Supplemental Material

A fragile record of fleeting water on Mars

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Table S1

Crater	THEMIS	CRISM	HiRISE	HRSC	CTX DTM
Becquerel	I36610007	frs000281b6 hrl00006290 frt00003245	PSP_005845_2015 PSP_004368_2015 ESP_024503_2015 ESP_038916_2015 PSP_001546_2015 ESP_017211_2015 ESP_016077_2015	h3231_0001	
Danielson	I36922027 I38694008	hrl0000434e frt000049ca	PSP_004434_1885 PSP_002733_1880 PSP_008930_1880 ESP_019835_1880 ESP_044507_1880 ESP_060331_1880 ESP_027615_1880 ESP_055069_1880 PSP_002522_1880 ESP_054858_1885		G21_026349_ 1880_G22_02 6626_1880
Firsoff	I36997007 I36685006	frt00017697 frt000236de frt0002437e	ESP_020534_1825 ESP_049531_1820 ESP_030292_1820 ESP_054568_1820 ESP_020679_1820 ESP_026270_1820 ESP_047184_1820 ESP_065276_1825 ESP_034907_1825 ESP_064643_1825 ESP_036674_1825 ESP_017132_1825 ESP_036041_1830 ESP_048753_1830 ESP_046617_1830 ESP_025769_1830 ESP_040393_1830 ESP_054146_1830 ESP_057918_1830 ESP_059553_1830 ESP_028024_1830	h1965_0000	G19_025703_ 1827_G19_02 5769_1827

			ESP_012385_1825 ESP_036819_1825 PSP_008772_1830 PSP_004355_1830		
Jiji	I37109006 I54411009	frt00016834	ESP_016657_1890 ESP_058788_1890		B18_016657_ 1881_B19_01 7013_1881
Kaporo	I62903011 I36210011 I02615005	hrs00003ec8	ESP_037794_1795 ESP_036238_1795 PSP_002429_1800 PSP_002508_1800 PSP_002930_1800		F04_037372_ 1798_P04_00 2429_1799
Kozova	I36510012 I30808002	frt0000b084 frt0000960f frt0000b710 hrl00008c6b	PSP_009154_1835 PSP_006860_1840 ESP_026573_1835 ESP_017580_1840 ESP_044309_1840 PSP_008798_1840 PSP_002390_1840 ESP_017224_1835	h5242_0000	
Sera	I39430005 I44984002	frt000037f7 frt00020385	ESP_022116_1890 PSP_002047_1890 ESP_055227_1890 ESP_037834_1890 ESP_017936_1890 ESP_047421_1890		P02_001902_ 1889_P03_00 2047_1889
Trouvelot	I45259009	frt00023898 hrl0000abb4		h3286_0000 h3275_0000	
Unnamed 1 (3° E, 2° N)	I39954007	frt00012fc3 frt00006445		h1976_0000 h1965_0000	
Unnamed 2 (2.9° E, 4.4° N)	I46382006	frt00009272 frt00009b5a frt0000a553		h1965_0000	

Supplementary Table 1. Identifiers of THEMIS, CRISM, HiRISE, and HRSC data used in this study. CTX data from the Murray Lab was used as a base map (NASA/JPL/MSSS/The Murray Lab). All underlying data is publicly available in the Geosciences Node of NASA's Planetary Data System (PDS). Derived THEMIS TIs, classification stamps, and crater count shape files are available at <https://openknowledge.nau.edu/id/eprint/5610/>. Derived CTX Digital Elevation Models (DEMs) and corresponding map projected CTX images are available at <https://doi.org/10.5281/zenodo.4663102>.

MINERAL MAPPING

We identified mineral trends and classified surfaces using CRISM targeted full and half resolution imagery (frt, frs, hrl, and hrs) in the long wavelength range (1.03-3.92 μm) as well as with CRISM MTRDR data, where available (Murchie et al., 2007). We employed multiple tools to reduce spiking and striping noise and subtract atmospheric contributions to CRISM TRDR data

This included standard ENVI CRISM Analysis Tools (CAT) methods as well as fitting methods outlined in Itoh and Parente, (2021). The Itoh and Parente (2021) method was employed on CRISM images: hrs00003ec8_07_if174l (Kaporó Crater), frt00020385_07_if165l (Sera Crater), frt000236de_07_if165l (Firsoff Crater), and frt00016834_07_if165l (Jiji Crater). The remaining CRISM images were processed using ENVI CAT version 7.4 and empirically optimized atmospheric correction by McGuire et al. (2009) volcano scan scaling. Regions of interest were identified after full image ratioing and image color stretching (see Pan et al., 2017). Characteristic absorption features (Figure 1B) were then identified by ratioing spectra in regions of interest to spectrally bland locations in the same image columns. We compared the resultant spectral shapes to spectra measured in the laboratory and previously identified in CRISM data in the region (e.g., Niles and Michalski, 2009; Wiseman et al., 2010; Michalski and Niles, 2012; Pondrelli et al., 2015; Zolotov and Mironenko, 2016; Bishop et al., 2018; Quinn and Ehlmann, 2019). We ultimately produced maps from the spectral band parameters outlined in Viviano-Beck et al. (2014) as inputs to our classification pipeline.

THERMOPHYSICAL PROPERTIES

Thermal inertia (TI) is a material property defined by the square root of the product of thermal conductivity, density and heat capacity ($\sqrt{k\rho c}$). TI controls a material's cadence in gaining or losing heat through time, thus defining the shape of daily and seasonal surface temperature curves. On Mars, where the atmosphere is thin and the heat capacity of geological materials varies relatively little, TI depends most strongly on thermal conductivity (Piqueux and Christensen, 2009). Because solids are much more efficient heat conductors than gases, small changes in the

amount of inter-grain cements can lead to large changes in thermal conductivity and TI (Presley et al., 2009).

We derive TI by pairing nighttime THEMIS images (~ 100 m/px) (Christensen et al., 2004) with a numerical thermal model, KRC (Ferguson et al., 2006a; Kieffer, 2013; Edwards et al., 2018), that considers slope angle, aspect, elevation, albedo, emissivity, estimated component specific heat, estimated component thermal conductivities, and other observation time parameters (e.g., season, local time, solar azimuth, dust opacity, and modeled air temperatures) (Putzig et al., 2005; Kieffer, 2013). In KRC's 'one-point' mode, models are run independently for each THEMIS pixel. We use an updated application of KRC (Edwards et al., 2018) that is distinct from other implementations (e.g., Ferguson et al., 2006; Kieffer, 2013) in its use of higher resolution model inputs, and here the model is allowed to run over a longer time span (two Mars years) to achieve a seasonally-stable output.

Surface TIs in the study regions range in value from ~ 50 to $\sim 700 \text{ Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$ and sulfate-bearing layered deposits (SBLDs) tend to have elevated TIs relative to surrounding units. However, due to instrumental, atmospheric and seasonal differences between different THEMIS images that are difficult to correct, there can be small (e.g., $\sim 50 \text{ Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$, corresponding to a ~ 3 K temperature error) systematic differences in TI values derived (Ferguson et al., 2006b; Edwards et al., 2011). Pixel-to-pixel uncertainty within a single image, however, is only ~ 0.5 K, equating to $< 15 \text{ Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$. Therefore, the relative difference in TI between two regions in the same image were considered more strongly than absolute TI for the purpose of delineating units.

The reported TI values for each crater site in this work (Figure S1) were taken either from a single image per crater or from images with overlapping areas demonstrating those images contain similar pixel values. The differences in TI values between surface units at each crater site were

calculated from the ANOVA difference in means between those units, with 95% confidence interval as the plus minus error. The reported overall differences in TI values between surface units (see section: *The Effect of Dust on Thermal Inertia*) and associated uncertainties are the means of the differences in TI at each individual crater site and associated uncertainties weighted by number of pixels used to calculate each mean.

CORRELATION ANALYSIS

For each study region, datasets were spatially aligned with tie pointing and resampled to 18 or 36 m/px (full or half CRISM resolution). Pixel values for CRISM band parameters and TI were then plotted by kernel density in order to roughly identify the number of clusters present. Normalized data was then assessed using principal components analysis. Effective classification was achieved by weighting normalized TI and band parameter data according to their relative component importance. Specifically, starting from the first principal component, the set of principal components that cumulatively explained more than 50% of the normalized sample variance were identified. Then, the key variables ($|\text{coefficient}| > 0.30$) within each of these principal components were identified, and only those variables were used in clustering.

We found that both this unsupervised method and human observers independently identified similar band parameters as having high variance (SINDEX2, LCPINDEX2, HCPINDEX2, OLINDEX3, BD1900_2, BD2100_2, and D2300). However, principal components that explained large amounts of variance in the data sometimes were a result of topographic effects in areas with high relief at multiple aspects. We prevented data subject to this effect from being incorporated in clustering by excluding band parameters with high positive or high negative (>0.4) correlations to the R1330 albedo proxy in regions of topography-dependent illumination.

Using the refined set of variables, we ultimately fit a gaussian mixture model to the data to determine a cluster probability for each data point. Data points with less than a 90% probability of being in a single cluster were assigned to cluster mixtures and not counted in cluster statistics. Our clustering method identified a total of six major surface types, each with distinct patterns in TI and CRISM band parameters. Although classification was achieved for regions covering each CRISM stamp independently, we grouped units among the different sites on the basis of consistent similarities in spectral trends, TI trends and morphology. These units are not necessarily stratigraphically-linked geologic units, but rather regions with highly correlated surface properties. The units are: I) moderate to relatively high TI SBLDs which show distinct light-toned layers and fractured textures at HiRISE image scales, II) relatively low TI SBLDs with moderate-toned and sometimes subdued layering structures and smooth textures at HiRISE image scales, III) relatively low TI pyroxene-rich cratered surfaces, sometimes hosting phyllosilicates with smooth textures at HiRISE image scales, IV) relatively high TI pyroxene-rich cratered surfaces sometimes hosting phyllosilicates and smooth to lightly fractured textures at HiRISE image scales, V) moderate TI, dark-toned, olivine-rich dunes, and VI) relatively low TI moderate-toned pyroxene-rich dunes, ripples and colluvium. In Figures 2 and S1, unit I is red, unit II is orange, unit III is green, unit IV is blue, unit V is purple, and unit VI is not shown but reflects similar attributes as unit III. As explained in the main text, we interpret that unit II can be defined as more dust-covered exposures of unit I.

We note that it was decided after scrutinization that none of the targeted CRISM imagery within Trouvelot crater captures SBLDs even though layered deposits appear to be present within the crater nearby.

REGIONAL TRENDS

A couple outcomes of classification that were not included in the main text are worth mentioning here. After classifying units, we found the highest unit median TI was 394.51 (Interquartile Range [IQR]: 359.36-428.44) $\text{Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$ for the SBLDs in Firsoff crater, at the western edge of our study area. The lowest median TI for a classified unit was 69.42 (IQR: 62.78-76.05) $\text{Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$ for apparently dusty terrain with dominantly mafic spectral characteristics in and around Kaporo crater, at the eastern edge of our study area. If related to aeolian activity, these numbers would reflect the broad-scale variability in the magnitude of aeolian erosion that is predicted by global climate models (e.g., Toigo et al., 2012) (Figure 3) and indicated by dust cover index maps (Ruff and Christensen, 2002) (Figure 1).

Another spatial trend is the abundance of phyllosilicate spectral signatures adjacent to or within SBLDs in the southernmost and highest elevation portion of the study area (Figure 1). There are also anomalously high TIs ($>500 \text{ Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$) in and around these regions. If this pattern reflects the distribution of aqueous activity, it would be consistent with the higher density of inferred fluvial channels (Hynek and Di Achille, 2017; Davis et al., 2019) present in the southern portion of the study region but inconsistent with predictions for groundwater depth from hydrological models (Andrews-Hanna et al., 2010).

AEOLIAN ACTIVITY ANALYSIS

Figure S2. MOLA Colorized elevation globe of Mars (NASA/JPL) centered at 0° , 0° . White arrows show grid of mesoscale model-derived wind direction (Toigo et al., 2012). Panels from near black point depict inferred local wind directions (red arrows) determined from A) yardang orientations, B) wind streaks, C) ripple orientations (paired with additional context), and D) dune orientations.

In order to assess the potential control that aeolian erosion has on apparent TI, we estimated where in the study areas aeolian erosional stress is likely to be strongest. First, we estimated prevailing regional and localized wind directions using a combination of gridded mesoscale model wind vectors (Toigo et al., 2012) and new observations of wind direction indicators (Figure S2). We then considered the crater geometries and compared them to surfaces studied in 2-D slope wind erosion models (e.g., Kite et al., 2013; Steele et al., 2018) and wind tunnel experiments (e.g.,

Day et al., 2016). Complex geometries that did not match those in the modeling studies prevented us from conducting detailed analysis with confidence in the majority of craters. However, we were able to make comparisons between TI and anticipated aeolian stress along profiles through Sera Crater (Figure 4).

CRATER DENSITY ESTIMATES

In order to gauge the relationship between crater retention and TI, as well as to assess exhumation rates, we conducted crater counts in Becquerel, Danielson, Firsoff, Jiji, Kaporo, Sera and Kozova Craters using the JMARS crater counting tool. We identified craters larger than ~20 m using CTX 6 m/px imagery (Malin et al., 2007) from the Murray Lab global CTX mosaic and HiRISE 30-60 cm/px targeted observations (McEwen et al., 2007), where available (images listed in Table S1). To mitigate observation bias, HiRISE images were only used to confirm whether certain types of features observed at CTX scales were craters (as opposed to circular dune forms or other erosional features). These decisions were then also applied to areas without HiRISE coverage.

As a means to quantitatively compare crater retention and TI, we chose to rasterize crater counts at THEMIS pixel resolution. Crater counts were loaded into ArcGIS Pro, where a kernel density estimation tool was used to assign values representative of the nearby number of craters to each pixel. Search area was defined using the bandwidth approach, which depends on the spatial spread in points, but ultimately maintains consistently-scaled density estimates from region to region. This method is explained in Silverman (1986).

In order to ensure craters from one surface unit were not counted in adjacent ones, we set boundaries by extrapolating the contacts of the units identified through clustering and calculated

crater densities for each surface unit separately. Also, instead of counting each crater as 1 point, craters were weighted by size using the formula: $\text{weight} = \log_{10}(\text{diameter}/10)$, which makes results more comparable to standard cumulative crater density studies and normalized to 100 m diameter craters. In essence, the x-values in Figure 2C can be considered an approximation of the number of 100 m craters per square km.

To estimate possible erosion rates in the SBLDs, we followed a method outlined in Kite and Mayer (2017) that measures the divergence of size-dependent cumulative crater densities from the anticipated isochron slopes of crater production functions (Figures 3D and S3). Given our focus on areas where dust cover appears to be limited as well as the absence of sediment-filled craters in these areas, we assume steady exhumation, or erosion, dominates over diffusive processes, such as sediment infilling. Under this assumption, an algebraic equation yields a rough estimate of erosion rate at each crater size bin,

$$E_D \approx 0.2D\phi / \left(\frac{N_D}{f_D a} \right)$$

where E_D is the erosion rate at a given crater size bin, D , N_D is the number of craters counted of that size range, a is the area of the region counted, and f_D is the predicted crater flux for the bin from Hartmann and Daubar, 2017) with the chronology function of (Hartmann, 2005) modified by (Michael, 2013). ϕ is the fraction of crater depth that must be obliterated before the crater is no longer recognizable, and, as in Kite and Mayer (2017), is taken to be 0.5. The 0.2 multiplier is an approximation of the depth-to-diameter ratio for craters in the size range of interest.

Kite and Mayer (2017) calculated uncertainties on the mean of erosion rates derived from multiple images. We did not specify uncertainty since only one estimate was made for each crater, rather than estimates from each image, producing an already imprecise range. However, the exhumation rate range was limited at the upper end by the identifiability limit of craters <20 m (~22 m

in $2^{1/2}$ bin space). The lower limit on the range was selected based on inflection points at 177 m present in the cumulative distributions for each region, suggesting a different cratering or preservation regime for craters of that size. Also, the cumulative number of identified craters with diameters larger than 177 m drops off dramatically, generating large uncertainties on the density of those craters. For these reasons, we exclude the exhumation rates derived from those larger crater size bins from our estimate.

Figure S3. Cumulative crater densities of sulfate-bearing layered deposits in A) Sera and B) Jiji Craters plotted in $2^{1/2}$ bin spacing using Craterstats 2.0 (Michael, 2013 and sources therein) with modeled crater production function isochron (solid line, (Hartmann, 2005; Michael, 2013; Hartmann and Daubar, 2017) shown for reference. Divergence of crater counts from modeled isochron may result from exhumation or infilling. Survey incompleteness cut-offs at 22 m diameter for both

regions are shown with black arrows. Cyan arrows show inflection points, suggesting regime change with larger crater preservation. The estimates of exhumation rates at each bin for C) Sera and D) Jiji Craters for the range stated in the study are shown in black; incompleteness cut-offs in gray; and the excluded larger crater regimes are in cyan.

SLOPE STABILITY ANALYSIS

Estimates of cohesion were achieved using slope stability analysis. We located eight sites where recent wedge failure is observed and 15 sites where recent circular slope failure is visible at HiRISE image resolutions (~100 m in size) within areas covered by overlapping HiRISE stereo pairs (Tables S2 & S3, Figures 4, S4, & S5). These features were identified by their characteristic failure trace morphologies (see Wyllie and Mah, 2004) as well as by large debris piles located directly down slope of the failure sites. We followed the methods outlined in Wyllie and Mah (2004), which is based on theoretical work from Hoek et al. (1973), for backward derivation of cohesion. For these calculations, we made a few simplifying assumptions: that pore pressure and liquid content of the materials are negligible, that no external forces are present, that the cohesive strength and internal friction angle of all slip faces are approximately the same, and that cohesion at the failure site is representative of overall deposit cohesion.

At sites of wedge failure, we measured the geometry of the wedge trace surfaces in HiRISE image-derived digital terrain models using the qgSurf tool (Alberti, 2019) in QGIS, which fits planes to a group of DTM point inputs. By setting the factor of safety to 1, the point at which a slope is unstable, we were able to equate the downward force of the weight of the wedge to the resistant cohesive force at the failure surfaces and derive an upper estimate for cohesion at the failure site. In this scenario cohesion can be calculated as:

$$C = \frac{1 - (A+B)\tan\phi}{(X+Y)} \cdot \frac{\rho_{rock}g_{Mars}H}{3}$$

Where ϕ is the rock's angle of internal friction, ρ_{rock} is the density of the rock, g_{Mars} is Mars' gravity (3.711 m/s^2), H is the measured height of the wedge, X , Y , A , and B are dimensionless factors that depend on the wedge failure plane geometries (see Wyllie and Mah, 2004, Ch. 7). In our calculations we used a range of internal angles of friction (10 - 60°) and density (1200 - 2000 kg/m^3), which are representative of sediments and sedimentary rocks observed on Mars (Golombek et al., 2008).

Figure S4. Geometries of the failed wedge trace in Figure 4 plotted on a stereonet with a cartoon depicting a simple model of the wedge and the balancing gravitational and cohesive forces during failure.

Table S2

Crater + Failure Site ID	Latitude (°N)	Longitude (°E)	HiRISE DTMs	Plane A Strike and Dip (°)	Plane B Strike and Dip (°)	H (m)	Derived Cohesion Range (kPa)
Sera W1	8.764	-1.112	ESP_065856_1890_ ESP_066067_1890*	167.3, 35.9 W	212.5, 39.2 W	39.5	2.30-11.25
Firsoff W1	2.171	-9.041	ESP_026270_1820_ ESP_027747_1820	333.7, 46.3 E	274.1, 48.9 E	117.9	1.65-5.65
Firsoff W2	2.164	-9.040	ESP_026270_1820_ ESP_027747_1820	119.3, 37.0 W	97.3, 37.2 W	55.3	0.65-4.56
Firsoff W3	2.156	-9.057	ESP_026270_1820_ ESP_027747_1820	82.8, 39.1 E	145.9, 45.2 W	63.0	2.79-17.90
Firsoff W4	2.146	-9.063	ESP_026270_1820_ ESP_027747_1820	347.1, 36.1 W	283.6, 43.4 W	26.2	1.39-12.79
Firsoff W5	2.153	-9.169	ESP_039404_1820_ ESP_047184_1820	208.3, 45.6 W	303.4, 61.4 E	53.6	1.41-9.50

Firsoff W6	2.113	-9.127	ESP_039404_1820_ ESP_047184_1820	286.4, 40.9 E	358.3, 53.5 E	68.9	2.18-8.10
Firsoff W7	2.186	-9.143	ESP_039404_1820_ ESP_047184_1820	261.7, 41.5 W	352.0, 53.2 E	26.4	0.13-15.90

Supplementary Table 2. Wedge failure sites analyzed for cohesion derivation. *All DTMs are available from the Geosciences Node of NASA's Planetary Data System (PDS) except for ESP_065856_1890_ESP_066067_1890, which is available from <https://doi.org/10.5281/zenodo.4681216>.

At sites of circular failure, we measured the height and slope angle. To estimate heights and slope angles, we also used digital elevation models derived from HiRISE images and extracted two dimensional slope profiles. Then, using Figure 8.6 from Wyllie and Mah (2004), with a factor of safety set to 1 and the same tested range of internal angles of friction, we estimated a range of possible values for $C/\rho_{\text{rock}}g_{\text{Mars}}H$, where H is now the height of the failed portion of the slope. We then estimated a maximum cohesion at each site from the same range of internal angle of friction and density combinations. A similar method to this has been applied on smaller scales to study materials at multiple Mars lander sites (e.g., Moore et al., 1987; Golombek et al., 2020). Resultant cohesion estimates were slightly higher for the circular failure style estimates. In both cases, lower internal angles of friction and higher densities correspond to higher cohesion.

Table S3

Crater + Failure Site ID	Latitude (°N)	Longitude (°E)	HiRISE DTMs	H (m)	Slope angle (°)	Upper Limit on Cohesion (kPa)
Sera C1	8.767	-1.092	PSP_002047_1890_ PSP_001902_1890	36.8	37.85	27
Sera C2	8.730	-1.068	PSP_002047_1890_ PSP_001902_1890	29.9	37.77	22
Jiji C1	8.842	-1.720	ESP_016657_1890_ ESP_017013_1890*	25.6	45.25	25
Jiji C2	8.906	-1.659	ESP_016657_1890_ ESP_017013_1890*	81.0	36.27	60
Danielson C1	8.202	-6.864	PSP_002878_1880_ PSP_002733_1880	58.1	35.77	43
Danielson C2	8.044	-6.790	PSP_002878_1880_ PSP_002733_1880	28.2	27.98	17
Firsoff C1	2.166	-9.043	ESP_026270_1820_ ESP_027747_1820	47.1	34.43	35
Firsoff C2	2.264	-9.056	ESP_026270_1820_ ESP_027747_1820	37.3	40.74	33
Firsoff C3	2.166	-9.062	ESP_026270_1820_ ESP_027747_1820	34.3	45.31	33

Firsoff C4	2.140	-9.063	ESP_026270_1820_ ESP_027747_1820	46.0	35.19	34
Firsoff C5	2.166	-9.043	ESP_026270_1820_ ESP_027747_1820	61.9	38.52	51
Firsoff C6	2.187	-9.009	ESP_026270_1820_ ESP_027747_1820	57.6	32.72	41
Firsoff C7	2.150	-9.108	ESP_039404_1820_ ESP_047184_1820	36.4	46.11	35
Firsoff C8	2.148	-9.109	ESP_039404_1820_ ESP_047184_1820	72.1	36.99	54
Firsoff C9	2.137	-9.397	ESP_029514_1820_ ESP_030292_1820	57.2	34.78	42

Supplementary Table 3. Circular failure sites analyzed for cohesion derivation. Plane A is always to shallower dipping plane. *All DTMs are available from the Geosciences Node of NASA's Planetary Data System (PDS) except for ESP_016657_1890_ESP_017013_1890, which was is available from <https://zenodo.org/record/3378969>.

Figure S5. Example of A) circular slope failure (Firsoff C6, -9.009°E 2.187°N, ESP_26270_1820) and B) wedge slope failure (Firsoff W6, -9.127°E 2.113°N, ESP_039404_1820) from Firsoff Crater.

DATA PROCESSION TOOLS

Data Processing was completed using JMARS, ENVI/IDL, ArcGIS Pro, QGIS, Matlab, Python, and the Ames Stereo Pipeline (Beyer et al., 2018). All data products used in this work are derived from data available at the Geosciences Node of NASA's Planetary Data System (PDS) (<https://pds-geosciences.wustl.edu/>). Derived products not directly available from that archive are available from <https://openknowledge.nau.edu/id/eprint/5610/>. Modeling was conducted using the KRC (<http://krc.mars.asu.edu>) thermal model in Davinci (<http://davinci.asu.edu>;

<https://github.com/NASA-Planetary-Science/krc>) and the Monsoon computing cluster at NAU. A statistical summary of TI data derived for this study is available in a separate supplementary Excel file.

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