
2131 solid state greenhouse effect (§3.1; *Kieffer*, 2007), and the patterns/rates of material they spew  
2132 out could be useful analogues for the following:

- 2133 • Sublimation processes proposed to cause to global-scale contrasting albedo regions on  
2134 Iapetus and Ganymede (*Giese et al.*, 2008; *Prockter et al.*, 1998; *Spencer and Denk*, 2010).
- 2135 • CO<sub>2</sub> ice signatures found on the trailing hemispheres of the Uranian satellites (*Cartwright*  
2136 *et al.*, 2015; *Grundy*, 2003; *Grundy et al.*, 2006), notably visible as a bright deposit inside  
2137 Wunda Crater on Umbriel (*Sori et al.*, 2017). Severe sublimation of CO<sub>2</sub>-ices is thought to  
2138 explain the pinnacle terrain on Callisto (*Howard et al.*, 2008; *White et al.*, 2016) (Figure  
2139 21h).
- 2140 • Solar-driven jets of materials found on other bodies. The solar-driven theory was actually  
2141 proposed to explain plumes on Triton (*Soderblom et al.*, 1990) before it was applied to  
2142 Mars’ geysers (*Kieffer et al.*, 2006), although later Triton observations suggested a  
2143 cryovolcanic origin might instead be responsible (*Waite et al.*, 2017).

2144 Additionally, interactions between wind and sublimation dynamics can be explored on Mars. For  
2145 example, a “sublimation wave” model of dynamics at the interface between an icy substrate and  
2146 a turbulent boundary layer flow may explain certain icy landform periodicities on Mars and  
2147 Earth (*Bordiec et al.*, 2020). Bodies such as Pluto, Ceres, and Jovian and Saturnian icy moons  
2148 are also hypothesized to have surface sublimation and winds, so similar dynamics could be  
2149 expected there (*Bordiec et al.*, 2020).

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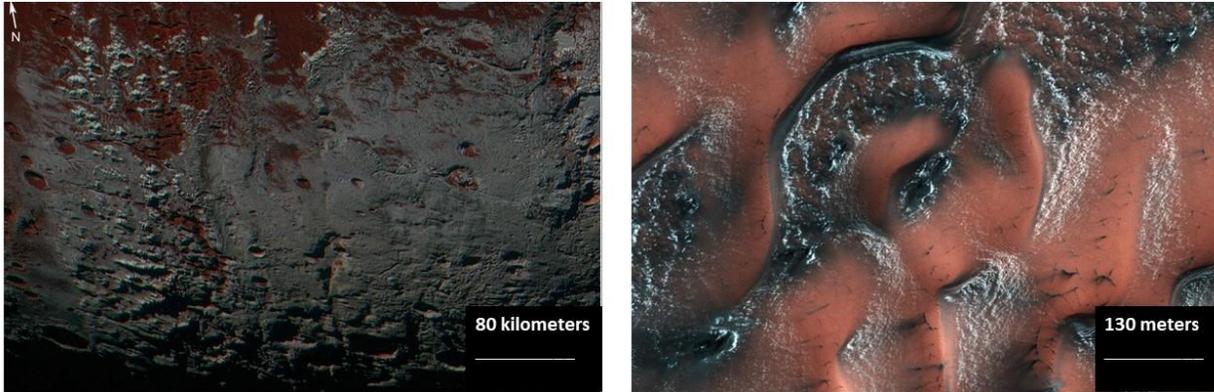


2151  
 2152 Figure 21. Geomorphic features on multiple planetary bodies, thought to be formed through  
 2153 surface frost sublimation and potentially analogous to features on Mars. (a) Swiss cheese terrain  
 2154 on Mars HiRISE image ESP\_057828\_0930. (b) Hollows on Mercury in Scarlatti impact basin,  
 2155 MDIS NAC image EN1051805374M. (c) Pits on Pluto, New Horizons LORRI image  
 2156 0299179742. (d) Gullies on Mars, HiRISE image. (e) Mass wasting gullies on Mercury in  
 2157 Nathair Facula, MDIS NAC image EN1059620367M. (f) Mass wasting gullies on Vesta in  
 2158 Cornelia Crater, DAWN Framing Camera image FC21B0025747. (g) Image of Saturn's Moon  
 2159 Helene taken by Cassini-Huygens ISS NA camera. Lit terrain is on the leading hemisphere of  
 2160 Helene measuring ~33 km across and North is down. (h) Pinnacle terrain on Callisto, Galileo  
 2161 image PICNO (Picture number) 30C0003. Image processing for MDIS and Galileo was  
 2162 performed using ISIS3 via the U.S. Geological Survey PILOT and POW systems.  
 2163

2164 In addition to sublimation, studies of the present-day accumulation and evolution of frost and  
2165 ice on Mars may provide analog information about the types and interactions of frost and ice on  
2166 other bodies. For example, one model of methane snow on Pluto (*Witzke*, 2015;  
2167 <https://www.nasa.gov/feature/methane-snow-on-pluto-s-peaks>; Figure 22) indicates that it  
2168 forms due to a circulation-induced high-altitude enrichment of gaseous methane, a process  
2169 different from those forming high-altitude snowpacks on Earth (*Bertrand et al.*, 2020).  
2170 Comparisons with H<sub>2</sub>O and CO<sub>2</sub> snow on Mars may provide a better comparative planetology  
2171 starting point to understand precipitation and volatile transport on planets and dwarf planets with  
2172 tenuous atmospheres. The surface pressure of the martian atmosphere is 2–3 orders of magnitude  
2173 less than that of the Earth, and Pluto’s atmosphere is another ~3 orders of magnitude lower,  
2174 providing a large physical range for future modeling that applies to a full suite of planetary  
2175 atmospheres in the Solar System and elsewhere. Laboratory studies of H<sub>2</sub>O and CO<sub>2</sub> ice (e.g.,  
2176 *Chinnery et al.*, 2018; *Kaufmann and Hagermann*, 2017; *Pommerol et al.*, 2019; *Portyankina et*  
2177 *al.*, 2019; *Yoldi et al.*, 2021), as well as how evolution of such materials is altered through  
2178 interaction between the ices and dust, coupled with Mars ice and environment observations  
2179 provides the current best route for formulating and calibrating models of these strange ices under  
2180 extraterrestrial conditions. Even if not providing a direct analog, study of martian ices may also  
2181 help ground truth models and demonstrate how to interpret spacecraft observations and connect  
2182 them with terrestrial experiments involving exotic ices.

2183 Finally, while not directly related to present-day frost accumulation/sublimation or  
2184 observable activity, the creep of martian glaciers also likely presents a useful analog for studies  
2185 of outer Solar System bodies. The balance between sublimation/ablation, deposition, and flow  
2186 rates is thought to be significantly different in martian vs. terrestrial glaciers. For example, due to  
2187 an overall lower surface temperature, water-ice glaciers on Mars exhibit different dynamics from  
2188 most terrestrial glaciers, i.e., without basal melting or basal sliding (*Head and Marchant*, 2003;  
2189 *Marchant et al.*, 1993); there may be evidence of past CO<sub>2</sub> glaciers (*Kreslavsky and Head*,  
2190 2011). Both valley and piedmont glaciers on Pluto have been identified in the region of Sputnik  
2191 Planitia (*Moore et al.*, 2016). Some show evidence of bulk flow, with basement material or  
2192 nunatuks protruding above the mobile material (*Stern et al.*, 2015).

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Figure 22. Snow on Pluto and Mars. The image on the left from the Multispectral Visible Imaging Camera on the New Horizons spacecraft shows possible methane snow on mountains in the southern hemisphere of Pluto. The image on the right is a false color image from the High Resolution Imaging Science Experiment (HiRISE) camera on NASA's Mars Reconnaissance Orbiter shows CO<sub>2</sub> frost on martian dunes at a northern latitude of 76° (north is down). Picture credits left: NASA/JHU APL/SwRI (discussed in NASA press-release 03-03-2016, <https://www.nasa.gov/feature/methane-snow-on-pluto-s-peaks>). Right: NASA/JPL/UA, HiRISE ESP\_050703\_2560.

### 7.3 Planetary bodies with variable-density atmospheres

Although in this study we primarily focused on known or hypothesized present-day surface activity and related landforms, as discussed above, studies of the present-day Mars provides a key to interpret the archive of past Mars' surface processes and climate conditions. In particular, both aeolian processes and sublimation dynamics will be influenced by atmospheric density, which has varied on Mars over seasonal to much longer timescales. Studies of observable surface activities in the present (including variations in activity rates correlated to seasonal or interannual environmental variations) enables testing of models that then are extrapolated back to past martian climates, or to other bodies that may experience analogous cyclic variations and/or atmospheric collapse.

As discussed in §3.1, the CO<sub>2</sub> atmosphere of Mars is in vapor pressure equilibrium with surface ice; seasonally CO<sub>2</sub> sublimates and condenses, changing the atmospheric density by >25% in the present climate (e.g., *Forget et al.*, 1998; 1999; *Hartogh et al.*, 2005; *Leighton and Murray*, 1966; *Pollack et al.*, 1990; 1993). The dynamics of this process modulate the global circulation and drive local sublimation winds, such as katabatic winds that are thought to play an important role in the formation of polar troughs (*Spiga and Smith*, 2018). Similar surface-atmosphere processes act on other planetary bodies where the atmosphere is in vapor pressure equilibrium with surface ice, such as on Triton, Pluto, and KBOs (*Bertrand et al.*, 2020; *Hansen*

2222 *et al.*, 2018; *Zalucha and Michaels*, 2013). Thus, Mars' processes and climate cycles may  
2223 present a good analog for interpreting the integrative geomorphological result of atmosphere-  
2224 surface processes on these bodies, including sublimation-driven formation of surface features  
2225 and aeolian-driven processes (*Moore et al.*, 2017; *Young*, 2012).

2226 Due to cycles in various orbital parameters (such as obliquity), the Mars atmospheric density  
2227 may cycle through a range of 1–12 mbar over thousands to millions of years timescales (*Buhler*  
2228 *et al.*, 2020; *Manning et al.*, 2019). Derivation of present-day martian surface activity models  
2229 that quantitatively connect landform morphologies to driving environmental conditions will  
2230 enable improved interpretation of relict features and reduce uncertainty when extrapolating  
2231 activity models through past Mars climates. Such developments will also provide a testable basis  
2232 for generation of similar models on other bodies that also experience large, cyclic changes in  
2233 atmospheric density, such as on Pluto (*Betrand et al.*, 2018; *Forget et al.*, 2017; *Hansen and*  
2234 *Paige*, 1996) and Triton (*Hansen and Paige*, 1992; *Trafton*, 1984; *Yelle et al.*, 1995). As Earth's  
2235 atmosphere has not gone through comparable large swings in atmospheric density during the  
2236 portion of Earth's history when most of the Earth's observable rock record was formed, Mars  
2237 provides important "ground truth" for this type of extrapolative analysis and integration of  
2238 predicted geologic records through different atmospheric pressures.

2239 For example, on Mars, both ancient surface and stratigraphic features and modern active  
2240 processes can be directly observed and measured. This enables models of sedimentary processes  
2241 to be investigated through different climate conditions. In particular, the morphologies of large  
2242 martian ripples have been proposed to provide a way to constrain atmospheric density changes  
2243 within Mars' climate history (*Lapôtre et al.*, 2016). As previously discussed (§2.1.3), the  
2244 wavelength of large martian ripples appears to be a function of atmospheric density (*Lapôtre et*  
2245 *al.*, 2016; *Lorenz et al.*, 2014). The wavelength of old ripples can be read in inactive ripple fields,  
2246 but also within the cross-stratification left behind by bedforms (e.g., *Rubin*, 1987; *Rubin and*  
2247 *Carter*, 2006). Thus, provided that bedform dimensions can be extrapolated from the martian  
2248 aeolian record (e.g., *Banham et al.*, 2018; *Grotzinger et al.*, 2005; *Lapôtre et al.*, 2016) and with  
2249 a mechanistic understanding of how atmospheric density controls bedform size (e.g., through  
2250 kinematic viscosity, specific sediment density, and possibly wind shear velocity; *Lapôtre et al.*,  
2251 2016; 2017), one should be able to reconstruct the history of atmospheric density from the  
2252 aeolian rock record. Such results, especially coupled with terrestrial-based sedimentary process

2253 models, could advance studies of analogous sedimentary deposits on other planetary bodies and  
2254 enable even more climatological and geologic history to be interpreted from limited  
2255 observations.

2256 Another example is about how erosive potential of basal sublimation from CO<sub>2</sub> ice slabs is  
2257 affected by the thickness of the seasonal ice layer or insolation conditions, leading to the  
2258 formation of araneiforms. As discussed in §3.3.2, some studies of these features suggested that  
2259 araneiforms may be active at very slow rates (*Piqueux and Christensen, 2008*), but repeat high-  
2260 resolution imaging of these features has not yielded any discernible changes in topography over  
2261 the last decade. Based on lab experiments of the CO<sub>2</sub> ice sublimation activity over granular  
2262 materials, it has instead been proposed that some of the araneiforms (especially the largest and  
2263 those displaying a non-radial network) may be relicts of a past climate when the frost depth or  
2264 insolation amount was different, leading to more energetic sublimation (*Mc Keown et al., 2021*).  
2265 Determination of the environmental controls on the basal sublimation rates and resultant erosion  
2266 potential of the escaping gas would enable improved interpretation of the ice layer  
2267 thickness/strength needed to form these features. Should the needed ice layer be more than those  
2268 forming in the present climate, then the araneiforms could be interpreted as direct records of past  
2269 wintertime conditions. Such results provide constraints on models of the pressures attained via  
2270 basal sublimation—a distinctively non-terrestrial process that would be applicable towards  
2271 studies of jets and substrate erosion on other bodies.

2272 In parallel, but out of phase with variable density atmospheres, the surface deposition of  
2273 meteoric ice will result in layered and likely stratified volatile deposits with impurities. On Mars,  
2274 impurities likely include dust, lithic fragments from volcanic eruptions or ejecta, fine salt grains,  
2275 trapped gasses, and isotopologues (*ICE-SAG, 2019; I.B. Smith et al., 2020*). Mars is not the only  
2276 planetary body to experience partial atmospheric collapse (*Soto et al., 2015*). Pluto (*Bertrand et*  
2277 *al., 2018; 2019; Hansen and Paige, 1996; Olkin et al., 2015*) and Titan (*Lorenz et al., 1997*)  
2278 likewise have strong seasonal atmospheric cycles (lasting hundreds of Earth years) and orbital  
2279 variations that could cause similar ice layering as is found on Mars, and atmospheric collapse has  
2280 been proposed for tidally locked planets around TRAPPIST-1 (*Turbot et al., 2018*). Earth, with  
2281 anthropomorphic influences, abundant biology, and liquid phases, does not provide a good  
2282 analog for such layered ice deposits or climate models.

2283

## 2284 **8 Lessons learned from planetary geomorphological studies**

2285 Based on recurrent challenges and some of the key science advancements within studies of  
2286 martian present-day activity, we identify pitfalls and strategies that may benefit future planetary  
2287 and terrestrial geomorphological studies. First, a key lesson is that geomorphological similarity  
2288 to terrestrial landforms may present a good starting point for a hypothesis of similar formation  
2289 process and driving environmental conditions, but geomorphological similarity alone is not  
2290 sufficient to conclude parallel evolution. One needs to consider other observations and datasets  
2291 to determine if there is “system”-level consistency with processes or environmental conditions  
2292 similar to those on the Earth (e.g., the timing of activity, geologic context, compositional  
2293 constraints, and contemporaneous environmental characterization). This applies both to  
2294 comparisons between features on the same planet (e.g., martian gullies (§3.2.1) versus dune  
2295 alcoves (§3.2.2)) and to comparisons between features on different planets (e.g., gullies (§3.2.1)  
2296 on Mars, Mercury, and Vesta (§7.2)).

2297 A second key lesson is that many interactions and controls are nonlinear, so there are often  
2298 complications both in scaling an analog process or landform under new environmental conditions  
2299 and in trying to separate out the influences of multiple processes on a planetary surface. For the  
2300 first, laboratory/field experiments and modeling studies are crucial for testing proposed  
2301 relationships and even seeing what the process looks like under exotic conditions; for example,  
2302 due to the low surface pressure, liquid water would flow and boil on the present-day martian  
2303 surface, creating small “flow” morphologies different from those observed on Earth (*Herny et*  
2304 *al.*, 2019; *Massé et al.*, 2016; *Raack et al.*, 2017); until these experiments were run, levitating  
2305 sand pellets were not expected or taken into account in theories. For the second, looking at a  
2306 range of activity and landform types across a planetary surface, as well as mapping where a  
2307 process seems to be active and where it appears to not be active (e.g., Figure 2), can help  
2308 detangle processes and driving environmental conditions.

2309 A third key lesson is that long-term observation of change is needed to fully characterize a  
2310 process and its expression and rate(s), as moderated by changes in driving environmental  
2311 conditions. Activity levels can vary dramatically from year to year (e.g., as is currently being  
2312 investigated with the 2018 Mars PEDE).

2313 Finally, to increase the science value of new observations and enable a holistic look at Mars  
2314 present-day phenomena, the international space agencies and Mars exploration programs along

2315 with an active and connected Mars science community have been instrumental in enabling  
2316 strategic linkages between observations, especially between orbital and in situ assets. Having  
2317 such community communication/coordination and data accessibility is clearly key for the  
2318 “system” science generally involved in investigations of geomorphological processes. Related, a  
2319 research and analysis program that supports both data analysis and fundamental research studies  
2320 helps scientists collaborate and combine different types of study (e.g., Mars’ “natural laboratory”  
2321 observations, laboratory/wind tunnel experiments, field analog studies, and physical/numerical  
2322 models) to robustly test and calibrate models describing the observed activity, and then  
2323 extrapolate from observed conditions to past or more exotic environments. As such work is  
2324 inherently cross-disciplinary, cross-target, and diverse in scope, it is critical also that the  
2325 community foster an interdisciplinary, diverse, equitable, inclusive, and accessible environment  
2326 so that a wide range of people and perspectives can interact, communicate, and then contribute  
2327 towards understanding the active surface processes and improve science advancement.

2328

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2349

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