Thermophysical and Compositional Analyses of Dunes at Hargraves Crater, Mars

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Abstract

We analyze thermal emission spectra using the 2001 Mars Odyssey Thermal Emission Imaging System and the Mars Global Surveyor Thermal Emission Spectrometer to characterize grain size and mineralogical composition of dunes at Hargraves crater, Mars. Thermal inertia and bulk composition of the dunes were compared to inferred provenances from the thermal infrared response of surface constituent materials. We use a Markov Chain Monte Carlo technique to estimate the bulk amount of mineralogy contributed by each inferred provenance to the dune field composition. An average thermal inertia value of $238 \pm 17 \text{ J m}^{-2} \text{K}^{-1} \text{s}^{-0.5}$ was found for the dunes, corresponding to a surface composed of an average effective grain size of $\sim 391 \pm 172 \mu m$. This effective particle size suggests the presence of mostly medium sand-sized materials mixed with fine and coarse grain sands. The dunes are likely composed of a weakly indurated surface mixed with unconsolidated materials. Compositional analysis specifies that the dunes are composed of a mixture of feldspar, olivine, pyroxene, and relatively low bulk-silica content. Dune materials were likely derived from physical weathering, especially aeolian erosion, predominantly from the crater ejecta unit at the crater, mixed with a small amount from the crater floor and crater rim and wall lithologies—indicating that the dune materials were likely sourced locally.

Unified Astronomy Thesaurus concepts: Mars (1007); Solar system (1528); Solar system astronomy (1529); Solar system planets (1260)

1. Introduction

Despite a variety of environments involved, dune fields provide an important insight into aeolian transport regimes (both past and present) of terrestrial planets (Greeley & Iversen 1987). The surface of Mars is rife with aeolian dune fields in a variety of locations, e.g., the circum-north polar region (Edgett & Christensen 1991). Dunes on the Martian surface were first detected during the early 1970s (Sagan et al. 1972). The advent of higher-resolution orbital images further confirmed the presence of dunes during the late 1990s (Edgett & Malin 2000) and showed similarities with terrestrial dunes (Bandeira et al. 2010). Recently, a global Martian dune database called the Mars Global Digital Dune Database (MGDD; Haydard et al. 2007b, 2010, 2012) has been prepared by planetary scientists. The database contains spatial and morphological characteristics of identified dune fields of $>1 \text{ km}^2$ (Hayward et al. 2007a, 2014). Understanding of Martian dune materials and aeolian processes opens a window for constraining the surface-atmosphere interactions, weather, climate, and climatic evolution of the planet (Greeley et al. 2001; Wilson & Zimbelman 2004). Over the past decades, studies were carried out to better understand Martian surface processes and global-scale atmospheric dynamics from analysis of various aeolian processes and wind-related features (e.g., Ward et al. 1985; Fenton 2005; Hayward et al. 2009; Silvestro et al. 2010; Gardin et al. 2012; Sefton-Nash et al. 2014).

Throughout its geologic history, the surface of Mars has been shaped by aeolian processes. Thus, analyses of thermophysical, grain-size, and mineral compositional distribution within dunes are regarded as important aspects for helping to constrain the geology and climate of the planet (Greeley & Iversen 1987; Greeley et al. 2001; Charles et al. 2017; Fenton et al. 2019b). Local- and regional-scale analysis of grain-size distributions is important in classifying sedimentary environments and understanding sediment transport dynamics. Particle grain size provides important information about shear stress applied by the medium of transport to initiate and sustain particle movement (Abuodha 2003). Particle grain size in dunes is actively influenced by the nature of source materials, surface topography, transport medium, and the distance traveled from source to sink (Abuodha 2003). On Mars, grain-size distribution has a direct relationship to thermophysical characteristics of constituent materials (Kieffer et al. 1977; Haberle & Jakosky 1991). Aeolian dunes of Mars are likely composed of particle sizes of homogenous materials within a dune field, and therefore grain-size determination from thermal inertia is considered a reliable method (Edwards et al. 2018).

The approximate bulk particle sizes of dunes on Mars have been determined successfully using thermal inertia from the Thermal Emission Imaging System (THEMIS; Christensen et al. 2004 images (Fenton & Mellon 2006; Fergason et al. 2006, 2012). Edwards et al. (2018) estimated the average grain size for the Bagnold dune field to be $\sim 250 \pm 85 \mu m$, with a minimum of $148 \mu m$ and a maximum of $968 \mu m$ based on derived thermal inertia values. Using data from Mars Science Laboratory (MSL) Curiosity, Ehlmann et al. (2017) reported the presence of very fine to medium-sized ($\sim 45–500 \mu m$) sand particles in the active Bagnold dune field, THEMIS-derived thermal inertia of $\sim 250–410 \text{ J m}^{-2} \text{K}^{-1} \text{s}^{-0.5}$ was found across the majority of the Gale crater landing site of MSL, which corresponds to dunes composed of an indurated surface, likely mixed with unconsolidated materials (Fergason et al. 2012). In addition, Mawrth Vallis has an average thermal inertia of $310 \text{ J m}^{-2} \text{K}^{-1} \text{s}^{-0.5}$, indicating a mixture of bedrock, bedform,
indurated surface, and unconsolidated materials (Fergason et al. 2012). Analysis of thermal inertia and local dune morphology variation provides insight into near-surface composition and local to regional climate characteristics (Courville et al. 2016), for instance, the presence of subsurface ice or volatiles. Besides thermal inertia derived from THEMIS nighttime thermal infrared (TIR) measurements, higher-resolution images compared to THEMIS, such as the Context Camera (CTX; Malin et al. 2007) and the High Resolution Imaging Science Experiment (HiRISE; McEwen et al. 2007), can be used as a visual survey tool for interpretation of thermal inertia (Fergason et al. 2012) and analyzing local tonal variability of dune fields. Compositional analysis of dune materials is key to better understanding sediment source and transport history. Bulk composition of dunes can reveal the extent of similarities (and dissimilarities) between aeolian sands and surrounding surface compositions, provenances (sources) of dune materials, transport behavior of aeolian medium, aeolian activity level, and alteration (if any) of aeolian sands versus their source materials (Fenton et al. 2019a, 2019b).

Fenton et al. (2019b) completed a complimentary science publication to MGD³ that describes mineralogy and morphologic stability of 79 large dune fields from TIR spectra at nearly global scale (80.7°S–41.7°N). Gullikson et al. (2018) expanded MGD³ to include thermal inertia and compositional information of dune fields in the equatorial and southern polar regions. However, the latest installments of MGD³ only included compositions for dune fields ≥300 km², and therefore compositional analysis for dune fields smaller than that threshold is warranted. Also, there is a need to compare dune field composition with surrounding materials to determine whether their potential provenance is at a local scale versus regional or global. The outlines of dune fields in MGD³ are not always accurate and overlook the recognition of smaller dune forms (Emran et al. 2020), likely due to the low-resolution THEMIS imagery used and the possibility of human error when manually outlining the dune fields. Thus, extracting information from an accurate dune outline is a challenge. With increasing image resolution, the accuracy of outlining dune fields also increases. Recently, Emran et al. (2020) accurately outlined an MGD³ dune field at Hargraves crater from CTX data using a semiautomated method to include small dune forms present within the crater that were overlooked by the MGD³-defined outline. Their research provides an excellent opportunity to study a small dune field (area of <100 km²) and compare it with potential provenances. The crater is located in the Nili Fossae region of the northwest Isidis basin (Figure 1). The region has received substantial attention for planetary exploration missions and has been extensively studied over the

Figure 1. (a) Hargraves crater is located east of Nili Fossae and northwest of Isidis Basin and is considered to be part of the Jezero watershed (Goudge et al. 2015). The background mosaic consists of MOLA colorized elevation data overlain onto a THEMIS image mosaic. The enlarged location (panel (a)) is shown in THEMIS daytime infrared images, where the yellow star indicates the location of the dune field within the crater. (b) The dune field in CTX image resolution. (c) Dune polygon layer as derived by Emran et al. (2019a). The red polygons highlight the dunes within the crater. The figure is recreated from Emran et al. (2020). North is up.
past decade (Salvatore et al. 2018, and references therein). Jezero crater, the landing site for Mars 2020 Perseverance rover, is also located in this region. Hargraves crater is considered to be part of the Jezero watershed (Goudge et al. 2015), though relatively little attention has been paid to this crater in comparison to Jezero crater and Nili Fossae.

We determine thermophysical and mineralogical characteristics of the dunes within Hargraves. The dune field outline in Emran et al. (2020) was chosen as the boundary of dune materials in this study (the dune database is found in Emran et al. 2019a). We analyze the surface grain-size distribution and bulk compositional (mineral abundances) characteristics across the dune field. Active dunes preclude the possibility of cementation and vertical variation of constituent materials, and therefore thermal inertia measurements can be used for grain-size analysis (Edwards et al. 2018). Grain-size analysis derived from THEMIS thermal inertia is considered to be a reliable measurement in this study because the dune field within Hargraves is potentially active (Emran et al. 2020) and covers a large area, i.e., its spatial extent spans across numerous THEMIS pixels.

We use CTX (∼6 m pixel$^{-1}$) and HiRISE (25–60 cm pixel$^{-1}$) images as a visual survey tool for analyzing tonal variability and the presence of induration morphology, i.e., degradational characteristics, such as erosional features (for example, mass wasting) or impact cratering, as well as surface mantling within the dune field. We determine morphologic stability of the dune field that measures the degree of modification by non-aeolian processes responsible for erosion and stabilization of the dunes (Fenton et al. 2019b). Martian dunes are classified into six stability index (SI) categories based on an inferred scale from 1 to 6 constructed upon the presence (or lack) of superposed non-aeolian features (such as gullies, small pits, mass wasting features, etc.) and apparent level of degradation by non-aeolian processes, i.e., erosion (Banks et al. 2018; Fenton et al. 2019a). Assigning an SI is a subjective procedure; dunes with no apparent stability features (i.e., an active dune field) are assigned an SI of 1, whereas SI values from 2 to 6 indicate increasing non-aeolian modification (Fenton et al. 2019a). Refer to Fenton & Hayward (2010) for details of dune stability classes and their characteristics.

We then compare the bulk mineral assemblage of the dune field to surrounding geologic units inside the crater (Goudge et al. 2015), which have been inferred as potential source materials. Martian dune sands are believed to have not traveled far from their sources and reflect varying mineralogy of surrounding local surface terrain (Fenton et al. 2019b). Thus, we hypothesize that the dunes at Hargraves are also locally sourced within the crater. Accordingly, we model spectral data from the Thermal Emission Spectrometer (TES; Christensen et al. 2001) instrument to determine the compositions and subsequently compare dune mineralogy with the surrounding geologic units at the crater.

2. Hargraves Crater

Hargraves crater is a 65 km diameter impact crater centered at 75.75°E, 20.75°N, east of Nili Fossae (Figure 1(a)), and formed during the late Noachian or early Hesperian (Ivanov et al. 2012). The impact event that created Hargraves crater involved an array of different target bedrock lithologies, including a diverse set of mineralogy, such as mafic minerals and phyllosilicates (Mangold et al. 2007). Based on tone, texture, and morphological characteristics, Goudge et al. (2015) identified a few major morphologic units inside the crater such as an alluvial fan (AF), crater central peak (Cep), crater floor materials (Cfm), crater ejecta (Ce), and surficial debris cover (Ac) (Figure 2). There is an outside-crater ejecta unit, which we hereafter refer to as the Ce (Outer) unit. A dune field, visible at CTX and THEMIS resolutions, is located on the western side of the crater floor (Figure 1(b)). The dune field is labeled as surficial debris cover (Ac) and is characterized by aggregates of dunes (Goudge et al. 2015). The MGD5 and previous studies (Emran 2019; Emran et al. 2019b, 2020) report that the dune field has barchan (B) and barchanoidal (Bd) dune types (see Figure 3(b); McKee & United States, National Aeronautics and Space Administration, Geological Survey (U.S.) 1979). Northwest of the dune field is an alluvial fan that was emplaced by fluvial activities along the crater rim (Mangold et al. 2007).

3. Observations and Methods

The bulk mineralogy was derived from TES data at spatial resolution roughly 3 × 6 km$^2$. We use THEMIS (∼100 m pixel$^{-1}$) data for additional compositional analysis. Though TES has lower spatial resolution compared to THEMIS, it has many more spectral bands than THEMIS, which allows bulk mineral assemblages to be modeled. THEMIS images were used to examine grain-size distribution using thermal inertia, as well as decorrelation stretch (DCS) as an additional compositional component.

3.1. Thermal Emission Spectrometer

TES, on board the Mars Global Surveyor (MGS) spacecraft, has an infrared (5.8–50 μm) interferometric spectrometer, a broadband thermal (5.1–150 μm) radiometer, and a visible/near-infrared (NIR; 0.3–2.9 μm) radiometer (Christensen et al. 2001). We use TES data recorded between 1651 and 201 cm$^{-1}$ with a sampling interval of 10 cm$^{-1}$, resulting in 143 channels. To ensure the best-quality data used in this study, we extract TES data following the method described in Rogers & Bandfield (2009) and Salvatore et al. (2018). TES spectra were extracted from early in the mapping orbit phase (orbits 1–5317) that belong to high surface temperatures (≥260 K), low total ice (<0.04), and low total dust cover (<0.15). This quality control practice has successfully been applied in previous studies (e.g., Rogers & Christensen 2007; Rogers et al. 2007; Salvatore et al. 2014, 2016) and helps to ensure the accuracy of the true spectral characteristics (Salvatore et al. 2018).

Individual spectra were collected only from TES footprints that fall over the dune field, hereafter referred to as dune unit (Du). For compositional comparison to inferred provenances, spectra were also collected over the crater floor materials (Cfm), crater ejecta (Ce), crater central peak (Ccp), and crater rim and wall materials (Crw) units. To ensure that the spectra represented individual geologic units, data were only collected when TES footprints were fully within the bounds of a unit. We limit our spectra collection to between orbit counter keeper (OCK) 1583 and 7000 to account for instrumental error. Based on measured slipface direction, Emran et al. (2020) reported that the prevailing wind inside the crater flows to the west–northwest. Since the alluvial fan (AF) is located downwind of the dune field, we did not consider the alluvial fan unit (AF) as
a potential source for the dunes. A list of TES spectra used for each geologic unit is given in the Appendix (Table 3).

3.1.1. Spectral Data Processing

TES spectral data were extracted using JMARS (Christensen et al. 2009). The data were then processed and analyzed with the DaVinci programming environment (Edwards et al. 2015). Prior to averaging the spectra for each representative morphologic unit, an atmospheric correction is applied to each orbital group (OCK) spectra using the deconvolution algorithm of Bandfield et al. (2000), Smith et al. (2000), and Rogers & Aharonson (2008). We used a library of atmospheric end-members (Bandfield et al. 2000) and treated each spectrum individually for atmospheric correction, which involves extracting surface emissivity through subtracting atmospheric contributions from measured emissivity spectra (Salvatore et al. 2018). Upon atmospheric removal, the surface emissivity spectra for each representative morphologic unit were averaged. The average surface composition and mineral assemblage were then modeled using an unmixing algorithm (Ramsey & Christensen 1998; Rogers & Aharonson 2008) with a selected mineral end-member library (see Table 1).

The algorithm uses a nonnegative least-squares minimization routine (Rogers & Aharonson 2008) between the wavelength range of 230 and 1305 cm$^{-1}$. The nonnegative least-squares routine employs all spectra in design matrix and remains available to the algorithm until a final nonnegative solution is reached (Rogers & Aharonson 2008). As the solution of the routine converges toward the best fit, a positive or zero spectral coefficient is maintained in the algorithm. However, the algorithm allows atmospheric and blackbody (BB) end-member coefficients to be negative. Details of the algorithm can be found in Rogers & Aharonson (2008). Using the spectral unmixing routine, we generated eight mineral groups (Rogers & Fergason 2011) for each unit with a $\sim$10% accuracy and detection threshold for mineral abundance (Feely & Christensen 1999; Christensen et al. 2000; Bandfield 2002; Rogers et al. 2007). The final outputs of spectral unmixing include modeled abundance mineral groups (feldspar, high-silica phases, pyroxene, olivine, hematite, sulfate, carbonate, and quartz), error estimation, and an individual end-member assemblage (Rogers & Fergason 2011).

3.1.2. Regression Models

We use linear regression to model the relationship between the bulk composition of each surrounding geologic unit and the dune field by fitting a linear equation to the mineral abundance. Since spectral unmixing generates abundances of eight mineral groups, i.e., end-member abundance with a mean ($\mu$) and uncertainty (1$\sigma$) value (Rogers & Fergason 2011), we resample each end-member assuming the bulk composition as a Gaussian scatter. We draw a total of 1000 normal random samples (125 for each end-member) from each surrounding geologic unit centered around $\mu$ within the limit imposed by the 1$\sigma$ of the end-member abundance. In our linear regression model, we assume the abundance of minerals in the dune field as the response ($y$) and each surrounding geologic unit as an explanatory variable ($x$). For Gaussian scatter, the linear regression model uses the equation

$$y = ax + b,$$

(1)

where the slope of the line is $a$, and $b$ is the intercept or offset. Both $a$ and $b$ are the parameters to estimate in the model. Using
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Table 1
The List of Thermal Infrared Spectral End-members (Both a Library of Atmospheric End-members and a Library of Surface End-member Spectra) Used for Unmixing TES Data in This Study

| Mineral Group | Spectrum/End-member | Name | Mineral ID | Reference |
|---------------|---------------------|------|------------|-----------|
| Quartz        | Quartz              | BUR-4120 | [1]        |
| Alkali feldspar| Microcline          | BUR-3460 | [1]        |
| Plagioclase   | Albite              | WAR-0235 | [1]        |
|               | Oligoclase          | BUR-060D | [1]        |
|               | Andesine            | WAR-0024 | [1]        |
|               | Labradorite         | BUR-3080A | [1]    |
|               | Bytownite           | WAR-1384 | [1]        |
|               | Anorthite           | BUR-340 | [1]        |
|               | Shocked anorthosite at 17 GPa at 25.5 GPa at 27 GPa at 38 GPa at 56.3 GPa | None | [2]        |
| Orthopyroxene | Bronzite            | NMNH-93527 | [1] |
|               | Enstatite           | HS-9.4B | [1]        |
|               | Hypersthene         | NMNH-B18247 | [1] |
| Low-Ca clinopyroxene | Average Lindsley pigeonite | None | [3] |
| High-Ca clinopyroxene | Diopside          | WAR-6474 | [1]        |
|               | Augite              | NMNH-9780 | [1]        |
|               | Augite              | NMNH-122302 | [1] |
|               | Hedenbergite manganoan | DSM-HED03 | [1] |
| Olivine       | Forsterite          | BUR-3720A | [1]        |
|               | Fayalite            | WAR-RGFAY01 | [1] |
|               | Olivine F0.35       | KI 3362 | [4]        |
|               | Olivine F0.38       | KI 3115 | [4]        |
|               | Olivine F0.35       | KI 3373 | [4]        |
|               | Olivine F0.38       | KI 3008 | [4]        |
| Phyllosilicates| Illite Imt-1 <0.2 μm (pellet) | Imt-1 | [5]        |
|               | Ca-montmorillonite solid | STx-1 | [1]        |
|               | Saponite (Eh-1) <0.2 μm (pellet) | None | [6]        |
|               | SWy-1 <0.2 microns (pellet) | None | [5]        |
| Glass         | K-rich glass        | None | [3]        |
|               | SiO2 glass          | None | [3]        |
| Amorphous silica | Opal A             | 01-011 | [7]        |
|               | Al-Opa              | None | [8]        |
| Oxide         | Average Meridiani and Aram Hematite (TT derived) | None | [9] |
| Sulfate       | Anhydrite           | ML-S9 | [1]        |
|               | Gypsum              | ML-S6 | [1]        |
| Carbonate     | Kieserite           | None | [10]       |
|               | Calcite             | C40 | [1]        |
| Zeolite       | Dolomite            | C20 | [1]        |
|               | Crystalline heulandite | None | [11]     |
|               | Crystalline stilbite | None | [11]     |
| Atmosphere    | Low-opacity dust    | None | [12]       |
|               | High-opacity dust   | None | [12]       |
|               | Water ice (small)   | None | [12]       |
|               | Water ice (large)   | None | [12]       |

| Mineral Group | Spectrum/End-member | Name | Mineral ID | Reference |
|---------------|---------------------|------|------------|-----------|
|               | Synthetic CO₂       | None | [12]       |
|               | Synthetic water vapor | None | [12] |

References. (1) Christensen et al. 2000; (2) Johnson et al. 2002; (3) Wyatt et al. 2001; (4) Koeppen and Hamilton 2008; (5) Michalski et al. 2006; (6) Michalski et al. 2005; (7) Michalski et al. 2003; (8) provided by M. D. Kraft/cited in Rogers and Ferguson 2011; (9) Glotch et al. 2004; (10) Baldrich 2007; (11) Ruff 2004; (12) Bandfield et al. 2000.

This model, we estimate the posterior probability distribution of the model parameters considering the uncertainty (1σ value) of the response (dune unit) using the Bayesian inference approach. The posterior probability distribution of the parameters of the linear model tells us the degree of similarity of the bulk mineralogy between the composition of each surrounding geologic unit and the dune field. Thus, the posterior probability distribution of the parameters indicates the likelihood of contribution of the bulk mineralogy from each geologic unit to the dune field’s bulk composition.

We employ a multiple linear regression model to estimate how much of each surrounding unit mixing with other units could produce the dune field composition. For that, we assume the volumetric (bulk) percent contribution of mineral compositions from each of the surrounding units to the dune field. We adopt this approach because it is the simplest and most physically possible scenario for the compositional mixing of geologic materials on Mars. In the case of volumetric percent contribution, the bulk compositions of the surrounding individual units (Xi) are summed with weights equal to the fractional contributed by each unit (Fi), as given by

$$Y = \sum_{i} F_i X_i + Z_0; 0 \leq F_i \leq 1,$$

where Y is the abundance of mineral end-member in the dune field, Z0 is the additive offset, and i represents the surrounding individual unit. In our model, the fractions contributed by each unit (Fi) and Z0 are the parameters to estimate considering the uncertainty (1σ) of Y. We estimate the posterior probability distribution of these parameters of the multiple linear model, which tells us the fractional contribution of bulk mineralogy from each surrounding unit to the dune field composition. The model uses a nonnegative linear combination (mixing) strategy, i.e., Fi must render a positive value.

We use a Markov Chain Monte Carlo (MCMC) simulation technique (Hogg & Foreman-Mackey 2018) to estimate the parameters of our models. Inherently the MCMC employs a Bayesian inference approach for sampling the probability density function, performing probabilistic inferences, and fitting models to data (Hogg & Foreman-Mackey 2018). For details of the mathematical foundation of the MCMC technique, refer to Goodman & Weare (2010) and Hogg & Foreman-Mackey (2018). We use emcee (Foreman-Mackey et al. 2013), a Python routine for MCMC simulation proposed by Goodman & Weare (2010), for the implementation of our models. With the MCMC routine, we use 1000 iterations to estimate the model parameters. We report the median
(including the 16% quantile as a lower 1σ error and the 84% quantile as an upper 1σ error) fraction contribution of the bulk composition of each surrounding unit using a corner plot diagram.

3.2. Thermal Emission Imaging System (THEMIS)

THEMIS consists of two multispectral imagers on board the 2001 Mars Odyssey spacecraft, enabling a useful analysis of the thermophysical, composition, and physical properties of the Martian surface (Christensen et al. 2004). The THEMIS sensor consists of visible (VIS), NIR, and TIR imagers. The TIR sensor has 10 channels ranging from 6.78 to 14.88 μm at a spatial resolution for daytime and nighttime images at 100 m pixel−1 (Christensen et al. 2004). We used THEMIS daytime IR to analyze surface composition and nighttime IR to analyze thermophysical characteristics.

Surface compositional and thermal inertia data can be extracted from representative surface units through overlapping THEMIS daytime and nighttime IR images. This allows an assessment of the relationship between composition and grain-size distribution (Bandfield et al. 2011). THEMIS daytime radiance data are atmospherically corrected, calibrated, and converted to surface emissivity using the methods described in Christensen et al. (2004), Bandfield et al. (2004a), and Edwards et al. (2011). We used the THEMIS I02781003 image acquired under an average surface temperature of 247 K, water ice opacity of 0.072, and low atmospheric dust opacity of 0.052 at 9 μm. THEMIS images with a similar specification of average surface temperature and low atmospheric dust and water ice opacity have successfully been used in previous studies (e.g., Bandfield et al. 2004b).

3.2.1. Thermal Inertia

Thermal inertia is considered as a primary physical property measuring the resistance of materials to a temperature change and can be used as an indicator of surface geological characteristics for the upper few centimeters (Jakosky et al. 2000; Mellon et al. 2000; Putzig et al. 2005). Thermal inertia is influenced by thermal conductivity and surface physical structure such as particle grain size, induration, porosity, rock abundance, bedrock exposure, and compaction of surface materials (Kieffer et al. 1977; Putzig et al. 2005; Ferguson et al. 2006; Piqueux & Christensen 2011; Putzig et al. 2014). Thermal inertia of material, \( I = \sqrt{(k/ρc)} \), can be defined as

\[
I = \sqrt{(k/ρc)}, \tag{3}
\]

where \( k \) is the thermal conductivity (W m\(^{-1}\) K\(^{-1}\)), \( ρ \) is the density (kg m\(^{-3}\)), and \( c \) is the specific heat (J kg\(^{-1}\) K\(^{-1}\)). On Mars, the value of \( ρc \) is assumed to be \( 1 \times 10^6 \) J m\(^{-3}\) K\(^{-1}\) (Neugebauer et al. 1971; Ferguson et al. 2006).

Ideally, higher thermal inertia values are associated with compacted rock and/or larger particle-size materials corresponding to a mechanically strong surface, such as well-indurated (cemented) rock. In contrast, weakly lithified rocks and/or unconsolidated (weakly indurated) materials are associated with lower thermal inertia values (Williams et al. 2018). Thus, lower thermal inertia \((< 350 \text{ Jm}^{-2} \text{K}^{-1} \text{s}^{-0.5})\) represents loose and fine-grained surface particles (e.g., Piqueux & Christensen 2009a, 2009b) whereas higher thermal inertia \((> 1200 \text{ Jm}^{-2} \text{K}^{-1} \text{s}^{-0.5})\) indicates igneous bedrock (Edwards et al. 2009). We extracted quantitative thermal inertia from the THEMIS global thermal inertia product (Christensen et al. 2013) produced using the method of Ferguson et al. (2006). The overall accuracy and precision of THEMIS-derived thermal inertia is around \(~20\%\) (Ferguson et al. 2006).

Effective grain size can be interpreted from thermal inertia values (Kieffer et al. 1973). We calculated the effective particle-size diameter of the dune materials using the following equation:

\[
k = C*P^{0.6}\times d^{-0.11} \log (\xi), \tag{4}
\]

where \( C \) and \( K \) are constants, with values of 0.0015 and \( 8.1 \times 10^4 \) torr, respectively; \( P \) is the atmospheric pressure in torr; \( d \) is the average particle diameter measured in μm; and \( k \) is the thermal conductivity in Wm\(^{-1}\)K\(^{-1}\), which was derived using Equation (3) (Presley & Christensen 1997b, 2010). Mars’s average surface pressure of 3.9 torr (5.2 mbar or 520 Pa) at \( L_s = 0^\circ \) (Smith & Zuber 1998) was used as the reference atmospheric pressure in this study. Using Equation (4), the derivation of particle diameter is valid for thermal inertia values less than \(~350\), and an expected precision of the measurements in particle size is \(~10\%~15\%\) (Presley & Christensen 1997b; Ferguson et al. 2006). Thermal inertia values higher than 350 are difficult to interpret and may be indicative of a solid rock outcrop mixed with fine-grained material within the footprint of thermal data (Presley & Christensen 1997b).

3.2.2. Decorrelation Stretch

Decorrelation is a contrast stretch method that displays enhanced information from multispectral TIR images. Refer to Gillespie et al. (1986) for details. We use DCS images (Edwards et al. 2011) derived as a standard THEMIS IR radiance product (Bandfield et al. 2004b) to highlight surface spectral and compositional variability at a spatial resolution much higher than TES. DCS images are typically rendered with three-band combinations such as 8-7-5, 9-6-4, and 6-4-2, displayed in red, green, and blue channels, respectively. The appearance of yellow in DCS images using bands 8-7-5 and 9-6-4 and of magenta using bands 6-4-2 tends to correspond with an elevated bulk-silica content (Amador & Bandfield 2016). Materials that appear purple in both 8-7-5 and 9-6-4 DCS images and cyan in 6-4-2 typically represent olivine-bearing basalts (Bandfield et al. 2011; Edwards et al. 2011; Amador & Bandfield 2016; Salvatore et al. 2016). We primarily focus on DCS image stretch using bands 8-7-5 in red, green, and blue channels, respectively, to highlight compositional variability present within the dune field and surrounding areas, because this band combination exhibits most variability. The color variability in DCS images is scene specific, and the spectral variability of DCS within a particular scene should be interpreted for that specific scene only because of its temperature dependence (Rogers & Ferguson 2011).

4. Results

The result of the grain-size distribution for the dune materials was presented first, followed by the results of dune morphology and stability, and bulk mineralogy of the dune field and the potential provenances. Then, the result of bulk mineralogy from the TES analysis was revisited using THEMIS-derived analysis of DCS image.
4.1. Grain-size Distribution

THEMIS nighttime data are more reliable than daytime data when deriving thermal inertia and estimating grain-size distribution because nighttime observation minimizes the effects of slope and albedo (Edwards et al. 2018). Quantitative thermal inertias were extracted using the THEMIS global thermal inertia mosaic by Christensen et al. (2013) for the dune field. The estimated average thermal inertia across the dune field is $238 \pm 17^3 \text{Jm}^{-2} \text{K}^{-1} \text{s}^{-0.5}$, with a minimum value of $200 \text{Jm}^{-2} \text{K}^{-1} \text{s}^{-0.5}$ and a maximum of $426 \text{Jm}^{-2} \text{K}^{-1} \text{s}^{-0.5}$. This value is consistent with the average thermal inertia measured in the Bagnold dune field using multiple THEMIS observations (Edwards et al. 2018). The thermal inertia map (Figure 3(a)) shows a concentration of relatively similar values within the outline of the dune field. Visual inspection indicates a comparatively lower thermal inertia across the dune field when compared to the surrounding geologic units. For instance, the crater floor materials (Cfm) on the east side of dunes have an average thermal inertia value of $\sim 400 \text{Jm}^{-2} \text{K}^{-1} \text{s}^{-0.5}$.

We calculate the average particle size across the dune field using Equations (3)–(4). We convert each THEMIS thermal inertia pixel to particle size instead of converting average thermal inertia to average particle size. We adopted this method because the conversion of thermal inertia to particle size is nonlinear (Edwards et al. 2018). The average grain-size values were calculated after converting each thermal inertia pixel to grain size, which resulted in an average particle size of $\sim 391 \pm 172 \mu m$. This particle-size diameter corresponds to mostly medium sand-sized materials mixed with fine and coarse grain sands (Presley & Christensen 1997a).

4.2. Morphology and Stability

Like the typical dunes elsewhere on Mars (Hayward et al. 2007b, 2010, 2012; Fenton et al. 2019b), the dune field on the floor of Hargraves crater is darker in tone when compared to surrounding materials (Figures 3(b)–(c)). A consistent tone across the dune field suggests that dune surface materials are relatively homogenous. Visual inspection of high-resolution morphology from CTX and HiRISE images finds an absence of in situ bedrock surface within the dune field, except exposure of a light-toned feature in the southeastern fringe (Figure 4(a)). The feature has a semiconical shape and appears to be layered. The layers dip away from the central point and have been eroded to the extent that they show the layered nature of this feature. A central vent, which is in line with some well-defined fractures, shows lighter shaded materials residing below the eroded layers that enshroud the feature. The feature can be interpreted to be a “possible” intrusion of igneous origin, developed from small (isolated) volcanos (Richardson et al. 2021), supported by the nearby presence of the Syrtis Major volcanic province (Hiesinger & Head 2004). However, this interpretation is based on our visual inspection, and other possible interpretations of this feature may also be plausible. The physical relationship of this feature to surrounding dunes does not, unambiguously, reveal the relative age of the feature.

We investigate fine-scale induration morphologies, non-aerological modification (or erosion), and degradation characteristics within the dune field, such as erosional features or impact cratering from HiRISE images ESP_030302_2010 and ESP_030091_2010 (Figure 5). An investigation of these images
reveals a sand apron on the western boundary of the dune field. We also find potential mantling on the boundary between the dunes and sand apron, as well as impact cratering on the mantling material. The presence of potential mantling along the boundary that appears to expand into the dune field by filling in space between the dunes, small impact craters, and an average thermal inertia value of $\sim 250 \text{ J m}^{-2} \text{K}^{-1} \text{s}^{-0.5}$ in the dune field may be an indication of weak induration or lithification of the dune materials (Fergason et al. 2012). We did not find other non-aeolian or erosional features like gullies or mass wasting features. However, the dunes on the southeastern boundary of the dune field show a different scenario. The CTX (see Figures 1(b) and 4(b)) image reveals thin dark streaks emanating from the dunes on the southeastern boundary of the dune field. There are features atop several dune crests and upper windward sides, such as sets of straight or very nearly straight
Table 2

| Mineral Group       | Dune (Du) | Crater Floor Material (Cfm) | Crater Ejecta (Ce) | Central Peak (Cp) | Central Wall (Ccw) | Crater Rim and Wall (Crw) |
|---------------------|-----------|-----------------------------|-------------------|------------------|--------------------|--------------------------|
| Feldspar            | 36 ± 7    | 24 ± 6                      | 39 ± 4            | 27 ± 6           | 36 ± 7             |                          |
| Pyroxene            | 22 ± 6    | 29 ± 4                      | 29 ± 4            | 21 ± 5           | 21 ± 5             |                          |
| Olivine             | 12 ± 5    | 7 ± 5                       | 3 ± 2             | 8 ± 4            | 7 ± 3              |                          |
| High-silica phase Carbonate | 16 ± 7    | 23 ± 5                      | 15 ± 4            | 22 ± 6           | 21 ± 5             |                          |
| Sulfate             | 5 ± 1     | 3 ± 1                       | 4 ± 1             | 4 ± 1            | 3 ± 1              |                          |
| Hematite            | 3 ± 2     | 10 ± 1                      | 7 ± 1             | 12 ± 3           | 9 ± 2              |                          |
| Quartz              | 4 ± 1     | 4 ± 1                       | 4 ± 1             | 6 ± 2            | 6 ± 1              |                          |
| Blackbody           | 1 ± 2     | 0 ± 0                       | 0 ± 0             | 0 ± 0            | 1 ± 1              |                          |
| RMSE                | 0.25      | 0.22                        | 0.28              | 0.26             | 0.17               |                          |

Note. Reported here are the average areal abundances of each mineral group (%) along with the calculated model error (RMSE). All values were normalized to the BB percentage (%).

grooves, akin to aeolian ripples, of several different widths, orientations, and spacings. All the various sets of grooves have in common that they are roughly perpendicular to the crest of the dunes on which they reside. There are also features on the dunes including curved and scoured grooves that follow trellis and dendritic patterns (Figure 5(b)). These curved and scoured features are closer to the crests of the dunes on which they reside than the nearby straight or nearly straight grooves, and in some instances they persist down the leeward (avalanche) face of the dune. Both of these groove features exhibit shallow penetration into the dune structure and reveal lighter color sands. These bright-toned grooves are interpreted to be Martian aeolian mega-ripples, often referred to as transverse aeolian ridges (TARs; Silvestro et al. 2020).

We assign the dune field within Hargraves crater an SI of 2, since it has a partial apron that exists along the western side of the dune field (Fenton & Hayward 2010). Note that the dune field may overlap the morphological characteristics between multiple SI classes and assignment to an appropriate class is a subjective procedure. Southern high-latitude dunes show increased stability and inactivity (Fenton et al. 2019a) due to the presence of subsurface (near-surface ground) water ice that could stabilize dunes, and their thermal model generally fits the rock/ice thermal signature (Gullikson et al. 2018). Unlike the southern high-latitude dune fields (Fenton & Hayward 2010; Gullikson et al. 2018; Fenton et al. 2019a), the dunes within Hargraves crater may not be affected by subsurface ice or volatiles (Emran et al. 2020). We presume this because the dunes at the crater are potentially active and deprived of stability features and morphologies—indicating the absence of influences from subsurface (near-surface ground) water ice that would otherwise stabilize the dunes. Subsurface ice extends as far equatorward as 45°S between 40°E and 140°E longitudes (Fenton & Hayward 2010). Martian water-equivalent hydrogen-rich deposits of >20% (by mass) at poleward of ±50° latitudes and less rich deposits near equatorial latitudes were reported from neutron data observed using the Neutron Spectrometer aboard 2001 Mars Odyssey (Feldman et al. 2004). Thus, residing in the equatorial region (~20°N), the dunes at Hargraves are less likely to be affected by subsurface ice or volatiles.

4.3. Bulk Composition

We model bulk mineralogy across the dune field (Du) and surrounding geologic units within the crater. We then compare the dune composition with potential provenances: crater floor materials (Cfm), crater ejecta (Ce), central peak (Ccp), and crater rim and wall (Ccw) units. The bulk composition analyzed here consists of different mineral groups (Rogers & Ferguson 2011) based on the spectral library used in the unmixing analysis from TES emissivity data. Spectral unmixing generates eight mineral groups, such as feldspar (both plagioclase and alkali feldspar), pyroxene (both high-Ca clinopyroxene [HCP] and low-Ca clinopyroxene [LCP]), high-silica phase (HSP; including amorphous silica and phyllosilicates), sulfate, olivine, hematite, carbonate, and quartz. Our compositional modeling results are presented in Table 2. Individual mineral end-member abundances are listed in the Appendix (Tables 4, 5, 6, 7, and 8). We report the unmixing result of model abundances to nearest integer value and model error to the hundredths decimal place as similarly reported by previous studies (e.g., Rogers & Christensen 2007). Though the model produces mineral abundances out to several decimal places, TES unmixing results are prone to ~10% uncertainty (Rogers et al. 2007) regardless of the actual model error reported. Furthermore, the model error in TES unmixing results is strongly tied to mineral end-member selection. The precision of mineral abundances and model error we report in this study are based on what the unmixing model has produced.

Our unmixing results reveal that the dune field is mainly composed of a mixture of feldspar, pyroxene, olivine, and other silicate minerals. The Du unit has both the highest concentration of olivine (12% ± 5%) and a lower concentration of...
high-silica phase minerals (16% ± 7%) compared to surrounding units. For example, the Ccp unit has the second-highest olivine abundance, modeled at 8% ± 4%, but a higher concentration of high-silica phase minerals (22% ± 6%). The Ce unit has the highest concentration of feldspar (39% ± 4%) but has the lowest abundance of olivine (3% ± 2%). The Cfm unit has the highest abundances of high-silica phases (23% ± 5%) but the lowest feldspar concentration (24% ± 6%). The average surface emissivity spectra and modeled spectra for the five geomorphic units within the crater are shown in Figure 6.

We model the relationship between the bulk mineralogic composition of the geological units within the crater and the dune field. Our results from the linear regression show the posterior probability distribution of slope and offset (intercept) of the model fit for each geologic unit (Figure 7). The values for the posterior probability distribution of the slope parameter in Figure 7 indicate the degree of similarities, and thus the likelihood of contribution by, each of the surrounding units has to the dune field. The Ce unit has the highest median slope value of 0.72 ± 0.02, followed by the Crw (0.64 ± 0.02), Cfm (0.60 ± 0.01), and Ccp (0.54 ± 0.01) units. Based on the median slope values from the posterior probability distribution, we suggest that the likelihood of contribution of the bulk mineralogy in the Ce unit to the bulk composition of the dune field is the highest, whereas the Ccp has the lowest likelihood within the crater.

We model the bulk compositions of the geological units within the crater to estimate the possible fraction amount contributed by each unit to the dune field. We show the result of the posterior probability distribution of the parameters of our multiple linear regression model in Figure 8. That is, the values for the posterior probability distribution for each of the units indicate the fractional contribution from each of the surrounding units to the dune field. The results of our modeled parameters show that the posterior probability distribution of the Ce unit has dominantly the highest median fraction of
0.66$_{+0.02}^{-0.02}$, followed by a disproportionally lower fraction contributed by the Crw (0.07$_{-0.03}^{+0.03}$) and Cfm (0.01$_{-0.00}^{+0.00}$) units. Our model did not predict any fraction contribution by the Ccp unit as reported in Figure 8. This indicates that the dune materials are, likely, predominantly sourced from the adjacent Ce unit within the crater—consistent with the linear model result described above. The Crw and Cfm units within the crater contribute to the dune composition mixture at lower fractions.

To assess the possibility of potential contribution from the geological units outside the crater, we further run MCMC simulation by including the outside-crater ejecta, Ce (Outer), unit located on the eastern and southeastern sides of the crater (see Figure 2). We consider the Ce (Outer) unit because it is the largest geological unit, extending to around ~30 km from the crater rim and wall (Crw), outside the crater. The selection of this unit is also based on the dune field’s raw slipface direction, indicating wind movement, and its upwind position with respect to the dune field. To minimize possible spectrum selection-induced bias, we used a maximum number (24 observations) of TES spectra from the Ce (Outer) unit. We adopt the same spectral unmixing procedure to estimate the bulk composition of the Ce (Outer) unit.

We first fit the linear model to the bulk mineralogy of the Ce (Outer) and dune field units (Figure 9). The result shows a median slope value of 0.52$_{-0.01}^{+0.01}$, the lowest likelihood unit to dune materials among the units within and beyond the crater. Using the same multiple linear regression model, we repeat the procedure of MCMC simulation using bulk mineralogy of the geologic units within the crater along with the Ce (Outer) unit. Our modeled results of the posterior probability distribution did not confirm any contribution from the Ce (Outer) unit. The median fraction contribution from the Ce (Outer) unit reports no value and did not alter the median fraction contribution by geological units within the crater as reported above (Figure 10). However, this presumption is based on the modeled results using the MCMC technique adopted here and may not be true.

Note that, even though the outside-crater ejecta unit has the pyroxene and high-silica phases that are very similar to the dune field, our model is based on the overall bulk composition of the geologic units, and therefore the result of the linear models is not determined by individual mineral end-members (phases) alone or a combination of a few. Thus, although multiple mineral end-member phases have similar abundances, the overall bulk mineralogy of the geologic units determines the fractional contribution to the dune field owing to the nature of the model used. The bulk mineralogy from TES spectra...
shows compositional disparities between the ejecta on the floor of the crater (Ce) and the outside-crater ejecta blanket, Ce (Outer) units. An interpretation of reasons of the differences between the bulk mineralogy of the ejecta on the floor of the crater and the crater’s ejecta blanket is beyond the scope of the current study. A further study can be carried out to understand the reasons for these disparities by considering the mechanism of crater formation and subsequent physical and chemical weathering processes, for instance.

For reference, the list of all TES spectra used for the Ce (Outer) unit and its bulk mineralogy are reported in the Appendix (Tables 9 and 10). We adopt this strategy because our models predict that the composition of the Ce (Outer) unit is less significant in interpreting the mineralogy of the dune field at the crater, and thus we restrict our subsequent analyses and discussion to the geological units within the crater units only.

The composition of the dune field was qualitatively revisited using DCS images at different band combinations, primarily of 8-7-5 bands, from THEMIS daytime IR observations (Figure 11). We used the DCS image because it can be used as a proxy tool for identifying olivine-bearing basalt and silica content (Edwards et al. 2011; Amador & Bandfield 2016). The dune field does not show dominance of a single color or distinctive spectral feature in DCS combinations. The 8-7-5 DCS has a mixture of dark purple, bright magenta, and yellow colors stretching across the dune field, suggesting a mixture of olivine, pyroxene, and Fe/Mg-smectite, which is a typical mineral assemblage for the Nili Fossae area (Goudge et al. 2015 and references therein). The color variations of the DCS band combinations of 9-6-4 and 6-4-2 are in agreement with the interpretation of the band combination 8-7-5. For instance, the dark purple appearance on crater floor materials in DCS 8-7-5 and 9-6-4 bands is consistent with the appearance of cyan color in DCS 6-4-2 bands, representing olivine-bearing basalts (Bandfield et al. 2011; Edwards et al. 2011; Amador & Bandfield 2016; Salvatore et al. 2016). The dune field also shows a mixture of different colors in DCS band combinations of 9-6-4 and 6-4-2, indicating a mixture of minerals.

5. Discussion

The thermal inertia results suggest that the dunes are composed of mostly medium sand-sized materials mixed with fine and coarse grain sands. Martian aeolian dune fields are
likely composed of particle size of homogeneous materials within a dune field that reduces the possibility of surface-mixture-induced anisothermality (Edwards et al. 2018). Thus, grain-size analysis from THEMIS thermal inertia characteristics for this dune field is considered ideal. We use the thermal inertia values across the dune field to estimate the grain-size distribution. However, thermal inertia can be an incredibly ambiguous metric of physical properties of Martian surface materials, and its interpretation is challenging owing to several features such as cementation, mixing of different particle sizes, presence of duricrust, ice exposure, sub-pixel-scale slope, etc. (Fergason et al. 2006; Piqueux & Christensen 2009a). For instance, low thermal inertia (< 100 J·m⁻²·K⁻¹·s⁻¹/²) can result from unconsolidated dust mantling the surface of the dune field (Presley & Christensen 1997b; Fergason et al. 2012). There are also possibilities where both vertical (layering) and lateral (horizontal mixing) mixtures (e.g., Fergason et al. 2006; Presley & Christensen 1997b; Piqueux & Christensen 2011) could also produce a low thermal inertia value. We did not incorporate these alternative explanations for the interpretation of thermal inertia. Examination of these alternative explanations needs a future study.

The presence of potential mantling, impact cratering on the mantling surface, and thermal inertia values in the dune field can be an indication of material induration (Fergason et al. 2012). Dunes within the crater are likely composed of weakly indurated surfaces, where there might be a consolidation that occurred in some degree to the surface and mixed with unconsolidated materials. This interpretation is consistent with the fact that a thermal inertia (~200 s) surface likely represents an aeolian bedform overlying an indurated regolith surface (Fergason et al. 2012). However, the presence of thin dark streaks on the southeastern boundary suggests a different scenario. Based on the relation between dark streaks, dunes, and subjacent terrain, the dark streaks are interpreted as the result of wind activities (Edgett 2002). Gusts of wind can blow the materials out and result in streaks of deposited sands (Thomas et al. 1981; Geissler et al. 2008) toward the downwind direction (west—northwest) into the dune field. Thus, the simplest explanation of these dark streaks can be the indication of the prevalence of wind activities (Edgett 2002) within the crater and mobilization of the dune materials. Along the dune faces, near the crests on the western boundary, appear to be mega-ripples (i.e., TARs) spaced several meters apart. Bright-toned TARs in equatorial areas, for example, Syrtis Major near Hargraves crater, suggest an active ripple migration and the presence of strong wind at the surface (Silvestro et al. 2020). Thus, we hypothesize that the dune field at Hargraves is potentially active—consistent with the study of Emran et al. (2020).

Based on the unmixing results, regression model results, and proximity of available sources of erodible materials (supported further by morphology and stability section), we suggest that the dune materials were likely derived from physical weathering, especially aeolian erosion, predominantly from the crater ejecta unit at the crater, mixed with a small amount from the...
crater floor and crater rim and wall lithologies. This is inferred owing to their similar compositions and proximity to the dune field, situated at upwind locations with respect to the dune field. Our regression model infers a possible combination of the fraction contributions (as shown in the posterior probability distribution) from the geological units within the crater to the dune field. Although the similarities of bulk compositions of surrounding geological units and the dune materials give the impression that the dunes are likely locally sourced, the use of regression models quantitatively strengthens that claim, because the regression models estimate the degree of similarities and the combination of fractional contribution by the geologic units to the dune field. However, we do not rule out the possibility that the dunes may be sourced from either one location or a varying combination of multiple provenances, e.g., the mixing of two units versus input from all units within the crater. Furthermore, there is also a possibility that the dune field was sourced from only one unit that may have undergone alteration or chemical weathering that could have altered the dune field’s composition.

Erosion and transport processes typically modify the relative modal abundances of constituent grains, such as the local example here, where olivine is apparently enriched in the dunes. The higher concentration of olivine in the dunes versus provenance sites cannot be attributed only to the density of the olivine grains (2.9 gm cm\(^{-3}\)), which is slightly less than the companion pyroxene grains (3.2–3.5 gm cm\(^{-3}\); Lapotre et al. 2017, 2018). Lapotre et al. (2017) suggested grain size and/or grain shape as mitigating factors in understanding compositional analysis of other Martian dunes. One of these factors likely causes olivine grains to be preferentially left behind as a lag deposit, particularly on the windward side of dunes such as those in the study area and the interdune areas (Greeley et al. 1999, 2002; Jerolmack et al. 2006; Mangold et al. 2011; Hooper et al. 2012). Rather than being a pervasive lag deposit, olivine may be sequestered specifically in small coarse-grained ripples and other fine-scale bedforms of the windward sides and interdunes, which are very common in dune areas of Mars (Hooper et al. 2012; Lapotre et al. 2017, 2018).

Olivine grain shape cannot be assessed directly, but the lack of cleavage in olivine and the pervasive presence of strong cleavage in pyroxenes may result in more angular and more nearly equidimensional grains as fundamental crystallographic characteristics of the olivine grain population (Klein & Philpotts 2016). Wind drag across coarse-grained Martian ripples has been suggested to cause infiltration of finer, less equidimensional grains (e.g., pyroxenes; Hooper et al. 2012). These factors, combined with the slight tendency for olivine grains to be slightly larger than companion pyroxenes in Mars aeolian systems (Lapotre et al. 2017, 2018), suggest that olivine may compose a smaller amount of surficial grains in the dune-mantling ripple deposits envisioned here. However, the presumption that the olivine-pyroxene grain size and shape differences lead to pyroxene filling pore spaces is based on our modeled results, and other possible explanations may also be plausible.

Figure 11. Three-panel DCS images using bands 8-7-5 (left), 9-6-4 (middle), and 6-4-2 (right) for the THEMIS image I02781003 overlain on the radiance image of the same scene. The geomorphic units are labeled on the DCS images: dune (Du), central peak (Ccp), floor materials (Cfm), crater ejecta (Ce), and rim and wall materials (Crw) in the color combination across all three stretches. A mixture of dark purple, bright magenta, and yellow colors in DCS 8-7-5 bands stretching across the dune field indicates a mixture of olivine, pyroxene, and Fe/Mg-smectite. The dark-purple appearance on crater floor materials in DCS 8-7-5 and 9-6-4 bands is consistent with the appearance of cyan color in DCS 6-4-2 bands. The yellowish appearance (left panel) in the alluvial fan corresponds to an elevated bulk-silica content.
Orbiter-based observations from spectroscopic measurements suggest that active dune fields and sand sheets are largely composed of mafic minerals, such as plagioclase feldspar, high and low calcium pyroxene, and olivine (Achilles et al. 2017; Cousin et al. 2017; Ehlmann et al. 2017; Lapotre et al. 2017; Rampe et al. 2018). A strong olivine signature has also been reported elsewhere on Mars in barchanoidal-type dunes along the upwind margins of dune fields, for instance, in the Bagnold dune field as observed by high-resolution VSWIR orbital data from the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM; Seelos et al. 2014; Lapotre et al. 2017; Rampe et al. 2018). The barchanoidal-type dunes at Hargraves crater may be another instance of a dune field with elevated olivine content. We model average spectra extracted over the entire dune field (not separately from upwind and downwind margins) from a comparatively lower resolution TES footprint, and thus we cannot confirm whether there is olivine enrichment along the upwind margins. However, this may be the scenario where enriched olivine content along the upwind margins of the barchanoidal dunes at Hargraves contributes to an enhanced olivine abundance in our modeled composition.

A lower abundance of high-silica phase is reported in the dune field. Martian sands sourced from local to deeper craters typically contain a proportionally lower abundance of high-silica phase (Fenton et al. 2019b). Dune fields with a lower abundance of high-silica phase are typically found on deeper crater floors, whereas those with an elevated abundance (>25%) are found on very shallow crater floors or intercrater plains (Fenton et al. 2019b). This statement is consistent with the geographical position of Hargraves crater and the relatively low abundance of high silica within the dune field. Abundance of high-silica phase can be correlated with morphological characteristics such as stabilization features and bedform activity. Dune fields with elevated abundances of high-silica phase are found in southern high latitudes (60°-70°S), where morphological characteristics of increasing stability are common (Fenton et al. 2019b). With an SI of 2, the dune field within Hargraves shows a relatively lower abundance of high-silica content. The yellowish appearance in DCS 8-7-5 for the alluvial fan unit represents a higher concentration of bulk-silica content (Amador & Bandfield 2016). We assume that the differences in silica content between the alluvial fan and the dune field are due to the different geologic processes that are responsible in forming these features, as well as their source material. The alluvial fan is likely sourced from the crater rim or outside the crater and deposited on the crater floor through fluvial processes.

Having a paucity of high-resolution observations from the CRISM (Murchie et al. 2007) across the crater floor covering all geologic units, we could not analyze reflectance spectra for the surface mineralogy. We did not use reflectance data from the Observatoire pour la Minéralogie, l’Eau, les Glaces et l’Activité (OMEGA; Bibring et al. 2004) sensor because the instrument has a spatial resolution of 300 m to 5 km pixel⁻¹, which is comparatively lower spatial resolution than found in both CRISM and THEMIS.

Bulk mineralogy in this area from TES data by Salvatore et al. (2018) indicates that the crater floor unit (combining the Cfm and Ccp units) has a roughly comparable phyllosilicate and amorphous component (PAC) to nearby Syrtis basalt. Their study also reported a comparable amount of plagioclase, pyroxene (both high and low calcium), and olivine to this study. There is a slight discrepancy between our result and Salvatore et al. (2018) when describing the bulk mineralogy of crater floor materials (e.g., olivine abundance of 7% ± 5% vs. 7% ± 3%, respectively), which is likely because they averaged spectra from the central peak and crater floor units into a single unit, as well as using different unmixing algorithms. Due to a scarcity of quality CRISM data across the crater floor, we cannot assess finer-scale individual mineral distributions for the units independently. However, the DCS image at different band combinations and the result from bulk mineralogy of TES data provide substantial information about the constituent materials of the dune field and its surrounding morphologic units.

6. Conclusion

We analyzed the TIR response of the dune materials to identify the grain-size distribution and surface mineral composition of the dune field at Hargraves crater. We hypothesize that the surrounding units within the crater are the provenances for the dune field, and this paper provided modeled compositions for these units. Accordingly, we compared the dune mineralogy with the composition of surrounding geological units.

We found the average thermal inertia for the dune materials to be $238 \pm 17 \text{ J m}^{-2} \text{ K}^{-1} \text{s}^{-0.5}$, indicating a surface composed of an average effective grain size of $\sim 391 \pm 172 \mu m$. This size range indicates that the dune field is likely composed of mostly medium sand-sized materials mixed with fine and coarse grain sands. The dunes are likely composed of weakly indurated surfaces, mixed with unconsolidated materials. In this study, the determination of particle size was based on the derived thermal inertia of surface dune materials. A future study can be carried out to investigate the possibility of vertical (layering) and lateral (horizontal mixing) mixtures in the dune field.

The results from our TES compositional modeling and THEMIS DCS image analysis suggest that the dune materials are composed of a mixture of feldspar, olivine, and pyroxene, and with relatively lower bulk-silica content than the surrounding geologic units. Compositional information of the dune materials and the surrounding geologic units suggest that the dunes were likely sourced locally within the crater, predominantly from the crater ejecta, mixed with a small amount from the crater floor and crater rim and wall lithologies. The erosion and transport process likely have modified the constituent grains to some degree. We consider that the dune materials were sourced through physical weathering of the inferred provenances, based on our TES unmixing results. It is likely that through erosional and transport processes constituent grains were modified; however, a future study is needed to investigate the possibility of alteration or chemical weathering of source materials.

Comparison of bulk mineralogy of dunes and surrounding geologic units provides useful clues on local versus distant provenances of aeolian materials. Moreover, the identification of the source provenances paves the way for a better understanding of the climate dynamics in the area of interest. For instance, the dunes at Hargraves crater are likely sourced from the geologic units within the crater—indicating the dominance of local-scale wind movement inside the crater and
responsible for the deposition of sands in the dune field. It also provides insight into the material transport mechanism and erosion pattern within the crater. Since the dunes are likely locally sourced, the influence of regional- and global-scale dust materials is minimal in the dune field. This result, coupled with the dune morphology and slipface orientation of the dunes, will help in future climatic studies and geologic processes acting on this region. On top of that, the understanding of the transport mechanism of dune materials from source to sink may provide information on recent climate changes in the region (Fenton 2005).

This study expands our knowledge in understanding the structure, composition, and aeolian transport regime of the dune field at Hargraves crater through investigating grain size, induration, modification by non-aeolian processes, and dune field stability. In addition, our compositional modeling helped to constrain likely provenances and whether the dune field was sourced locally, regionally, or globally. We interpret an active and weakly indurated, simultaneously, Martian dune field and showed the avenues for analyzing dunes that appear to exist on a spectrum exhibiting features of both. Our approach provides a systematic guide for interpretation and helps in expanding our understanding of both active and weakly indurated aeolian environments from the observations of sands on Mars.

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Appendix

A list of all used TES observations (A.1.) and spectral unmixing results of individual mineral end-members (A.2.) for the geologic units within the crater and a list of all TES spectra used for the Ce (Outer) unit and its bulk mineralogy (A.3.) are reported in this appendix.

A.1.

The list of TES spectra used for each geologic unit within the crater is given in Table 3.

### Table 3

| Unit                      | TES Observations |
|---------------------------|------------------|
| Dune (Du)                 | OCK  | ICK  | DET |
|                           | 3094 | 1950 | 3   |
|                           | 3094 | 1950 | 6   |
|                           | 3094 | 1949 | 3   |
| Crater ejecta (Ce)        | 3094 | 1947 | 6   |
|                           | 3094 | 1947 | 5   |
|                           | 3094 | 1947 | 3   |
| Crater central peak (Ccp) | 3421 | 1942 | 2   |
|                           | 3421 | 1942 | 1   |
|                           | 3421 | 1941 | 4   |
|                           | 3421 | 1942 | 6   |
| Rim and wall (Crw)        | 3421 | 1937 | 5   |
|                           | 3421 | 1938 | 1   |
|                           | 3421 | 1938 | 2   |
|                           | 3421 | 1938 | 3   |
|                           | 3094 | 1946 | 1   |
|                           | 3421 | 1937 | 6   |
|                           | 3094 | 1946 | 2   |
|                           | 3421 | 1938 | 6   |
| Crater floor materials (Cfm)| 3421 | 1940 | 1   |
|                           | 3421 | 1939 | 5   |
|                           | 3421 | 1940 | 2   |
|                           | 3421 | 1940 | 4   |
|                           | 3421 | 1940 | 5   |
|                           | 3421 | 1939 | 6   |

Note. For each unit, all TES observations were averaged and treated as a single TIR spectrum throughout the entirety of the study (see text for more information regarding the analysis of TES observations). OCK = Orbit Counter Keeper; ICK = Incremental Counter Keeper; DET = Individual Detector.

A.2.

Individual mineral end-members abundances for the geologic units within the crater are listed as Du (Table 4), Ce (Table 5), Ccp (Table 6), Cfm (Table 7), and Crw (Table 8). Reported here are the average areal abundances of end-members (%) and normalized for BB % along with model error (RMSE).
### Table 4
Dune (Du) Unit

| End-member                  | Abundance (%) | Normalized for BB (%) |
|-----------------------------|---------------|-----------------------|
| Bronzite NMNH-93527         | 10.20         | +/- 3.79              |
| Shocked An 25.5 GPa        | 7.90          | +/- 8.67              |
| Bytownite WAR-1384          | 7.43          | +/- 6.95              |
| Swy-1 <0.2 mic             | 6.28          | +/- 5.29              |
| KI 3362 Fo60               | 5.45          | +/- 2.93              |
| Anorthite BUR-340          | 4.07          | +/- 5.31              |
| Dolomite C20               | 2.99          | +/- 0.52              |
| Avg. Lindsley pigeonite    | 2.61          | +/- 4.15              |
| Average Martian            | 2.55          | +/- 0.86              |
| Hematite                   |               |                       |
| Microcline BUR-3460        | 1.98          | +/- 2.12              |
| Kieserite Kieserite        | 1.84          | +/- 1.11              |
| Crystalline heulandite (z) | 1.77          | +/- 5.49              |
| 02-011 Opal A              | 1.70          | +/- 1.50              |
| Quarz BUR-4120             | 0.80          | +/- 1.07              |
| KI 3115 Fo68               | 0.75          | +/- 6.68              |
| Fayalite WAR-RGFAYO1       | 0.66          | +/- 2.62              |
| KI 3373 Fo35               | 0.50          | +/- 5.23              |
| Augite NMNH-122302         | 0.32          | +/- 3.78              |
| Oligoclase BUR-060D         | 0.15          | +/- 5.11              |
| Enstatite HS-9.4B          | 0.08          | +/- 3.04              |

**Sum (BB included) = 99.99**  
**Sum (BB normalized) = 100.00**  
**Blackbody abundance = 19.77 +/−2.51**  
**RMS error (%) = 0.28**

### Table 5
Crater Ejecta (Ce) Unit

| End-member                  | Abundance (%) | Normalized for BB (%) |
|-----------------------------|---------------|-----------------------|
| Labradorite EIBUR-3080A     | 13.72         | +/- 7.70              |
| Avg. Lindsley pigeonite     | 12.05         | +/- 3.55              |
| Bronzite NMNH-93527         | 11.28         | +/- 1.85              |
| Bytownite WAR-1384          | 10.91         | +/- 7.98              |
| Crystalline heulandite (z) | 10.78         | +/- 4.96              |
| Anorthite BUR-340           | 6.34          | +/- 3.79              |
| Gypsum (Satin spar) IS6     | 3.29          | +/- 1.21              |
| Average Martian             | 3.05          | +/- 0.73              |
| Hematite                    |               |                       |
| Dolomite C20                | 2.92          | +/- 0.37              |
| Kieserite Kieserite         | 2.39          | +/- 0.81              |
| KI 3362 Fo60                | 1.50          | +/- 2.69              |
| Crystalline stilbite (zeo)  | 0.76          | +/- 2.33              |
| KI 3373 Fo35                | 0.62          | +/- 2.86              |
| 02-011 Opal A               | 0.62          | +/- 0.93              |

**Sum (BB included) = 99.99**  
**Sum (BB normalized) = 100.00**  
**Blackbody abundance = 19.77 +/−2.51**  
**RMS error (%) = 0.28**

### Table 6
Crater Central Peak (Cep) Unit

| End-member                  | Abundance (%) | Normalized for BB (%) |
|-----------------------------|---------------|-----------------------|
| Anorthite BUR-340           | 9.83          | +/- 4.53              |
| Bronzite NMNH-93527         | 8.94          | +/- 2.22              |
| Bytownite WAR-1384          | 7.43          | +/- 6.95              |
| Swy-1 <0.2 mic             | 6.28          | +/- 5.29              |
| Gypsum (Satin spar) IS6     | 5.45          | +/- 9.73              |
| Anorthite BUR-340          | 4.07          | +/- 5.31              |
| Dolomite C20               | 2.99          | +/- 0.52              |
| Avg. Lindsley pigeonite    | 2.61          | +/- 4.15              |
| Average Martian            | 2.55          | +/- 0.86              |
| Hematite                   |               |                       |
| Kieserite Kieserite        | 3.09          | +/- 1.08              |
| KI 3115 Fo68               | 3.00          | +/- 1.74              |
| Dolomite C20               | 2.36          | +/- 0.42              |
| KI 3373 Fo35               | 1.64          | +/- 2.46              |
| Microcline BUR-3460        | 1.05          | +/- 2.34              |
| Crystalline stilbite (zeo) | 0.73          | +/- 3.69              |
| saponite < 0.2 mic         | 0.66          | +/- 1.04              |
| 02-011 Opal A              | 0.39          | +/- 1.08              |
| Anhydrite S9               | 0.12          | +/- 0.80              |

**Sum (BB included) = 99.88**  
**Sum (BB normalized) = 100.00**  
**Blackbody abundance = 42.56 +/−2.63**  
**RMS error (%) = 0.26**

### Table 7
Crater Floor Material (Cfm) Unit

| End-member                  | Abundance (%) | Normalized for BB (%) |
|-----------------------------|---------------|-----------------------|
| Bytownite WAR-1384          | 11.84         | +/- 4.63              |
| Bronzite NMNH-93527         | 11.23         | +/- 1.84              |
| Crystalline heulandite (z) | 8.71          | +/- 4.12              |
| Augite NMNH-122302          | 3.75          | +/- 2.64              |
| Kieserite Kieserite         | 3.35          | +/- 0.66              |
| KI 3115 Fo68               | 3.21          | +/- 1.55              |
| Swy-1 <0.2 mic             | 3.09          | +/- 4.16              |
| Anorthite BUR-340          | 3.02          | +/- 3.83              |
| Avg. Lindsley pigeonite    | 3.02          | +/- 3.36              |
| Gypsum (Satin spar) IS6     | 2.69          | +/- 0.97              |
| Average Martian            | 2.28          | +/- 0.88              |
| Hematite                   |               |                       |
| Crystalline stilbite (zeo) | 1.67          | +/- 2.45              |
| Dolomite C20               | 1.56          | +/- 0.40              |
| Fayalite WAR-RGFAYO1       | 0.92          | +/- 4.25              |
| K-rich Glass               | 0.40          | +/- 2.24              |
| KI 3373 Fo35               | 0.25          | +/- 3.64              |
| KI 3008 Fo10               | 0.10          | +/- 4.52              |

**Sum (BB included) = 100.00**  
**Sum (BB normalized) = 100.00**  
**Blackbody abundance = 38.92 +/−2.22**  
**RMS error (%) = 0.22**
The list of all TES spectra used (Table 9) and modeled bulk mineralogy (Table 10) for the Ce (Outer) unit.

### Table 8
Crater Rim and Wall (Crw) Unit

| End-member                          | Abundance (%) | Normalized for BB (%) |
|-------------------------------------|---------------|-----------------------|
| Bytownite WAR-1384                  | 10.21         | 22.27                 |
| Crystalline heulandite (*z*)        | 6.54          | 14.26                 |
| Shocked An 21 GPa                   | 4.97          | 10.85                 |
| Bronzite NMNH-93527                 | 4.79          | 9.45                  |
| Avg. Lindsley pigeonite             | 4.33          | 5.45                  |
| Swy-1 <0.2 mic                      | 3.32          | 7.23                  |
| Average Martian                     | 2.72          | 5.94                  |
| Hematite                            |               |                       |
| Kieserite Kieserite                 | 2.27          | 4.95                  |
| Gypsum (Satin spar) IS6             | 1.68          | 3.67                  |
| Dolomite C20                        | 1.40          | 3.05                  |
| Anorthite BUR-340                   | 1.12          | 2.43                  |
| KI 3362 Fo60                        | 0.91          | 1.99                  |
| Quartz BUR-4120                     | 0.50          | 1.08                  |
| KI 3115 Fo68                        | 0.47          | 1.03                  |
| Microcline BUR-3460                 | 0.30          | 0.65                  |
| Enstatite HS-9.4B                   | 0.29          | 0.62                  |
| Albite WAR-0235                     | 0.02          | 0.04                  |
| Anhydrite S9                        | 0.01          | 0.03                  |

Sum (BB included) = 99.97

Sum (BB normalized) = 100.00

Blackbody abundance = 54.12 ± 2.32

RMS error (%) = 0.17

Note. All TES observations were averaged and treated as a single TIR spectrum throughout the entirety of the study (see text for more information regarding the analysis of TES observations). OCK = Orbit Counter Keeper; ICK = Incremental Counter Keeper; DET = Individual Detector.

### Table 9
A List of All TES Observations Used in This Investigation for the Ce (Outer) Unit

| Unit               | TES Observations |
|--------------------|------------------|
| Ce (Outer)         |                  |
|                    | OCK  | ICK | DET | OCK | ICK | DET | OCK | ICK | DET |
| 3836               | 1944 | 1   |     | 3836| 1944| 4   | 3836| 1943| 5   |
| 3836               | 1944 | 3   |     | 3836| 1942| 4   | 3836| 1945| 4   |
| 3836               | 1942 | 3   |     | 3836| 1944| 6   | 3836| 1945| 6   |
| 3836               | 1942 | 2   |     | 3836| 1942| 6   | 3836| 1948| 1   |
| 3836               | 1944 | 5   |     | 3836| 1945| 1   | 3836| 1947| 1   |
| 3836               | 1942 | 5   |     | 3836| 1945| 3   | 3836| 1949| 1   |
| 3836               | 1949 | 2   |     | 3836| 1949| 5   | 3836| 1948| 5   |
| 3836               | 1948 | 4   |     | 3836| 1948| 6   | 3836| 1949| 3   |

Note. All TES observations were averaged and treated as a single TIR spectrum throughout the entirety of the study (see text for more information regarding the analysis of TES observations). OCK = Orbit Counter Keeper; ICK = Incremental Counter Keeper; DET = Individual Detector.

### Table 10
Our Spectral Unmixing Result for Ce (Outer) Unit

| Mineral Groups | Feldspar | Pyroxene | Olivine | High-silica Phase | Carbonate | Sulfate | Hematite | Quartz | BB   | RMSE  |
|----------------|---------|----------|---------|-------------------|-----------|---------|----------|--------|------|-------|
| %             | 28 ± 4  | 25 ± 4   | 8 ± 2   | 17 ± 6            | 3 ± 1     | 9 ± 2   | 8 ± 2    | 2 ± 1  | 0.26 ± 3 | 0.21  |

Note. Reported here are the average areal abundances of each mineral group (%) along with the calculated model error (RMSE). All values were normalized to the BB percentage (%).
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