Evidence for fluvial and glacial activities within impact craters that excavated into a Noachian volcanic dome on Mars

Harish Harish, S. Vijayan, N. Mangold

► To cite this version:

Harish Harish, S. Vijayan, N. Mangold. Evidence for fluvial and glacial activities within impact craters that excavated into a Noachian volcanic dome on Mars. *Icarus*, 2021, 361, pp.114397. 10.1016/j.icarus.2021.114397 . hal-03376416

HAL Id: hal-03376416

<https://hal.science/hal-03376416>

Submitted on 13 Oct 2021

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

Evidence for fluvial and glacial activities within impact craters that excavated into a Noachian volcanic dome

Harish^{1,2}, S Vijayan¹, and N Mangold³

¹Planetary Science Division, Physical Research Laboratory, Ahmedabad 380009, India

²Indian Institute of Technology, Gandhinagar 382355, India

³Laboratoire de Planétologie et Géodynamique, Nantes, CNRS UMR6112, Université de Nantes, Université d'Angers, 44322 Nantes, France

Corresponding author:

S Vijayan

email: vijayan@prl.res.in, vijayansiva@gmail.com

24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46

Highlights:

1. Impact craters Degana and Degana-A excavated into a Noachian volcanic dome
2. New observations of multiple alluvial fans, moraine-like ridges within these craters
3. Olivine and LCP exposed by these impact craters reveal the composition of volcano
4. Snow/ice is the likely source for post-Noachian fluvial and glacial activities

Abstract:

Impact craters on Mars preserve diverse records of volcanic, fluvial, and glacial activities. Enigmatically, the preservation of these major activities or records altogether within impact craters is rare. We report one such new observation of impact craters that formed on a volcanic dome studied using data from the Mars Reconnaissance Orbiter's (MRO) Context Camera (CTX), High-Resolution Imaging Science Experiment (HiRISE), and Compact Reconnaissance Imaging Spectrometer for Mars (CRISM), Mars Global Surveyor's Mars Orbiter Laser Altimeter (MOLA), and Mars Express' High-Resolution Stereo Camera (HRSC). A ~20 km diameter impact crater, informally named as Degana-A, is formed within the ~50 km diameter impact crater Degana. Mineralogical analysis reveals exposures of low-calcium pyroxene and olivine deposits, which occupy Degana-A's eastern walls, leading to the idea that pristine Noachian bedrocks might be exposed from beneath the volcanic dome. Degana-A floor is completely covered by alluvial fans from all sides with distributaries. Within Degana crater, multiple fans are observed along its eastern to southern side only. The most likely source of water was the accumulation of snow on Degana crater walls, which possibly melted as a result of the impact of Degana-A. We observed a ~1 km wide breach on the eastern wall of Degana-A and the estimated maximum flow velocity is ~2 m/s and a run-off ~2.25 mm/hr. Over the south-facing walls, multiple moraine-like ridges superposed the fans, which suggests overprinting by glacial activities. The presence of fans and superposed moraine-like ridges located at the mid-latitudes (~23°S) implies atmosphere-derived snow/ice precipitation was possible. Chronologically, the dome is of Noachian age, whereas Degana crater formed in the Hesperian period and crater retention on the fans indicates late Hesperian to Amazonian ages. Overall, the preserved Noachian crustal material underneath a volcanic dome is rarely exposed in its pristine context, which offers a rare window into early igneous processes. This intriguing location also witnessed a climatic transition as implied by water/ice derived landforms formed by non-coeval events.

1. Introduction

Sedimentary deposits on Mars retain the record of ancient environments and prevailing climatic conditions (Moore and Howard, 2005; Kraal et al., 2008; Blair and McPherson, 2009; Grant and Wilson, 2011; Morgan et al., 2012; Morgan et al., 2014; Wilson et al., 2021). Sedimentary deposits are more developed in the Late Noachian and Early Hesperian epochs, where they are frequently associated with widespread mineralogical assemblages as observed in-situ at Gale crater (e.g., Bristow et al., 2018) and from orbit (e.g., Carter et al., 2013). However, among the various sedimentary deposits observed on Mars, mid-latitude alluvial fans date predominantly from the Late Hesperian-Early Amazonian epochs, so they are of great interest as they represent fluvial activity that occurred after a potential climatic optimum of the Noachian (Moore and Howard, 2005; Kraal et al., 2008; Grant and Wilson, 2011; Morgan et al., 2012; Mangold et al., 2012a,b; Kite et al., 2017; Wilson et al., 2021). Although many studies have described alluvial fans, their origin remains debated among several hypotheses: snowmelt versus rainfall (e.g., Milton, 1973; Hoke and Hynek, 2009; Di Achille and Hynek, 2010a,b; Wordsworth et al., 2013; Kite et al., 2013), transient or stable climatic conditions (Moore and Howard, 2005; Armitage et al., 2011; Grant and Wilson, 2011,2012; Kite et al., 2017), the role of impact or volcanic activity in melting water ice (e.g., Gulick, 2001; Hauber et al., 2005; Williams et al., 2008; Ansan and Mangold, 2013). Thus, finding key locations to better understand these processes is an important challenge to provide answers on Martian climate change.

In this context, we present new observations of fluvial and glacial landforms in Degana crater (23.72°S, 314.50°E), an impact crater that formed on a volcanic dome (Xiao et al., 2012) (Figure 1). Xiao et al. (2012) mapped almost all the ancient volcanoes on Mars, excluding the Tharsis region. Their study identified the region over which Degana crater is located as a huge Noachian-age volcanic dome. They suggested these domes are mostly conical in shape and lack calderas. However, no direct observation has yet been made to show that they host fluvial deposits and glacial landforms. Likewise, no mineralogical analysis has been carried out over this volcanic dome to decipher the Noachian mineralogy. Thus, this region offers a unique chance to observe volcanic, impact, glacial, and fluvial landforms in the same location, assessing whether the processes that formed them occurred in concert or not. In addition, the composition of such Noachian volcanoes is critical to understand the evolution of ancient volcanic crust on Mars (Xiao

et al., 2012). In this study, we will provide new information about the geological context and bedrock composition and characterize the morphology of the fluvial and glacial landforms. We established the relative stratigraphic relationships between the volcanic dome, impact craters, fluvial deposits, and glacial landforms within both impact craters. Then, we performed quantitative analysis on the fluvial and glacial landforms to constrain their extent and timing. Also, using crater-size frequency analysis, we determined the apparent ages of these processes. At last, we discuss the formation hypothesis for the fluvial and glacial deposits and the climatic evolution of Mars.

2. Data and Methods

For the geomorphological analysis, we used ~6 m/px Mars Reconnaissance Orbiter (MRO)-Context Camera (CTX), and ~25 cm/px MRO-High-Resolution Imaging Science Experiment (HiRISE) images (Malin et al., 2007; McEwen et al., 2007). For the topographic analyses, we used ~463 m/px Mars Global Surveyor (MGS)-Mars Orbiter Laser Altimeter (MOLA) DEM (Smith et al., 2001), and the ~200 m/px Mars Express (MEX)-High-Resolution Stereo Camera (HRSC) – MGS MOLA blended DEM (Ferguson et al., 2018). We also used high-resolution CTX DEM and HiRISE DEMs (wherever available), generated using the open-source pipeline of MarsSI (Quantin-Nataf et al., 2018).

For the mineralogical analysis, we used data by the MRO-Compact Reconnaissance Imaging Spectrometer for Mars (CRISM), a hyperspectral camera (Murchie et al., 2007). CRISM used two detectors with 544 channels in the visible to the near-infrared spectral region (0.4 to 4.0 μm) (Murchie et al., 2007). In this study, CRISM data analysis was performed using both S-detector (VNIR) and L-detector (IR) datasets (Murchie et al., 2007). The short wavelength (VNIR) data ranges from 0.36 to 1.05 μm , while the long-wavelength (IR) data ranges from 1.0 to 3.92 μm . The analysis was performed using the CRISM Analysis Toolkit (CAT) plug-in added to ENVI 4.5. CRISM observations were converted into I/F reflectances by applying corrections for instrumental artifacts (Murchie et al., 2007). Degana crater includes only one full resolution targeted (FRT00023F32), which was processed to correct for photometric, and atmospheric noise in the spectral data using the method suggested by Murchie et al. (2007) and Flahaut et al. (2010).

For identifying the composition, we used the spectral parameters OLINDEX2, LCPINDEX, and HCPINDEX mentioned in Viviano-Beck et al. (2014). OLINDEX2 is the modified olivine parameter that measures olivine's 1.0 μm band depth in the presence of spectral slopes (Salvatore

et al., 2010). LCPINDEX and HCPINDEX are the low Calcium-pyroxene and high Calcium-pyroxene parameters, respectively, that measure the pyroxene's 1.0 μm and 2.0 μm band depth (Pelkey et al., 2007). The regions highlighted in spectral parameters were examined in more detail to get a diagnostic spectral signature. Areas with strong spectral absorption signatures were used to determine the region of interest (ROI) and spectra were taken from ROIs of several pixels. We used a minimum of 3*3 pixels ROI for each average spectrum considered in this study (Table 1). We used spectral data in the wavelength range from 0.5 to 2.6 μm by integrating the VNIR and IR data to confirm the mineral absorptions of the Fe-bearing phases, which generally turned up around 1 μm (Sun and Milliken, 2015). The selected ROI spectra were ratioed with the neutral spectra, mostly a dust-rich area obtained from the same column (Ehlmann et al., 2009) to enhance the spectral features. We calculated the ratioed spectra for each detector (VNIR and IR) separately and then integrated them. We also removed the overlapping spectral data from 1.0 to 1.05 μm in the S-detector and L-detector datasets as suggested by Murchie et al. (2007).

3. Observations and Interpretations

3.1 Geologic context and bedrock composition

3.1.1 Geologic context and volcanic dome

The study area (Figure 1a) is located at the easternmost side of the Coprates quadrangle in the southern highlands of Mars (McCauley, 1978). Figure 1a shows the geologic context of the area of study, which shows an elevated area similar to a dome. This dome extends more than 200 km in diameter (Figure 2a). Many impact craters of varying diameters superimpose the dome (Figure 1). The largest crater on the western flank of the volcanic dome is the Degana crater whose diameter and depth are ~50 km and ~2 km, respectively. Another impact within the Degana crater leads to form a ~20 km diameter crater, and depth of ~0.7 km, which is named in this study as 'Degana-A' (Figure 2b).

Degana crater and its surrounding region are mapped as Mid-Noachian highland units (Figure 1b), which hosts undifferentiated materials formed by volcanic and impact activity (Scott and Tanaka, 1986; Tanaka et al., 2014). Also, this unit is identified as moderately to heavily degraded, heavily cratered, hosts dense valleys, grabens, and wrinkle ridges (Scott and Tanaka, 1986; Tanaka et al., 2014). Xiao et al. (2012) categorized this dome structure as a shield volcano. The dome is

embayed on its edge by Late Noachian volcanic plains, which contain wrinkle ridges as typically observed in Late Noachian and Hesperian plains of Coprates Planum (Tanaka et al., 2014). Tanaka et al. (2014) have mapped two valley networks in the region. The valley network in the south-west is Her Desher Vallis, whereas the other valley network has been mapped on the eastern flank of the dome (Figure 1b).

An elevation profile AA' (north to south) taken along this volcanic dome shows typical domical topography (Figure 2a). The elevated dome stands ~1.2 km above the adjoining plain surface (Figure 2a), whereas, the craters Degana (Figure 3a) and Degana-A (Figure 3b) have excavated ~1.5 km below the regional baseline defined by the reference elevation of the plain surface adjacent to the dome. Overall, the two craters in combination penetrated ~2.7 km deep into the dome (Figure 2b). Considering the surrounding plains, the surface elevation at the time of dome formation might be ~500 m. It means Degana-A reached a terrain deeper than the surface that exists before dome formation, i.e. an outcrop could be exhumed from the primitive crust below the volcanic dome or from a crystallized magma chamber beneath the dome. If the surrounding plains are thicker than 1.5 km, the outcrop would be exhumed from the deep part of the volcanic dome. In both cases, this observation provides an opportunity to analyze the pristine bedrock of Mars, which is preserved from the middle-Noachian epoch (Xiao et al., 2012). Apart from Degana and Degana-A craters, the other circular features atop the dome (Figure 1) are unnamed craters marked in Robbins and Hynes (2012)'s impact crater database. These four large craters are superposed over the volcanic dome, as shown in (Figure 3). The diameter of these craters ranges from ~20 km to 40 km. In this study, the four craters are named unnamed craters A, B, C, and D (hereafter referred to as UA, UB, UC, and UD) (Figure 1a). Figure 3c-f shows the four individual craters with their raised rims and relatively distinguishable ejecta from two of them (Figure 3e,f). Based on their raised rims and ejecta, we interpret these craters as impact craters (Figures 1a,3) and not as volcanic features.

3.1.2 Bedrock mineralogy

The exposed bedrock material from the two superposed craters is analyzed using CRISM data. Before analyzing the individual CRISM spectra, we derived the CRISM color composite map or spectral parameter map according to Viviano-Beck et al. (2014) by using the OLINDEX2,

LCPINDEX, and HCPINDEX. This derived color composite image is overlaid over the CTX-DEM (**Figure 4a**) to assess the compositional variation observed within the crater. This RGB combination of spectral parameters (OLINDEX2, LCPINDEX, and HCPINDEX) shown in **Figure 4a** highlights potential olivine-rich regions in red color, low-calcium pyroxene (LCP) regions as green/cyan and high-calcium pyroxene (HCP) regions as blue color as suggested by **Viviano-Beck et al. (2014)**.

From these highlighted regions, we have chosen spectra from 9 ROIs (**Figure 4a**), which are mostly located on the northeastern and southeastern wall/rim of Degana-A. The locations of all the numerator spectra, denominator spectra, and the ROI sizes are given in **Table 1**. These are the locations where the diagnostic absorption features in the spectra are distinguishable. Spectra signature from the northeastern rim of Degana-A shows (**Figure 4b**) a broad absorption near 1 μm and 2 μm wavelength. Absorptions around 1 and 2 μm are spectrally identified as pyroxene, where the band centers strongly dependent on the Ca content (**Adams, 1974; Cloutis et al., 1986; Cloutis and Gaffey, 1991; Klima et al., 2007**). All the colored spectra from the green/cyan region (**Figure 4a,b**) exhibit a strong absorption feature between 0.8-1.0 μm , indicating that the ferrous phase is plausibly low-calcium pyroxene (LCP) (**Huang et al., 2012**). An additional broad absorption before 2 μm (**Figure 4b**) implies the presence of low-calcium pyroxene as distinguished elsewhere on Mars (**Mustard et al., 2005; Skok et al., 2012; Clenet et al., 2011b**). We have observed that LCP is widespread over this part of the north-eastern rim (**Figure 4a,b**). Among the six identified locations of LCP detections, spectra P1, P2, and P6 are located on the northeastern rim. LCP identified in locations P3 and P4 is exposed along the eroded walls of the valleys, which are connected to the east rim of Degana-A. The LCP spectra obtained from P5 are located on the south-eastern wall of Degana-A crater (**Figure 4a**). The LCP ratioed CRISM I/F reflectance spectra obtained from the six ROIs matches with RELAB (**Viviano-Beck et al., 2015**) representative spectra of low-calcium pyroxene (**Figure 4b**).

The other diagnostic spectra are obtained from the south-eastern wall of Degana-A crater. **Figure 4c** shows a broad absorption around 1 μm and no other diagnostic absorption at other wavelengths. A broad absorption around 1 μm in the near-infrared region and a lack of absorption around 2 μm is indicative of olivine where the band center is dependent on the relative content of Fe and Mg (**Burns, 1970; King and Ridley, 1987; Sunshine and Pieters, 1998**). The spectra shown in

Figure 4c indicate absorption around shorter wavelengths ($<1\ \mu\text{m}$), which is due to an increase in Mg content (Ody et al., 2013). In addition, the absence of strong absorption between $1.2\text{--}1.4\ \mu\text{m}$ (**Figure 4b**) also indicates that the olivine is richer in Mg (e.g., Huang et al., 2012). This suggests that this potential olivine outcrop is consistent with Mg-rich olivine (**Figure 4c**). Spectra O1 and O2 in **Figure 4c** exhibit olivine signatures within Degana crater, whereas, the spectra shown in black is the lab spectrum of Mg-rich olivine (Forsterite) from RELAB (Viviano-Beck et al., 2015). We have observed similar absorptions at location O3 (**Figure 4a**). Although the olivine mineral exposures are spatially prevalent on the south-eastern wall of Degana-A, isolated patches have also been observed on the north-eastern wall in the color composite image (**Figure 4a**). We have also used the CRISM CAT tool associated Modified Gaussian Model (MGM) to estimate the cation compositions of the pyroxene mineral (Sunshine et al., 1990). Using MGM we determined the normalized LCP/(LCP+HCP) ratio, which remains high ($>70\%$) in most of the locations, as shown in **Figure S1**. This analysis suggests that the proportion of LCP is high within Degana crater in comparison to other mafic minerals. Overall, the minerals identified using CRISM are dominated by primary mafic minerals (e.g. Viviano-Beck et al., 2015). However, we have not detected any hydrated minerals within the fan deposits or along the eroded walls of valleys. The complete lack of hydrated minerals is possibly caused by dust mantling on the floor, or by the limited area covered by the single CRISM cube, so the presence of hydrated minerals cannot be ruled out.

Figure 5 shows the HiRISE RGB and greyscale images locations where CRISM mineral spectra were obtained. **Figure 5a,b** shows the locations of LCP exposures in false-color images by HiRISE, as indicated by cyan coloring (Sun and Milliken, 2015). We have also observed distinct spectral absorptions over the light-toned outcrop (**Figure 5c**), which is identified as a possible olivine exposure (**Figure 4c**) on the inner crater wall. In addition, olivine exposures were identified along the inner crater wall of Degana-A (**Figure 5d**) in the greyscale HiRISE images. Interestingly, all the LCP and olivine detected within the craters are linked to either the inner crater wall or the proximal part of the valley. Overall, the notable mafic signatures are observed along the rim of Degana-A crater, where the slope is high enough ($>30^\circ$) to expose these signatures. This suggests that the elevated region of the inner crater wall and the adjoining valley wall exposed the mafic minerals due to their steep slope and possible erosion. It also suggests that a significant portion of the mafic minerals is blanketed by alluvial deposits in other areas. The presence of LCP on the northeastern and southeastern wall of Degana-A crater indicates a continuous layer along the rim

that was exposed by the impact. This small exposure, although dust-covered, allows us to infer the mineral composition exposed within this dome. Thus, we infer these exposed mineral units to be pristine Noachian crust uplifted by volcanic dome formation.

3.2 Morphology of fan deposits

We mapped the occurrence and extent of fan deposits, valleys, and ridges within the crater (**Figure 6**). The four fan deposits within Degana-A crater cover the entire floor area (**Figure 6a,b**). The fan deposits are also observed on the floor of Degana crater, especially in the southwestern quadrant of the crater (**Figure 6b**). **Figure 6b** shows the geomorphic map of the impact craters and their diverse features. **Figure 7** shows six prominent fan deposits formed within Degana-A and Degana craters, and their corresponding area, fan length, and slope gradient are given in **Table 2**.

We observed several channels with positive relief on the fans (**Figure 7**). These channels are consistent with original distributary, where the positive relief likely formed as a result of the finer surrounding material being removed by the wind. The fans within Degana crater (**Figure 7**) host several distributaries. Some of these distributaries present on the fans' surface are marked by white arrows (**Figure 7**). We observed that these distributaries are mostly radiating away from the fan apex (**Figure 7**). The toe regions of the fans located within Degana-A fused with each other (**Figures 6,7**). Within fans (f1-f4), multiple distributaries are observed and they vary in their extent and width. These distributaries are few kilometers long and a couple of meters wide (**Figure 7**). The fans within the Degana-A are of low gradients in the range of 2° to 4° (**Table 2**), and the four fans have almost equal areas (**Table 2**). In almost every fan deposit (**Figure 7a-d**), near the apex region, the distributaries appear in a cluster, whereas away from the apex region towards the downstream, they are apart from each other and observed to follow the typical topographic driven flow. The distributaries are oriented downslope, and they are aligned along with the fan orientation (**Figure 7d**). The fans located on the floor of Degana crater (**Figure 7e,f**) did not show as many sharp distributaries as observed on the fans located within Degana-A. The distributaries are observed only at the termini of fans f5 and f6 (**Figure 7 e, f**), and they are mantled by a dust cover. Interestingly, **Figure 7e** shows the possible coalescing of two fans in the toe region. Dust cover and transverse aeolian ridges (TAR) (Wilson and Zimbelman, 2004; Berman et al., 2011) were observed to fill the intermediate troughs within the fan deposits.

Figure 8a shows the HiRISE DEM of surface features of fan f1, where the prominent distributaries are radiating from the fan apex. The elevation gradient along the distributaries can be seen in the HiRISE DEM. Though all the distributaries are sourced from the apex points, some of them are discontinuous. **Figure 8b** shows a topographic profile with varying elevations along a distributary within fan f1, which has a relief of ~10 m. **Figure 8c** shows distributaries on fan f1, which are often stacked, exhibiting crosscutting and superposition relationships. We have also observed that the width of these distributaries decreases as we move farther from the apex (**Figure 8c**). These distributaries over the Degana fans are up to 10s of meters thick. The observed crosscutting of the distributaries is the most prominent intersection at an acute angle that occurred within fan f1 (**Figure 8c**). The distal portion of this distributary is discontinuous (**Figure 8b**) and terminates abruptly.

Apart from distributaries, the fans in Degana-A, particularly fans f1 and f3 (**Figure 8c,d**) exhibit surface textures with noticeable bedding/layered sediments. Fan f1 hosts many distinguishable stratigraphies (**Figure 8c,d**) observed at the distributary termini and along the intermediate region between the distributaries. **Figure 8d** also shows that the layering is evident along the flanks of the distributaries located on the fan f1. These layers are 100s of meters long, and the entire sequence is 10s of meters thick. Also, the source area of fan f3 on the western crater wall (**Figure 9**) shows well-exposed layers along the flanks of the distributary. The distributary flank hosts notable small boulders resolvable in HiRISE color images (**Figure 9**).

On observing fan f1 apex region, a significant part of the crater rim was breached and this breach is associated with five valleys, which originated from Degana's crater wall (**Figure 10a**). Over this part of the Degana-A crater rim, we observed that the rim height is ~800 m towards the northern side, whereas it is ~920 m towards the southern side. We observed ~800 m of the wall has been eroded and made as an inlet for the fan f1 deposits. Nearly all five valleys converge outside the eastern wall of Degana-A crater (**Figure 10a**) and breach this part to form the f1 fan deposits. **Figure 10b** shows one of the valleys from the Degana eastern wall with exposed layering in the HiRISE color image. This we observe as the earlier deposited sediments over the eastern floor of Degana were incised, possibly during the Degana-A rim breaching.

On observing the distal fan f2, we found that it has coalesced with the toe part of fan f1 (**Figure 11a**). Layers are observed at the toe region of both the fans f1 and f2, whereas at their coalescing

part, hummocky materials are observed (**Figure 11a**). The region of coalescing can be identified by a rough-textured surface with an absence of distributaries (**Figure 11a**). **Figure 11b** shows distributaries (marked by black arrow) and layering (marked by white arrow) on fan f2. Fan f3 also shows evidence of distributaries (**Figure 11b**) and layering that are distinguishable (**Figure 9**). All the four fans within Degana-A have coalesced at their toe region or by their sides.

We also observed fan deposits over the walls of Degana crater (f5 to f9), which mostly terminate at the rim of the Degana-A crater (**Figure 6b**). The slopes of these fans are in the range of 2° to 3°. Apparently, they have formed in the south and southeast of Degana-A and connected with the valleys presented on the wall of Degana (**Figure 6b**). Out of fans f5-f9, the smallest fan (f8) is ~4 km long and covers an area of ~10 km², while the largest fan (f6) is ~12 km long and covers an area of ~85 km². Fan f5 is the second-largest fan within Degana crater though it is formed by deposits from multiple valleys (**Table 2**).

Based on Blair and McPherson's (1994, 2009) definition, alluvial fans are semi-conical depositional landforms that develop through a channel, and branching spreads the sediments across the surface. In general, alluvial fans are hundreds of meters to few kilometers long along their major axis and fed by constrained catchments. This definition applies well to the observed fan deposits (**Figure 7**) inside Degana and Degana-A craters.

3.3 Ridges

In Degana-A crater, we observed arcuate or wide U-shaped ridges located over fans f2 and f3 (**Figure 13**). Photo-geological observations and profiles across these ridges indicate positive relief features with knobs (**Figures 13,14**). They were observed to drape over the eroded wall and fan of the Degana-A crater (**Figure 13**). These ridges are 10s of kilometers long and 100s of meters wide. Two ridges in the center over fan f2 have merged and formed a thick ridge towards the west (**Figure 13**). Further to the west, we observed a discontinuity along this merged ridge (**Figure 13**). The ridges in Degana-A appear more degraded towards the eastern side (**Figure 13**). This ridge is relatively thicker in its center and the thickness decreases towards both of its frontal lobes (**Figure 13**). We have also observed that fan f2 is oriented from northwest to southeast, whereas these ridges are oriented in a northeast-southwest direction (**Figure 13**). Similar, but only 5 km long, ridges are also observed on the alluvial fan f3 (**Figure 13**). Interestingly, the ridges superposed the fan deposits and are observed on the pole-facing wall of Degana crater. The ridges in Degana and

their characteristics such as convex geometry, an arcuate shape, and knobs are consistent with other moraines observed on Mars (Arfstrom and Hartmann, 2005; Shean, 2010; Scanlon et al., 2015).

We analyzed elevation profiles along north-south directions using the CTX DEM across the ridges located on fan f2 (no HiRISE DEM is available for this location). Four profiles were examined for which we determined the average thicknesses. The limited resolution of the CTX DEM does not allow extracting precise thicknesses for some of the ridges (Figure 14). These measured thicknesses vary from 8 m (BB') to 51 m (CC') (Figure 14, Table 4), and we have taken these values as a range for the possible ridge thickness. Further, we followed the approach by Hartmann et al. (2014), to estimate the ratio of basal stress (ζ) to density (ρ), which provides information about the nature of a glacier. To calculate this ratio, we used Paterson's (1994) relation, which suggests a relationship between basal stress and the density given as:

$$\zeta = \rho g h \tan \alpha \quad (7)$$

where g is the gravity (3.72 m/s^2), h is the height/thickness, and α is the slope in degrees (Paterson, 1994; Hartmann et al., 2014). The average and maximum thickness (h) measured for moraine-like ridges over fan f2 are estimated as $\sim 22 \text{ m}$ and 51 m , respectively. The average and the maximum values are used because this region could have undergone erosion after their formation. Slope (α) values were also determined, along with all the profiles that are considered for thickness measurements (Table 3). We obtained the average ratio of basal stress to density for Degana glacier(s) as $\sim 4 \text{ m}^2/\text{s}^2$ (Table 3).

3.4 Small polygonal patterns

Within the Degana crater and on the fan deposits, several polygonal-shaped features are observed (Figure 15). Polygonal patterns are observed, especially on the fans in Degana-A, and also on the inner and outer eastern wall of Degana-A crater (Figure 15). The majority of polygons' boundaries or fractures observed on the fans appear irregular (Figure 15 a-c), whereas polygons over the Degana-A walls visually appear regular or organized to each other (Figure 15 d-f). These polygons are 10s of meters wide and the fracture length varies accordingly. Figure 15b shows polygons over the fan f1, with fractures cutting through several layers. The polygonal fractures are well distinguishable over the fan f1 and within the vertically exposed stratigraphy (Figure 15b). We

have observed that over the fans (**Figure 15 a,b**) and the wall (**Figure 15 d-e**), most of the polygons are negative relief features. In contrast, **Figure 15c** shows a distinct style of polygons with filled ridges and positive reliefs in some areas on fan f1. Polygons in the wall of Degana-A (**Figure 15d**) are distinct from the polygons observed on the fans (**Figure 15a-c**) in terms of the slope at which they formed. Polygons shown in **Figure 15a-c** have formed on the relatively level floor ($<10^\circ$), whereas polygons shown in **Figure 15d** have formed on the more inclined wall ($>30^\circ$).

Overall, polygons on the fans suggest a potential involvement of volatiles potentially ice or adsorbed water in clay minerals, in their formation and, which would imply a periglacial environment postdating the deposition of the pile of sediments.

3.5 Crater size-frequency distributions

Xiao et al. (2012) estimated the age of the volcanic dome as ~ 4 Ga. In this study, we used a Poisson timing analysis (**Michael et al., 2016**) to estimate the absolute model ages. We mapped the best visible ejecta boundary of Degana crater and counted the superposed craters over the ejecta (**Figure 16a**). Our crater count statistics over the ejecta (area $\sim 6546 \text{ km}^2$) resulted in a total of 362 craters from ~ 100 m to 20 km. We obtained the best fit model age by fitting 17 craters, whose diameter is greater than 1.3 km (**Figure 16b**). From this, the age of Degana crater is ~ 3.7 Ga (early Hesperian) (**Michael, 2013**). Further, we analyzed the ejecta of Degana-A; however, it is very difficult to distinguish the ejecta, which is degraded by mass movement processes within the host crater Degana. The floor of Degana-A is almost entirely covered by sediment deposits (**Figure 6**) from all around the wall, occupying an area of $\sim 313 \text{ km}^2$. Hence, we counted all the possible superposed and embedded craters (**Figure S2**) over these alluvial fans (f1-f4) and determined cumulative (**Figure 16c**) and differential (**Figure 16d**) crater size-frequency distributions. The cumulative and differential fan-derived crater size-frequency distribution for Degana-A is early Amazonian (**Michael, 2013**). This age is not the formation age of the fan deposits within Degana-A; rather, it is estimated from the craters retained after their formation. Also, the diverse erosion of fan ridges observed (**Figure 11**) shows the difficulty in determining the model fan age. We want to clarify here that this age is the crater retention age for all the alluvial fans and can only suggest a possible lower bound to the deposits within Degana-A. In addition, we estimated cumulative (**Figure 16c**) and differential (**Figure 16d**) fan-derived crater size-frequency distributions for the four fans (f5 to f9) located over the walls of Degana crater. The obtained age is also early

Amazonian. This suggests that the fan deposits observed within both craters (Degana and Degana-A) are likely Amazonian.

4. Discussion

4.1 Evidence for exposure of pristine Noachian crust

The presence of mafic minerals in well-exposed outcrops within Degana crater indicates igneous lithology or basaltic/volcanic subsurface layers (e.g., Skok et al., 2012) exposed by the Degana and Degana-A impact. The following mechanisms have been proposed for the origin of olivine and LCP bearing rocks on Mars: Noachian crust/volcanism (Mustard et al., 2005; Hamilton and Christensen, 2005; Tornabene et al., 2008; Poulet et al., 2009a,b; Flahaut et al., 2012; Quantin et al., 2012; Skok et al., 2012; Rogers et al., 2018; Viviano-Beck et al., 2019) or impact melting from large impact basins (Mustard et al., 2007; Palumbo and Head, 2017). Noachian/early crust on Mars can contain evidence for its formation and crustal modification afterward (Skok et al., 2012). Mg-rich olivine got exposed on the wall/rim of Degana-A crater. McSween et al. (2006) suggested that Mg-enriched olivine on Mars likely indicates ancient primitive magmatic rock. Thus, the exposure of more mafic rocks in this region has implications for the magmatic history of the region. The exposed olivine and LCP in the uplifted rim of Degana-A provide a piece of direct evidence for the materials present in the interior of the volcanic dome. This strongly reveals that the materials exposed by the Degana outcrops represent either pristine Noachian crust and/or volcanic deposits sourced by a primitive mantle (McSween et al., 2006; Mustard et al., 2009; Skok et al., 2010; Flahaut et al., 2012; Quantin et al., 2012; Baratoux et al., 2013).

4.2 The fate of water at Degana crater

Earlier studies suggest regional glacial deposits in the southern highlands poleward of 30°S (Kargel and Strom, 1992; Head and Pratt, 2001; Ghatan and Head, 2002; Fastook et al., 2012; Scanlon et al., 2014; Head and Marchant, 2014; Fastook and Head, 2015; Kress and Head, 2015; Scanlon et al., 2015). While the glacial deposits around the Tharis volcanoes have got attention for the Amazonian equatorial ice (Sean et al., 2005; Forget et al., 2006), several craters located in low to mid-latitude (~5° to 60°) regions on Mars host young landforms on their floors and walls, which potentially indicate the involvement of ice-rich material with lobate

tongues and arcuate ridges (Arfstrom and Hartmann, 2005; Head et al., 2005; Shean, 2010; Scanlon et al., 2015; Sinha and Vijayan, 2017; Hartmann et al. 2002, 2003).

Thus like other craters, the local deposition of atmospheric ice occurred within the Degana. The moraines observed within Degana-A crater, which is situated at 23°S latitude, provide crucial evidence of equatorward glacial deposits during the Amazonian period. Global climate models for the equatorial highlands region of Mars show the possibility of snow accumulation and melting over the low and mid-latitudes, which would enable fan formation by snowmelt hypothesis (Wordsworth et al., 2013; Kite et al., 2013). However, there is no unambiguous observation of glacial deposits connected to the alluvial fans. While the interpreted glacial landforms postdate the alluvial fans at Degana, the fact that glacial ice was able to be deposited during the Amazonian shows that this location was once favorable to snow precipitation in the past. The average ratio of basal stress to density for Degana glacier(s) (i.e. $\sim 4 \text{ m}^2/\text{s}^2$) in comparison to terrestrial values of pure ice glaciers (40-100 m^2/s^2) is one order of magnitude lower, but not so much lower than values of 3.7 to 44.5 m^2/s^2 found by Hartmann et al. (2014) at Greg crater. These low values are likely explained by the fact that these moraines underestimate the actual thickness of the past glaciers, but are consistent with pure-ice glaciers, while rock glaciers would show higher values and a partial filling by the residual material (Hartmann et al., 2014).

Morgan et al. (2014) documented distributaries within Saheki crater' fans as observed on Degana' fans (Figure 7), which are downslope oriented and show radiating outward patterns from the fan apex. Many other studies (Moore and Howard, 2005; Irwin et al., 2018; Boatwright and Head, 2019) reported similar ridges over the fans within many martian craters, which were interpreted as inverted channels or fan distributary ridges. Morgan et al. (2014) documented layered sediments within the Saheki crater and interpreted that overbank flows, which might have been sourced from multiple flows within the distributaries, have deposited these layered sediments. Distributaries have spread across the fans in Degana-A crater (Figure 7), and they were the potential source for layered sediments, as suggested by Morgan et al. (2014). The flanks of the discontinuous distributaries (Figures 8d,9) with layers are records of possible long-duration flow events in the region. Additionally, Morgan et al. (2014) suggested that such layers were exposed by aeolian erosion. Exposures of layers on both sides of the distributary (Figure 9) indicate that erosion occurred possibly due to wind-driven activity. In the Saheki crater, a similar distributary with

exposed layers and bluish clasts in the HiRISE RGB-images suggested the presence of different materials (**Morgan et al., 2014**). Within fan f3 of Degana-A, the HiRISE image shows bluish clasts on the distributary (**Figure 9**) that are possibly comparable to Saheki crater ridge. However, this part of fan f3 lacks mineralogical data (CRISM) to assess the composition.

The rim breaching of Degana-A crater and fan f1 formation is closely associated with the valleys over the eastern wall of Degana crater. In addition, the valleys that incise the outer rim of Degana-A (black arrows in **Figure 10**) cut through a layered wall material on the eastern side of Degana (**Figure 10b**). This layered material might be residual deposits from an initial fan that formed on the outer rim of Degana-A, similar to fans f5 to f9. Then, the breach through the rim produced a change in the base level of the deposits, starting to build fan f1. In this scenario, the fact that a breach formed here rather than for the other fans f5 to f9 is likely due to the large watershed (**Figure 12**). An important implication of this scenario is that the fans f5-f9 would have formed coevally to the fans f1 to f4 inside Degana-A crater (see section 3.5). We also identified embedded craters on the fans (**Figure S2**), these possibly display evidence of interbedding within the crater (**Kite et al., 2017**). We use observations of fan f1 to estimate the flow velocity and discharge quantitatively (see **supplementary file**). Following **Morgan et al. (2014)**, we determined the value of maximum discharge as $\sim 84 \text{ m}^3/\text{s}$ (**Table S1**). Based on this, we also determined the supply rate of water, which is the maximum discharge per total area of source catchment, and found that the source catchment would have supplied the water at a rate of $\sim 2.25 \text{ mm/hr}$.

Studies on the deposition of large equatorial alluvial fans and deltas show the role of enhanced precipitation by rainfall or snow/ice melting followed by runoff (**Moore and Howard, 2005; Di Achille and Hynek, 2010a,b; Armitage et al., 2011; Grant and Wilson, 2011, 2012, Wilson et al., 2021**). Although many alluvial fans are evident within Degana (**Figure 6**), due to limited CRISM coverage, we cannot rule out hydrated mineral formation within the craters. The polygonal patterns display characteristics consistent with the patterns observed in aqueous sedimentary layers containing clay minerals that formed by contraction either from desiccation or from enhanced cold temperatures during their exhumation (**Elmaarry et al., 2012; Oehler et al., 2016**). The major hypotheses for polygon formation on Mars are thermal contraction and desiccation of sediments either enabled by ground ice or by sediments containing absorbed water such as in clay minerals (**Neal et al., 1968; Levy et al., 2009; Elmaarry et al., 2010, 2012; Oehler et al., 2016**). Although

the polygons were observed on alluvial fans, which are located within the volcanic dome, we do not interpret them as lava-related cooling cracks. Polygon patterns are generally defined by four factors: shape, size, boundary type, and intersection pattern (Yoshikawa, 2003). They can be characterized by an elevated ridge boundary or a depressed trough along the boundary. The pattern of intersection can be orthogonal, random orthogonal, or random (Yoshikawa, 2003; Mangold, 2005; Oehler et al., 2016). The rectangular shape and size of the polygon in our study area form an orthogonal pattern indicative of desiccation cracks (Neal et al., 1968; Elmaarry et al., 2010, 2012; Oehler et al., 2016) and/or thermal contracted sediments, i.e., potentially ice-wedge or sand-wedge polygons (Mangold, 2005; Hallet et al., 2013; Kerber et al., 2017; Brooker et al., 2018). Desiccation generally forms nested polygons that display many secondary fractures (Oehler et al., 2016). Therefore desiccation is less likely than thermal contracted ice-wedge polygons over potential periglacial terrain (Soare et al., 2014) formed by ice/snow deposition as indicated by the presence of moraines. Two types of ice-wedge polygons were suggested based on thermal stability: 1) a low centered polygon and 2) a high centered polygon (Washburn, 1980; Mackay, 2000). It was suggested that differences in their morphology potentially reflect a change in climatic conditions (Lachenbruch 1963; Mackay 2000; Jorgenson et al. 2006; Mellon et al. 2009; Abolt et al. 2018). In contrast to high-centered polygons, polygons in Figure 15c show ridges along the boundary, implying low-centered polygons. Those polygons do not have a regular shape and follow a more random geometry. They are consistent with networks of veins (also named box work deposits, Siebach and Grotzinger, 2014), which are formed by groundwater circulation and precipitations along the fractures. Their presence may imply a period of groundwater circulation during or after the alluvial fan formation period. Alternatively, another possible reason can be deposition and cementation of sand/dust in cracks formed by thermal contraction and further erosion of the material inside polygons (Oehler et al., 2016). On the other hand, the preservation of olivine-rich outcrops means that no prolonged aqueous activity took place at this location, or outcrops have been covered and therefore protected from alteration until relatively recently. Thus, the presence of alteration minerals cannot be conclusively demonstrated, but a well-developed chemical alteration seems unlikely. To decipher the fate of water-related deposits within Degana we are assuming that the crater was once covered by ice before the formation of Degana-A. With this assumption, three scenarios may explain our observations: (A) An episodic climatic warming, perhaps related to obliquity variations (e.g., Kite et al., 2017), (B) fluvial activity related to

volcanic heat, possibly related to Alba Patera or Ceraunius Tholus (e.g., **Gulick, 2001; Hauber et al., 2005; Ansan and Mangold, 2013**), or (C) the role of the Degana-A impact whose heat release may have helped to melt snow and trigger fluvial activity. Scenario (A) requires to have favorable climatic warming relatively recently in the Amazonian era. This is possible according to recent studies (e.g., **Kite et al., 2017**), but speculative, lacking other evidence for warming, at least regionally. For instance, the older crater located in the east of Degana (**Figure 3c**) shows only ancient erosion by valleys, but no pristine alluvial fans as reported in Degana, while its size and wall hillslopes look quite similar. This comparison suggests that the alluvial fans formed at Degana are local. Scenario (B) is unlikely because Degana-A is far younger than the volcanic activity, and no volcanic landforms are observed inside either Degana and Degana-A craters. Regarding hypothesis (C), it has been shown that hot ejecta emplaced over an ice-rich surface could have generated local melting (**Mangold, 2012; Mangold et al., 2012c; Weiss and Head, 2016**) possibly explaining some of the Amazonian age fluvial activity inside craters (e.g., **Williams et al., 2009**). Ejecta temperatures for craters of diameter range between 5 km to 150 km were determined by **Weiss and Head (2016)** to be up to 490K. Snow melting due to ejecta emplacement produced by the Degana-A impact would have deposited sediments for fan f1. We also determined maximum and minimum values for flow velocity and discharge rate for fan f1 in Degana crater. The maximum value for the flow velocity is ~2 m/s whereas the discharge rate is estimated as ~2.25 mm/hr. These values are lower than the estimated flow velocity and discharge rate by **Kite et al. (2013)** and **Morgan et al. (2014)**. These results support our interpretation regarding snow melting and the formation of fan f1. The breach of the ~1 km wide and ~900 m high eastern rim wall of Degana-A crater also indicates the long-lasting melting of snow/ice due to Degana-A impact. At least five valleys (**Figure 10a**) converge around the eastern wall region of Degana-A, which might have provided sufficient sediments to breach the crater rim. **Figure 12** shows the catchment area associated with these five valleys. Large catchment area (**Figure 12**) and positive relief of the dome lead us to envisage a high possibility of atmospheric snow/ice precipitation. This suggests that sufficient ice melt occurred to form the fan deposits within the craters. Thus, we interpret local atmospheric snow/ice precipitation as the most likely source for fluvial activity within the craters.

4.3 Geological history of the Degana crater region and its implications for the geological history of Mars

The overall geologic interpretation and the sequence of events are based on stratigraphic and chronological relationships. The sequence of geological events that occurred in the Degana region aids in deciphering the Martian surface processes, and climatic conditions that prevailed over this region. A schematic scenario is shown in **Figure 17**, which represents the activities that took place in this region.

(1) Apparent model ages derived from crater size-frequency measurements by **Xiao et al. (2012)** suggest the volcanic dome formed around 4 Ga ago, during the Early to Mid-Noachian epochs (**Figure 17a**).

(2) Degana impact crater formed on the volcanic dome and exposed the dome interior down to ~2 km. Apparent model ages suggest that Degana crater formed in the early Hesperian 3.7 Ga $\pm 0.04/-0.16$ or close to the Noachian-Hesperian boundary (**Figure 16b**). No volcanic-origin flows were observed within the crater. Therefore, we suggest that dome-related volcanic activity had ceased around 3.7 Ga.

(3) Degana-A crater superposed Degana crater after an unknown period. It is very difficult to estimate the age of Degana-A's formation because a lot of its ejecta were likely sequestered inside Degana crater. Degana-A further exposed the interior of the dome or the underlying Noachian crust by another ~0.7 km. In total, both craters excavated ~2.7 km deep into the volcanic dome. However, the floor of the crater is covered by four large depositional fans (**Figure 6**) whose tentative crater retention epoch is early Amazonian.

(4) Multiple valleys on the wall, and floor of Degana crater, and several fan deposits within Degana-A (**Figures 6b,18**) provide strong evidence for fluvial activity. Although the length of time over which the fans formed is uncertain, we evaluate two scenarios. The orthogonal coalesced alluvial fans (**Figure 6b**), inverted channels/distributaries, and the discharge rate for fan f1 suggest that the fans were formed by steady discharges over a very long period, possibly millions of years (**Moore et al., 2003**). We suggest that the widespread occurrence of fans within Degana required snow/ice precipitation (**Moore and Howard, 2005; Grant and Wilson, 2011, 2012**), followed by snowmelt within Degana to help the rim breaching of Degana-A. This suggests that sufficient ice deposition by mantling has occurred at least since the formation of the Degana crater. On Degana-A southern floor, possibly all the fans coalesced at their toe region and eroded (**Figures 6b, 18**). Thus, we propose that Degana-A was affected by snow/ice cover over Degana emplaced at some

point after its formation around ~3.7 Ga. We infer the ejecta from Degana-A played a role in melting the deposited snow/ice over the shattered dome, a scenario that had already been proposed for other locations on Mars (**Mangold, 2012**).

(5) At the last stage, during the Amazonian or post fan formation (**Kite et al., 2017**), moraine-like ridges are formed. The glacial activities in lower latitudes were reported to be more likely on the pole-facing walls of the craters (**Williams et al., 2009; de Haas et al., 2015**). The presence of moraine-like ridges on the pole facing northern wall of Degana (**Figures 13, 18**) provides comprehensive evidence for the glacial activity that occurred up to lower mid-latitude (**Shean, 2010**). Moraine-like ridges superposed over fans f2 and f3 (**Figure 13**) imply their post fan formation. This also suggests that the atmosphere-derived snow/ice was possible up to this latitude during the Amazonian period (**Figures 17, 18**).

Overall, the major implications from our study are that the craters excavated into a Noachian volcanic edifice and exposed pristine early Martian minerals, which is similar to earlier studies at different location of Mars (**Mustard et al., 2009; Skok et al., 2010; Quantin et al., 2012; Baratoux et al., 2013**). Degana crater also hosts significant evidence for post-Noachian fluvial activities. This means that they formed under climatic conditions more favorable to sustain runoff, which was capable to breach the Degana-A rim. Notably, no delta deposits like those formed in the Jezero crater (**e.g., Goudge et al., 2017**) were observed in the Degana. This implies that the fans in Degana-A were not deposited into a standing body of water (**Fassett and Head, 2008**). This indicates that the climatic conditions were suited for alluvial fan deposits only which might be due to limited liquid water availability due to low temperatures, lack of moisture, and/or insufficient atmospheric pressure. However, this climatic condition varied over the Hesperian-Amazonian. This suggests that during the Hesperian-Amazonian, the ice accumulation at this latitude considerably reduced, which enabled moraine-forming glacial activity during the Amazonian, but no more (glacio-)fluvial activity. Thus, this region is one of the locations on Mars, which witnessed the transition from fan-forming fluvial activity in the Hesperian to moraine-building glacial activity in the Amazonian.

Conclusions:

We carried out a detailed topographical, morphological, mineralogical, and chronological analysis of impact craters Degana and Degana-A. Our results reveal the diverse geologic activities, which are:

- Degana and Degana-A craters formed on a ~1.2 km high Noachian volcanic dome. Both impacts excavated up to ~2.7 km deep, i.e., ~1.5 km deep in the Noachian crust located below the base of the volcanic dome. These impacts exposed pristine Noachian mafic minerals (Mg-rich olivine and low-calcium pyroxene) on the walls of Degana-A crater. **Xiao et al., 2012** mentioned that the true composition of these ancient Noachian-aged volcanoes is not known, as most of these volcanoes are heavily dust-covered and no in-situ observations were taken yet. Thus, this study provides the likely composition of one of the ancient volcanoes on Mars.
- Both crater floors are covered by depositional fans, and especially Degana-A crater fans show distinguishable ridges, stratigraphic layers, and coalesced deposits. A notable ~1 km wide rim breach on the eastern rim of Degana-A crater and the hydrological estimation over this region suggest water flowed with a velocity of ~2 m/s and with a discharge rate of ~2.25 mm/hr. This is in agreement with **Kite et al., 2019** that reported a discharge rate of 3 mm/hr for globally present channels on Mars.
- Moraine-like ridges linked to glacial activities superposed the alluvial fans. Such moraine-like ridges are observed only on the pole-facing walls of Degana crater. The alluvial fan deposits and the glacial moraine ridges are non-coeval events. They strongly indicate evidence for snow/ice deposition towards the lower latitudes of Mars during Hesperian-Amazonian epochs suggesting local climate change on Mars. Thus, Degana is one of the important locations that report lower latitudes (23°S) glacial activities on Mars (**Shean, 2005**).
- Based on the dome elevation, the formation of fluvial/glacial deposits within the impact craters, and the latitudinal location, we suggest atmospheric snow/ice precipitation as the possible source.

Thus, Degana is a unique location on Mars that has mid-Noachian aged mafic minerals, Hesperian-Amazonian aged fluvial and glacial deposits.

Acknowledgment: We thank Hannes Bernhardt of Arizona State University and an anonymous reviewer for the constructive suggestions. We thank Physical Research Laboratory, Department of Space, India, for supporting our work. NM acknowledges the support from the Centre National d'Etudes Spatiales (CNES), France. We thank the CRISM, HiRISE and CTX team for the availability of the required datasets used in this study, which were downloaded from the PDS (<https://ode.rsl.wustl.edu/mars/>).

Figure captions:

Figure 1: Location map of Degana crater, Mars. Image is MOLA DEM. **a)** Topographic map of Degana crater and its surrounding region. **b)** Geologic map from Tanaka et al. (2014) indicates the terrain is Noachian in age.

Figure 2: MOLA DEM-derived topographic profiles **a)** profile is taken across the dome, which has a bell curve shape. This dome stands ~1.2 km high from its surroundings. **b)** Profile taken across Degana and Degana-A craters. Both impacts occurred on the dome and they have excavated ~1.5 km deep material from the base of the dome. (For profiles refer to **Figure 1a**)

Figure 3: Impact craters located on top of the volcanic dome. **a)** Degana crater, the largest on the dome. **b)** Degana-A crater, which superposed Degana crater. **c)** Unnamed crater- UA, with no distinguishable ejecta. **d)** Unnamed crater-UB located to the southeast of Degana. **e)** Unnamed crater-UC, with distinguishable raised rim and another smaller crater superposing its eastern wall. **f)** Unnamed crater-UD, located at the flank of the dome. The location of all the impact craters is marked in **Figure 1a**. (CTX mosaic by **Dickson et al. (2018)** is used and north is up in all the images)

671 **Figure 4:** a) MRO-CRISM derived summary parameters overlaid on CTX mosaic around the
672 eastern wall of Degana-A crater. Summary parameters are calculated for CRISM image
673 FRT00023F32. The summary parameter colors indicate: Red - Olivine, Cyan/green – LCP, Blue
674 – HCP (Viviano-Beck et al., 2014). b) CRISM ratioed spectra for low- Calcium pyroxene
675 identified with diagnostic 1 μm and 2 μm spectral features. A laboratory spectra from the RELAB
676 library is shown below for comparison. Locations P1 to P6 indicate from where the spectras are
677 extracted. c) CRISM ratioed spectra for olivine, identified with diagnostic 1 μm spectral features.
678 RELAB laboratory-derived spectrum for forsterite is given below for comparison. Locations O1
679 to O3 indicates from where the spectras are extracted.

680

681

682 **Figure 5:** Examples of locations in HiRISE false color image for **a)** pyroxene on the elevated rim,

683 **b)** pyroxene along topmost part of the eroded valley (HiRISE ESP_026522_1560). Example

684 location for **c)** olivine detected along the eastern wall with light-toned exposure and d) olivine

685 signature along the eastern wall (HiRISE ESP_16540_1560).

Figure 6: **a)** CTX mosaic of Degana and superposed Degana-A crater. **b)** Geomorphological map of Degana and Degana-A craters overlaid at 80% transparency on the CTX mosaic. The predominant mapped units are fan deposits, valleys over the crater rim, and moraine-like ridges.

Figure 7: Distribution of fan deposits within Degana-A crater, **a)** Fan f1, located on the eastern wall, whose apex region is situated on the breached rim. Individual ridges and dendritic distributaries are present over the fan surface. **b)** Fan f2, from the northern wall. **c)** Fan f3 from the northwestern side, with distinguishable ridges at its toe side. **d)** Fan f4 from the southwest side, with deposits and distinguishable ridges. Distribution of fan deposits within Degana crater, **e)** Fan f5, the largest fan within both craters. The toe region of this fan merges with another adjacent small fan deposit. **f)** Fan f6, from the southeastern wall with broad fan-shaped deposits at the toe region. The toe region of the fan deposits terminated at the walls of Degana-A crater. Colored lines over the fans indicate the traverses of topographic profile **Figure S3**, White arrows - distinguishable distributaries over the fans. (CTX IDs **a-d:** G02_018874_1561_XN_23S045W, and **e-f:** D19_034592_1579_XN_22S045W)

Figure 8: **a)** HiRISE stereo pair (ESP_016540_1560 and ESP_016039_1560) derived DEM image for fan f1. **b)** Elevation profile (AA') taken along a discontinuous distributary ridge, note the height variation within the ridge **c)** HiRISE false-color image shows some parts of the discontinuous ridge. At the downslope side, crosscutting ridges are observed with exposed stratigraphic deposits (HiRISE ID: ESP_016540_1560). This false-color image also highlights the multiple distinguishable layers within the fan deposits. The black arrows display the locations marked on the elevation profile in **Figure 7b**. **d)** HiRISE false-color image (ESP_016540_1560) over the toe region of fan f1. The eroded ridges expose multiple stratigraphic layers located at different elevations (marked by white arrow). These layers likely hint at multiple stages of deposition within Degana-A crater. (White dot shows elevation value at that particular location.)

Figure 9: HiRISE color image ESP_018874_1560 reveals details of ridges with interbedded fine sediments and exposed stratigraphy. The layers are exposed all around the ridge and they are individually distinguishable. The ridge is surrounded by aeolian deposits, TARs, and mantled by dust showcasing the role of wind in eroding the ridge and exposing layers. Towards the toe side of the ridge, it is narrow and exposed more distinct light-toned layers, which are ~3-5 m thick.

Figure 10: **a)** The apex region of fan f1 located over the breached rim region of Degana-A. Towards the north, the rim height is ~800 m, whereas, towards the south, it is much higher at ~920 m. This suggests that the ~900 m high and ~ 1 km wide rim is breached by fan f1 deposits. Over this part, nearly five valleys from the Degana crater wall converge, which suggests the possible source for the rim breaching. Black arrows display the valleys, which fed fan f1. **b)** HiRISE color image ESP_026522_1560_COLOR of one of the eroded walls of the valley shows distinguishable layers. We interpret this exposed stratigraphy as earlier deposits within the Degana eastern wall, which later got breached by the valley associated with the rim breaching.

Figure 11: Example locations of intersecting fan deposits within Degana-A crater. **a)** Region of convergence of fan f1 and f2, with marked distributaries (black arrow, aligned in the direction of the fan). Layers were identified at the front scarp of both the fans. The hummocky patch located at the termini region of both the fans (f1 and f2) is ~20 m elevated. On both sides of this hummocky surface, the fan deposits expose the layers at their front scarp (HiRISE ID ESP_016039_1560 and ESP_035159_1560). **b)** Region of convergence between fan f2 and f3, with marked distributaries (black arrows, aligned in the direction of the fan). Layers were identified at the termini region of fan f2. The dust mantled the trough in between the fan deposits. The orientation of both fan deposits are different, which crosscut/superpose each other nearly orthogonal (HiRISE ID ESP_018874_1560).

Figure 12: MGS MOLA- MEX HRSC blended DEM overlaid on the CTX image in the eastern side of Degana crater. This part is the possible catchment area for the five valleys, which played a role in breaching the Degana-A eastern rim. The elevation of the catchment area ranges from -511 to 2087 m. The boundary of fan f1 is marked with a dotted gray line. The estimated catchment area is larger in comparison to the area of fan f1.

Figure 13: **a)** CTX image (G02_018874_1561_XN_23S045W) of the northern wall of Degana crater showing moraine-like ridges (marked with black arrow). These moraine-like ridges occur on the pole-facing wall. These ridges superposed over two fan deposits (f2 and f3). **b)** four distinguishable moraine-like ridges formed on fan f2. The thickness and the relative height of the ridges vary.

Figure 14: Example elevation profiles used to determine the thickness of moraine-like ridges extracted from CTX DEM (stereo-pair CTX_016540_1579_034592_1579). Red is the location of ridges. The height of some of the ridges is unresolvable in the CTX DEM.

Figure 15: Examples of polygons within the Degana-A fan deposits. **a)** Trough boundary type polygons on fan f1. **b)** Polygons formed over the layered ridges. The size and extent of the polygons vary around the ridge. **c)** Ridge boundary type polygon with hollows at the center and raised ridges at the boundaries. Examples of polygons within the walls of Degana crater. **d)** Polygons over the raised northern outer wall (OW) of Degana-A crater with varying sizes. **e)** Polygons with dust mantling over the OW of Degana-A crater. **f)** Polygons on the inner walls (IW) of Degana crater. All the polygons on the walls are observed along the topmost part of the rim only. (HiRISE ID ESP_016540_1560).

Figure 16: a) Degana crater and its ejecta boundary with superposed craters considered for crater chronology. Counts (red circles) were conducted using CTX images. b) Relative ages were determined using the production function of **Ivanov (2001)** and chronology function of **Hartmann and Neukum (2001)**. The apparent model age of Degana crater is ~ 3.7 Gyr. c) Cumulative crater size-frequency distribution for all the fans within both craters. d) Differential crater size-frequency distribution for all the fans within both craters.

Figure 17: Schematic perspective outlining the tentative sequence of events. **a)** Volcanic dome formation at ~4.0 Ga with an elevation of ~1.2 km above the surrounding surface (baseline of the dome). **b)** Formation of Degana crater at ~3.7 Ga excavating ~2 km of the volcanic dome. **c)** Formation of Degana-A in Degana crater and additional ~0.7 km of excavation. In total, both the craters excavate ~2.7 km. The crater size-frequency distribution for all the fans suggests an Amazonian crater retention age.

Figure 18: Schematic diagram of the current state of Degana and Degana-A. Locations of mafic minerals, fluvial and glacial landforms are marked within the exposed surface.

Supplementary Figure S1: MGM map prepared using CRISM image FRT00023F32. The scale has been adjusted to convey the differences. The MGM map shows that the region is enriched in LCP.

Supplementary Figure S2: Example of embedded craters identified on Degana-A crater fan deposits (HiRISE ID: ESP_035159_1560 (a), ESP_018874_1560 (b,d), ESP_016540_1560 (c)).

Supplementary Figure S3: CTX DEM-derived elevation profiles taken along the major axis of four fans (f1-f4, **Figure 7a-d**) located within Degana-A crater.

Tables:

Table 1: Details of center pixel and the number of pixels used for the spectra shown in figure 4b,c.

Labels	Center Pixel/Pixels	Number of Pixels
Figure 4b		
P1 (numerator); (denominator)	X:551, Y:344; X:551, Y:301	5*5
P2 (numerator); (denominator)	X:370, Y:358; X:370, Y:149	3*3
P3 (numerator); (denominator)	X:575, Y:338; X:575, Y:282	5*5
P4 (numerator); (denominator)	X:323, Y:283; X:323, Y:96	3*2
P5 (numerator); (denominator)	X:581, Y:101; X:581, Y:83	3*3
P6 (numerator); (denominator)	X:524, Y:410; X:524, Y:100	5*5

Figure 4c		
O1 (numerator); (denominator)	X:615, Y:117; X:615, Y:10	6*3
O2 (numerator); (denominator)	X:547, Y:99; X:547, Y:8	5*5
O3 (numerator); (denominator)	X:581, Y:42; X:581, Y:12	9*7

Table 2: Details of all the fans mapped within Degana crater with length, area, and slope.

Fan Name	Approx. Fan length (along major axis) (in km)	Approx. Area (in km ²)	Slope (degrees)
f1	13.39	70.06	2.62
f2	9.09	60.36	2.47
f3	10.21	76.00	2.71
f4	10.10	52.11	3.81
f5	11.10	79.22	2.32
f6	11.98	85.05	2.50
f7	5.67	25.65	2.98
f8	4.22	10.11	2.75
f9	7.63	27.44	2.59

Table 3: Geometric characteristics of moraine-like ridges on the northern wall of Degana-A.

Profile ID	Maximum height of moraine-like ridges (m)	Slope (in degrees) α	Basal stress/density (m ² /s ²)
AA'	20	3.49	4.5
BB'	8	3.0	1.6
CC'	51	2.74	9.1
DD'	8	2.24	1.2
Average	22		4.1

References

812 Abolt, C. J., Young, M. H., Atchley, A. L., and Harp, D. R., 2018. Microtopographic control on
813 the ground thermal regime in ice-wedge polygons. *The Cryosphere*, 12, 1957–1968,
814 <https://doi.org/10.5194/tc-12-1957-2018>.

815 Adams, J. B., 1974. Visible and near-infrared diffuse reflectance: Spectra of pyroxenes as applied
816 to remote sensing of solid objects in the Solar System. *J. Geophys. Res.*, 79, 4829–4836,
817 DOI:10.1029/JB079i032p04829.

818 Ansan, V., and Mangold, N., 2013. 3D morphometry of valley networks on Mars from
819 HRSC/MEX DEMs: Implications for climatic evolution through time. *Journal of Geophysical*
820 *Research: Planets*, 118, 1873–1894. <https://doi.org/10.1002/jgre.20117>.

821 Arfstrom, J., and Hartmann, W. K., 2005. Martian flow features, moraine-like ridges, and gullies:
822 Terrestrial analogs and interrelationships. *Icarus*, 174 (2), 321– 335.
823 <https://doi.org/10.1016/j.icarus.2004.05.026>

824 Armitage, J. J., Warner, N. H., Goddard, K., and Gupta, S., 2011. Timescales of alluvial fan
825 development by precipitation on Mars. *Geophys. Res. Lett.*, 38, L17203,
826 DOI:10.1029/2011GL048907

827 Baratoux, D., Toplis, M. J., Monnereau, M., and Sautter, V., 2013. The petrological expression of
828 early Mars volcanism. *J. Geophys. Res. Planets*, 118, 59–64, DOI:10.1029/2012JE004234.

829 Berman, D. C., Balme, M. R., Rafkin, S. C. R., and Zimbelman, J.R., 2011. Transverse Aeolian
830 Ridges (TARs) on Mars II: Distributions, orientations, and ages. *Icarus*, 213, 116– 130,
831 doi:10.1016/j.icarus.2011.02.014.

832 Blair, T.C. and McPherson, J.G., 1994. Alluvial fans and their natural distinction from rivers based
833 on morphology, hydraulic processes, sedimentary processes, and facies assemblages. *Journal of*
834 *sedimentary research*, 64(3a), pp. 450-489.

835 Blair, T.C., and McPherson, J.G., 2009. Processes and forms of alluvial fans. In: Parsons, A.J.,
836 Abrahams, A.D. (Eds.), *Geomorphology of Desert Environments*. Springer Science+Business
837 Media B.V., pp. 413–466.

838 Boatwright, B. D., & Head, J. W. 2019. Fluvial Geology of the Northwest Hellas Region, Mars:
 839 Evidence for Localized Drainage and Terrain Inversion. In *Lunar and Planetary Science*
 840 *Conference* (p. 2688).

841 Brooker, L. M. et al., 2018. Clastic Polygonal Networks Around Lyot Crater, Mars: Possible
 842 Formation Mechanisms from Morphometric Analysis. *Icarus*, 302, 386 – 406.

843 Bristow, T. F., Rampe, E. B., Achilles, C. N., Blake, D. F., Chipera, S. J., Craig, P., et al., 2018.
 844 Clay mineral diversity and abundance in sedimentary rocks of Gale crater, Mars. *Science*
 845 *Advances*, 4(6), eaar3330. <https://doi.org/10.1126/sciadv.aar3330>.

846 Burns, R. G., 1970. Crystal field spectra and evidence of cation ordering in olivine minerals.
 847 *American Mineralogist*, 55, 1608–1632.

848 Carter, J. et al., 2013. Hydrous minerals on Mars as seen by the CRISM and OMEGA imaging
 849 spectrometers: Updated global view. *J. Geophys. Res. (Planets)*, 118 (2013), pp. 831-858

850 Clenet, H., Pinet, P. C., Daydou, Y., Heuripeau, F., Rosemberg, C., Baratoux, D., and Chevrel,
 851 S., 2011. A new systematic approach using the Modified Gaussian Model: Insight for the
 852 characterization of chemical composition of olivines, pyroxenes, and olivine–pyroxene mixtures.
 853 *Icarus*, 213, 404– 422, DOI:10.1016/j.icarus.2011.03.002.

854 Cloutis, E. A., Galley, M. J., Jackowski, T. L., and Reed, R. L., 1986. Calibrations of phase
 855 abundance, composition, and particle size distributions for olivine-orthopyroxene mixtures from
 856 reflectance spectra. *J. Geophys. Res.*, 91, 11,641–11,653, DOI:10.1029/JB091iB11p11641.

857 Cloutis, E. A., and Gaffey, M. J., 1991. Pyroxene spectroscopy revisited: Spectral composition
 858 correlations and relationship to geothermometry. *J. Geophys. Res.*, 96, 22,809–22,826,
 859 DOI:10.1029/91JE02512.

860 Costa, J.E., 1988. Rheologic, geomorphic, and sedimentologic differentiation of water floods,
 861 hyper-concentrated flows, and debris flows. In: Baker, V.R. et al. (Eds.), *Flood Geomorphology*.
 862 Wiley-Interscience, pp. 113–122.

863 Crosta, G.B. and Frattini, P., 2004. Controls on modern alluvial fan processes in the central Alps,
 864 northern Italy. *Earth Surface Processes and Landforms*, 29, 267–293.

865 Davies, T.R. and McSaveney, M.J., 2008. Principles of sustainable development on fans. *Journal*
866 *of Hydrology (New Zealand)*, 47, 43–65.

867 de Haas, T., Hauber, E., Conway, S.J., van Steijn, H., Johnsson, A., Kleinhans, M.G., 2015. Earth-
868 like aqueous debris-flow activity on Mars at high orbital obliquity in the last million years. *Nat.*
869 *Commun.* 6. DOI:10.1038/ncomms.

870 De Scally, F.A. and Owens, I.F., 2004. Morphometric controls and geomorphic responses on fans
871 in the Southern Alps, New Zealand. *Earth Surface Processes and Landforms*, 29, 311–322.

872 Di Achille, G., and Hynek, B. M., 2010a. Deltas and valley networks on Mars: Implications for a
873 global hydrosphere, in *Lakes on Mars*, edited by N. Cabrol and E. Grin. Chap. 8, Elsevier, New
874 York.

875 Di Achille, G., and Hynek, B.M., 2010b. Ancient ocean on Mars supported by global distribution
876 of deltas and valleys. *Nat. Geosci.*, 3, 459– 463, doi:10.1038/NGEO891.

877 Dickson, J.L., Head, J.W., and Fassett, C.I., 2012. Patterns of accumulation and flow of ice in the
878 mid-latitudes of Mars during the Amazonian. *Icarus*, 219, pp. 723-732,
879 10.1016/j.icarus.2012.03.010.

880 Dickson, J.L., Kerber, L.A., Fassett, C.I., and Ehlmann, B.L., 2018. A global, blended CTX mosaic
881 of Mars with vectorized seam mapping: A new mosaicking pipeline using principles of non-
882 destructive image editing. *Proc. Lunar Planet. Sci. Conf.* 49th, abstract#2480.

883 Ehlmann, B. L., et al., 2009. Identification of hydrated silicate minerals on Mars using MRO-
884 CRISM: Geologic context near Nili Fossae and implications for aqueous alteration. *J. Geophys.*
885 *Res.*, 114, E00D08, DOI:10.1029/2009JE003339.

886 Elmaarry, M.R., Markiewicz, W.J., Mellon, M.T., et al., 2010. Crater floor polygons: Desiccation
887 patterns of ancient lakes on Mars? *J. Geophys. Res.-Planets* 115, E10 0 06. DOI: 10.1029/2010JE0
888 03609.

889 Elmaarry, M.R., Kodikara, J., Wijessoriya, S., Markiewicz, W.J., Thomas, N., 2012. Desiccation
890 mechanism for formation of giant polygons on Earth and intermediate-sized polygons on Mars:
891 Results from a pre-fracture model. *Earth Planet. Sci. Lett.* 323–324, 19–26.

892 Fassett, C.I., and Head, J.W., 2008. The timing of martian valley network activity: constraints from
893 buffered crater counting. *Icarus*, 195, pp. 61-89, 10.1016/j.icarus.2007.12.009

894 Fastook, J. L., Head, J. W., Marchant, D. R., Forget, F., and Madeleine, J.-B., 2012. Early Mars
895 climate near the Noachian–Hesperian boundary: Independent evidence for cold conditions from
896 basal melting of the south polar ice sheet (Dorsa Argentea Formation) and implications for valley
897 network formation. *Icarus*, 219(1), 25– 40. <https://doi.org/10.1016/j.icarus.2012.02.013>.

898 Fastook, J.L., and Head, J.W., 2015. Glaciation in the Late Noachian Icy Highlands: Ice
899 accumulation, distribution, flow rates, basal melting, and top-down melting rate sand patterns.
900 *Planetary and Space Science*, 106, 82-98.

901 Ferguson, R., 2007. Flow resistance equations for gravel- and boulder-bed streams. *Water Resour.*
902 *Res.* 43, W05427. <http://dx.doi.org/10.1029/2006WR005422>

903 Ferguson, R. L, Hare, T. M., Laura, J., 2018. HRSC and MOLA Blended Digital Elevation Model
904 at 200 m v2, Astrogeology PDS Annex, U.S. Geological Survey.
905 http://bit.ly/HRSC_MOLA_Blend_v0.

906 Flahaut, J., Mustard, J.F., Quantin, C., Clenet, H., Allemand, P., and Thomas, P., 2010. Dikes of
907 distinct composition intruded into Noachian-aged crust exposed in the walls of Valles Marineris.
908 *Geophysical Research Letters*, 38, 15.

909 Flahaut, J., Quantin, C., Clenet, H., Allemand, P., Mustard, J.F., Thomas, P., 2012. Pristine
910 Noachian crust and key geologic transitions in the lower walls of Valles Marineris: Insights into
911 early igneous processes on Mars. *Icarus*, 221, 420-435.

912 Forget, F., Haberle, R. M., Montmessin, F., Levrard, B., and Head, J. W., 2006. Formation of
913 glaciers on Mars by atmospheric precipitation at high obliquity. *Science*, 311(5759), 368– 371.

914 Ghail, R.C., and Hutchison, J.E., 2003. An alluvial fan at Apollinaris Patera, Mars. LPSC 2003.
915 #Abstract 1775.

916 Ghatan, G.J., and Head, J.W., 2002. Candidate subglacial volcanoes in the south polar region of
917 Mars: Morphology, morphometry, and eruption conditions. *Journal of Geophysical Research*, 107,
918 E7.

919 Goldspiel, J.M., and Squyres, S.W., 2011. Groundwater discharge and gully formation on martian
920 slopes. *Icarus*, 211, 238-258.

921 Grant, J. A., and Wilson, S. A., 2011. Late alluvial fan formation in southern Margaritifer Terra,
922 Mars. *Geophys. Res. Lett.*, 38, L08201, DOI:10.1029/2011GL046844.

923 Grant, J. A., and Wilson, S. A., 2012. A possible synoptic source of water for alluvial fan formation
924 in Southern Margaritifer Terra, Mars. *Planet. Space Sci.*, 72(1), 44– 52.

925 Gulick, V. C., 2001. Origin of the valley networks on Mars: A hydrological perspective,
926 *Geomorphology*, 37, 241– 268, doi:10.1016/S0169-555X(00)00086-6.

927 Hallet, B., Sletten, R.S., Stewart, W., et al., 2013. Fracture Networks, Gale Crater, Mars. In:
928 *Proceedings of 44th LPSC. #Abstract No. 3108.*

929 Hamilton, V. E., and Christensen, P. R., 2005. Evidence for extensive, olivine-rich bedrock on
930 Mars. *Geology*, 33(6), 433–436. <https://doi.org/10.1130/G21258.1>

931 Hartmann, W. K., and Neukum, G., 2001. Cratering chronology and the evolution of Mars. *Space*
932 *Sci. Rev.*, 96, 165–194, DOI:10.1023/A:1011945222010.

933 Hartmann, W.K., Thorsteinsson, T., Sigurdsson, F., 2002. Comparison of Icelandic and martian
934 hillside gullies. *Lunar and Planet. Sci. XLIII. Abstract #1904.*

935 Hartmann, W.K., Thorsteinsson, T., Sigurdsson, F., 2003. Martian hillside gullies and Icelandic
936 analogs. *Icarus* 162, 259–277.

937 Hartmann, W. K., Ansan, V., Berman, D. C., Mangold, N., and Forget, F., 2014. Comprehensive
938 analysis of glaciated Martian crater Greg. *Icarus*, 228, 96– 120.
939 [https://doi.org/10.1016/j.icarus.2013.09.016.](https://doi.org/10.1016/j.icarus.2013.09.016)

940 Hauber, E., Gasselt, S.V., Ivanov, B., Werner, S., Head, J.W., Neukum, G., Jaumann, R., Greeley,
941 R., Mitchell, L. K., Muller, P., and The HRSC Co-Investigator Team, 2005. Discovery of a flank
942 caldera and very young glacial activity at Hecatus Tholus, Mars. *Nature* 434, 356-361.
943 <https://doi.org/10.1038/nature03423>

944 Head, J. W., & Pratt, S., 2001. Extensive Hesperian-aged south polar ice sheet on Mars: Evidence
 945 for massive melting and retreat, and lateral flow and ponding of meltwater. *Journal of Geophysical*
 946 *Research*, 106(E6), 12275– 12299. <https://doi.org/10.1029/2000JE001359>.

947 Head, J. et al., 2005. Tropical to mid-latitude snow and ice accumulation, flow, and glaciation on
 948 Mars. *Nature* 434, 346–351.

949 Head, J.W. and Merchant, D.R., 2014. The climate history of early Mars: insights from the
 950 Antarctic McMurdo Dry Valleys hydrologic system. *Antarctic Science* 26(6), 774–800.

951 Hoke, M. R. T., and Hynek, B. M., 2009. Roaming zones of precipitation on ancient Mars as
 952 recorded in valley networks, *J. Geophys. Res.*, 114, E08002, doi:10.1029/2008JE003247.

953 Huang, J., Edwards, C. S., Horgan, B. H. N., Christensen, P. R., Kraft, M. D., and Xiao, L., 2012.
 954 Identification and mapping of dikes with relatively primitive compositions in Thaumasia Planum
 955 on Mars: Implications for Tharsis volcanism and the opening of Valles Marineris. *Geophys. Res.*
 956 *Lett.*, 39, L17201, DOI:10.1029/2012GL052523.

957 Irwin, R. P., Howard, A. D., Craddock, R. A., Moore, and J. M., 2005. An intense terminal epoch
 958 of widespread fluvial activity on early Mars: 2. Increased runoff and paleolake development, *J.*
 959 *Geophys. Res.*, 110, E12S15, doi:10.1029/2005JE002460.

960 Irwin, R. P., Wray, J. J., Mest, S. C., & Maxwell, T. A. (2018). Wind-Eroded Crater Floors and
 961 Intercrater Plains, Terra Sabaea, Mars. *Journal of Geophysical Research: Planets*, 123(2), 445–
 962 467. <https://doi.org/10.1002/2017JE005270>

963 Ivanov, B. A., 2001. Mars/Moon cratering rate ratio estimates. *Space Sci. Rev.*, 96, 87–104,
 964 doi:10.1023/A:1011941121102.

965 Jorgenson, M. T., Shur, Y. L., and Pullman, E. R., 2006. Abrupt increase in permafrost degradation
 966 in Arctic Alaska. *Geophys. Res. Lett.*, 33, L02503, DOI:10.1029/ 2005GL024960.

967 Kargel, J.S., and Strom, R.G., 1992. Ancient glaciation on Mars. *Geology*, 20 (1), 3-7.

968 Kerber, L., Dickson, J.L., Head, J.W., Grosfils, E.B., 2017. Polygonal ridge networks on Mars:
 969 Diversity of morphologies and the special case of the Eastern Medusae Fossae Formation. *Icarus*
 970 281, 200–219.<https://doi.org/10.1016/j.icarus.2016.08.020>.

971 King, T. V. V., and Ridley, W. I., 1987. Relation of the spectroscopic reflectance of olivine to
 972 mineral chemistry and some remote sensing implications, *J. Geophys. Res.*, 92, 11,457–11,469,
 973 DOI:10.1029/JB092iB11p11457.

974 Kite, E. S., Harvey, I., Kahre, M. A., Wolff, M. J., and Manga, M., 2013. Seasonal melting and
 975 the formation of sedimentary rocks on Mars, with predictions for the Gale crater mound. *Icarus*,
 976 223, 131–210, DOI:10.1016/j.icarus.2012.11.034.

977 Kite, E. S., Sneed, J., Mayer, D. P., and Wilson, S. A., 2017. Persistent or repeated surface
 978 habitability on Mars during the late Hesperian – Amazonian. *Geophys. Res. Lett.*, 44, 3991–3999.
 979 DOI:10.1002/2017GL072660.

980 Kleinhans, M.G., 2005. Flow discharge and sediment transport models for estimating a minimum
 981 timescale of hydrological activity and channel and delta formation on Mars. *J. Geophys. Res.* 110,
 982 E12003. [http://dx.doi.org/ 10.1029/2005JE002521](http://dx.doi.org/10.1029/2005JE002521).

983 Klima, R. L., Pieers, C. M., and Dyar, M. D., 2007. Spectroscopy of synthetic Mg-Fe pyroxenes
 984 I: Spin-allowed and spin-forbidden crystal field bands in the visible and near-infrared. *Meteorit.*
 985 *Planet. Sci*, 42, 235–253, DOI:10.1111/j.1945-5100.2007.tb00230.x.

986 Kostaschuk, R.A., MacDonald, G.M. and Putnam, P.E., 1986. Depositional process and alluvial
 987 fan-drainage basin morphometric relationships near Banff, Alberta, Canada. *Earth Surface*
 988 *Processes and Landforms*, 11, 471–484.

989 Kraal, E. R., Asphaug, E., Moore, J. M., Howard, A., and Brecht, A., 2008. Catalogue of large
 990 alluvial fans in Martian impact craters. *Icarus*, 194, 101–110, DOI:10.1016/j.icarus.2007.09.028.

991 Kress, A. M., and Head, J. W., 2015. Late Noachian and early Hesperian ridge systems in the south
 992 circumpolar Dorsa Argentea Formation, Mars: Evidence for two stages of melting of an extensive
 993 late Noachian ice sheet. *Planetary and Space Science*, 109-110, 1–20.
 994 <https://doi.org/10.1016/j.pss.2014.11.025>

995 Lachenbruch, A.H., 1963. Contraction theory of ice-wedge polygons: A qualitative discussion. In:
 996 *Proceedings of International Permafrost Conference, Proceedings, Lafayette, Indiana. National*
 997 *Academy of Sciences, National Research Council Publication No. 1287, pp. 63–71.*

998 Lamb, M.P., Dietrich, W.E., Sklar, L.S., 2008. A model for fluvial bedrock incision by impacting
 999 suspended and bedload sediment. *J. Geophys. Res.* 113, F03025.
 1000 <http://dx.doi.org/10.1029/2007JF000915>.

1001 Levy, J.S., Head, J., Marchant, D., 2009. Thermal contraction crack polygons on Mars:
 1002 Classification, distribution, and climate implications from HiRISE observations. *J. Geophys. Res.*
 1003 114. doi: 10.1029/2008JE003273.

1004 Mackay, J. R., 2000. Thermally induced movements in ice-wedge polygons, western arctic coast:
 1005 A long-term study. *Geogr. Phys. Quaternaire*, 54,41–68.

1006 Malin, M.C., and Edgett, K.S., 2000. Sedimentary rocks of early Mars. *Science*, 290, 1927-1937.

1007 Malin, M. C., et al., 2007. Context Camera investigation onboard the Mars Reconnaissance
 1008 Orbiter. *J. Geophys. Res.*, 112, E05S04, doi:10.1029/2006JE002808.

1009 Mangold, N., 2005. High latitude patterned grounds on Mars: Classification, distribution, and
 1010 climatic control. *Icarus* 174, 336–359.

1011 Mangold, N., Adeli, S., Conway, S., Ansan, V., Langlais, B., 2012a. A chronology of early Mars
 1012 climatic evolution from impact crater degradation. *J. Geophys. Res.-Planets*, 117, E4,
 1013 doi:10.1029/2011JE004005.

1014 Mangold, N., Carter, J., Poulet, F., Dehouck, E., Ansan, V., Loizeau, D., 2012b. Late Hesperian
 1015 aqueous alteration at Majuro crater, Mars. *Planetary and Space Science*, 72, 18-30.

1016 Mangold, N., 2012. Fluvial landforms on fresh impact ejecta on Mars, *Planetary and Space*
 1017 *Science*, 62, 69-85.

1018 Mangold N., Kite, E.S., Kleinhans, M.G., Newsom, H., Ansan, V., Hauber, E., Kraal, E., Quantin,
 1019 C., Tanaka, K., 2012c. The origin and timing of fluvial activity at Eberswalde crater, Mars. *Icarus*,
 1020 220, 530-551.

1021 Mattern, E., Matas, J., Ricard, Y., Bass, J., 2005. Lower mantle composition and temperature from
 1022 mineral physics and thermodynamic modeling. *Geophys.J. Int.* 160, 973–990.

1023 McCauley, J.F., 1978. Geologic map of the Coprates quadrangle of Mars. DOI 10.3133/i897.

1024 McEwen, A. S., et al., 2007. Mars Reconnaissance Orbiter's High-Resolution Imaging Science
 1025 Experiment (HiRISE). *J. Geophys. Res.*, 112, E05S02, DOI:10.1029/2005JE002605

1026 McSween, H. Y., Jr., et al., 1999. Chemical, multispectral, and textural constraints on the
 1027 composition and origin of rocks at the Mars Pathfinder landing site. *J. Geophys. Res.*, 104, 8679–
 1028 8715, DOI:10.1029/98JE02551.

1029 McSween, H. Y., et al., 2006. Characterization and petrologic interpretation of olivine-rich basalts
 1030 at Gusev Crater, Mars. *J. Geophys. Res.*, 111, E02S10, DOI:10.1029/2005JE002477.

1031 Melton, M.A., 1965. The geomorphic and paleoclimatic significance of alluvial deposits in
 1032 southern Arizona. *Journal of Geology*, 73, 1–38.

1033 Mellon, M. T., et al., 2009. Ground ice at the Phoenix Landing Site: Stability state and origin. *J.*
 1034 *Geophys. Res.*, 114,E00E07, DOI:10.1029/2009JE003417.

1035 Michael, G.G., 2013. Planetary surface dating from crater size-frequency distribution
 1036 measurements: Multiple resurfacing episodes and differential isochron fitting. *Icarus*, 226, 885–
 1037 890.

1038 Michael, G.G., Kneissl, T., Neesemann, A., 2016. Planetary surface dating from crater size-
 1039 frequency distribution measurements: Poisson timing analysis. *Icarus*, 227, 279-285.

1040 Milton, D. J. (1973), Water and processes of degradation in the Martian landscape, *J. Geophys.*
 1041 *Res.*, 78, 4037– 4047.

1042 Moore, J. M., Howard, A. D., Dietrich, W. E., and Schenk, P. M., 2003. Martian Layered Fluvial
 1043 Deposits: Implications for Noachian Climate Scenarios. *Geophys. Res. Lett.*, 30 (24), 2292,
 1044 DOI:10.1029/2003GL019002.

1045 Moore, J. M., and Howard, A. D., 2005. Large alluvial fans on Mars. *J. Geophys. Res.*, 110,
 1046 E04005, doi:10.1029/2004JE002352.

1047 Morgan, A.M., Beyer, R.A., Howard, A.D., and Moore, J.M., 2012. The alluvial fans of Saheki
 1048 crater. *Lunar Planet. Sci.* 43. Abstract #2815.

1049 Morgan, A.M., Howard, A.D., Hopley, D.E., Moore, J.M., Dietrich, W.E., Williams, R.M., Burr,
 1050 D.M., Grant, J.A., Wilson, S.A. and Matsubara, Y., 2014. Sedimentology and climatic

1051 environment of alluvial fans in the Martian Saheki crater and a comparison with terrestrial fans in
 1052 the Atacama Desert. *Icarus*, 229, pp.131-156.

1053 Murchie, S., et al., 2007. Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) on
 1054 Mars Reconnaissance Orbiter (MRO). *J. Geophys. Res.*, 112, E05S03,
 1055 doi:10.1029/2006JE002682.

1056 Mustard, J.F., Poulet, F., Gendrin, A., Bibring, J.-P., Langevin, Y., Gondet, B., Mangold,
 1057 N., Bellucci, G., Altieri, F., 2005. Olivine and pyroxene diversity in the crust of Mars. *Science*,
 1058 307, 1594-1597.

1059 Mustard, J. F., Poulet, F., Head, J. W., Mangold, N., Bibring, J.-P., Pelkey, S. M., et al., 2007.
 1060 Mineralogy of the Nili Fossae region with OMEGA/Mex data: 1 ancient impact melt in the Isidis
 1061 basin and implications for the transition from the Noachian to Hesperian. *Journal of Geophysical*
 1062 *Research*, 112, E08S03.

1063 Mustard, J. F., Ehlmann, B. L., Murchie, S. L., Poulet, F., Mangold, N., Head, J. W., Bibring, J.P.,
 1064 and Roach, L. H., 2009. Composition, Morphology, and Stratigraphy of Noachian Crust around
 1065 the Isidis basin. *J. Geophys. Res.*, 114, E00D12, DOI:10.1029/2009JE003349.

1066 Neal, J.T., Langer, A.M., Kerr, P.F., 1968. Giant desiccation polygons of Great Basin playas. *Geol.*
 1067 *Soc. Am. Bull.* 79 (1), 69–90.

1068 Neukum, G. and Ivanov, B.A., 2001. Crater Production function on Mars. *Proc. Lunar Planet. Sci.*
 1069 *Conf.* 32th, Abstract#1757.

1070 Ody, A., Poulet, F., Bibring, J.P., Loizeau, D., Carter, J., Gondet, B., Langevin, Y., 2013. Global
 1071 investigation of olivine on Mars: Insights into crust and mantle compositions. *Journal of*
 1072 *Geophysical Research*, 118, 234-262.

1073 Oehler, D.Z., Mangold, N., Hallet, B., Fairén, A.G., Le Deit, L., Williams, A.J., Sletten, R.S. and
 1074 Martínez-Frías, J., 2016. Origin and significance of decameter-scale polygons in the lower Peace
 1075 Vallis fan of Gale crater, Mars. *Icarus*, 277, pp.56-72.

1076 Palumbo, A., and Head, J.W., 2017. Impact cratering as a cause of climate change, surface
 1077 alteration, and resurfacing during the early history of Mars. *Meteoritics and Planetary Science*.
 1078 <https://doi.org/10.1111/maps.13001>.

1079 Paterson, W.S.B., 1994. The Physics of Glaciers, third ed. Pergamon Press, New York. Phillips,
 1080 L.O., Schumm, S.A., 1987. Effect of regional slope on drainage networks. *Geology* 15, 813–816.

1081 Pelkey, S. M., Mustard, J. F., Murchie, S., Clancy, R. T., Wolff, M., Smith, M., Milliken, R.,
 1082 Bibring, J.-P., Gendrin, A., Poulet, F., Langevin, Y., & Gondet, B., 2007. CRISM multispectral
 1083 summary products: Parameterizing mineral diversity on Mars from reflectance. *Journal of*
 1084 *Geophysical Research*, 112, E08S14.

1085 Poulet, F., Bibring, J.P., Langevin, Y., Mustard, J.F., Mangold, N., Vincendon, M., Gondet, B.,
 1086 Pinet, P., Bardintzeff, J.M., Platevoet, B., 2009a. Quantitative compositional analysis of martian
 1087 mafic regions using the MEx/OMEGA reflectance data 1. Methodology, uncertainties and
 1088 examples of application. *Icarus*, 201, 69-83.

1089 Poulet, F., Mangold, N., Platevoet, B., Bardintzeff, J.-M., Sautter, V., Mustard, J.F., Bibring, J.-
 1090 P., Pinet, P.C., Langevin, Y., Gondet, B., Aléon-Toppani, A., 2009b. Quantitative compositional
 1091 analysis of martian mafic regions using the MEx/OMEGA reflectance data 2. Petrological
 1092 implications. *Icarus*, 201, 84–101.

1093 Quantin-Nataf, C., Lozach, L., Thollot, P., Loizeau, D., Bultel, B., Fernando, J., Allemand, P.,
 1094 Dubuffet, F., Poulet, F., Ody, A., Clenet, H., Leyrat, C., & Harrisson, S., 2018. MarsSI: Martian
 1095 surface data processing information system. *Planetary and Space Science*, 150, 157– 170.

1096 Quantin, C., Flahaut, J., Clenet, H. Allemand, P., Thomas, P., 2012. Composition and structures
 1097 of the subsurface in the vicinity of Valles Marineris as revealed by central uplifts of impact craters.
 1098 *Icarus*, 221, 436-452.

1099 Robbins, S.J., and Hynek, B.M., 2012. A new global database of Mars impact craters ≥ 1 km:
 1100 Global crater properties and regional variations of the simple-to-complex transition diameter.
 1101 *Journal of Geophysical Research*, 117, E06001.

1102 Rogers, A. D., Warner, N. H., Golombek, M. P., Head, J. W., and Cowart, J. C., 2018. Areal
 1103 extensive surface bedrock exposures on Mars: Many are clastic rocks, not lavas. *Geophysical*
 1104 *Research Letters*, 45, 1767–1777. <https://doi.org/10.1002/2018GL077030>

1105 Salvatore, M. R., Mustard, J. F., Wyatt, M. B., and Murchie, S. L., 2010. Definitive evidence of
 1106 Hesperian basalt in Acidalia and Chryse planitiae. *J. Geophys. Res.*, 115, E7,
 1107 DOI:10.1029/2009JE003519.

1108 Scanlon, K. E., Head, J. W., Wilson, L., and Marchant, D. R., 2014. Volcano–ice interactions in
 1109 the Arsia Mons tropical mountain glacier deposits. *Icarus*, 237, 315– 339.
 1110 <https://doi.org/10.1016/j.icarus.2014.04.024>.

1111 Scanlon, K. E., Head, J. W., and Marchant, D. R., 2015. Volcanism-induced, local wet-based
 1112 glacial conditions recorded in the Late Amazonian Arsia Mons tropical mountain glacier deposits.
 1113 *Icarus*, 250, 18– 31. <https://doi.org/10.1016/j.icarus.2014.11.016>.

1114 Scott, D. H., and Tanaka, K. L., 1986. Geologic map of the western equatorial region of Mars.
 1115 U.S. Geol. Surv. Misc. Invest. Ser. Map, I-1802-A.

1116 Shean, D.E., Head, J.W., Marchant, D.R., 2005. Origin and evolution of a cold-based tropical
 1117 mountain glacier on Mars: The Pavonis Mons fan-shaped deposit. *J.Geophys. Res.* 110, E05001.

1118 Shean, D. E., 2010. Candidate ice-rich material within equatorial craters on Mars. *Geophys. Res.*
 1119 *Lett.*,37,L24202, DOI:10.1029/2010GL045181.

1120 Skinner, J. A., Hare, T. M., and Tanaka, K. L., 2006. LPSC XXXVII, Abstract #2331.

1121 Skok, J. R., Mustard, J. F., Murchie, S. L., Wyatt, M. B., and Ehlmann, B. L., 2010. Spectrally
 1122 distinct ejecta in Syrtis Major, Mars: Evidence for environmental change at the Hesperian-
 1123 Amazonian boundary. *J. Geophys. Res.*, 115, E00D14, DOI:10.1029/2009JE003338.

1124 Skok, J. R., Mustard, J. F., Tornabene, L. L., Pan, C., Rogers, D., and Murchie, S. L., 2012. A
 1125 spectroscopic analysis of Martian crater central peaks: Formation of the ancient crust. *J. Geophys.*
 1126 *Res.*,117, E00J18, DOI:10.1029/2012JE004148.

1127 Sinha, R.K. and Vijayan, S., 2017. Geomorphic investigation of craters in Alba Mons, Mars:
 1128 Implications for Late Amazonian glacial activity in the region. *Planetary and Space Science*, 144,
 1129 32-48. <https://doi.org/10.1016/j.pss.2017.05.014>.

1130 Smith, D. E., et al., 2001. Mars Orbiter Laser Altimeter: Experiment summary after the first year
 1131 of global mapping of Mars. *J. Geophys. Res.*, 106 (E10), 23689–23722,
 1132 DOI:10.1029/2000JE001364.

1133 Soare, R.J., Conwat, S.J., Dohm, J.M., 2014. Possible ice-wedge polygons and recent landscape
 1134 modification by “wet” periglacial processes in and around the Argyre impact basin, Mars. *Icarus*,
 1135 233, 214–228. <https://doi.org/10.1016/j.icarus.2014.01.034>.

1136 Sun, V. Z., and Milliken, R. E., 2015. The geology and mineralogy of Ritchey crater, Mars:
 1137 Evidence for post-Noachian clay formation. *Journal of Geophysical Research: Planets*, 119, 810–
 1138 836. <https://doi.org/10.1002/2013JE004602>.

1139 Sunshine, J. M., Pieters, C. M. and Pratt, S. F., 1990. Deconvolution of mineral absorption bands:
 1140 An improved approach. *J. Geophys. Res.*, 95, 6955–6966, DOI:10.1029/JB095iB05p06955.

1141 Sunshine, J. M., and Pieters, C. M., 1998. Determining the composition of olivine from reflectance
 1142 spectroscopy. *J. Geophys. Res.*, 103, 13,675–13,688, DOI:10.1029/98JE01217.

1143 Tanaka, K.L., Skinner, J.A., Dohm, J.M., Irwin, R.P., Kolb, E.J., Fortezzo, C.M., Platz, T.,
 1144 Michael, G.G., and Hare, T.M., 2014, Geologic map of Mars: U.S. Geological Survey Scientific
 1145 Investigations Map 3292, scale 1:20,000,000, pamphlet 43 p., <https://dx.doi.org/10.3133/sim3292>.

1146 Goudge, T.A., Milliken, R. E., Head, J. W., Mustard, J. F., Fassett, C.I., 2017. Sedimentological
 1147 evidence for a deltaic origin of the western fan deposit in Jezero crater, Mars and implications for
 1148 future exploration. *Earth and Planetary Science Letters*, 458, 357–365.
 1149 <https://doi.org/10.1016/j.epsl.2016.10.056>

1150 Tornabene, L. L., Moersch, J. E., McSween, H. Y., Hamilton, V. E., Piatek, J. L., and Christensen,
 1151 P. R., 2008. Surface and crater-exposed lithologic units of the Isidis Basin as mapped by coanalysis
 1152 of THEMIS and TES derived data products. *Journal of Geophysical Research*, 113, E10001.
 1153 <https://doi.org/10.1029/2007JE002988>

1154 Ventra, D., and Clarke, L. E., 2018. Geology and Geomorphology of Alluvial and Fluvial Fans:
 1155 Terrestrial and Planetary Perspectives. Geological Society, London, Special Publications, 440, 1-
 1156 21.

1157 Viviano-Beck, C. E., et al., 2014. Revised CRISM spectral parameters and summary products
 1158 based on the currently detected mineral diversity on Mars. *J. Geophys. Res. Planets*, 119, 1403–
 1159 1431, DOI:10.1002/2014JE004627.

1160 Viviano-Beck, C. E. et al., 2015. MRO CRISM Type Spectra Library, NASA Planetary Data
 1161 System. <http://crismtypespectra.rsl.wustl.edu>

1162 Viviano-Beck, C.E., Murchie, S.L., Daubar, I.J., Morgan, M.F., Seelos, F.P., Plescia, J.B., 2019.
 1163 Composition of Amazonian volcanic material in Tharsis and Elysium, Mars, from MRO/CRISM
 1164 reflectance spectra. *Icarus* 328, 274-286. <https://doi.org/10.1016/j.icarus.2019.03.001>.

1165 Washburn, A.L., 1980. A survey of periglacial processes and environments. *Geocryology*, The
 1166 Blackburn Press, New Jersey.

1167 Weiss, D.K., Head, J.W., 2016. Impact ejecta induced melting of surface ice deposits on Mars.
 1168 *Icarus* 280, 205–233.doi:10.1016/j.icarus.2016.07.007.

1169 Williams, K., Toon, O., Heldmann, J., and Mellon, M., 2009. Ancient melting of mid-latitude
 1170 snowpacks on mars as a water source for gullies. *Icarus*, 200, 418–425.

1171 Wilson, L., Ghatan, G.J., Head, J.W., Mitchell, K.L., 2004. Mars outflow channels: A reappraisal
 1172 of the estimation of water flow velocities from water depths, regional slopes, and channel floor
 1173 properties. *J. Geophys. Res. E: Planets* 109, E09003, 1–10.

1174 Wilson, S.A., Zimbelman, J.R., 2004. Latitude-dependent nature and physical characteristics of
 1175 transverse aeolian ridges on Mars. *J. Geophys. Res. E: Planets* 109, E10003.
 1176 <http://dx.doi.org/10.1029/2004JE002247>.

1177 Wilson, S. A., Grant, J. A., and Howard, A. D., 2013. Inventory of equatorial alluvial fans and
 1178 deltas on Mars. *Lunar and Planet. Sci. Conf.*, Abstract 2710.

1179 Wilson, S.A., Morgan, A.M., Howard, A.D., and Grant, J.A., 2021. The global distribution of
 1180 craters with alluvial fans and deltas on Mars. *Geophysical Research Letters*, DOI:
 1181 <https://doi.org/10.1029/2020GL091653>

1182 Wordsworth, R., Forget, F., Millour, E., Head, J. W., Madeleine, J. B., and Charnay, B., 2013.
 1183 Global modeling of the early Martian climate under a denser CO₂ atmosphere: Water cycle and
 1184 ice evolution. *Icarus*, 222 (1), 1– 19, DOI:10.1016/j.icarus.2012.09.036.

 1185 Xiao, L., Huang, J., Christensen, P. R., Greeley, R., Williams, D. A., Zhao, J., and He, Q., 2012.
 1186 Ancient volcanism and its implication for thermal evolution of Mars, *Earth Planet. Sci. Lett.*, 323–
 1187 324, 9–18.

 1188 Yoshikawa, K., 2003. Origin of the polygons and the thickness of Vastitas Borealis Formation in
 1189 Western Utopia Planitia on Mars. *Geophys. Res. Lett.*, 30 (12), 1603,
 1190 DOI:10.1029/2003GL017165.