

Martian Dune Fields: Aeolian Activity, Morphology, Sediment Pathways, and Provenance

A Dissertation Presented for the
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Matthew Chojnacki
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DEDICATION

This dissertation is dedicated
To my family, for their support and patience while I pursued this degree.

To my Father, Bob who encouraged my interest in space and technology,
but always let me choose my own path.

To my Mom, Rosie who cheered me to win on many cold ski slopes,
all the while hoping I wouldn't get hurt.

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ABSTRACT

Wind has likely been the dominant geologic agent for most of Mars' history. The widespread nature of sand dunes there shows that near-surface winds have commonly interacted with plentiful mobile sediments. Early studies of these dunes suggested minimal activity, dominantly unidirectional simple dune morphologies, and little variations in basaltic sand compositions. This dissertation examines martian sand dunes and aeolian systems, in terms of their activity, morphologies, thermophysical properties, sand compositions, geologic contexts, and source-lithologies using new higher-resolution orbital data. Although previous evidence for contemporary dune activity has been limited, results presented in Chapter II show substantial activity in Endeavour Crater, Meridiani Planum. The translation and erosion of dunes there constitutes the largest contemporary movement of sand-sized sediment reported on Mars to date and demonstrates that Endeavour crater has been subject to wind profiles exceeding the threshold velocity at the surface (daily/seasonally and/or episodically) in the recent past. Global mapping has shown dune fields to dominantly occur in topographically benign locations (e.g., craters, polar basins), where the largest exception is the Valles Marineris (VM) rift system. Chapter III documents multiple occurrences of "wall dunes" found several kilometers above the canyon floor. These relatively unique dune morphologies show that wind blown sediment has interacted with local and regional topography and are relevant for understanding aeolian sediment flux, sediment sources, and wind directions. Chapter IV provides results of a low- to mid-latitude survey of all martian dune fields in comparison to the extensive dune population in VM to test the effect of local and regional environments on duneform properties. That study found VM dune fields to be qualitatively and quantitatively distinct from other dune populations, most readily attributed to the rift's unusual setting. Sources and pathways of the martian sand have largely been uncertain. Chapter V documents likely sediment sources and pathways for the VM dune populations, where local and regional derivation of dune sand has occurred. We find in some chasmata dune sand is dominantly derived from Noachian-aged (altered and unaltered) igneous wall materials, whereas in other chasmata dunes are sourced from Early Hesperian-aged sulfate-bearing sedimentary layered deposits.

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CHAPTER I: INTRODUCTION

1.0 Sand Dunes and Aeolian Systems

Planetary sand dunes are important sedimentary components of geologic environments that can both modify the landscape and respond to changes in climate. For bodies possessing sustained atmospheres, aeolian bedforms are likely one of the default depositional mechanisms that record the deep-time of past sediments in the geologic record. We know aeolian bedforms occur on four planetary bodies in our Solar System (Venus, Earth, Mars, and Titan) and are hypothesized to exist on Pluto. The presence, volume, morphology, texture, and composition of planetary aeolian bedforms provide valuable clues as to the hosting planetary bodies past geologic history. For example, the dearth of resolvable extensive dune fields on Venus suggests the thick CO₂-atmosphere was incapable of sustained near-surface transport of particulate material and/or the sand supply was limited. On Earth, Pleistocene and Holocene lithified dunes preserve evidence of Quaternary climate change and the pre-history exploits of early humans. Mars hosts a diversity of sand compositions, which indicate variability in crustal composition and/or chemical alteration. Extensive sand seas or ergs on Titan imply vast reserves of sediments, likely water ice-organics, are present on the surface and maybe the partial products of fluvial erosion and atmospheric chemistry. The potential discovery of dunes on Pluto would indicate aeolian processes could operate on worlds with tenuous atmospheres and frigid temperatures, and would broaden the reach of known aeolian phenomenon to the fringes of the Solar System. These vastly different planetary bodies each come with specific boundary conditions that create variations in the production, transport, accumulation, and composition of aeolian sediments, yet duneform morphologies and their collective patterns are largely analogous. Thus, planetary sand dunes and aeolian systems provide clues to sedimentary processes, crustal compositions, alteration processes, and near-surface atmospheric flow.

I choose the topic of martian sand dunes for my dissertation based on their potential implications for climate change, surface activity, and clues to crustal compositions. Additionally, I chose dunes after an extensive literature search on terrestrial and martian aeolian systems, which revealed several areas needing of study. Low-albedo dunes on Mars are largely dust-free surfaces that may allow sensitive spectroscopic and thermophysical measurements to be preformed without exterior contamination by dust, a prevalent component of the martian environment. These measurements can provide insight into the compositions and physical properties of dune sediment. Likewise, the presence and morphology of dune populations can yield information on Mars regional or global climatic trends. Analysis of bedform patterns may allow for the reconstruction of formative wind regimes, which are critical in our understanding sand pathways and climate change. There is additional excitement in investigating a morphology that has the potential for coming “alive” with activity on a planet that was widely held to be “dead” in many respects. I’ll also point out, many results and discoveries described herein would not have been possible without the emergence of new data, some of which were targeted for this investigation with instruments onboard the Mars Reconnaissance Orbiter (MRO), which went into martian orbit shortly before the onset of my research (section 4).

2.0 Historical Perspective I: Pre-MGS

Prior to the Mars Global Surveyor (MGS) orbiter arrival to Mars in 1997, the Mariner 9 and Viking Orbiters provided the first broad-scale views of martian aeolian phenomena. Many of these early missions observed global-encircling dust storms, suggesting a vast supply of light-toned, fine-grained particles ($<20\ \mu\text{m}$) was present in the atmosphere [Greeley *et al.*, 1992]. More importantly, these dust storms indicate aeolian sedimentary processes are active under

contemporary atmospheric conditions despite Mars having a surface pressure less than a hundredth of terrestrial values (~6 millibars). Studies using data from these earlier orbiters were able to report dark-toned aeolian deposits in terms of their gross morphology, color, activity, and physical properties. Low-albedo “splotches”, wind streaks, and dune fields were progressively discovered, categorized, and studied from orbit [*Sagan et al.*, 1973; *Arvidson*, 1974; *Breed et al.*, 1979; *Kieffer et al.*, 1992]. For example, dark-toned, intracrater duneforms and deposits possessed broad-scale thermophysical properties consistent with sand sizes [*Christensen*, 1988a; *Edgett*, 1994]. Variability in dark streak orientation and dune field albedo provided indirect evidence for saltation events, but no evidence of dune migration was detected [*Arvidson*, 1974; *Christensen*, 1988a]. Because of limitations in instrument resolutions and wavelengths, fundamental questions persisted, with respect to the hierarchy of aeolian bedforms, extent of dune field environments, sand compositions, and relations to sand sources.

3.0 Historical Perspective II: Pre-MRO

With the arrival of a fleet of orbiters between 1997 and 2003 and their suite of instruments, great progress has been achieved in the field of Mars aeolian science. For example, studies using thermal infrared data onboard MGS described low-albedo dune field compositions as largely basaltic with subtle variations in basaltic minerals and glasses [*Aben*, 2003; *Christensen et al.*, 2008; *Stockstill-Cahill et al.*, 2008]. Studies using visible and near-infrared reflectance (VNIR) data acquired from instruments onboard Mars Express (MEx) confirmed the prevalence of mafic dune-forming minerals (e.g., pyroxene, olivine), but also detected gypsum-bearing dunes in the high northern latitudes [*Langevin et al.*, 2005; *Tirsch et al.*, 2011]. However, distinguishing between dunes and intradune area using the relatively low spatial

resolutions of these early spectrometers made conclusions vague in terms of sand mineralogical variability and evidence for alteration.

Photogeologic analysis using images from cameras onboard Viking and MGS detected no dune migration between the 14 Mars year time span between missions [*Edgett and Malin, 2000b*]. Hints at activity occurred with reports of slip face activity and dune deflation [*Fenton, 2006; Bourke et al., 2008*], but at the start of this dissertation (2008) no evidence of bulk dune migration were reported.

Although aeolian processes appear to be the dominant geologic process in Mars' recent history, many questions remain regarding the level of activity, range of morphologies, and possible sediment sources.

For example:

- Why isn't aeolian bedform change more commonly observed from orbit?
- What factors are controlling aeolian bedform movement?
- What is the extent of dune fields in locations of high-roughness and variable topography?
- Do sand compositions vary greatly with region and is there evidence for extensive alteration products?
- When did the sand dunes form?
- What are the sources of Mars dune sand? Are sources local, regional, or distant?
- What factors control sand availability?
- How can sand composition and thermophysical properties be used to determine sand provenance and chemical alteration?

These are just some of the questions I hope to provide insight into with my dissertation.

4.0 Purpose of Study

This dissertation consists of five chapters, where four of which are self-contained research projects (Chapters II–V). These chapters are intended to broaden the context by which we view martian aeolian systems, in terms of dune activity, morphologies, thermophysical properties, compositions, and source-lithologies. For Chapter II, I present a serendipitous discovery for evidence of substantial dune activity in Endeavour Crater, Meridiani Planum, the current site of the Opportunity rover. For Chapter III, I provide an introduction to the Valles Marineris rift system and the unique topographically-related dune morphologies of climbing and falling dunes. In Chapter IV, I discuss the results of a low- to mid-latitude survey of all martian dune fields in comparison to those found in Valles Marineris to test the effect of local and regional environments on duneform properties. Finally, for Chapter V, I investigate possible sediment sources and pathways for the Valles Marineris dune field populations.

5.0 Authorship

Portions of this dissertation have multiple authors. Collaborating authors are identified at the start of those relevant chapters. In all cases, coauthors were aware of their contributions. I wrote the entirety of Chapters I, and III-IV. Timothy Michaels, an expert in planetary atmospheric circulation models, was the primary author of sections 2.4, 3.6, and 4.4 in Chapter II. James Wray and Briony Horgan, both experts in VNIR spectroscopy and Mars mineralogy, assisted with data interpretation for Chapters IV-V. I collected, processed, and analyzed all data reported herein except where noted.

References

- Aben, L. K. (2003), Compositional and thermophysical analysis of Martian aeolian dunes, M.S. Thesis, Ariz. State Univ, Tempe, AZ.
- Arvidson, R. E. (1974), Wind-blown streaks, splotches, and associated craters on Mars: Statistical analysis of Mariner 9 photographs, *Icarus*, 21(1), 12–27, doi:10.1016/0019-1035(74)90086-4.
- Bourke, M., K. Edgett, and B. Cantor (2008), Recent aeolian dune change on Mars, *Geomorphology*, 94(1-2), 247–255, doi:10.1016/j.geomorph.2007.05.012.
- Breed, C. S., M. J. Grolier, and J. F. McCauley (1979), Morphology and Distribution of Common “Sand” Dunes on Mars: Comparison With the Earth, *J. Geophys. Res.*, 84(B14), 8183–8204, doi:10.1029/JB084iB14p08183.
- Christensen, P. (1988), Eolian intracrater deposits on Mars: Physical properties and global distribution, *Icarus*, 56(3), 496–518, doi:10.1016/0019-1035(83)90169-0.
- Christensen, P. R., J. L. Bandfield, A. D. Rogers, T. D. Glotch, V. E. Hamilton, S. W. Ruff, and M. B. Wyatt (2008), Global Mineralogy Mapped from the Mars Global Surveyor Thermal Emission Spectrometer, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell, pp. 195–220, Cambridge University Press, Cambridge.
- Edgett, K. S. (1994), The Sand Component of the Modern Martian Aeolian Sedimentary System, Ph.D. Dissertation, Ariz. State Univ., Tempe, AZ.
- Edgett, K. S., and M. C. Malin (2000), New views of Mars eolian activity, materials, and surface properties: Three vignettes from the Mars Global Surveyor Mars Orbiter Camera, *J. Geophys. Res.*, 105(E1), 1623–1650, doi:10.1029/1999JE001152.
- Fenton, L. K. (2006), Dune migration and slip face advancement in the Rabe Crater dune field, Mars, *Geophysical Research Letters*, 33, doi:10.1029/2006GL027133.
- Greeley, R., N. Lancaster, S. W. Lee, and P. C. Thomas (1992), Martian Aeolian Processes, Sediments and Features, in *Mars*, edited by H. H. Kieffer, B. M. Jakosky, C. W. Snyder, and M. S. Matthews, pp. 730–766, University of Arizona Press, Tucson.
- Kieffer, H. H., B. M. Jakosky, C. W. Snyder, and M. S. Matthews (Eds.) (1992), *Mars*, Space science series., University of Arizona Press, Tucson.
- Langevin, Y., F. Poulet, J.-P. Bibring, and B. Gondet (2005), Sulfates in the North Polar Region of Mars Detected by OMEGA/Mars Express, *Science*, 307(5715), 1584–1586, doi:10.1126/science.1109091.
- Sagan, C. et al. (1973), Variable Features on Mars, 2, Mariner 9 Global Results, *J. Geophys. Res.*, 78(20), 4163–4196, doi:10.1029/JB078i020p04163.
- Stockstill-Cahill, K. R., F. S. Anderson, and V. E. Hamilton (2008), A study of low-albedo deposits within Amazonis Planitia craters: Evidence for locally derived ultramafic to mafic materials, *J. Geophys. Res.*, 113(E7), doi:10.1029/2007JE003036.
- Tirsch, D., R. Jaumann, A. Pacifici, and F. Poulet (2011), Dark aeolian sediments in Martian craters: Composition and sources, *J. Geophys. Res.*, 116(E3), doi:10.1029/2009JE003562.

CHAPTER II: ORBITAL OBSERVATIONS OF CONTEMPORARY DUNE ACTIVITY IN ENDEAVOUR CRATER, MERIDIANI PLANUM, MARS

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Abstract

Although aeolian landforms are pervasive on Mars, evidence for contemporary activity has been limited. The next major campaign for the Mars Exploration Rover “Opportunity” is the investigation of the ~20-kilometer-diameter Endeavour crater, ~6 km to southeast of the rover’s position as of December 2010. We present evidence from orbital images that eight aeolian bedforms (~14,000 m²) in Endeavour crater have been active within the past decade (2001–2009), at a spatial scale that should be directly observable by Opportunity from the crater rim. Two dunes appear to show translational migration (~10–20 m), but all dunes indicate erosion to be the dominant process with up to 100% sediment removal. Thermophysical properties of these dunes are consistent with very fine to fine sand-sizes, the particle sizes most easily moved by the martian atmosphere. The dunes that show the most surface change have a rippled appearance without well-defined slip faces. Based on their morphology (elliptical-shape), we classify them as dome dunes. Mesoscale atmospheric modeling is employed to provide insight into the atmospheric forcing of this aeolian system. The major wind regimes from modeling are consistent with observations of wind streaks, sand streamers, ripples, and slip faces of regional dune fields although modeled wind speeds are insufficient to move sand. The translation and erosion of these dunes constitutes the largest contemporary movement of sand-sized sediment reported on Mars to date and demonstrates that Endeavour crater has been subject to wind profiles exceeding the threshold velocity at the surface (daily/seasonally and/or episodically) in the recent past.

1.0 Introduction and Study Area

Despite a dynamic atmosphere and plentiful sediment supply, orbital detection of dune movement on Mars has been elusive. For more than a century, telescopic observations of Mars showed temporal albedo variations associated with regions such as (what is now known as) Meridiani Planum [*Flammarion*, 1892]. Spacecraft data from Mariner 9 and Viking suggested aeolian deposits are common and dynamic on Mars, with wind-related albedo changes occurring during these missions [e.g., *Sagan et al.*, 1973; *Greeley et al.*, 1992]. *Geissler* [2005] used three decades of orbital data to document changes in albedo, and suggested that more than a third of the martian surface had brightened or darkened by roughly 10% in that time by wind-related processes. Albedo changes have been attributed to the aeolian deposition and raising of the light-toned dust particles [e.g., *Sagan et al.*, 1973; *Geissler*, 2005; *Cantor*, 2007]. The calculated saltation threshold friction speeds (verified in wind tunnel experiments under martian atmospheric conditions) are much greater for dust than for sand-size particles because of the cohesive effects of interparticle forces [*Greeley et al.*, 1980]. Winds that are strong enough to directly mobilize single dust particles should also be of sufficient magnitude to initiate saltation of most sand sizes. Yet somewhat paradoxically, although lofting of dust from the martian surface is amply documented, the orbital detection of sand dune modification has been lacking [*Edgett and Malin*, 2000; *Malin and Edgett*, 2001; *Bridges et al.*, 2007]. The Mars Global Surveyor's (MGS) Mars Orbiter Camera (MOC) experiment was used to look for dune migration that may have occurred between the time of the Viking and MGS missions [*Edgett and Malin*, 2000; *Malin and Edgett*, 2001]. No dune migration was observed over that 14 Mars year time span [*Edgett and Malin*, 2000; *Malin and Edgett*, 2001].

However, *Bourke et al.* [2008, 2009a] observed the gradual disappearance of two small ($\sim 1000 \text{ m}^2$) dome dunes and $\sim 85\%$ deflation of a third over a five year time span (1999-2004) in MOC images. That study reported no bedform migration. More recently, orbital images from the Mars Reconnaissance Orbiter (MRO) High Resolution Imaging Science Experiment (HiRISE) [*McEwen et al.*, 2007] have identified unambiguous meter-scale modification of ripples and dune edges in less than a martian season [*Silvestro et al.*, 2010a, 2010b]. These events demonstrate that the threshold wind speed for sand entrainment was exceeded under current conditions in these locations.

Evidence for aeolian activity, both ancient and contemporary, has been detected many times in surface-based observations. The combination of abundant sand-sized particles and erosional surfaces (e.g., ventifact flutes, wind streaks, etc.) at the Viking and Pathfinder landing sites suggested that abrasion due to repeated saltation has been commonplace [*Bridges et al.*, 1999; *Greeley et al.*, 2002], at least relative to geologic timescales. The Mars Exploration Rover (MER) Spirit has encountered a number of aeolian features and phenomena at Gusev crater, including: ventifacts, ripples, wind streaks, the El Dorado ripple field, and active dust devils [*Greeley et al.*, 2006; 2008; *Sullivan et al.*, 2008]. Evidence for geomorphically effective wind events occurred during the 2007 global dust storm, with Spirit observing rover track modification and ripple migration [*Sullivan et al.*, 2008]. In addition to observed ripple modifications with associated sand movement, sand-sized grains were found on the 66-cm-tall rover deck [*Greeley et al.*, 2006], supporting the notion of contemporary sand saltation.

In the last decade, Meridiani Planum has become one of the most intensely studied regions on Mars. The orbital detection of crystalline hematite associated with regional-scale layered deposits [*Christensen et al.*, 2001] contributed to Meridiani being selected as the landing

site for the Opportunity rover [Golombek *et al.*, 2003; Squyres *et al.*, 2004]. At the Opportunity landing site, ripples (of mixed particle sizes) and occasional wind streaks (dark and light) are pervasive [Sullivan *et al.*, 2005; Ferguson *et al.*, 2006a]. The last major ripple migration episode along the Opportunity traverse is estimated from crater statistics to have occurred between ~50 ka and ~200 ka [Golombek *et al.*, 2010], and is thought to be related to the atmospheric conditions during periods of higher axial obliquity [Arvidson, *et al.*, 2010]. Evidence for modern aeolian activity at the rover landing site was found in the form of the deposition and erosion of basaltic sand observed by Opportunity in a dark streak at Victoria Crater [Geissler *et al.*, 2008]. However, modeled near-surface mean winds for the landing site at $L_s \sim 320^\circ$ [Rafkin and Michaels, 2003] were below the minimum threshold for saltating sand on Mars [*cf.* Greeley *et al.*, 1980], and no wind-related bedform change has been detected in orbital observations.

The apparent disagreement between surface and orbital observations at Meridiani Planum, and the limited orbital detection of motion in general on Mars raise some pertinent questions concerning aeolian processes on Mars: (1) Why isn't bedform change more commonly observed from orbit? (2) Why do some bedforms show changes from orbit, but not others? (3) What factors are controlling bedform movement?

The next major campaign planned for Opportunity will be an investigation of Endeavour crater (2.3°S, 5.3°W; Fig. 2.1¹), which reaches deeper into the stratigraphy and geologic history of this region than any of the other craters previously visited by the rover. Endeavour crater is a degraded primary crater with a diameter of ~20 km. This crater predates the deposition of the

¹ All Tables and Figures in a given chapter are located in the chapter appendix.

sulfate-bearing layered deposits and is Noachian in age [*Wray et al.*, 2009; *Arvidson et al.*, 2010]. Crater rim segments are in various preservation states: rim segments to the north, west, and east have been rounded and degraded, whereas rims to the northwest and south have been completely removed or buried. Endeavour's floor shows heightened spectral contrast for basalt, hematite, and possibly hydrated sulfates as compared with the surrounding plains [*Chojnacki et al.*, 2010a]. This enhancement was attributed to the intracrater dust coverage being lower than outside the crater (i.e., more frequent dust cleaning events), rather than to a greater mineral abundance. As of December 2010, Opportunity is ~6 km from the western rim of Endeavour crater. The rim of Endeavour has been found to include phyllosilicate minerals [*Wray et al.*, 2009], a mineralogy not yet examined in situ on Mars and a driving motivator for Opportunity's exploration there.

In this paper, we present evidence for contemporary dune deflation, modification, and translation in Endeavour Crater. First we provide a summary of the methods used to characterize martian dunes and likely atmospheric conditions with associated spacecraft data and mesoscale atmospheric modeling. Next we present an overview of Endeavour crater and the characterization of aeolian activity there using a suite of remote sensing datasets and mesoscale atmospheric modeling. We then discuss the factors operating to create or reveal such an apparently high level of local aeolian activity and the implications of dune change on Mars. We conclude with a brief summary of our findings.

2.0 Methods and Data

2.1 Visible Wavelength Images

This study utilized visible-wavelength images from several spacecraft to search for surface changes. Narrow-angle images acquired between 2001 and 2003 from the MGS MOC instrument [Malin *et al.*, 1992] provide visible wavelength data at ~ 3 m/pixel. Additionally, we utilized images acquired by MRO during the period 2007 through 2009 with both the Context Camera (CTX) [Malin *et al.*, 2007] at 5-6 m/pixel and the HiRISE instrument at up to 25 cm/pixel. Images of Endeavour crater from all three instruments were examined for photogeologic evidence of aeolian bedforms. We utilized MOC images acquired in 2001 and 2003 to compare with MRO images taken in 2008, a time period of 3.3 Mars years (6.3 Earth years) and 2.3 Mars years (4.3 Earth years), respectively (see Table 2.1 for image identification and resolution). We examined other MRO images in addition to those listed in Table 2.1 for evidence of subsequent or intermediate surface change. Images were processed using U.S. Geological Survey (USGS) Integrated Software for Imagers and Spectrometers (ISIS) software [Gaddis *et al.*, 1997] for radiometric calibration, and spatially registered to each other using ArcGIS and ENVI software, allowing dune morphometry to be measured.

Complications in comparing data from the three imagers arise from differences in atmospheric conditions, in viewing geometries, and, most significantly, in spatial resolution. These differences have been mitigated through the superior resolution and high signal-to-noise ratio of HiRISE data and the co-registration of small features (e.g., craters). Additionally, we have applied an empirically-derived contrast enhancement to the MRO images to match the dynamic range of pixel brightness in the MOC images. This method for modifying the contrast

takes the (MOC) image near-minimum-maximum pixel values and ties them to the (MRO) near-minimum-maximum pixels values linearly. One example of the result of this method is shown in Figs. 2.2a and 2.2b where we have also down-sampled the HiRISE image to match the original MOC resolution.

As a simple assessment of the accuracy of our area measurements for the different image combinations, we made area measurements of several (presumably unchanging) impact craters of approximately the same scale as the dunes that were the subject of this study (1000-8000 m²). We estimate minimum threshold values for detectability of 6%, 4%, and 4% change in area (i.e., the amount of dune area that would have to change in order to be detected by comparing image pairs) for the MOC-HiRISE, MOC-CTX, and CTX-CTX combinations, respectively.

2.2 Thermal Inertia

Thermal inertia is a key surface property controlling diurnal temperature variations of the near-surface and provides insight into the physical properties (e.g., grain size, degree of induration) of the martian surface. Thermal inertia maps were derived from Mars Odyssey's Thermal Emission Imaging System (THEMIS) nighttime infrared data [Christensen *et al.*, 2003; Fergason *et al.*, 2006b] using the thermal model of Putzig and Mellon [2007], as implemented in the "jENVI" software suite [<http://arsia.gg.utk.edu/~utmars/jenvi/>]. This method uses THEMIS Band 9 (12.57 μ m) nighttime brightness temperature to derive best-fit thermal inertia values by interpolation within a seven-dimensional lookup table using: season, time of sol (Mars day), latitude, thermal inertia, albedo, elevation, and visible dust opacity [Putzig and Mellon, 2007]. THEMIS thermal inertia values were converted to effective particle sizes via the relationships determined experimentally by Piqueux and Christensen [2009]. The effective

particle size describes the upper centimeters of an unconsolidated surface and is a function of mixing, cementation, porosity, and sub-surface layering [*Presley and Christensen, 1997*]. Although thermal inertia is largely controlled by particle size, several factors can complicate the relationship. Ambiguities in the interpretation can arise from particle size mixing, grain cementation, subsurface layering, and atmospheric effects [*Ferguson et al., 2006b; Putzig and Mellon, 2007; Piqueux and Christensen, 2009*]. However, some of these issues are mitigated when dealing with well-sorted sedimentary deposits (i.e., dunes and aeolian bedforms) [*Ferguson et al., 2006a; 2006b; Putzig and Mellon, 2007*]. The latter point is related to the fine component properties, rather than those of the rock components, in controlling the bulk nighttime thermal inertia within the field of view [*Christensen, 1986*]. Additionally, bedforms that are thicker than several diurnal skin depths reduce the chance of subsurface vertical heterogeneities or layering effects (e.g., sand over bedrock) [*Putzig and Mellon, 2007; Piqueux and Christensen, 2009*].

2.3 Topography

To quantitatively investigate surface morphology and estimate the underlying slopes digital elevation models (DEMs) were used. DEMs created from Mars Express High Resolution Stereo Camera (HRSC) data [*Neukum et al., 2004*] were acquired from the HRSC data explorer at <http://hrscview.fu-berlin.de> and then reprojected and reformatted using ISIS software.

2.4 Mesoscale Atmospheric Modeling

Mesoscale climate modeling can provide insight into wind direction, speed, and its ability to move sediment on a planetary surface. The mesoscale Mars atmospheric model used in this work was the Mars Regional Atmospheric Modeling System (MRAMS) [*Rafkin et al., 2001*].

This model is a regional (vs. global) three-dimensional, nonhydrostatic atmospheric model, enabling relatively high spatial resolution (grid-spacing 10s of kilometers or less) over timescales of sols at a targeted location. The output from a NASA Ames Mars General Circulation Model (MGCM) [Haberle *et al.*, 1993] simulation is used for initial states and time-dependent boundary conditions for MRAMS runs. In this work, a series of five nested computational grids (each with a successively smaller total area and horizontal grid-spacing) is used to achieve a grid spacing of ~ 2.5 km over Endeavour crater and its surroundings. In these simulations, approximately 10×10 horizontal model gridpoints are within Endeavour crater.

Seasonal MRAMS simulations were conducted for $L_s \sim 30^\circ$, 120° , 210° , and 300° , each ~ 4 sols in duration (first sol is spinup), to characterize basic seasonal differences in the aeolian environment at Endeavour crater. MRAMS and the MGCM both used latitudinally- and seasonally-varying atmospheric dust loadings based on MGS Thermal Emission Spectrometer (TES) dust opacity observations [e.g., Smith, 2004]. MRAMS bases its terrain and surface characteristics on $1/64$ deg/pixel (degree per pixel) gridded MGS Mars Orbiter Laser Altimeter (MOLA) topography [Smith *et al.*, 2001], $1/8$ deg/pixel gridded MGS TES albedo, and $1/20$ deg/pixel gridded MGS TES-based nighttime thermal inertia [Putzig and Mellon, 2007]. The model-resolved topography of the crater is generally muted compared to what is known from higher resolution data (section 3.6). This issue is unavoidable, given the finite resolution of the computational grid and a numerical stability requirement that the model topography not contain structure at scales less than two times the horizontal grid spacing. Instantaneous snapshots of the model state were output every $1/3$ Mars-hour (simulated).

Previous attempts to identify correlations between mesoscale atmospheric modeling and Mars aeolian systems [e.g., Fenton *et al.*, 2005; Hayward *et al.*, 2009] have met with mixed

degrees of success. This study is different, however, in that it targets a dune system that is known to have changed over a relatively short period of time. The atmospheric modeling aspect of this work aims not only to determine whether sediment-mobilizing winds are occurring, but also to gain insight into when these winds occur, and into what causes these winds. This approach often involves identifying and characterizing the wind regimes for the area of interest, Endeavour crater. Wind regimes we define as distinct periods (> 30 Mars-minutes in duration) when strong and directionally-coherent near-surface mean winds occur at nearly the same time-of-sol for one or more seasons. The mix of wind regimes present may vary seasonally. Additionally, the modeling work aims to achieve a measure of model “validation” (i.e., gain confidence that MRAMS can satisfactorily predict winds relevant to aeolian surface processes), and provide information to help interpret future observations of Endeavour crater aeolian features and processes.

3.0 Results

3.1 Recent Aeolian Activity in Endeavour Crater

Endeavour crater (Fig. 2.1a) has three types of aeolian morphologies: two main populations of dune bedforms (Fig. 2.1b) and transverse aeolian ridges (TARs). Dune fields are superposed on the following units (in stratigraphic order, lowest to highest): a light-toned etched unit (described by *Hynek et al.* [2002]), an immobile mid-toned mantling unit (related to the hematite-bearing plains intercrater unit [*Christensen et al.*, 2001; *Hynek et al.*, 2002]), and TARs in a relatively sediment-starved landscape [*Chojnacki et al.*, 2010a]. Barchans and barchanoid dunes are found in the western portions of the crater (Fig. 2.1b). In contrast the eastern half of the crater is populated by a greater diversity of dune morphologies, including bedforms that may

be classified as transverse dunes, sand sheets, and small dome dunes (Figs. 2.1c, 2.2–2.3). The consistently darker-tone (lower DN values) on the down-sun side of these modest bedforms suggests muted topography (shallower slopes) compared to the western barchan dunes. High-resolution images reveal a rippled, “crosshatched” morphology with 2-5 meter wavelengths (e.g., Figs. 2.2b, 2.2d, 2.2f). Although these eastern dunes lack prominent slip faces, we suggest that they qualify as “dunes” as defined by *Bagnold* [1941]: mounds of windblown sand-sized sediment with a central higher-standing rise. We classify the bedforms in Figs. 2.2 and 2.3c–2.3d as dome dunes; that is, relatively small dunes, circular to elliptical in plan view, and without prominent slip faces [*cf. McKee*, 1979]. We use the more general term *bedforms* when describing the collective group of eastern aeolian morphologies (Fig. 2.1c).

Modeling and spectroscopy give some indication of a primary wind direction and propensity for moving sediment. Previously published results from the MRAMS described a generally easterly (towards the west) wind flow at ~10 m/s in the vicinity of Endeavour crater during afternoon at $L_s \sim 320^\circ$ [*Rafkin and Michaels*, 2003]. These modeled wind vectors are consistent with the north-south orientations of ripples on the smaller dome dunes (insets in Fig. 2.2) and dark streaks associated with some of the western barchans (Fig. 2.4), although not with the western group dunes slip face orientations (toward the south-to-southeast). MGS TES-derived dust cover index (DCI) [*Ruff and Christensen*, 2002] values of 0.96 to 0.98 in the southern interior of Endeavour crater indicate minimal to no surface dust at the time of data acquisition, consistent with a surface that has been scoured by aeolian processes or has not undergone dust deposition.

Our analysis of Endeavour crater orbital images from the past ~2–3 Mars years (~4–6 Earth years) shows several significant styles of dune and bedform erosion, modification, and

translation (Figs. 2.2–2.3). The eight dunes that show surface change are labeled i–viii in Fig. 2.1c. Eight eastern dunes appear in the 2001 and 2003 MOC images; however, in CTX and HiRISE images taken in 2008, two of the dunes (i and ii) have disappeared entirely (e.g., Figs. 2.2a–2.2d), whereas six others (iii–viii) have decreased in area (e.g., Figs. 2.2c–2.2f, 2.3c–2.3d; Attachment 2.1; Table 2.1). Dunes iv–vi are only imaged in the lower resolution MOC-CTX pair of images, making area estimates more difficult than for the other dunes, which are covered by the MOC-HiRISE pair of images.

3.2 Dune Deflation

There are two potential explanations for the disappearance of these dunes (dunes i–viii): (1) aeolian removal and/or deflation of dune sediment, or (2) dust deposition with consequent obscuration via an increase in albedo, effectively “blending” the dunes into the lighter-toned background plains. The latter explanation is less credible because the locations where the dark-toned features were originally present now lack (as of 2008) the crosshatched, rippled appearance seen in the other dark-toned dunes that are still present (see insets in Figs. 2.2b, 2.2d, 2.2f). Although the MOC images from 2001/2003 and CTX images are of insufficient resolution to distinguish this texture in the now-missing features, it is reasonable to assume that all the dark-toned features in this vicinity originally had the same rippled surface texture. If a mantle of bright dust were responsible for the disappearance of the dark-toned features the rippled texture would have been preserved, but it is not observed. Moreover, substantial dust mantlings are relatively uncommon on the surface of Meridiani Planum, as observed by Opportunity [Herkenhoff *et al.*, 2006]. Thus, we infer that these eight dunes were partially or completely deflated within the 2.3–3.3 Mars years (4.3–6.3 Earth years) or less, between the images.

The majority of dune or bedform change is found in the eastern group, although there are sand streamers or dark streaks emanating from one western dune that suggest periods of winds capable of saltating this material (Figs. 2.4, showing change in dominance of west-southwest and southeast orientations; Attachment 2.2). For dunes captured in the 2001/2003 MOC and 2008 MRO combinations, dune area reduction ranged from ~10% to 100% (Table 2.1). An additional CTX observation in 2009 supports the trend of deflation with dunes iii and vii reduced in area ~24% and ~55% respectively, when compared with CTX in 2008 (Table 2.1). For these eastern dunes we estimate that the total surface area from which sediment has been removed is greater than ~14,000 m² (Table 2.1). In comparison to dune fields previously visited by MER, the total area of Endeavour dunes removed is roughly one quarter the area of the total Victoria crater dune field (55,000 m²), or about the same area as Gusev crater's El Dorado ripple field (13,000 m²). Dune coverage area lost per unit time varies widely from dune to dune (Table 2.1), but the mean for all dunes is 750 m² per Mars year (400 m² per Earth year). These rates are averages assuming removal at a steady rate throughout the time between image acquisitions. The assumption of steady state may be incorrect, as some dune(s) may have been active more intermittently (section 4.2).

3.3 Dune Morphology Modification

The dunes presented in Fig. 2.2 show no evidence for downwind or southward migration. However, several larger dunes or bedforms ~500 m to the southeast (the inferred downwind direction) of dunes ii–iii do show surface modification (Fig. 2.1c, white arrows). Minor changes, presumably due to aeolian modification, are evident on the windward side of what appears to be a deflated barchan dune and a minor limb of a transverse dune (Figs. 2.3a 2.3b, white arrows)

and other locations (Figs 2.1c, white arrows). From this evidence, we suggest two likely outcomes for the sand involved in the dune changes between 2001 and 2009. The sand removed from the northern-most dome dunes may have been dispersed downwind too diffusely to be detected in HiRISE images, but some of this sand may have been absorbed and incorporated into the larger dunes to the south, including those that show shape modification (Figs. 2.3a–2.3b). Some combination of the two scenarios is also possible.

3.4 Dune Migration

In addition to dune deflation and the modification of dune morphology, several dunes noticeably translate from 2003 to 2008. Dune v, in addition to losing ~40% of its area, shows a ~20 m southeastward change of its centroid, defined as the location equidistant along the long and short axis (Figs. 2.3c–2.3d; blue to yellow polygon). This evidence is interpreted as southeastward dune migration (section 4.1). Likewise, dune vi also appears to have migrated ~10 m from its position in earlier images (Table 2.1). To our knowledge, this change is the first documentation of dune migration on Mars, although smaller ripples have been observed to migrate at landing sites [Sullivan *et al.*, 2008] and in HiRISE observations [Silvestro *et al.*, 2010a]. These estimates of dune translation are based on the available MOC-CTX pair of observations, and thus precision is limited. If migration continues, the dune centroid positions could be better constrained using future high-resolution observations.

3.5 Thermal Inertia and Topography

A THEMIS thermal inertia map of Endeavour crater (Fig. 2.5) exhibits values from 80–350 $\text{Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$ (hereafter referred to as “tiu”, or thermal inertia unit). The western barchan sand dunes exhibit a thermal inertia of 150–240 tiu, whereas the eastern dunes have a thermal

inertia range of 110–190 tiu. For unconsolidated sediments, these thermal inertia values are consistent with particle sizes of 100–600 μm (very fine to coarse sand-sizes) and 50–200 μm sediment (very fine to fine sand-sizes), respectively [Piqueux and Christensen, 2009]. The elevated thermal values could be the result of differences in dune thickness or sediment induration, instead of or in addition to differences in grain size.

A CTX visible-wavelength image colorized with HRSC elevation is shown in Fig. 2.6, illustrating the local topography of the Endeavour crater basin. Topographic profiles aligned parallel to the inferred primary wind direction indicate hummocky terrain dropping off southward across the eastern dunes and bedforms.

3.6 Mesoscale Atmospheric Modeling Results

The MRAMS simulations performed for this work do not have the spatial resolution (due to computational limitations usual for such modeling) needed to properly resolve boundary layer turbulence, although wind gusts generated by such turbulence are suspected of being very important to aeolian processes on Mars [e.g., Fenton and Michaels, 2010]. The MRAMS output fields analyzed here are effectively the spatial mean (within each $\sim 2.5 \times 2.5$ km grid cell) of the actual variability within those fields (a range that exists due to turbulence). One of those fields is the aerodynamic surface shear stress ($\tau = \rho_{atm} u_*^2$; with units of mN m^{-2} or mPa). Shear stress is an important quantity for estimating aeolian saltation potential because it includes the effects of both the local atmospheric density (ρ_{atm} ; varies significantly between night and day on Mars) and friction speed (u_* ; also varies significantly, especially when turbulence is present). The threshold aerodynamic surface shear stress (τ_t) needed to initiate saltation of sand-sized particles was also calculated for every MRAMS grid point at all output times using the semi-empirical expressions

of *Newman et al.* [2002, equation (5)], the local atmospheric density at the surface, and a particle mass density of 2800 kg m^{-3} (similar to weathered basalt).

Fig. 2.7a shows the model topographic contours on a portion of grid 5 (~20% of the full area), an area which includes Endeavour crater as well as Opportunity's traverse, overlaid on a mosaic of CTX images colored with HRSC topography. An area of 4x4 model gridpoints, where dune activity is known from photogeologic evidence, is analyzed in detail. This target area was selected to receive more detailed attention because of its relevance to the observed dune modifications discussed above (Fig. 2.7a, black box). Polar coordinate scatter plots were generated of τ/τ_t versus wind direction (Fig. 2.8) within that area, valid at each of the four seasons that were simulated ($L_s \sim 30^\circ, 120^\circ, 210^\circ$, and 300°). Values of τ/τ_t greater than or equal to one would indicate that the spatial mean winds at the associated model grid point(s) are likely capable of initiating the saltation of basaltic sand.

The regional scale atmospheric model (MRAMS) results are only broadly similar to the much coarser resolution GCM results (used as distant boundary conditions for the MRAMS runs discussed here), with both models suggesting that regional-scale winds should evolve seasonally, with a significant easterly wind component occurring often. The MRAMS results suggest that the study area within Endeavour crater is subject to three flow regimes (two major, one minor) that vary seasonally. The wind regime with the highest modeled winds (WR1) occurs for ~2–3 Mars-hours during the mid-morning at all seasons, with the direction and strength varying with season: from the east (with the strongest overall magnitudes) during the warmer seasons (e.g., $L_s \sim 210^\circ, 300^\circ$; Figs. 2.4a, 2.7b, 2.8), and from the southeast to south at during the southern fall and winter (e.g., $L_s \sim 30^\circ, 120^\circ$; Fig. 2.8). The strength and direction of this wind regime is generated by regional forces associated with both the atmospheric diurnal tides and the regional-scale

topography. The wind regime with the second highest modeled winds (WR2) occurs for ~ 1 Mars-hour in the early evening, and is characterized by winds primarily from the northwest. It is strongest in southern spring (e.g., $L_s \sim 210^\circ$; Figs. 2.3b, 2.4b, 2.7c, 2.8) and nearly absent during southern winter (e.g., $L_s \sim 120^\circ$). This northwest flow is initially generated by regional forces, but its magnitude is significantly enhanced through interaction with Endeavour crater. Furthermore, this enhancement only appears to occur in the northern half of the crater, with weak winds elsewhere within the crater. The third flow regime (WR3) is a minor one that is only significantly present during the southern winter (e.g., $L_s \sim 120^\circ$; Figs. 2.7d, 2.8). It is characterized by flow from the southeast to south-southeast into the southern third of Endeavour crater, and is a result of local flow modification due to the presence of the crater.

At all seasons, the modeled τ/τ_i has magnitudes below 0.5, and more commonly less than 0.25. Although the MRAMS-modeled τ/τ_i values are less than one (indicating winds nominally incapable of mobilizing sediment), MRAMS does not take into account possible turbulent gusts superposed on the mean wind speed. To achieve the needed increase in τ/τ_i of a factor of two to four, corresponding gust strength of $\sqrt{2}$ to $\sqrt{4}$ times the mean wind speed would be required. Such enhancements are plausible according to *Fenton and Michaels* [2010]. In discussing correlations between MRAMS wind vector orientations and the orientations of aeolian surface changes, we make the assumption that this level of gustiness is achieved. Therefore we suggest, in a relative sense, the results indicate that the southern spring and summer seasons (e.g., $L_s \sim 210^\circ, 300^\circ$; Fig. 2.8) are more likely to support aeolian activity within Endeavour crater than the other two seasons. According to the model results, winds would move, if of sufficient magnitude, aeolian features predominantly westward or southeastward across the basin with some minor reworking northwestward (Fig. 2.7). The results also suggest that southern autumn

($L_s \sim 30^\circ$) is the least likely time of year for significant aeolian activity within the crater. The modeling also indicates that some sol-to-sol variability occurs, but does not significantly change the general results presented above.

4.0 Discussion

4.1 Comparison with Other Martian and Terrestrial Examples

Our results have documented a high degree of dune activity in Endeavour crater during the relatively short time span of $\sim 2\text{--}3$ Mars years ($\sim 4\text{--}6$ Earth years). We suggest saltation, rather than creep or suspension, as the primary mechanism to be responsible for the erosion, modification, and migration of these dunes. A surface of coarse sand particles, which is less prone to saltation than to surface creep, is inconsistent with estimates from the thermal inertia data. Likewise, removal by suspension would require sub-sand sizes that would correspond with lower thermal inertia surfaces than estimated from THEMIS data (Fig. 2.5).

This evidence for widespread dune deflation implies local winds that are sufficient to initiate saltation, as observed elsewhere on Mars [Greeley *et al.*, 2006; Bourke *et al.*, 2008; Geissler *et al.*, 2008; Sullivan *et al.*, 2008; Silvestro *et al.*, 2010a, 2010b]. Indeed, Endeavour crater dune area removal rates are an order of magnitude larger than polar dunes that show surface change (see below). This sediment movement in Endeavour crater represents the most significant contemporary sand-sized aeolian sediment transport on Mars detected to date. This movement indicates that Endeavour crater experienced significant near-surface wind speeds, although the temporal nature of those winds (e.g., daily winds? episodic wind events?; section 4.2) is poorly constrained. Regardless of the time frame for movement, documentation of relatively large-scale ($>1000\text{ m}^2$) bedform activity may have been inevitable due to spacecraft

imaging campaigns with progressively greater spatial and temporal resolution, as well as areal coverage [Sullivan *et al.*, 2008].

Thorough image analysis campaigns looking for dune change have been ongoing, with comparisons being made between Viking-acquired images to MOC, MOC to MOC [for a review see Bourke *et al.*, 2008], and most recently MRO to MRO [Bridges *et al.*, 2007], yet have been limited in success. Endeavour crater, Nili Patera (where ripple migration has been observed for the first time from orbit [Silvestro *et al.*, 2010b]), and the small dome dunes seen to be modified in the north polar region [Bourke *et al.*, 2008], are the only such identifications thus far. An intriguing common thread between these three known anomalous sites is that the features with observed aeolian modification were much smaller in size (lesser sediment volume) than the more typical “larger” dunes across Mars that have yielded nearly no evidence of contemporary aeolian modification. This correlation supports the possibility that the contemporary aeolian sediment transport processes on Mars may be of a magnitude and/or intermittency which allows low volume bedforms to change detectably over a few Mars-years (or less), but may take Mars-decades to produce detectable changes on larger bedforms.

The average migration and removal rates we have deduced for the dunes in Endeavour crater provide an opportunity for comparison with terrestrial rates. The migration rates for the two martian dunes were ~4–9 meters per Mars year (~2–5 meters per Earth year; Table 2.1), which are generally lower rates than terrestrial studies of small dune (typically barchans) movement [Finkel, 1959; Pye and Tsoar, 1990; Jimenez *et al.*, 1999]. However, Antarctic hyper-arid polar dunes, thought to be analogous to at least some martian dunes, travel at an average rate of 1.5 meters per Earth year [Bourke *et al.*, 2009b]. Similarly, terrestrial studies of dome dune migration rates vary from 4–8 meters per Earth year [Dong *et al.*, 2000; Bristow and

Lancaster, 2004], which is similar to our estimates of 2–5 meters per Earth year for the Endeavour crater dome dunes.

The estimated areal removal fluxes vary by an order of magnitude from dune to dune, but the higher-resolution MOC-HiRISE comparisons suggest $\sim 350 \text{ m}^2$ per Mars year ($\sim 190 \text{ m}^2$ per Earth year). For comparison, we have also calculated areal removal rates for the deflating polar dome dunes (dunes i–iii) identified by *Bourke et al.* [2008]. We derive areal change rates for these dunes of $\sim 180 \text{ m}^2$ per Mars year (100 m^2 per Earth year), which is smaller than but comparable to the removal rates in Endeavour crater. The unusual parameter of areal removal rate is necessitated for the Mars dunes because no data are available on the thickness of the dunes (section 4.3). This parameter is not what is typically used in terrestrial studies, but *Dong et al.* [2000] do report several instances of small crescent and dome dunes deflating in the Taklimakan Desert, China. Dune morphometric data from these authors suggests rates of ~ 85 and 115 m^2 per Earth year for the collective study (including small crescent dunes) and dome dunes, respectively. Reports of terrestrial dunes experiencing 100% deflation are uncommon. However, one small ($\sim 45 \text{ m}$ diameter) dome dune in the Namib Sand Sea [*Bristow and Lancaster*, 2004] was reported to have dissipated 100% within three years (1999–2002) of that study [*C. S. Bristow*, personal communication, 2010]. This report results in a poorly constrained, areal removal flux of $\sim 1000 \text{ m}^2$ per Mars year ($\sim 550 \text{ m}^2$ per Earth year), clearly a much higher rate than has been observed on Mars to date.

4.2 Timing: Episodic and/or Gradual Aeolian Activity

Both episodic and gradual aeolian events are suspected to have taken part in the aeolian activity at Endeavour crater. Surface observations during operations at Mars Exploration Rover

landing sites led *Sullivan et al.* [2008] to conclude: (1) strong wind events, usually associated with regional or global dust storms, occur with some randomness, (2) the strongest events are also the rarest, and (3) these strong wind events contribute the most to surface change. A planet-encircling dust storm occurred in July-August 2007 and disrupted MER-surface operations at both landing sites. Several local changes associated with this storm were observed from the surface and from orbit. During that period, atmospheric opacity (τ) reached ~ 5.0 at Victoria crater, and eventually Opportunity's solar array performance dropped to record lows [*Geissler et al.*, 2010; *Vaughan et al.*, 2010]. Commonly associated with these sometimes sudden increases in dust opacity were episodic strong wind events, which stripped sediment from MER solar panels [*Geissler et al.*, 2010; *Vaughan et al.*, 2010]. The spatial extent of these exceptional wind events, and thus their potential effect on Endeavour crater bedforms, is unknown – in part because all of these dust-cleaning events occurred during long hiatuses between overlapping Endeavour crater imaging. During the course of the mission, several local changes associated with this storm were observed from the surface and from orbit [*Geissler et al.*, 2010]. Opportunity observed its rover tracks both gradually fade (brighten) due to dust deposition and be erased due to aeolian deflation, including some that were less than one Mars year old [*Geissler et al.*, 2010]. In contrast to the slow rate of dust deposition, episodic transport and deposition of basaltic sand by surface winds may produce rapid erasure, as suggested by the disappearance of the rover tracks and dark streaks around Victoria crater during the 2007 dust storm [*Geissler et al.*, 2008, 2010]. It is thus conceivable that much of the surface change observed herein is due to one major episode of high-magnitude winds associated with the 2007 planet-encircling dust storm. Because it is known only that changes occurred sometime between the acquisition of the two images showing dune transport and removal, these time spans for

change and the removal and migration rates shown in Table 2.1 are temporal averages that assume continuous activity, as no intermediate images of suitable quality are available.

Alternatively, some evidence exists that dune bedform surface change is more gradual with changes occurring on an inter- or even intra-seasonal basis. At the Opportunity landing site, landing rockets fired prior to the rover dropping during the landing sequence made bright scour marks easily resolvable in subsequent MOC images. The bright marks are likely created by the removal of sand and hematite-rich concretions by the rocket blast [Geissler *et al.*, 2010]. The blast marks were observed to gradually fade in several orbital images, presumably due to seasonal wind-driven deflation, in less than a martian year [Geissler *et al.*, 2010]. Contemporary “reversing” bright wind streaks (each feature having two streak orientations, northwest and southeast) that are associated with small craters are also common in this portion of Meridiani Planum [Sullivan *et al.*, 2005; Jerolmack *et al.*, 2006]. There is evidence from repeated visible-wavelength data that these dust streaks appear at predictable and repeatable times of the martian year (section 4.4) [Jerolmack *et al.*, 2006]. This evidence suggests some element of seasonal aeolian behavior controls the emplacement and erosion of dust, producing these features. Bourke *et al.* [2008, 2009a] documented polar domes that showed gradual deflation every 1–2 Earth years and some changes occurring over a single martian season. Other than the 2001–2007 time-step, the frequent MRO images coverage of Endeavour crater supports a gradual erasure of dune iii and possibly neighboring bedforms. Future observations will reveal whether or not areal removal and/or migration rates are relatively continuous in time.

4.3 Factors Contributing to Surface Change

Coordinated orbital and rover observations at Meridiani and Gusev crater made significant progress in addressing question 1, why bedform change is not more commonly observed from orbit [Geissler *et al.*, 2008; Sullivan *et al.*, 2008]. These studies have provided some insight for observations of other, less (remotely) characterized aeolian features on Mars and their potential for contemporary activity. For example, the temporal change of the Victoria crater dark-streaks are easily resolved at MOC-NA and HiRISE spatial resolutions, but required MER-B observations to determine that they were due to the saltation of dark sediment (versus the removal of a bright mantling material overlying a dark substrate) [Geissler *et al.*, 2008]. In contrast to the Victoria crater dark streaks the El Dorado ripple migration (~ 2 cm) [Sullivan *et al.*, 2008] would not have been discovered solely from orbital observations even with repeated HiRISE imaging. The recent discovery of dune mobilization in Endeavour crater and ripples migrating ~ 1.7 meters in Nili Patera over one martian season [Silvestro *et al.*, 2010b] was accomplished with the 25 cm/pixel resolution of HiRISE and the multiple of overlapping images.

The reasons some bedforms change and others do not as observed from orbit (question 2) and the factors controlling detectable movement (question 3) may be a combination of effects associated with: particle sizes, degree of sediment induration, areal extent and/or volume, position relative to associated topography, bedform morphology, and smaller-scale atmospheric flow structure. The thermophysical properties of the local sediment may give us some indication why some dunes show a greater propensity for change than others (Fig. 2.5). The THEMIS-derived mean grain size value of ~ 120 μm (50–200 μm) for the eastern dunes and bedforms encompasses the particle size (~ 115 μm) most easily mobilized by the martian atmosphere [Greeley *et al.*, 1980]. Particles are expected to be composed of basaltic-grains (low-calcium

pyroxene), rather than lower density dust aggregates based on MRO's Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) spectroscopic observations [Chojnacki *et al.*, 2010a]. This particle size range for basaltic grains, at the estimated Meridiani Planum atmospheric pressure (~ 6.7 mb), would require an friction threshold velocity (u_{*t}) of ~ 2 m/s [Greeley *et al.*, 1980] or surface wind velocity of ~ 17.5 m/s (at ~ 1 m height and an aerodynamic roughness value of 3 cm [Sullivan *et al.*, 2005]) for entrainment. The thermal inertia of these dunes and bedforms also supports the notion that on these dunes, like the El Dorado ripples [Sullivan *et al.*, 2008], saltation of hard sand-sized particles is winning its competition against surface induration.

Dune area removal rates appear relatively similar across the study area and between imaging epochs with the exception of dunes v-vii, where measurements were done with lower resolution MOC-CTX data. Volumetric sediment fluxes are difficult to estimate in Endeavour crater because no appropriate images currently exists that can be used to construct a DEM with sufficient resolution. However, we can make crude estimates of dome dune volumes, mass, and associated flux by looking at different dune heights. To obtain an upper limit for dune height, we note that these dome dunes have no slip faces and thus possess slopes below the angle of repose ($< 33^\circ$). This lack of slip faces can be used to put upper limits on dune heights. For example, dune iii at ~ 39 m wide could be ~ 9 m tall (assuming slopes of 25°) but not possess a slip face. This dune height is likely an overestimate, as terrestrial dome dunes have relatively modest slope angles when compared with other dune morphologies [Dong *et al.*, 2000; Bristow and Lancaster, 2004], so we will also consider a height of 2 m. Volume estimates from the 2008 dune iii measurement are estimated at ~ 900 m³ and 16,900 m³ [Monolithic Dome Institute, 2001], when using heights of 2 m and 9 m respectively. These estimates illustrate dunes that are shorter are more susceptible to deflation and surface change, because they contain less sediment.

Bedforms move proportional to the wind velocity and inversely to their size [*Bagnold*, 1941; *Finkel*, 1959; *Sauermann et al.*, 2000; *Hesse*, 2008]. Endeavour crater dunes show no clear inverse correlation between areal size and removal rate, as a proxy for transport rate (i.e., the smallest dunes do not appear to be removed faster than larger dunes in Table 2.1). However, this lack of an inverse correlation is likely an artifact of poor temporal resolution and we suspect the smallest dunes ($< 2000 \text{ m}^2$) were deflated in a shorter time span than larger ones.

Topography and degree of bedform exposure are additional factors influencing dune sediment transport rates and dune morphology [*Taylor et al.*, 1987; *Pye and Tsoar*, 1990]. All dunes (except vi) that show removal or reduction in size are located at the inferred up-wind perimeter of the dune field (Fig. 2.1c), making them most exposed to atmospheric flow and prone to deflation. Profiles across the eastern dune field (Fig. 2.6) parallel with what we have defined as the primary wind/transport direction indicate that many mobile dunes are superimposed on a surface that slopes downward to the southeast (with 200 m of relief). The majority of dunes that show erosion and transport are located at the summits of small hills or on gently southeastward sloping ($1\text{--}7^\circ$) surfaces (Fig. 2.6). Local topographic highs, as with the stoss slope leading up to the crest of an individual dune, are locations of increased wind velocity due to streamline convergence [*Taylor et al.*, 1987; *Pye and Tsoar*, 1990]. Saltation pathlengths are longer where wind vectors align with the downhill direction, enhancing the wind's capacity for transporting sediment [*Tsoar et al.*, 1996; *Pye and Tsoar*, 1990].

As described above (section 3.1) these bedforms are classified as simple dome dunes [*cf.* *McKee*, 1979]. Dome dunes are not as efficient at trapping sand as barchans and are known to be highly susceptible to wind erosion [*Bagnold*, 1941; *McKee*, 1979; *Bourke et al.*, 2008, 2009a]. Dome dunes may develop into barchan dunes given sufficient sand supply and wind velocity

[McKee, 1979; Bourke and Goudie, 2009] or develop from larger barchan limbs as part of a calving event [Bourke, 2010]: thus an evolutionary relationship between dome and barchan morphologies can exist. We suggest the changing dome dunes in Endeavour crater are an example of progressive degradation of barchan dunes leading to present-day dome dunes and the loss of slip faces. A survey of dune field morphology supports the idea of disequilibrium. Dune height monotonically decreases from the western-most barchan dunes toward the east, as inferred by using the technique of Bourke *et al.* [2006], in which the horizontal lengths of slip faces (where present) provide a proxy for dune heights.

All of the barchan dunes in western Endeavour crater are classified as fat barchans, based on the ratio of length of windward slope to horn-to-horn width [Long and Sharp, 1964]. Dunes in this subclass are common on Earth, but less so on Mars [Bourke and Goudie, 2009]. Models suggest fat barchans form in the presence of varying wind directions (up to $\sim 40^\circ$) [Reffet *et al.*, 2009], in locations that are topographically-influenced and/or on the margins of larger dune fields [Bourke, 2010]. There is evidence for all of these factors in the western dune field of Endeavour crater. Many dunes of the western group have maintained a crescent shape, but have greatly degraded slip faces (e.g., Fig. 2.4). In general, the presence of barchan dune morphologies commonly has been suggested to imply a relatively unidirectional wind regime and limited sand supply [Bagnold, 1941]. The sediment supply north of these dunes, presumably upwind, certainly is sparse according to our photogeologic and thermophysical analyses (Figs. 2.1b–2.1c, 2.4a, 2.5). The dune field, primarily the eastern group, shows evidence for a trend in dune degradation (from north to south) due to wind energy stripping material away (erosion) from the northern-most dunes and depositing sand southward as part of larger bedforms.

The remaining dunes exhibiting recent movement (e.g., dune iii) have similar dune morphologies to those eroding in the north polar region [Bourke *et al.*, 2008, 2009a]. Fig. 2.9 compares the meter-scale dune morphology of the polar dome dune iii from Bourke *et al* [2008] to the dune in Endeavour crater that we have labeled dune iii, presented at the same scale. Low-relief dome dunes with meter-scale rippled surface textures without major slip faces predominate in both cases. We suggest that, based on evidence for bedform movement from both polar and equatorial dome dunes, these small dome dunes are the best morphological candidates, with the available spatial and temporal resolutions of current spacecraft instruments, for discrete bedform surface change on the time-scale of orbital missions.

4.4 Aeolian Morphology vs. Modeled Winds

MRAMS predicts three modeled wind regimes (section 3.6, Figs. 2.7–2.8), none of which has sufficient strength to move sediment, but each of which correlates with some of the observed aeolian morphology orientations (Fig. 2.10). The disconnect between the modeled wind strength and evidence from spacecraft data for aeolian activity suggests episodic or gusty winds not resolved by MRAMS are responsible for surface change. This scenario would be consistent with terrestrial bedform activity, as sediment moves only rarely during unusually intense events [Conradsen *et al.*, 1984]. However, the seasonality of MRAMS-predicted wind regime is consistent with the timing of surface changes observed in orbital images. Fig. 2.10 summarizes the multiple aeolian morphologies of the region (dune slip face orientation, mega-ripple fields, sand streamers, dust streaks), which all align to three observed wind regimes: easterly, northwesterly, and southeasterly [Sullivan *et al.*, 2005; Jerolmack *et al.*, 2006; Geissler *et al.*, 2008; 2010; Arvidson *et al.*, 2010; Chojnacki *et al.*, 2010b].

The formation of the plains ripples [Sullivan *et al.*, 2005; Golombek *et al.*, 2010; Arvidson *et al.*, 2010], and ripples superimposed on dome dunes (Fig. 2.2; insets) are both consistent with winds from the east or west. Sand streamers (Fig. 2.4) indicate strong contemporary winds from both the east and northwest. Endeavour crater dune slip face orientations (Fig. 2.1b) suggest a wind direction from the northwest (Fig. 2.4). Indeed, all five craters similar in size (5– to 20–km–diameter) to and within ~100 km of Endeavour crater possess similar intracrater concentrations of dark-toned sediment forming dune fields. All of these intracrater dune fields have dune slip faces consistent with a wind direction at the time of formation from the northwest. Observed dune modifications within Endeavour crater (Figs. 2.2–2.4) indicate the presence of a contemporary wind regime, also from the northwest, that is able to mobilize sediment. Contemporary “reversing” bright wind streaks (each feature having two streak orientations, northwest and southeast) that are associated with small craters are also common in this portion of

Meridiani Planum [Sullivan *et al.*, 2005; Jerolmack *et al.*, 2006]. Additionally, ubiquitous crater-associated dark wind streaks, formed from south-southeasterly winds, are found on the plains adjacent to Endeavour crater, including those associated with Victoria crater [Geissler *et al.*, 2008].

Seasonal dust streaks were mapped out from 1999–2005 using 54 MOC images by Jerolmack *et al.* [2006] to infer two major wind regimes associated with their formation (dust streak azimuths of ~130° and ~320°). The dust streak azimuths were found to correspond with the seasons immediately preceding the MRAMS modeled WR1 (southeast, at $L_s \sim 120^\circ$) and WR2 (northwest, at $L_s \sim 210^\circ$), (Figs. 2.8, 2.10). The season when the Victoria crater dark streak

typically emerges has been found to be $\sim L_s$ 90° (with some variability) [Geissler *et al.*, 2010] and is broadly correlated with GCM and MRAMS predicted patterns (Fig. 2.10).

However, as previously discussed (section 3.6) MRAMS does not explicitly predict flow magnitudes that are greater than the threshold required for the initiation of sand saltation. Prior modeling work has suggested that Meridiani Planum is not a windy place by martian standards [e.g., Rafkin and Michaels, 2003]. Wind magnitudes predicted in other mesoscale atmospheric modeling work commonly are too low to move sediment [Fenton *et al.*, 2005; Hayward *et al.*, 2009]. This result may be due to insufficient model spatial and temporal resolution, such that convective turbulent gusts, likely important in aeolian processes on Mars [e.g., Michaels and Fenton, 2010; Fenton and Michaels, 2010], are not resolved. Alternatively (or additionally), energy from decaying northern autumn and winter baroclinic storms from southern Chryse Planitia (to the northwest) may episodically enhance the northwesterly flow so that it is more effective at mobilizing sediment than the easterly regime. Further atmospheric modeling work (particularly at higher resolutions) is needed to improve our understanding of the contemporary aeolian sediment transport processes at work in and near Endeavour crater. To conclude, modeled (mean) winds are insufficient to initiate saltation, but less frequent turbulent winds superimposed on these mean winds would provide sufficient shear stress to entrain sediment.

4.5 Possible Future Observations by Opportunity

Endeavour crater has been the future destination for Opportunity since leaving Victoria crater in 2008. Its rim has become increasingly apparent in panoramic images of the southern horizon (Fig. 2.11). Upon Opportunity's anticipated arrival at the northwestern rim of Endeavour, possibly as early as 2011, the rover will be ~ 6 km from the portion of the crater floor

where we have identified eroding dunes and bedforms. The average lateral extent (perpendicular to the line of sight from the northwest rim) of dunes that have experienced 100% deflation is ~60 m. Opportunity's Pancam instrument [Squyres *et al.*, 2003; Bell *et al.*, 2003] has a resolution of 0.28 mrad/pixel [Bell *et al.*, 2006] meaning that dunes of this size would span ~18 Pancam pixels (assuming 100% deflation) as viewed from the northwest rim. If Opportunity spends as much time at Endeavour crater as it spent at Victoria crater (>1 Mars year), the rover could detect the removal of additional dunes if more occur at the scale and rates we have observed in recent orbital data. Additionally, the MRAMS modeling work suggests that certain seasons (and even times of sol) would be optimal for documenting dune change in and near the crater (section 3.6, Figs. 2.7–2.8). Documentation of aeolian bedform movement from the surface in coordination with orbital observations would help clarify the apparent disconnect between the two observational scales that has been prevalent until now. If Opportunity is still healthy upon reaching the rim of Endeavour crater, an attractive long-term intracrater drive target might be to the vicinity of these changing dunes, allowing this active erosional process to be viewed *in situ* (or nearly so) for the first time on Mars.

5.0 Summary and Conclusions

Eight Endeavour crater dunes with a combined surface area of ~14,000 m² show activity over the past decade. These changes constitute the most extensive contemporary activity of sand-sized sediment observed to date. These dunes primarily show evidence for deflation, up to 100% in some cases, but two dunes have translated as part of bedform migration. The active dunes in Endeavour crater are dome dunes, the same morphologic class as the only other group of dunes known to be active on Mars, identified in the north polar region by Bourke *et al.* [2008].

The fact that these two groups share the same morphologic type despite the vastly different surface environmental conditions (e.g., temperature, carbon-dioxide frost cycle and water ice) implies that small dome dunes may be the best candidates for seeking out decadal-timescale surface dune change from remote sensing. The dune activity in Endeavour crater shows that it has experienced relatively strong atmospheric flow, which could have been relatively continuous or episodic. Mesoscale atmospheric modeling suggests that wind energy and possible sand transport within the crater may peak during the southern spring and summer seasons. The sand from these dunes has either been thinly dissipated and/or incorporated into larger dunes ~100s of meters downwind that also show evidence for modification. Future orbital observations of this dune field with greater temporal resolution may be used to document surface evolution and sand transport rates. Anticipated MER operations at the rim of Endeavour crater may provide a chance to document large-scale aeolian modification of the surface as it occurs

References

- Arvidson, R. E., et al. (2010), Opportunity Mars Rover Mission: Overview and Selected Results from Purgatory Ripple to Traverses to Endeavour Crater, *J. Geophys. Res.*, doi:10.1029/2010JE003746.
- Bagnold, R. A. (1941), *The Physics of Blown Sand and Desert Dunes*, 265 pp., Methuen, London.
- Bell, J. F., III, et al. (2003), Mars Exploration Rover Athena Panoramic Camera (Pancam) investigation, *J. Geophys. Res.*, 108(E12), 8063, doi:10.1029/2003JE002070.
- Bell, J. F., III, et al. (2006), In-flight calibration and performance of the Mars Exploration Rover Panoramic Camera (Pancam) instruments, *J. Geophys. Res.*, 111, E02S03, doi:10.1029/2005JE002444.
- Bourke, M. C. (2010), Barchan dune asymmetry: Observations from Mars and Earth, *Icarus* 205, 183–197.
- Bourke, M.C., and A. S. Goudie (2009), Varieties of barchan dunes in the Namib Desert and on Mars, *Aeolian Res.*, doi:10.1016/j.aeolia.2009.05.002.
- Bourke, M.C., Balme, M., R. Bayer, R., K. K. Williams, and J. R. Zimbelman (2006), A comparison of methods used to estimate the height of sand dunes on Mars, *Geomorphology*, 81, 440–452.
- Bourke, M. C., K. S., Edgett K. S., and B. A. Cantor (2008), Recent aeolian dune change on Mars, *Geomorphology*, 94, 247–255.
- Bourke, M.C., Philippoff, A., Bridges, N. (2009a), Studies of dune change on Mars combining MOC and HiRISE images, *Lunar Planet. Sci. Conf. XL* (Abstract 1748).
- Bourke, M. C., R. C. Ewing, D. Finnegan, and H. A. McGowan (2009b), Sand dune movement in the Victoria Valley, Antarctica, *Geomorphology*, 109, 148–160.
- Bridges, N. T., R. Greeley, A. F. C. Haldemann, K. E. Herkenhoff, M. Kraft, T. J. Parker, and A. W. Ward (1999), Ventifacts at the Pathfinder landing site, *J. Geophys. Res.*, 104, 8595–8615.
- Bridges, N. T., et al. (2007), Windy Mars: A dynamic planet as seen by the HiRISE camera, *Geophys. Res. Lett.*, 34, L23205, doi:10.1029/2007GL031445.
- Bristow, C. S., and N. Lancaster (2004), Movement of a small slipfaceless dome dune in the Namib Sand Sea, Namibia, *Geomorphology*, 59, 189–196.
- Cantor, B.A. (2007), MOC observations of the 2001 Mars planet-encircling dust storm, *Icarus* 186 (2007) 60–96.
- Chojnacki, M., J. E. Moersch, J. J. Wray and D. M. Burr (2010a), The Stratigraphy, Composition and Thermophysical Properties of Endeavour Crater, Meridiani Planum, Mars, from Orbital Remote Sensing, *Lunar Planet. Sci.*, XLI, abstract 2175.
- Chojnacki, M., D. M. Burr and J. E. Moersch (2010b), Recent Dune Changes at Endeavour Crater, Meridiani Planum, Mars, from Orbital Observations, *Lunar Planet. Sci.*, XLI, abstract 2326.
- Christensen, P. R. (1986), The spatial distribution of rocks on Mars, *Icarus*, 68, 217–238.
- Christensen, P. R., R. V. Morris, M. D. Lane, J. L. Bandfield, and M. C. Malin (2001), Global mapping of Martian hematite mineral deposits: Remnants of water-driven processes on early Mars, *J. Geophys. Res.*, 106, 23,873–23,885.

- Christensen, P. R., et al. (2003), The Thermal Emission Imaging System (THEMIS) for the Mars 2001 Odyssey mission, *Space Sci. Rev.*, 110, 85–130, doi:10.1023/B:SPAC.0000021008.16305.94.
- Conradsen, K., L. B. Nielsen, and L. P. Prahm (1984), Review of Weibull Statistics for Estimation of Wind-Speed Distributions, *Journal Of Climate And Applied Meteorology*, 23 (8), 1173-1183.
- Dong, Z., X. Wang, and G. Chen (2000), Monitoring sand dune advance in the Taklimakan Desert, *Geomorphology*, 35 (3–4), 219–231.
- Edgett, K.S., and M. C. Malin (2000), New views of Mars eolian activity, materials, and surface properties: three vignettes from the Mars global surveyor Mars orbital camera, *J. Geophys. Res.*, 105 (E1), 1623–1650.
- Fenton, L. K., A. D. Toigo, and M. I. Richardson (2005), Aeolian processes in Proctor Crater on Mars: Mesoscale modeling of dune-forming winds, *J. Geophys. Res.*, 110, E06005, doi:10.1029/2004JE002309.
- Fenton, L. K., and T. I. Michaels (2010), Modeling Aeolian Erosion Potential on Mars with the MRAMS LES, Second International Planetary Dunes Workshop, Alamosa, Colorado (USA), abstract 2026.
- Ferguson, R. L., P. R. Christensen, J. F. Bell III, M. P. Golombek, K. E. Herkenhoff, and H. H. Kieffer (2006a), Physical properties of the Mars Exploration Rover landing sites as inferred from Mini-TES–derived thermal inertia, *J. Geophys. Res.*, 111, E02S21, doi:10.1029/2005JE002583.
- Ferguson, R. L., P. R. Christensen, and H. H. Kieffer (2006b), High-resolution thermal inertia derived from the Thermal Emission Imaging System (THEMIS): Thermal model and applications, *J. Geophys. Res.*, 111, E12004, doi:10.1029/2006JE002735.
- Finkel, H. J. (1959), The barchans of southern Peru, *Journal of Geology*, 67, 614–647.
- Flammarion, C. (1892), *La Plane`te Mars et ses Conditions d’Habitabilite’* Vols 1 and 2, Gauthier Villars et Fils, Paris.
- Gaddis, L.R., et al. (1997), An Overview of the Integrated Software for Imaging Spectrometers, *Lunar Planet. Sci.*, XXVIII, Abstract, pp. 387-388.
- Geissler, P. E. (2005), Three decades of Martian surface changes, *J. Geophys. Res.*, 110, E02001, doi:10.1029/2004JE002345.
- Geissler, P. E., et al. (2008), First in situ investigation of a dark wind streak on Mars, *J. Geophys. Res.*, 113, E12S31, doi:10.1029/2008JE003102.
- Geissler, P. E., R. Sullivan, M. Golombek, J. R. Johnson, K. Herkenhoff, N. Bridges, A. Vaughan, J. Maki, T. Parker, and J. Bell (2010), Gone with the wind: Eolian erasure of the Mars Rover tracks, *J. Geophys. Res.*, 115, E00F11, doi:10.1029/2010JE003674.
- Golombek, M. P., et al. (2003), Selection of the Mars Exploration Rover landing sites, *J. Geophys. Res.*, 108(E12), 8072, doi:10.1029/2003JE002074.
- Golombek, M. et al., (2010), Constraints on Ripple Migration at Meridiani Planum from 3 Opportunity and HiRISE Observations of Fresh Craters, *J. Geophys. Res.*, 115, E00F08, doi:10.1029/2010JE003628.
- Greeley, R., R. Leach, B. R. White, J. Iversen, and J. Pollack (1980), Threshold windspeeds for sand on Mars: Wind tunnel simulations, *Geophys. Res. Lett.*, 7, 121–124.
- Greeley, R., S. Lee, and P. Thomas (1992), *Martian Aeolian Processes, Sediments and Features, Mars*, Edited by Kieffer, Jakosky, Snyder and Matthews.

- Greeley, R., et al. (2002), Terrestrial analogs to wind-related features at the Viking and Pathfinder landing sites on Mars, *J. Geophys. Res.*, 107, E15005, doi:10.1029/2000JE001481.
- Greeley, R., et al. (2006), Gusev crater: Wind-related features and processes observed by the Mars Exploration Rover Spirit, *J. Geophys. Res.*, 111, E02S09, doi:10.1029/2005JE002491.
- Greeley, R., et al. (2008), Columbia Hills, Mars: Aeolian features seen from the ground and orbit, *J. Geophys. Res.*, 113, E06S06, doi:10.1029/2007JE002971.
- Haberle, R. M., J. B. Pollack, J. R. Barnes, R. W. Zurek, C. B. Leovy, J. R. Murphy, J. Schaeffer, and H. Lee (1993), Mars atmospheric dynamics as simulated by the NASA Ames general circulation model, I. The zonal-mean circulation, *J. Geophys. Res.*, 98, 3093–3123.
- Hayward, R. K., T. N. Titus, T. I. Michaels, L. K. Fenton, A. Colaprete, and P. R. Christensen (2009), Aeolian dunes as ground truth for atmospheric modeling on Mars, *J. Geophys. Res.*, 114, E11012, doi:10.1029/2009JE003428.
- Herkenhoff, K. et al. (2006), Overview of Athena Microscopic Imager Results, *Lunar Planet. Sci.*, XXXVII, abstract 1816.
- Hesse, R. (2008), Do swarms of migrating barchan dunes record paleoenvironmental changes? A case study spanning the middle to late Holocene in the Pampa de Jaguay, southern Peru, *Geomorphology*, doi:10.1016/j.geomorph.2008.08.006
- Hynek, B. M., R. E. Arvidson, and R. J. Phillips (2002), Geologic setting and origin of Terra Meridiani hematite deposit on Mars, *J. Geophys. Res.*, 107(E10), 5088, doi:10.1029/2002JE001891.
- Jerolmack, D. J., D. Mohrig, J. P. Grotzinger, D. A. Fike, and W. A. Watters (2006), Spatial grain size sorting in eolian ripples and estimation of wind conditions on planetary surfaces: Application to Meridiani Planum, Mars, *J. Geophys. Res.*, 111, E12S02, doi:10.1029/2005JE002544.
- Jimenez, J. A., L. P. Maia, J. Serraà, and J. Morais (1999), Aeolian dune migration along the Ceara coast, north-eastern Brazil, *Sedimentology*, 46, 689–701.
- Long, J. T., and R. P. Sharp (1964) Barchan-dune movement in Imperial Valley, California, *Geol. Soc. Am. Bull.*, 75, 149–156.
- Malin, M. C., and K. S. Edgett (2001), Mars Global Surveyor Mars Orbiter Camera: Interplanetary cruise through primary mission, *J. Geophys. Res.*, 106, 23,429–23,570, doi:10.1029/2000JE001455.
- Malin, M. C., et al. (1992), Mars Observer Camera, *J. Geophys. Res.*, 97, 7699–7718.
- Malin, M. C., et al. (2007), Context Camera Investigation on board the Mars Reconnaissance Orbiter, *J. Geophys. Res.*, 112, E05S04, doi:10.1029/2006JE002808.
- McEwen, A. S., et al. (2007), Mars Reconnaissance Orbiter's High Resolution Imaging Science Experiment (HiRISE), *J. Geophys. Res.*, 112, E05S02, doi:10.1029/2005JE002605.
- McKee, E. D. (1979), Introduction to a study of global sand seas, in *A Study of Global Sand Seas*, edited by E. D. McKee, U.S. Geol. Surv. Prof. Pap., 1052, 3–17.
- Michaels, T. I., and L. K. Fenton (2010), Characterizing the Sensitivity of Daytime Turbulent Activity and Aeolian Erosion Potential on Mars with the MRAMS LES, *Second International Planetary Dunes Workshop*, Alamosa, Colorado (USA), abstract 1955.
- Monolithic Dome Institute (2001), Spherical dome formulas. Available at <http://www.monolithic.com/stories/advanced-dome-calculator>.
- Neukum G., R., et al. (2004), HRSC: the High Resolution Stereo Camera of Mars Express, *ESA SP-1240*, 17–35.

- Newman, C. E., S. R. Lewis, P. L. Read, and F. Forget (2002), Modeling the Martian dust cycle, 1, Representations of dust transport processes, *J. Geophys. Res.*, 107(E12), 5123, doi:10.1029/2002JE001910.
- Presley, M. A., and P. R. Christensen (1997), Thermal conductivity measurements of particulate materials: 1. A review, *J. Geophys. Res.*, 102(E3), 6535–6549.
- Piqueux, S., and P. R. Christensen (2009), A model of thermal conductivity for planetary soils: 1. Theory for unconsolidated soils, *J. Geophys. Res.*, 114, E09005, doi:10.1029/2008JE003308.
- Putzig, N. E., and M. T. Mellon (2007), Apparent thermal inertia and the surface heterogeneity of Mars, *Icarus*, 191, 68–94.
- Pye, K., and H. Tsoar (1990), *Aeolian Sand and Sand Dunes*, Unwin Hyman, London, 396 pp.
- Rafkin, S. C. R., R. M. Haberle, and T. I. Michaels (2001), The Mars regional atmospheric modeling system: Model description and selected simulations, *Icarus*, 151, 228–256, doi:10.1006/icar.2001.6605.
- Rafkin, S. C. R., and T. I. Michaels (2003), Meteorological predictions for 2003 Mars Exploration Rover high-priority landing sites, *J. Geophys. Res.*, 108, doi:10.129/20002JE0002027.
- Reffet, E., S. Courrech du Pont, P. Hersen, M. Fulchignoni, and S. Douady (2009), Solitary dunes under bimodal winds, American Astronomical Society, DPS meeting #40, #15.09. Bulletin of the American Astronomical Society, 557 pp.
- Ruff, S. W., and P. R. Christensen (2002), Bright and dark regions on Mars: Particle size and mineralogical characteristics based on Thermal Emission Spectrometer data, *J. Geophys. Res.*, 107(E12), doi:10.1029/2001JE001580.
- Sagan, C. et al (1973), Variable features on Mars: Preliminary Mariner 9 television results, *Icarus*, 22, 346–372.
- Sauermann, G. et al (2000), The shape of the barchan dunes of Southern Morocco, *Geomorphology*, 36, 47–62.
- Silvestro, S., L. K. Fenton, and D. A. Vaz (2010a), Ripple Migration and Small Modifications of Active Dark Dunes in Nili Patera (Mars), *Lunar Planet. Sci.*, XLI, abstract 1838.
- Silvestro, S., L. K. Fenton, D. A. Vaz, N. T. Bridges and G. G. Ori (2010b), Ripple migration and dune activity on Mars: Evidence for dynamic wind processes, *Geophys. Res. Lett.*, 37, L20203, doi:10.1029/2010GL044743.
- Smith, D. E., et al. (2001), Mars Orbiter laser altimeter: Experiment summary after the first year of global mapping of Mars, *J. Geophys. Res.*, 106, 23,689–23,722, doi:10.1029/2000JE001364.
- Smith, M. D. (2004), Interannual variability in TES atmospheric observations of Mars during 1999–2003, *Icarus*, 167, 148–165.
- Sullivan, R., et al. (2005), Aeolian processes at the Mars Exploration Rover Meridiani Planum landing site, *Nature*, 436, 58–61, doi:10.1038/nature03641.
- Sullivan, R., et al. (2008), Wind-driven particle mobility on Mars: Insights from Mars Exploration Rover observations at “El Dorado” and surroundings at Gusev Crater, *J. Geophys. Res.*, 113, E06S07, doi:10.1029/2008JE003101.
- Squyres, S. W., et al. (2003), Athena Mars rover science investigation, *J. Geophys. Res.*, 108(E12), 8062, doi:10.1029/2003JE002121.
- Squyres, S. W., et al. (2004), In situ evidence for an ancient aqueous environment at Meridiani Planum, Mars, *Science*, 306, 1709–1714, doi:10.1126/science.1104559.

- Taylor, P. A., P. J. Mason, and E. F. Bradley (1987), Boundary-Layer Flow Over Low Hills, *Boundary-Layer Meteorology*, 39, 107-132.
- Tsoar, H., B. R. White, and E. Berman (1996), The effect of slopes on sand transport - numerical modelling, *Landscape and Urban Planning*, 34, 17 1 - 18 1.
- Vaughan, A., J. R. Johnson, K. E. Herkenhoff, R. Sullivan, G. A. Landis, W. Goetz, and M. B. Madsen (2010). Pancam and Microscopic Imager observations of dust on the Spirit Rover: Cleaning events, spectral properties, and agglomerates, *Mars*, 5, 129–145, doi:10.1555/mars.2010.0005.
- Wray, J. J., E. Z. Noe Dobrea, R. E. Arvidson, S. M. Wiseman, S. W. Squyres, A. S. McEwen, J. F. Mustard, and S. L. Murchie (2009). Phyllosilicates and sulfates at Endeavour Crater, Meridiani Planum, Mars, *Geophys. Res. Lett.* 36, L21201.

Appendix II

Table 2.1 Morphometric Values Before and After Bedform Removal and/or Transport^a.

	Image ID, DOA, and L _s	Length (m)	Width (m)	Area (m ²)	Area Loss	Distance Moved (m)	Time Change Earth (years)	Time Change Mars (years)	Area Removal Rate Earth (m ² /years)	Area Removal Rate Mars (m ² /years)	Transport Rate Earth (m/years)	Transport Rate Mars (m/years)
Dune i	M1 12/8/2001 L _s 287°	58	24	1071								
	H1 3/30/2008 L _s 52°	N/A	N/A	0	100%	N/A	6.3	3.3	170	325	N/A	N/A
Dune ii	M1 12/8/2001 L _s 287°	63	29	1630								
	H1 3/30/2008 L _s 52°	N/A	N/A	0	100%	N/A	6.3	3.3	259	494	N/A	N/A
Dune iii	M1 12/8/2001 L _s 287°	71	43	2094								
	H1 3/30/2008 L _s 52°	47	30	1159	45%	N/A	6.3	3.3	148	283	N/A	N/A
	C1 3/30/2008 L _s 52°	45	31	1090								
	C2 7/18/2009 L _s 306°	41	29	827	24%	N/A	1.3	0.7	202	376	N/A	N/A
Dune iv	M2 12/30/2003 L _s 325°	62	39	1929								
	C1 3/30/2008 L _s 52°	48	34	1259	35%	N/A	4.3	2.3	156	291	N/A	N/A
Dune v	M2 12/30/2003 L _s 325°	109	45	4322								
	C1 3/30/2008 L _s 52°	69	50	2728	37%	20	4.3	2.3	371	693	4.7	8.7
Dune vi	M2 12/30/2003 L _s 325°	131	84	8666								
	C1 3/30/2008 L _s 52°	88	74	5786	33%	10	4.3	2.3	670	1252	2.3	4.3
Dune vii	M1 12/8/2001 L _s 287°	97	76	4718								
	H1 3/30/2008 L _s 52°	75	60	3103	34%	N/A	6.3	3.3	256	489	N/A	N/A
	C1 3/30/2008 L _s 52°	71	62	3185								
	C2 7/18/2009 L _s 306°	57	51	1422	55%	N/A	1.3	0.7	1356	2519	N/A	N/A
Dune viii	M1 12/8/2001 L _s 287°	129	96	7313								
	H1 3/30/2008 L _s 52°	117	82	6666	9%	N/A	6.3	3.3	103	196	N/A	N/A
	C1 3/30/2008 L _s 52°	122	83	7010								
	C2 7/18/2009 L _s 306°	110	81	6019	14%	N/A	1.3	0.7	762	1416	N/A	N/A
Total				14,059								

^aThe dates of image of acquisitions (DOA) are given. Image identification and resolution are the following: C1= P17_007849_1793_XN_00S005W at 5.37 m/pixel, C2=B11_013954_1780_XN_02S005W at 5.37 m/pixel, H1=PSP_007849_1775 at 0.25 m/pixel, M1=E1101328 at 3.61 m/pixel, and M2=R1203949 at 3.59 m/pixel, where C=CTX, H=HiRISE, and M=MOC. Values for dunes iii, vii and viii are based on two pairs of observations, M1–H1 and C1–C2

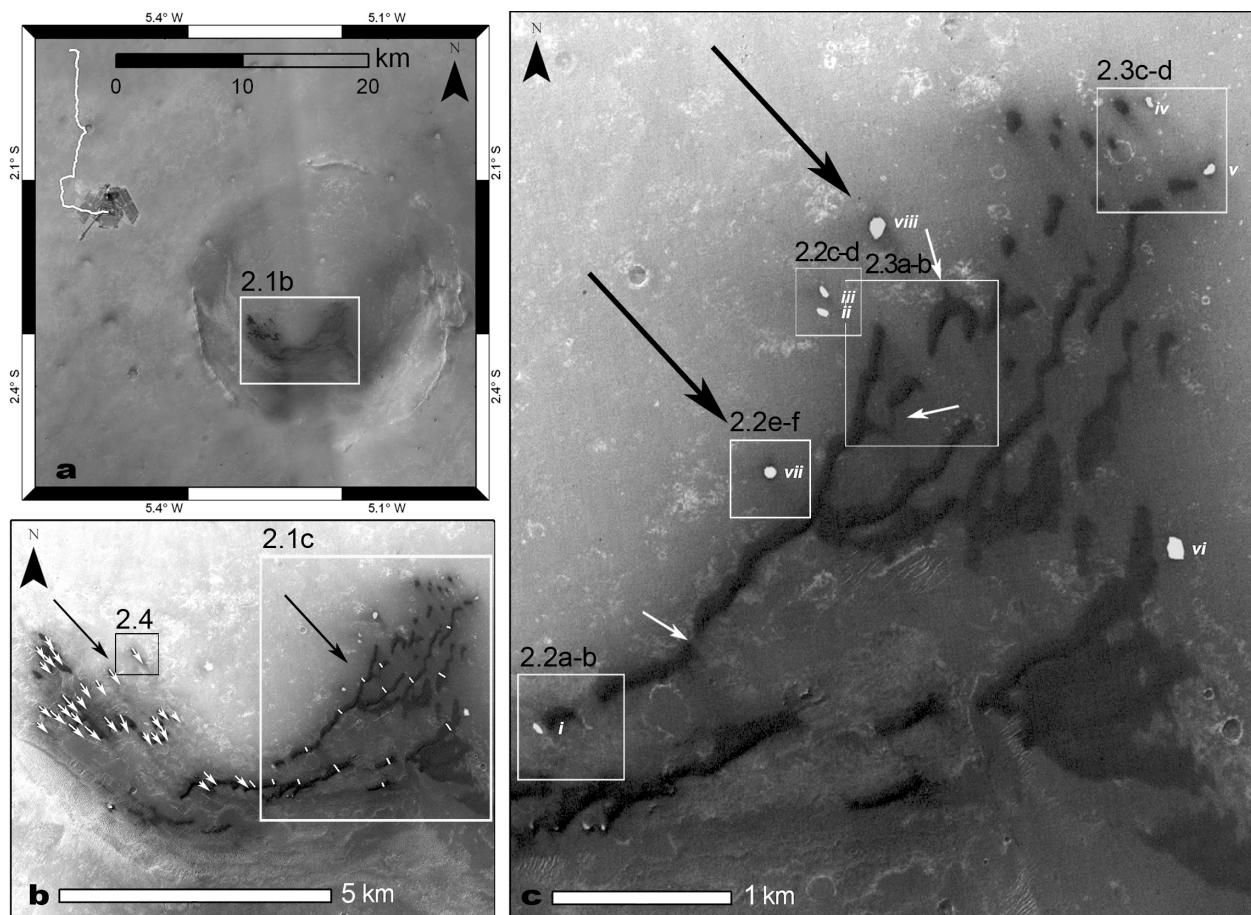


Figure 2.1 Context images showing dune fields in Endeavour crater, Meridiani Planum. (a) A CTX visible-wavelength image mosaic showing Endeavour crater and the location of the Opportunity rover as of December 2010. (b) CTX image P17_007849_1793_XN_00S005W showing the crater basin and the aeolian sand dunes forming two distinct groups to the east and west (black arrows). For the well-formed barchans in the west, white arrows are oriented perpendicular to dune slip faces (downwind). For the degraded dunes and bedforms in the east, white lines are oriented perpendicular to transverse dune orientation. (c) A closer view of the eastern dunes showing the bedforms that were observed to change from 2001/2003 to 2008/2009. Black arrows indicate inferred primary wind direction and white arrows show areas of dune morphological modification of larger dunes. North is towards the top of the figure in all images unless otherwise indicated.

Figure 2.2 Visible-wavelength images showing the deflation of dunes i-iii and vii. Blue polygons represent the extent of dunes in 2001 and yellow the extent in 2008. See Table 2.1 for all image numbers and resolutions. (a) MOC image and (b) a HiRISE image down-sampled to MOC resolution with approximately the same contrast enhancement of the scene. A closer view in the color HiRISE image illustrating that the texture of the eastern dune is no longer visible where dune i was previously located (inset with a ~80 m field of view). (c) MOC shows dune ii in 2001. (d) HiRISE shows the deflation for dune ii and the ~50% reduction in size of dune iii in 2008. Closer examination shows differences in surface texture from dune ii to dune iii and hints at a southward erosional direction (inset with a ~80 m field of view). (e) MOC image and (f) HiRISE shows a ~35% reduction in area for dune vii. Closer examination shows differences in surface texture from the previous extent (inset with a ~80 m field of view).

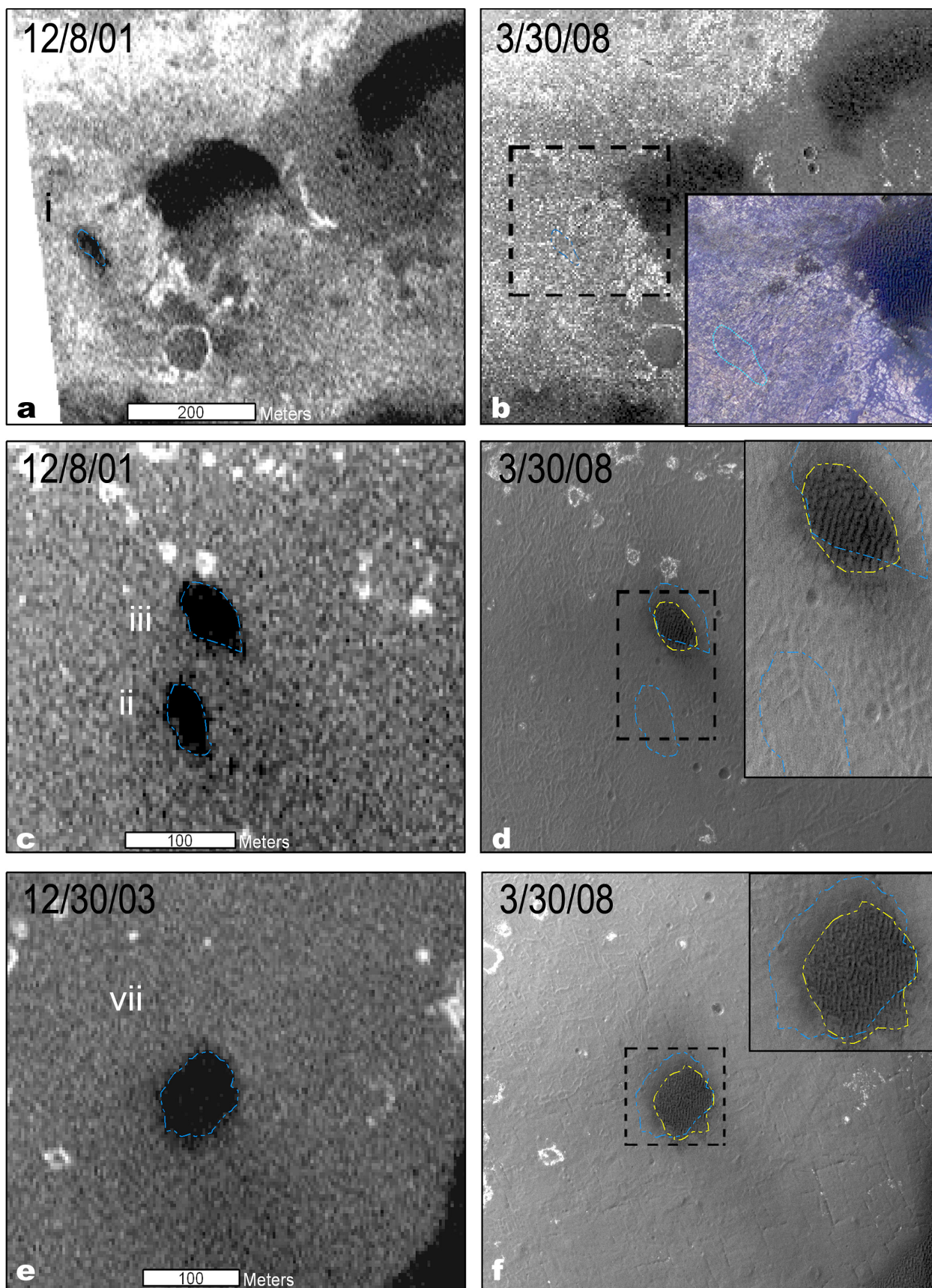


Figure 2.2

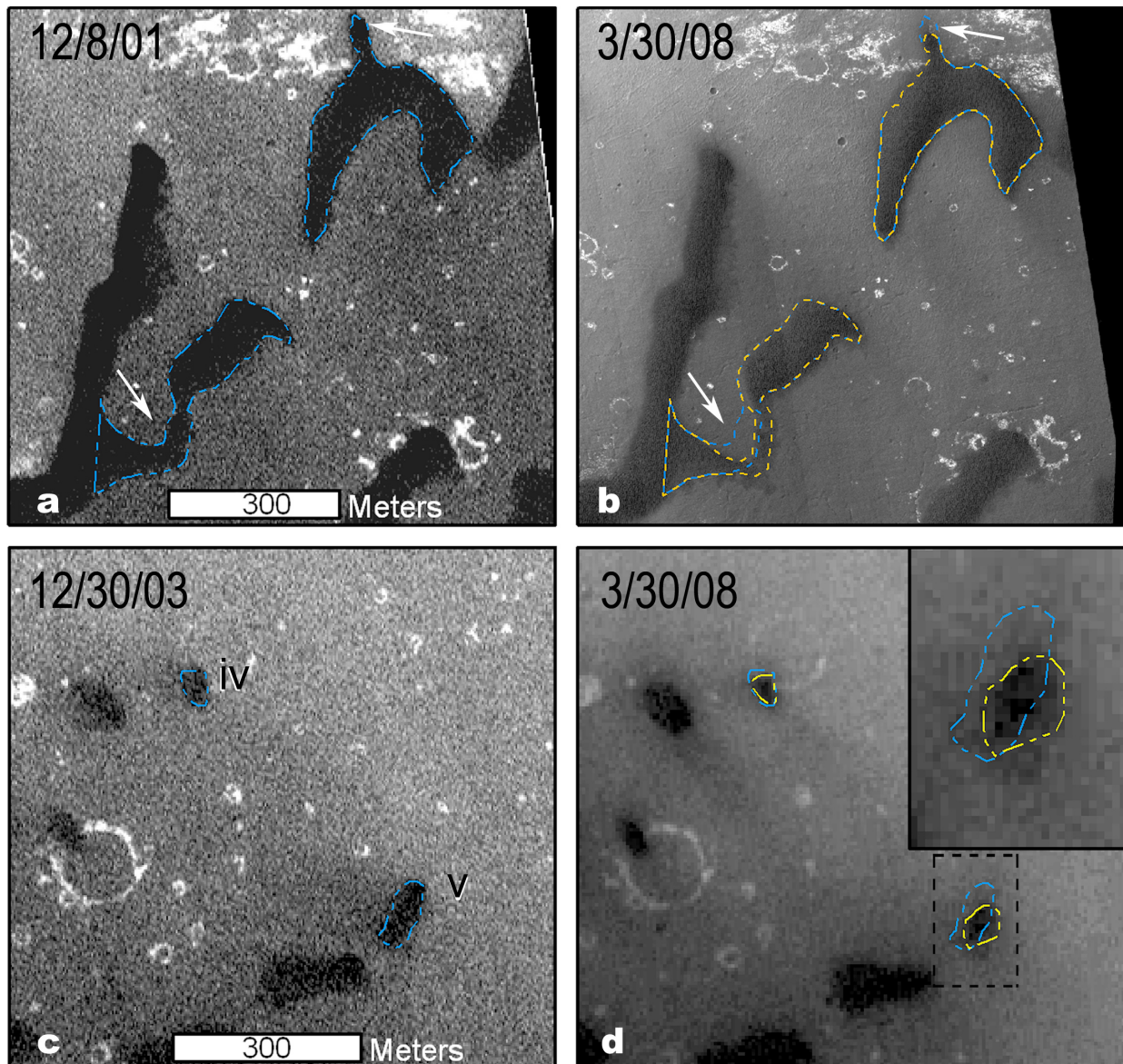


Figure 2.3 Deflation, modification and transport of Endeavour crater dunes and bedforms. Blue polygons represent the extent of dunes in 2001 and 2003 and yellow the extent of bedforms in 2008. Modifications (white arrows) of larger dunes (south of dune ii-iii) are shown in the (a) 2001 MOC and (b) 2008 HiRISE image. (c) A 2003 MOC image showing the easternmost dunes iv-v and their subsequent reduction in size over 2.3 Mars years in the (d) CTX image. Close examination of dune v suggests southward migration ~20 m in addition to sediment loss (inset with a ~150 m field of view).

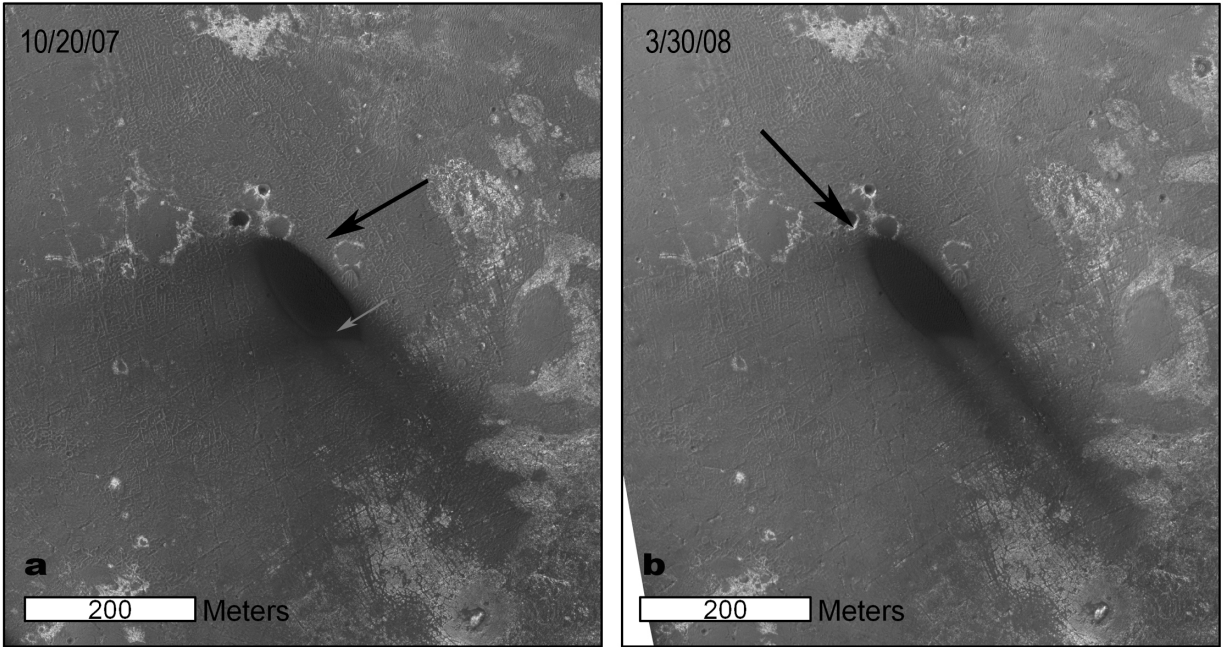


Figure 2.4 HiRISE images of a degrading barchan dune of the western dune field group. (a) PSP_005779_1775 taken on 20 Oct 2007 ($L_s \sim 334^\circ$) and (b) PSP_007849_1775 on 30 March 2008 ($L_s \sim 51^\circ$), are separated by one martian season (0.24 Mars year). Sand streamers appear both to the southeast and west-southwest in (a). However, the distribution appears to be more concentrated to the southeast and the dark-toned material, inferred to be sand, present in (a) appears to have been largely removed in the western portion of the second image (b). Also, note the degraded state of the slip face of this former barchan dune (gray arrow). Phase angle is $\sim 15^\circ$ greater for the second image.

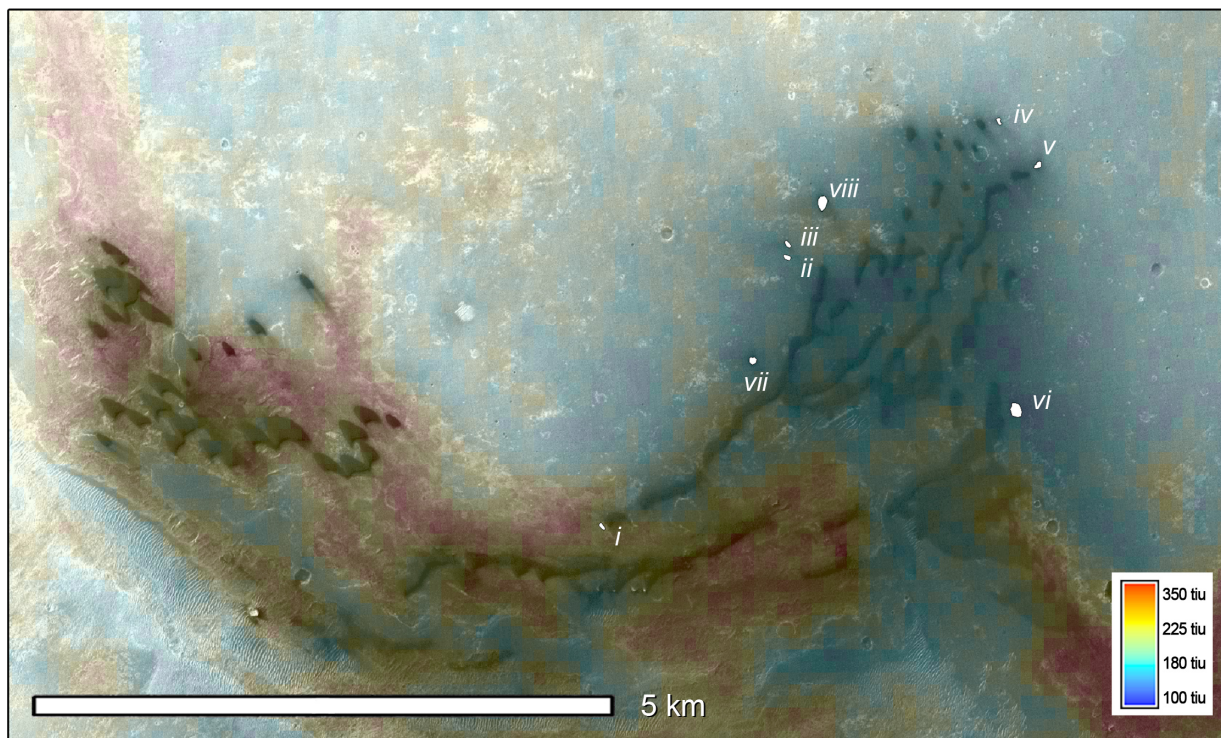


Figure 2.5 A CTX mosaic colorized with THEMIS thermal inertia of Endeavour crater. Dunes that show change in morphology are highlighted in white. The western dune fields are found to have thermal inertia values consistent with fine to medium sand-sizes, whereas eastern dunes that show change have thermal inertia values suggestive of very fine to fine sand-sizes. THEMIS image I22535006.

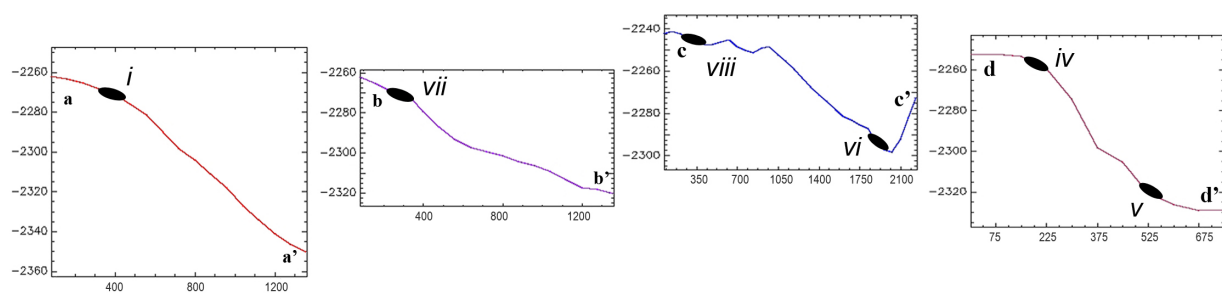
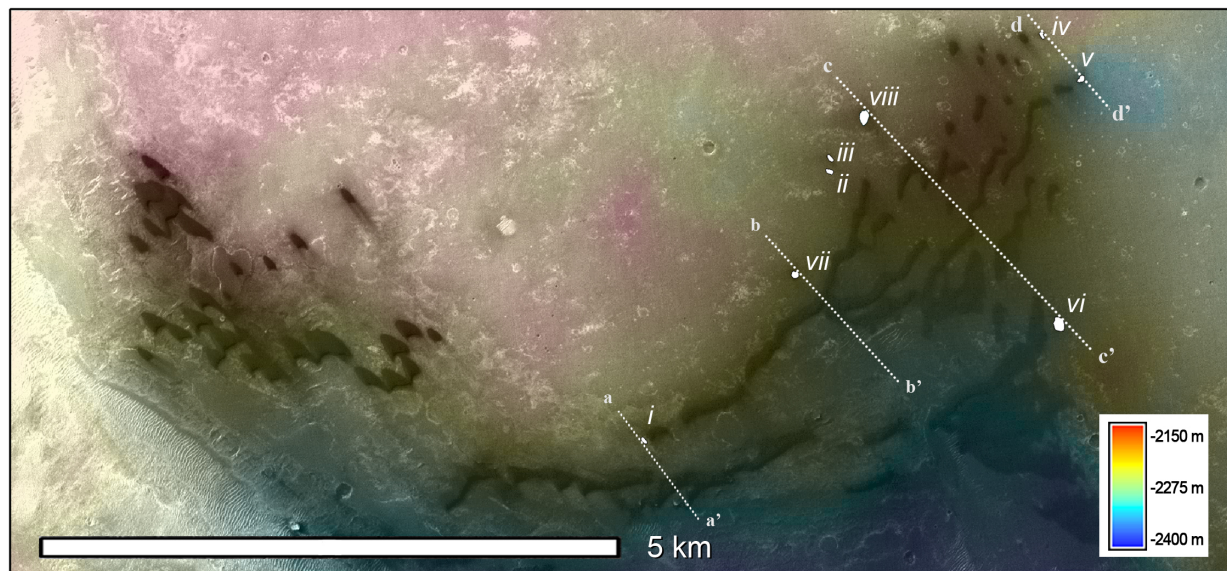


Figure 2.6 A CTX mosaic colorized with HRSC elevation of Endeavour crater.

The eastern dunes and bedforms are situated at the top of a ~200 m-tall hill, that drops off southward and downwind across the dune field. Four topographic profiles (in units of meters) with the same vertical spacing are given across dunes that show change (highlighted in white). Profiles are parallel to and lead downwind (inferred direction) of the dunes that show change (shown by black ellipse on the profiles). HRSC DEM H2064_0000_DT4 at 75 m/pixel.

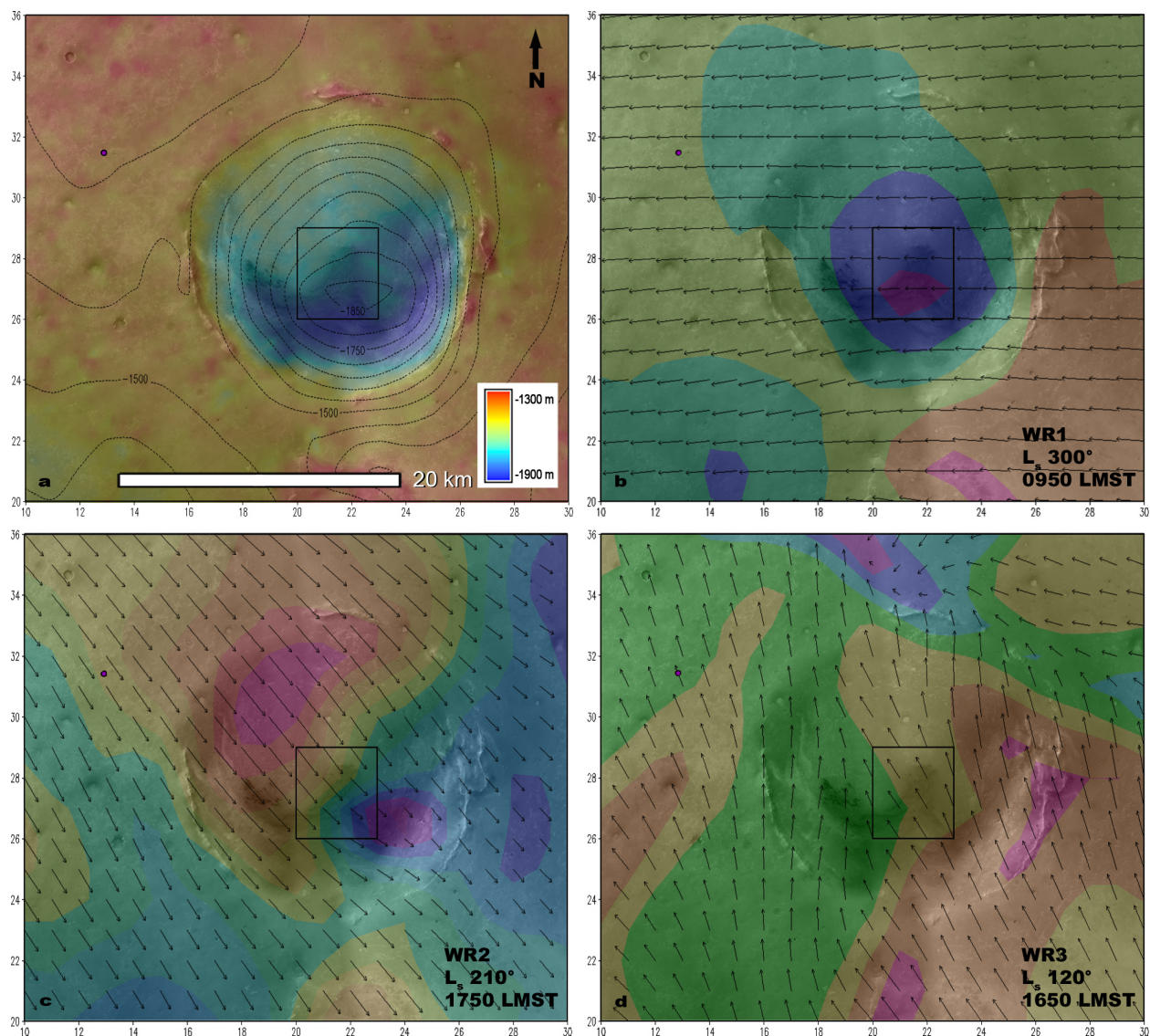


Figure 2.7 Topography and MRAMS maps over CTX mosaics of Endeavour crater. (a) CTX image mosaic of Endeavour crater and its vicinity, overlain by color-shaded HRSC elevation (100 m/pixel H1183_0000_DT4) and the MOLA-based elevation contours used by MRAMS. (b-d) CTX visible-wavelength image mosaic of the Endeavour crater vicinity with examples of flow regimes: (b) wind regime 1, MRAMS wind vectors and magnitude 1.3 m above ground level (AGL) (interval 0.5, purple < 6.5, red > 8.5 m/s) at $L_s \sim 300$ and 0950 LMST, (c) wind regime 2, MRAMS wind vectors and magnitude 1.3 m AGL (interval 0.5, purple < 1, red > 3.5 m/s) at $L_s \sim 120$ and 1650 LMST, and (d) wind regime 3, MRAMS wind vectors and magnitude at 1.3 m AGL (interval 0.3, light purple < 1.5, dark red > 4.5 m/s) at $L_s \sim 210$ and 1750 LMST. The magenta circle in the northwest quadrant of the figures represents the location of the Opportunity rover in December of 2010.

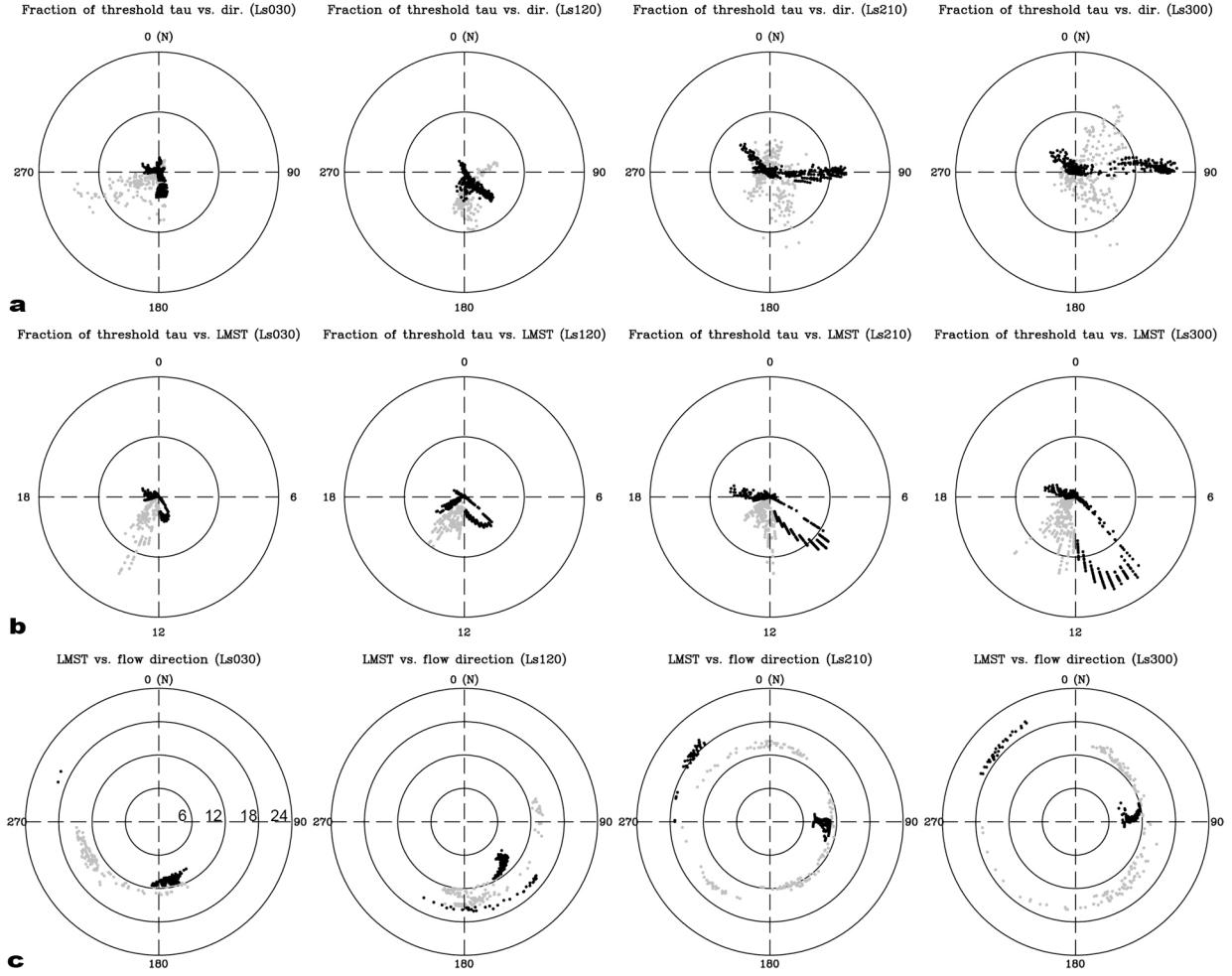


Figure 2.8 MRAMS polar coordinate scatter plot for four representative seasons.

Each of the 20 model grid points within the outlined area in Fig. 2.7 are plotted at each of the 72 model output times within in a single sol for $L_s \sim 30^\circ$, 120° , 210° , and 300° . Points with LMST between approximately 12 and 16 Mars-hours are plotted in gray to indicate that their magnitude and direction are contaminated (to varying degrees) by the occurrence within the model results of poorly resolved/realistic dry convective structures at those times. Inspection of the other (lower resolution) model grids strongly suggests that τ and direction during this time interval should not have a significantly different character than that in the late morning. Directions are given using the meteorological convention (i.e., the direction the flow is from). (a) Fraction of threshold aerodynamic surface shear stress (τ/τ_t) vs. wind direction (degrees clockwise from north), with a grid ring interval of 0.25, illustrating the flow directions with the greatest/least magnitudes. (b) Fraction of aerodynamic surface shear stress (τ/τ_t) vs. local mean solar time (Mars-hours), with a grid ring interval of 0.25, indicating the local times with the greatest/least flow magnitudes. (c) Local mean solar time vs. wind direction (degrees clockwise from north), with a grid ring interval of 6 Mars-hours and plotting only those points with $\tau/\tau_t > 0.0625$, illustrating when and for how long flow in various directions exists.

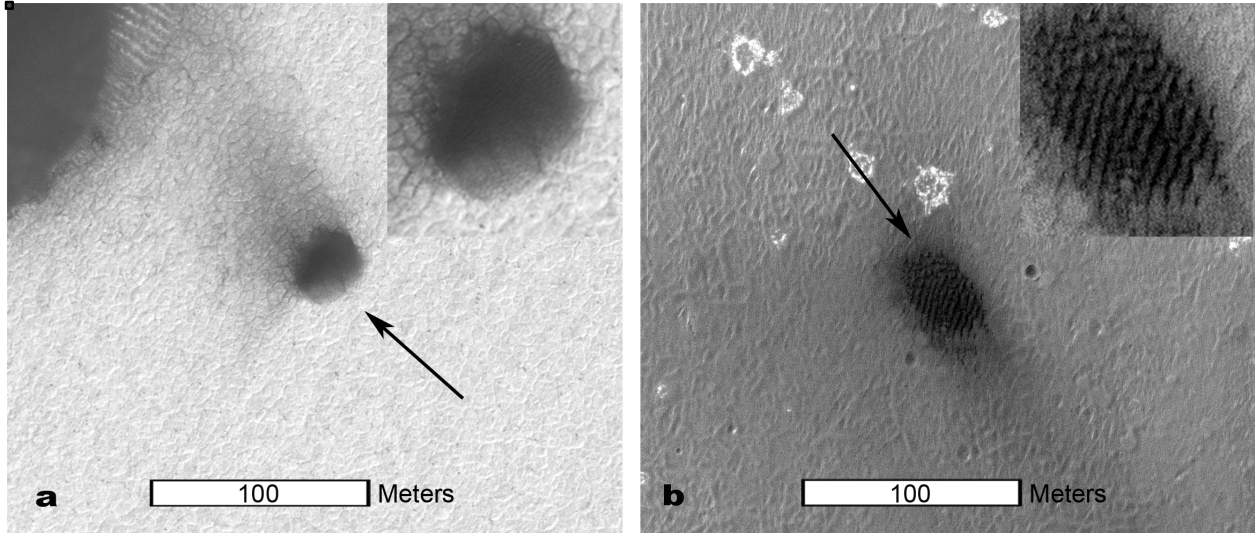


Figure 2.9 A comparison between (a) the north polar dome dune iii as described by *Bourke et al.* [2008] and the (b) Endeavour crater dune iii at the same HiRISE resolution. Note the similar morphologies of the different dunes that have both been areally deflated in the past decade. Arrows show inferred wind direction. North is towards the top of the figures. (a) PSP_009295_2565 and (b) PSP_007849_1775 where inset shows a ~90 m field of view.

Morphology	Location	Inferred wind direction (from)	Time (date of image acquisition or relative)	Figures	MRAMS Wind Regime and Season
Ripples on dome dunes ¹	Endeavour crater (E)	E or W 	Contemporary?	2.2	WR1 SH spring/summer (~ Ls 180°-360°) 
Dune sand streamers ¹	Endeavour crater (W)	E (~60°) 	10/20/2007 (L _s 334°)	2.4a	
Plains ripples ²	Meridiani plains	E 	Ancient?	na	
Dune modification ¹	Endeavour crater (E)	NW (~320°) 	Contemporary	2.2f, 2.3b	WR2 SH spring/summer (~ Ls 180°-360°) 
Dome dune transport ¹	Endeavour crater (E)	NW (~300°) 	Contemporary	2.3d	
Dune sand streamers ¹	Endeavour crater (W)	NW (~320°) 	3/30/2008 (L _s ~51°)	2.4b	
Barchan slip faces ¹	Endeavour crater (W)	NW (~320°) 	Ancient?	2.1b	
Barchanoid slip faces ¹	Craters surrounding Endeavour	NW (300°-330°) 	Ancient?	na	
Dust streaks ^{3,4}	Meridiani plains	NW (320°-340°) 	Contemporary (L _s 280°-350°)	na	
Dunes ⁵	Victoria crater	SE or NW 	Ancient?	na	WR2 or WR3 SH spring/summer or winter (~ Ls 180°-360° or 90°-180°)
Sand streaks ^{5,6}	Victoria crater	SSE (140°-160°) 	Contemporary (L _s 96°-142°)	na	WR3 SH winter (~ Ls 90°-180°) 
Dust streaks ^{3,4}	Meridiani plains	S to SSE (120°-150°) 	Contemporary (L _s 120°-150°)	na	

Figure 2.10 Table relating morphological evidence of aeolian activity to location, wind direction, time, figure, and MRAMS wind regime.

Abbreviations are as follows: WR, wind regime and SH, southern hemisphere. Here 1, this study; 2, *Arvidson et al.* [2010]; 3, *Sullivan et al.* [2005]; 4, *Jerolmack et al.* [2006]; 5, *Geissler et al.* [2008]; 6, *Geissler et al.* [2010].

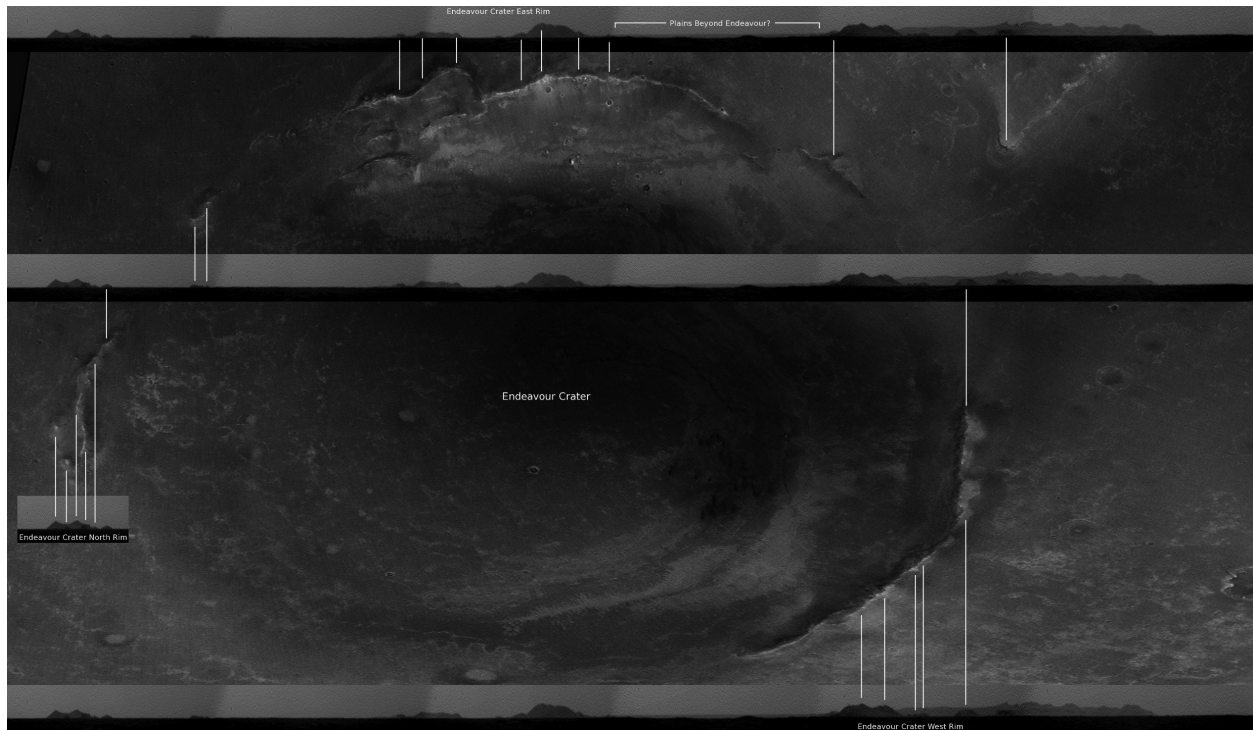


Figure 2.11 Opportunity's southward view (top, bottom, and middle Pancam mosaic insets) of Endeavour crater rim segments visible on the horizon, in comparison to HRSC orbital data. HRSC data have been re-projected into distance (vertical coordinate) and azimuth (horizontal coordinate) from the location where the Pancam mosaic was taken (South is towards the top of the figure). The azimuth coordinate is rescaled to match that of the Pancam resolution such that the same features in both ground and orbital data align vertically across the image. Pancam mosaics have been vertically stretched to exaggerate the horizon, and were taken near Concepcion crater (Fig. 2.1a) in early 2010. Image credit: NASA/JPL/Cornell University (Pancam), ESA/DLR/FU Berlin (HRSC), and processing courtesy of James Canvin at <http://www.nivnac.co.uk/mer/>.

Attachment 2.1 An animated time-step sequence of dunes ii-iii and viii between 2001 (MOC E1101328) and 2008 (HiRISE PSP_007849_1775). Compare with Figures 2.2c–d.

Attachment 2.2 An animated time-step sequence of a degrading barchan dune on the eastern edge of the western dune group. HiRISE images PSP_005779_1775 taken on 20 Oct 2007 and PSP_007849_1775 taken on 30 March 2008 are separated by one Martian season. Compare with Figures 2.4.

CHAPTER III: CLIMBING AND FALLING DUNES IN VALLES MARINERIS, MARS

This chapter is a reformatted version of a paper, by the same name, previously submitted and accepted to the *Geophysical Research Letters*. Matthew Chojnacki, Jeffrey E. Moersch, and Devon M. Burr are the authors. Matthew performed all data analysis.

Chojnacki, M., J. E. Moersch, and D. M. Burr (2010), Climbing and falling dunes in Valles Marineris, Mars, *Geophys. Res. Lett.*, 37, L08201, doi:10.1029/2009GL042263.

Abstract

Multiple occurrences of “wall dunes” are found several kilometers above the Valles Marineris canyon floor. Dune slip face orientation and bedform morphologies indicate transport direction and whether the wall dunes are climbing dunes or falling dunes. On Earth, these types of dunes form in a unidirectional wind regime and are strongly controlled by the local topography. Newly acquired Mars Reconnaissance Orbiter (MRO) images and topography of the walls of Valles Marineris show similar sand dune morphologies, as wind blown sediment has interacted with local and regional topography. Primarily found in Melas and Coprates chasmata, these climbing and falling dunes are relevant for understanding aeolian sediment flux, sediment sources, and wind directions. Falling dunes show photogeologic and thermophysical evidence of their sand being provided by adjacent outcrops.

1.0 Introduction

Aeolian activity has likely been the most consistent geologic process throughout Mars' history. Low-albedo, fine-grained bedforms are commonly found scattered at landing sites and in orbital images [*Herkenhoff et al.*, 2008]. Viking studies of Valles Marineris (VM) noted the dark mantles found on the chasm floors, interpreted as dunes [*Kieffer et al.*, 1992, and references therein]. These studies speculated that the putative dunes were composed of basaltic sediment derived from local outcrops. Although significant progress has been made over the past three decades in understanding aeolian processes and landforms on Mars, questions remain regarding dune composition, age, morphology, and sediment supply under present and past martian surface conditions.

As part of a near-global dune study, *Hayward et al* [2007] cataloged dunes widely distributed over a range of elevations and terrain types. Equatorial dune fields are typically found in classic low albedo features, such as Syrtis Major, Terra Meridiani, and Valles Marineris. Indeed, the majority of dune field area between 0° and 30° S are in VM, with 20 dune fields on VM trough floors [*Hayward et al.*, 2007]. Regional meteorology is largely influenced by the ~9 km of VM topography and the resulting atmospheric pressure gradient [*Rafkin and Michaels*, 2003].

As part of a regional study we mapped 50 dunes fields in VM (Fig. 3.1a) to identify significant trends in composition, thermophysical properties, morphology, and possible provenance [*Chojnacki and Moersch*, 2009]. That study found VM dune fields could be broadly divided into two classes: floor- and wall-related dune fields. Wall-related dune fields, or “wall dunes,” are found proximal to or on walls (Fig. 3.1). We have identified nearly two-dozen occurrences of wall dunes (sub-classified as either climbing or falling dunes) found high on the

spur-and-gully walls. Based on available data, these dunes are located preferentially on north-facing slopes in Melas and Coprates chasmata (Fig. 3.1b and 3.1c). While previous studies have described martian climbing and falling dunes within crater rims and troughs [Fenton *et al.*, 2003; Bourke *et al.*, 2006], the climbing and falling dunes found in VM are more abundant, areally larger, and are formed on greater relief. In addition, the climbing and falling dunes in VM possess distinct slip faces, unlike those previously described elsewhere on Mars. The VM climbing and falling dunes are relevant for understanding aeolian sediment flux, sediment sources, and wind direction at the time of their most recent activity. For example, dune slip face orientation preserves a record of the prevailing wind direction at the last time of major dune activity, which may be compared with atmospheric circulation models.

2.0 Climbing and Falling Dunes on Earth

As discussed in Pye and Tsoar [1990], sediment and airflow interactions with topography at all length scales in mountainous regions significantly affect dune location, size, shape, and orientation. Topography can accelerate and decelerate airflow, and create turbulence, often leading to enhanced erosion, deposition and/or transport of sediment. In areas with a near-unidirectional wind regime, topographically-related dunes can be divided into two classes depending on whether they form upwind or downwind of an obstacle. Upwind accumulations of sand include echo dunes (anchored) or climbing dunes (not anchored), whereas downwind accumulations form lee dunes (anchored) or falling dunes (not anchored). The morphology and geometry (steep cliffs $>50^\circ$) required for echo and lee dune fields [Tsoar, 1983; Liu *et al.*, 1999] are not observed in the local HRSC topography [Neukum *et al.*, 2004], and will be not be discussed further.

Climbing dunes are formed when migrating dunes encounter and ascend a substantial slope or cliff ($>10^\circ$), where there is no major wind flow blockage (Fig. 3.2) [Pye and Tsoar, 1990]. A wind direction aligned with the maximum gradient is also important in formation of climbing dunes [Liu et al., 1999]. Climbing dunes often cascade over the topography through local low points (i.e., passes) to become falling dunes as part of sediment pathways [Evans, 1962; Koscielniak, 1973].

Falling dunes, found on the downwind side of large topographic highs, are formed by unidirectional down slope winds and gravity in two main environments (Fig. 3.2). One example of falling dunes is found in Kuwait, where size and morphology are largely influenced by the size, slope, and orientation (with respect to the prevailing wind direction) of their host wadis, or desert valleys [Al-Enezi et al., 2008]. The slip face of these dunes may be parallel to wadi walls (resembling a linear dune morphology), arcuate, perpendicular to them (and crescent shaped), or absent altogether [Al-Enezi et al., 2008]. Bedforms encountering mountainous areas with greater relief than wadis may form the falling dune class [Liu et al., 1999]. If the cliff face is without a prominent protuberance, less streamline convergence, reduction of wind velocity, and sediment deposition may result in falling dunes [Pye and Tsoar, 1990].

Climbing and falling dunes have characteristic slopes and sediment pathways. The underlying topographic slopes for climbing dunes are generally shallower ($\sim 10^\circ$) than the slopes under falling dunes ($\sim 25^\circ$) [Evans, 1962; Tsoar, 1983; Liu et al., 1999]. Falling dunes preferentially are transported down gullies (or wadis) as the path of least resistance, whereas climbing dunes may encounter and ascend either gullies or spurs [Koscielniak, 1973; Al-Enezi et al., 2008]. In addition, falling dunes may act as feeders for other duneforms within the trough or

valley, whereas climbing dunes may allow sediment to escape from a topographic low [Bourke *et al.*, 2006].

The morphological characteristics that together uniquely define climbing and falling dunes are given in Fig. 3.2. Definitions of wall dune types are somewhat ambiguous in the terrestrial literature. In the present work, we define *climbing dunes* as bedforms with lee faces on the uphill side and *falling dunes* as bedforms with lee faces on the downhill side, both lying on topography $>8^\circ$ in slope (Fig. 3.2). Climbing dunes are expected lower on a wall and are often linked to floor dunes with similar slip face orientations, indicating a common transport direction. Also, dune horns of falling dunes may point upwind opposite of the transport direction, contrary to simple barchans.

3.0 Study Region and Methods

Valles Marineris is a geologically rich region with large landslides [e.g., Quantin *et al.*, 2004], aqueous mineralogical alteration [e.g., Gendrin *et al.*, 2005], and aeolian sedimentation [e.g., Hayward *et al.*, 2007]. The VM dune fields show a large range of dune morphologies, thermophysical properties, and potential sand sources [Chojnacki and Moersch, 2009]. The topography and meteorology of the expansive rift system is distinct from intracrater terrain (Fig. 3.1a) [Hayward *et al.*, 2007]. The VM aeolian morphologies are likewise distinct from those in intracrater plains, and include a variety of bedforms found several kilometers above the canyon floor on steep walls, within gullies, and atop landslide heads.

All of the observed VM wall dunes are located on much steeper ($\geq 10^\circ$) underlying topography than are the floor dunes in the region ($< 4^\circ$). Lee-side slip faces must be identifiable to determine dune class (i.e., climbing or falling), otherwise we use the more general term of

wall dunes. As discussed above, morphological evidence for transport directions, such as the presence of bedforms above or below the wall dune, near vertical topography and feasible sand pathways are also considered in our determination of whether a given dune is a wall dune. These wall dune fields are small (300 m–3 km-wide, 2–7 km-long) and have not been previously described in the literature, possibly due to the scarcity of high-resolution images.

With improved spatial resolution and areal coverage, many of these dunes can now be properly classified using data from the Mars Reconnaissance Orbiter (MRO). Images from both the Context Camera (CTX) [Malin *et al.*, 2007] and High Resolution Imaging Science Experiment (HiRISE) [McEwen *et al.*, 2007] were examined for photogeologic evidence of wall dunes. Thermal inertia maps were derived from Mars Odyssey’s Thermal Emission Imaging System (THEMIS) [Christensen *et al.*, 2004; Fergason *et al.*, 2006] data using the model of Putzig and Mellon [2007], as implemented in the “jENVI” software suite [Piatek and Moersch, 2006]. The Mars Express High Resolution Stereo Camera (HRSC) [Neukum *et al.*, 2004] level 4 digital elevation models (DEMs) were used to quantitatively investigate surface morphology (slope profiles, dune orientation) at 50–150 m/pixel. In addition, a DEM was produced from the one available HiRISE stereo pair of VM wall dunes for extraction of dune field slope profiles with ~1 m/pixel horizontal and vertical precision at 1 m/pixel spatial resolution [Kirk *et al.*, 2008].

4.0 Results

4.1 Valles Marineris Climbing Dunes

Climbing dunes, primarily in Melas Chasma (Fig. 3.1b), are only found on north-facing walls. Fig. 3.3a and 3.4a shows a series of climbing dunes extending ~5 km up a prominent

gully in the extreme southwest corner of the chasm. This dune field atop a 15° slope shows multiple slip face morphologies (crescent and linear), all indicating unidirectional, upslope gully wind. Another unambiguous group of climbing dunes is found in Coprates Chasma (Fig. 3.3b), but with a slightly different morphology. These streamlined dunes appear nearly halfway up the ~9-km-tall wall, with several minor slip faces and a prominent single sand ridge or slip face parallel to the cliff face above.

The most spatially extensive climbing dunes are found in a series of dark-toned bedforms draped over the preexisting topography of spur-and-gully walls in south Melas Chasma. Images of this area (Figs. 3.4b, 3.5a–3.5b) show a series of distinct slip faces on the uphill side of the dunes, both within gullies and atop spurs. A large dune field (Fig. 3.1b and 3.4b) is located on the floor of the chasma 10 km to the north and 1 km lower in elevation. Because of its proximity to the climbing dunes and its slip face orientations indicating a southward wind direction, this floor dune field is the probable sand source for the climbing dunes (Fig. 3.4b). The THEMIS mosaic shows a gradual increase in thermal inertia upslope, up to a thermal boundary (Fig. 3.5b, black arrows). This trend of increasing thermal inertia with elevation is interpreted as evidence for a mixed sand and dust surface lower in the sequence (Fig. 3.5a, blue) with a gradual decrease in dust and increase in grain-size uphill (Fig. 3.5a, cyan). Inspection shows that this boundary corresponds with the upper slip faces and higher-lying, dark-toned aeolian material. We infer it to mark the contact between the lower thermal inertia sand-sized sediment and the higher albedo (~0.15), higher thermal inertia ($>800 \text{ Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$, units hereafter referred to as “tiu” for thermal inertia units) wall material. A “halo” of dark-toned (possibly finer-grained) material blankets the wall rock above the slip faces, lowering its thermal inertia substantially (i.e., 900 tiu to 350 tiu). Together, these two observations suggest a high level of aeolian activity, with wind driving sand

(preferentially fine sand (100-150 μm) as the easiest entrained particle sizes [Kieffer *et al.*, 1992]) upslope and removing dust.

4.2 Valles Marineris Falling Dunes

Eastern Coprates Chasma (Fig. 3.1c) hosts a dense and varied population of aeolian morphologies, including: barchan, sand sheet, and climbing and falling dunes that are found within craters, canyon floors, reentrant depressions, and gullies several kilometers above the chasma floor [Chojnacki and Moersch, 2009]. The dunes found in depressions and gullies, primarily interpreted as falling dunes (Figs. 3.2d, 3.3c, 3.4c–3.4d), exhibit a range of sizes, shapes, and orientations. All falling dunes discussed here are found on the southern wall of the main chasm and both sides of an isolated ~200 km long eastern wall remnant (Fig. 3.1c; red polygons). Bedforms are not always oriented perpendicular to the local maximum slope, but instead are oriented diagonally across walls, indicating sediment transport in that direction (Fig. 3.3d). These dunes appear to be largely affected by mesoscale topography: a ~3-km-long spur between two dune fields suggests leeward deceleration and consequent deposition (Fig. 3.3d; eastern dune field).

One ~90 km stretch of the southern walls has at least eight wall dunes within neighboring gullies, 2-3 km above the ubiquitous floor dune fields (Fig. 3.1c). Many of these wall dunes have a similar symmetric “lobate” morphology of repetitive, slightly arcuate slip faces, with horns oriented uphill (Figs. 3.4c, 3.5c, 3.5e). On a flat surface, this barchan-like bedform would indicate a wind direction uphill. However, inspection of HiRISE DEMs shows material (inferred sand) at the angle of repose (32° – 34°) and likely slip faces on the downhill sides of the dunes (Figs. 3.5d–3.5e), indicating downward movement as falling dunes. Several falling dunes in the

U.S. Great Basin have strikingly similar arcuate, lobate morphologies and orientations.

Koscielniak [1973] found falling lobate dunes to be caused by dune crest overburden that, along with sand saltation, advances the dune crest faster than the flanks (Fig. 3.6). Additionally, the falling dunes in Figs. 3.5 and 3.6a are clearly restricted to the bottom of the gully and its thermal properties and tone are consistent with local provenance.

5.0 Discussion

The existence of climbing and falling dunes in VM has many broad implications for the VM-regional and martian-global aeolian regime. These intriguing aeolian morphologies are the result of a wind environment largely dictated by the micro- and mesoscale topography and large relief of VM. For example, falling dunes found in Coprates Chasma often have horns pointing upwind (Figs. 3.3c, 3.5c–3.5e, 3.6a). In contrast, the majority of terrestrial and martian crescent-shaped dunes that occur on relatively flat topography (i.e., barchans) have horns pointing downwind. This morphology suggests slip face advancement occurs at a greater rate towards the center of the dune as compared to the flanks, presumably due to a higher velocity and/or more consistent airflow over the center of the dune. Three-dimensional models of airflow down terrestrial wadis support this notion, with higher wind velocity vectors near the center of the wadis and decreased wind flow on the flanks [*Al-Enezi et al.*, 2008], potentially due to airflow separation of the walls.

Micro- and mesoscale topography has a direct consequence on sediment transport direction. In many cases (e.g., Coprates Chasma), the climbing and falling dune slip faces orientations are inconsistent with those of dunes found on the canyon floor [*Hayward et al.*, 2007], suggesting a more complex picture than simple up- or down-canyon winds. Contrasting

with this, south Melas Chasma climbing dunes indicate a strong, southward wind direction during periods of dune activity, which agrees with meteorological model predictions for the region under current conditions [Rafkin and Michaels, 2003].

To date, no dark-toned dunes have been located near the rim of VM, nor on the plateau surrounding the chasma system. The closest major dune field concentration, in Argyre Planitia, is more than 2000 km away [Hayward *et al.*, 2007]. Although minor light-toned transverse aeolian ridges (TARs) are common to adjacent plateaus, and no doubt over time some of this sediment has blown into the rift system, these TARs have inconsistent thermal inertia and albedo with dune material discussed herein [e.g., Figs. 3.5c, 3.6a; Chojnacki and Moersch, 2009]. All the falling dunes discussed here, which are located in east Coprates Chasma, appear isolated from extra-rift sand sources. The plateau above the south wall dunes (bottom left Fig. 3.1c) show a barrier (cantena) to sediment transport and currently lacks visible aeolian features. Climbing and falling dunes, on Earth and Mars, have been found to be part of larger aeolian sediment pathways across troughs or mountains [Lancaster and Tchakerian, 1996; Bourke *et al.*, 2006]. If the Mars dune sediment source were the plateau regions outside of VM, we would expect to find falling dunes proximal to chasma rims (i.e., on the upper half of canyon walls). Rather, falling dunes on lower reaches of canyon walls that have eroding outcrops above them and floor dunes below them (Figs. 3.1c, 3.5c–3.5e, 3.6a) provide evidence that dune material is locally derived. This dune provenance may mark a distinction between terrestrial and martian examples of falling dunes, as terrestrial falling dunes have not been reported to form *in situ*.

References

- Al-Enezi, A., K. Pye, R. Misak, and S. Al-Hajra (2008), Morphologic characteristics and development of falling dunes, northeast Kuwait, *Journal of Arid Environments*, 72, 423–439.
- Bourke, M. C., J. E. Bullard, and O. S. Barnouin-Jha (2006), Aeolian sediment transport pathways and aerodynamics at troughs on Mars, *J. Geophys. Res.*, 109, E07005, doi:10.1029/2003JE002155.
- Chojnacki, M., and Jeffrey E. Moersch (2009), Valles Marineris Dune Fields: Thermophysical Properties, Morphology, and Provenance, *Lunar Planet. Sci.*, XL, #2486.
- Christensen, P. R., et al. (2004), The Thermal Emission Imaging System (THEMIS) for the Mars 2001 Odyssey mission, *Space Sci. Rev.*, 110, 85–130, doi:10.1023/B:SPAC.0000021008.16305.94.
- Evan, R. (1962), Falling and Climbing Sand Dunes in the Cronese “Cat” Mountain Area, San Bernardino County, California, *J. Geol.*, 70, pp. 107–113.
- Fenton, L. K., J. L. Bandfield, and A. W. Ward (2003), Aeolian processes in Proctor Crater on Mars: Sedimentary history as analyzed from multiple data sets, *J. Geophys. Res.*, 108(E12), 5129, doi:10.1029/2002JE002015.
- Ferguson, R. L., P. R. Christensen, and H. H. Kieffer (2006), High-resolution thermal inertia derived from the Thermal Emission Imaging System (THEMIS): Thermal model and applications, *J. Geophys. Res.*, 111, E12004, doi:10.1029/2006JE002735.
- Gendrin, A., et al. (2005), Sulfates in Martian layered terrains: The OMEGA/Mars Express view, *Science*, 307, 1587–1591.
- Gorelick, N. S. and P. R. Christensen (2005), THEMIS Global Mosaics, *Eos Trans. AGU*, 86(52), Fall Meet. Suppl., Abstract P21C-0161.
- Hayward, R. K., et al. (2007), Mars Global Digital Dune Database and initial science results, *J. Geophys. Res.*, 112, E11007, doi:10.1029/2007JE002943.
- Herkenhoff, K. E., et al. (2008), In situ observations of the physical properties of the Martian surface, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell, pp. 451–467, Cambridge Univ. Press, New York, doi:10.1017/CBO9780511536076.021.
- Kieffer, H. H., and A. P. Zent (1992), Quasi-periodic climatic change on Mars, in *Mars*, edited by H. H. Kieffer et al., pp. 1180–1218, Univ. of Ariz. Press, Tucson.
- Kirk, R. L., et al. (2008), Ultrahigh resolution topographic mapping of Mars with MRO HiRISE stereo images: Meter-scale slopes of candidate Phoenix landing sites, *J. Geophys. Res.*, 113, E002.24, doi:10.1029/2007JE003000.
- Koscielniak, D. E. (1973), Eolian deposits on a volcanic terrain near Saint Anthony, Idaho, Master of Arts Thesis, State University of New York at Buffalo.
- Lancaster, N. and Tchakerian, V.P. (1996), Geomorphology and sediment of sand ramps in the Mojave Desert, *Geomorphology*, 17: 151, 165.
- Liu, X., S. Li and J. Shen (1999), Wind tunnel simulation of mountain dunes, *J. Arid Envir.*, 42, 49–59.
- Malin, M. C., et al. (2007), Context Camera Investigation on board the Mars Reconnaissance Orbiter, *J. Geophys. Res.*, 112, E05S04, doi:10.1029/2006JE002808.
- McEwen, A. S., et al. (2007), Mars Reconnaissance Orbiter's High Resolution Imaging Science Experiment (HiRISE), *J. Geophys. Res.*, 112, E05S02, doi:10.1029/2005JE002605.

- Neukum G., R., et al. (2004), HRSC: the High Resolution Stereo Camera of Mars Express, ESA SP-1240, 17-35.
- Piatek J. L., and J. E. Moersch (2006), A Strategy for Atmospheric Correction of THEMIS Infrared Data, *Lunar Planet. Sci.*, XXXVII, #1158.
- Putzig, N. E., and M. T. Mellon (2007), Apparent thermal inertia and the surface heterogeneity of Mars, *Icarus*, 191, 68-94.
- Pye, K., and H. Tsoar (1990), *Aeolian Sand and Sand Dunes*, Unwin Hyman, London, 396 pp.
- Quantin, C., P. Allemand, and C. Delacourt (2004), Morphology and geometry of Valles Marineris Landslides, *Planet. Space Sci.*, 52, 1011–1022.
- Rafkin, S. C. R., and T. I. Michaels (2003), Meteorological predictions for 2003 Mars Exploration Rover high-priority landing sites, *J. Geophys. Res.*, 108, doi:10.129/20002JE0002027.
- Tsoar, H. (1983), Wind tunnel modeling of echo and climbing dunes, In: M.E. Brookfield and T.S. Ahlbrandt (Editors), *Eolian Sediments and Processes*, *Developments in Sedimentology*, Elsevier, Amsterdam, pp. 247-259. Appendix III

Appendix III

Figure 3.1 THEMIS daytime infrared of Valles Marineris dune field (orange) distribution. Mosaics of (a) Valles Marineris, (b) southern Melas Chasma, (c) and central Coprates Chasma with climbing and falling dunes shown in red and indicated by red arrows. Context boxes for (b) and (c) are in white. Context boxes for Figs. 3.3–3.5 are also shown in panels (a-c), and black arrows represent viewing angles for Fig. 3.4. THEMIS image credit: *Gorelick and Christiansen* [2005].

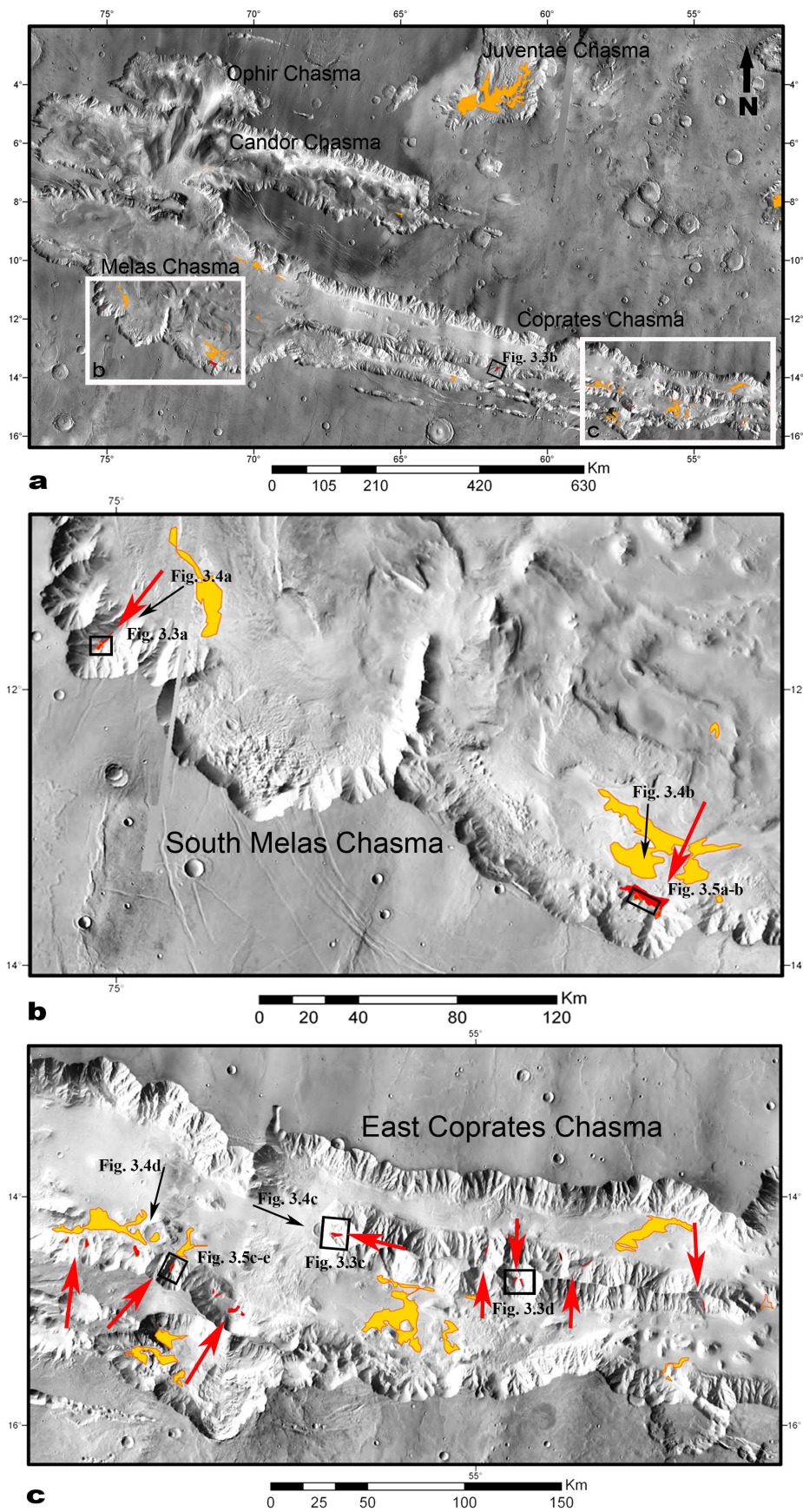


Figure 3.1

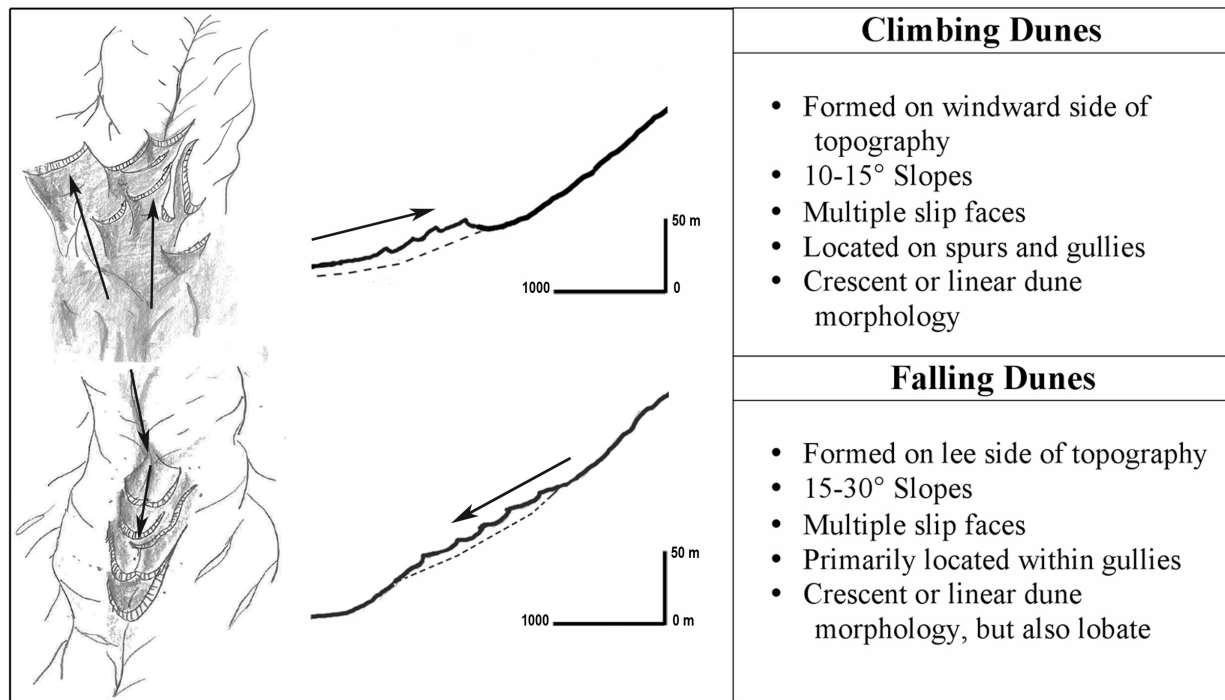


Figure 3.2 A schematic representation of climbing and falling dunes as seen from overhead and in cross-section, with characteristic properties listed. Cross sections are interpretive from terrestrial climbing and falling dune profiles and are largely based on *Lancaster and Tchakerian* [1996]. The dashed line distinguishes bedrock from aeolian sediment. Black arrows indicate inferred predominant wind direction/direction of sediment transport.

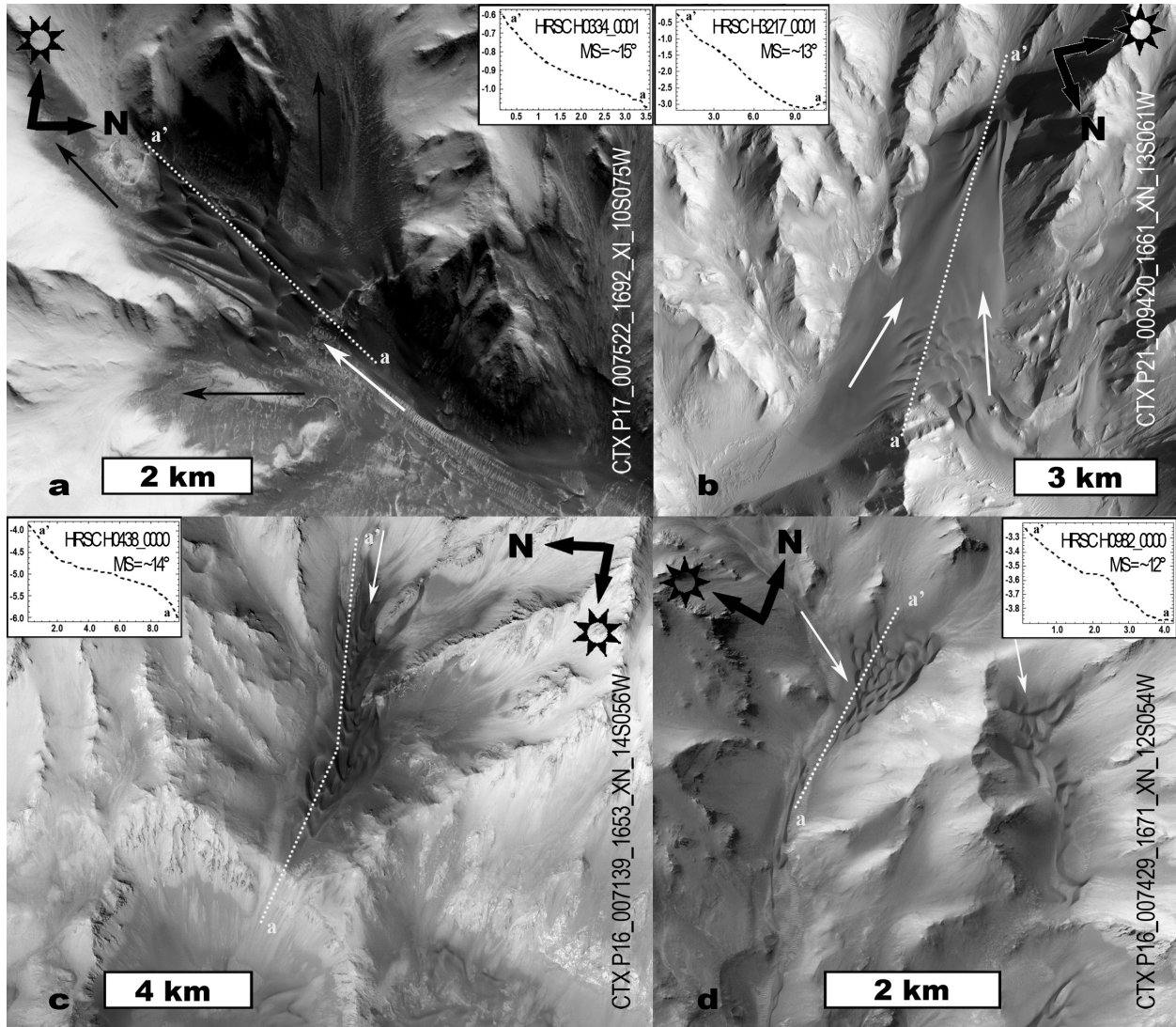


Figure 3.3 CTX images of Valles Marineris climbing and falling dunes.

Inferred wind directions are shown by white arrows. Images are oriented so lower elevations are toward the bottom of the images. HRSC DEM profiles (a to a') are provided with mean slope values (MS) and axes in units of kilometers. (a) Climbing dunes in southwest Melas Chasma. Notice multiple slip faces and dark-toned sediment superimposed on the higher albedo bedrock (black arrows). (b) Climbing dunes in central Coprates Chasma. Note the prominent slip face approximately parallel to the obstructing wall rock above. (c-d) Falling dunes in Coprates Chasma.

Figure 3.4 Oblique perspective views of Valles Marineris climbing and falling dunes. CTX images using HRSC elevation, where vertical exaggeration is 2–3, viewing orientations are shown in Fig. 3.1, and scale bars are provided in the nadir version of the image. Images and DEMs are the same as Fig. 3.3 except where noted. (a-b) Southward views of Melas Chasma climbing dunes seen in Fig. 3.3a and 3.2b. (c) Eastward view of falling dunes shown in Fig. 3.3c. (d) A southward view of CTX image (P12_005570_1650_XN_15S057W) on HRSC elevation (H1918_0000) showing falling dunes in multiple gullies.

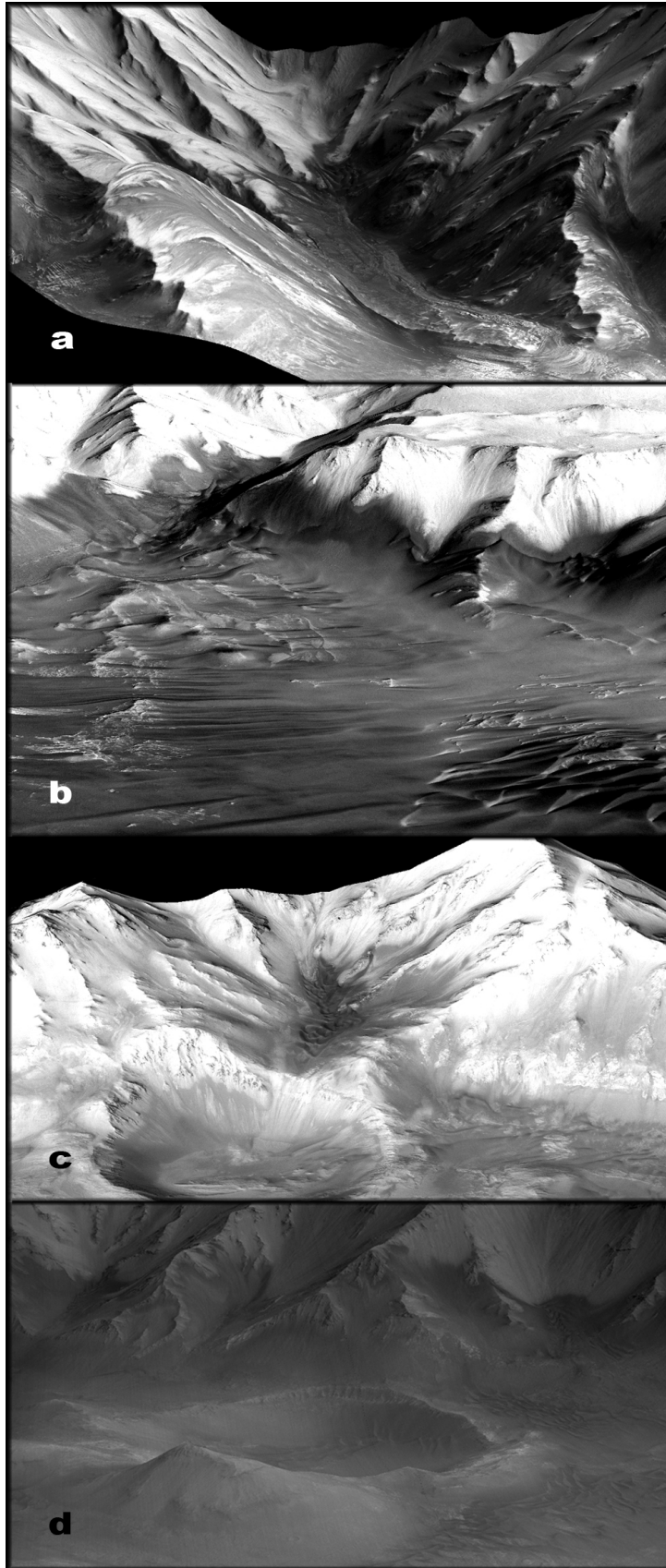


Figure 3.4

Figure 3.5 Visible-wavelength images, thermal inertia, and topography of Melas Chasma climbing dunes and Coprates Chasma falling dunes.

Multiple climbing dunes in southern Melas Chasma shown in (a) THEMIS thermal inertia and (b) a CTX image shows increasing thermal inertia southward and upslope to a distinct thermal boundary (black arrows), above which high thermal inertia wall rock is located. In visible-wavelength images (b) this same location marks the contact between the furthest extent of the slip faces and dark-toned sediment higher on the slope. A HRSC DEM profile (a to a') is provided with a mean slope value (MS) and axes in units of kilometers. (c-e) A series of falling dunes in southern Coprates Chasma shown in (c) a CTX image colorized with THEMIS thermal inertia data and (d-e) shows a HiRISE image. In (d) a distinct, arcuate lobate morphology common to other local wall dunes is observed. HiRISE DEM elevation (top) in meters and slope angle (bottom) profiles (a to a') are provided across the bottom three dunes. Closer examination (e), confirms slip faces are located on the downhill and flank sides of the dunes.

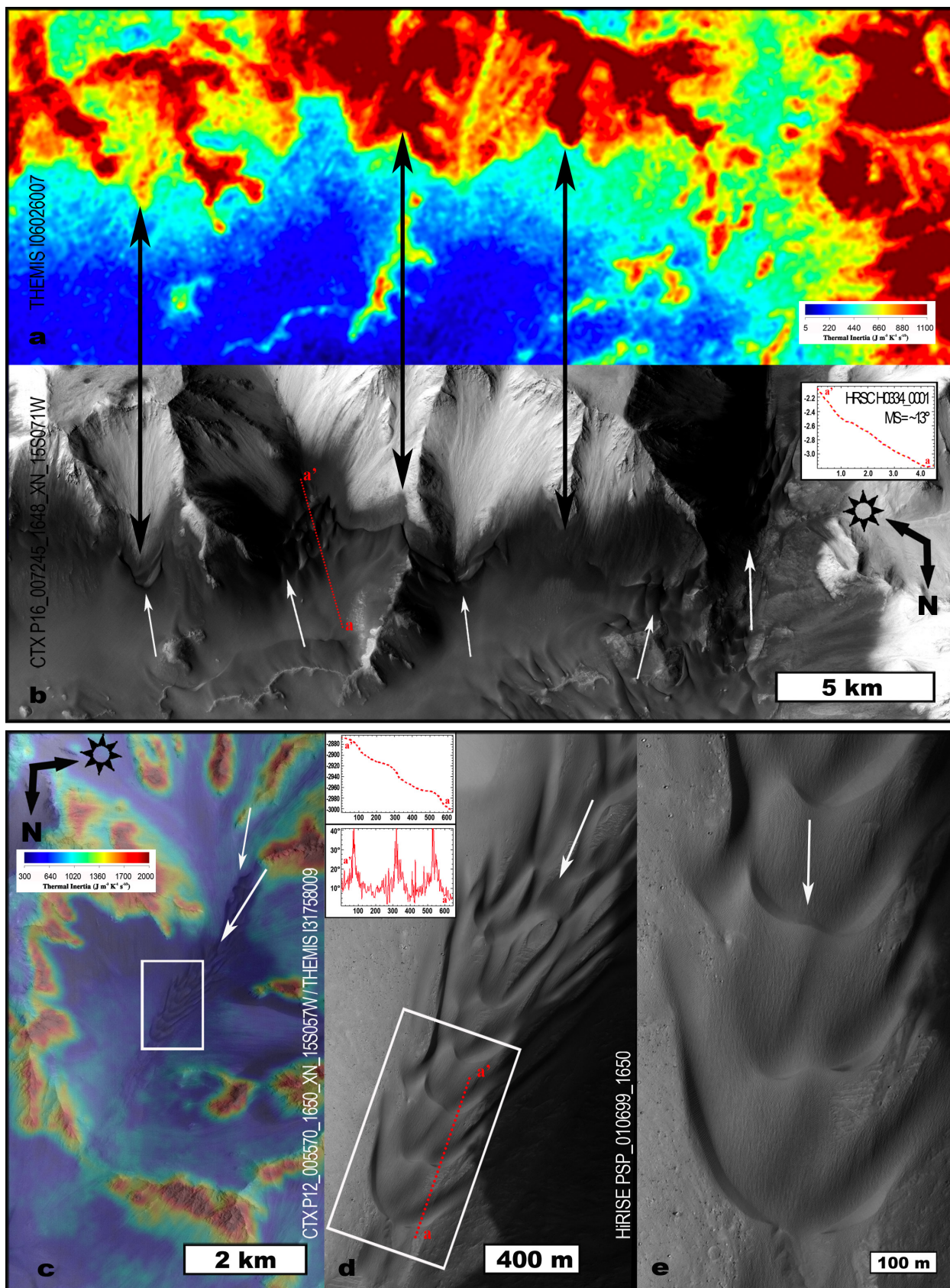


Figure 3.5

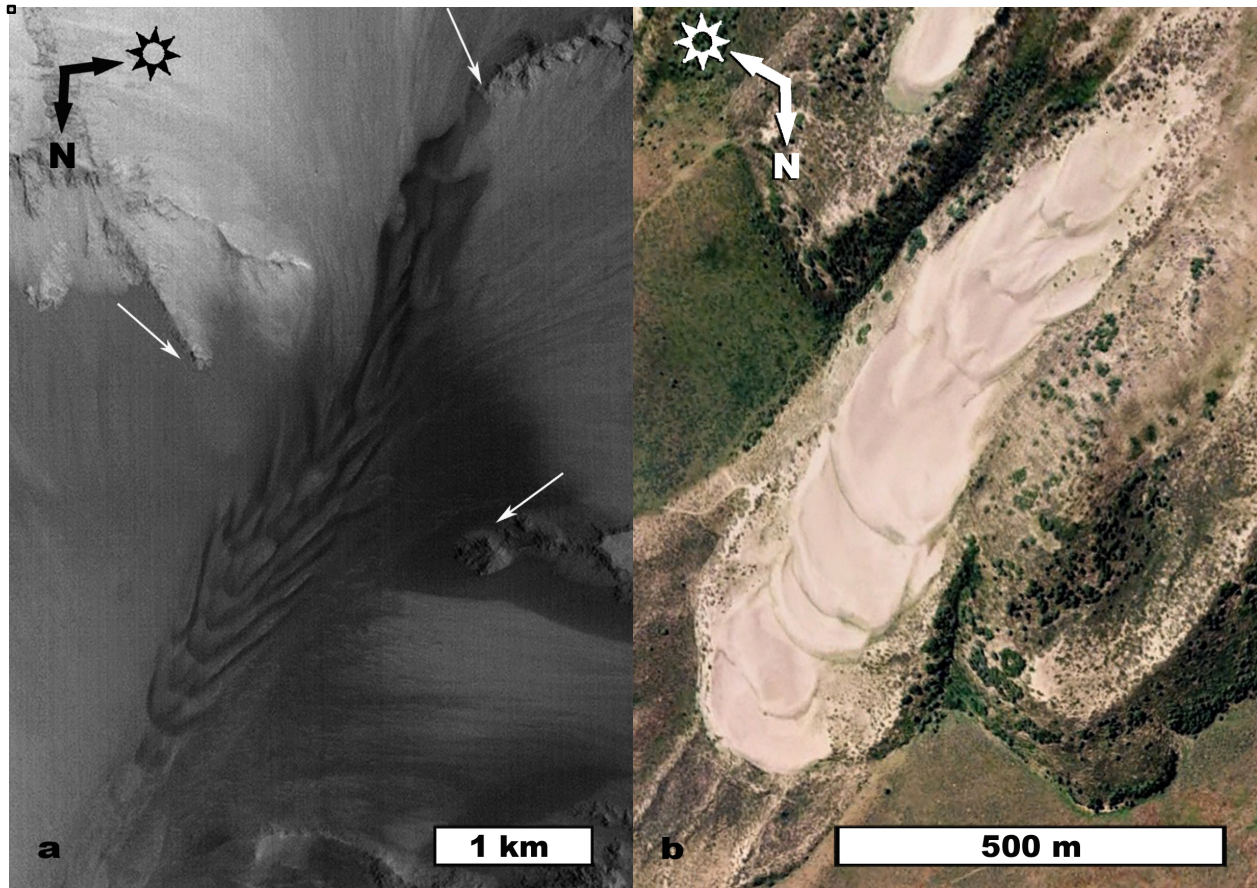


Figure 3.6 A comparison of martian (a) and Terrestrial (b) falling dune morphologies. Down slope and the transport direction is towards the bottom of the figure. (a) CTX image P12_005570_1650_XN_15S057W shows the same Coprates Chasma falling dunes found in Fig. 3.5c-e. White arrows indicate putative sand sources. (b) Falling dunes of Juniper Buttes, Idaho near 44.00 N 111.90 W. Image credit: Google Earth.

**CHAPTER IV: VALLES MARINERIS DUNE FIELDS AS COMPARED
WITH OTHER MARTIAN POPULATIONS: DIVERSITY OF DUNE
COMPOSITIONS, MORPHOLOGIES, AND THERMOPHYSICAL
PROPERTIES**

Abstract

Planetary dune field properties and their bulk bedform morphologies relate to regional wind patterns, sediment supply, climate, and topography. On Mars, major occurrences of spatially contiguous low-albedo sand dunes are primarily found in three major topographic settings: impact craters, high-latitude basins, and linear troughs or valleys, the largest being the Valles Marineris (VM) rift system. As one of the primary present day martian sediment sinks, VM holds nearly a third of the non-polar dune area on Mars. Moreover, VM differs from other regions due to its unusual geologic, topographic, and atmospheric setting. Herein, we test the overarching hypothesis that VM dune fields are compositionally, morphologically, and thermophysically distinct from other low- and mid-latitude (50°N–50°S latitude) dune fields. Topographic measurements of dune fields and their underlying terrains indicate slopes, roughnesses, and reliefs to be notably greater for those in VM. Variable VM dune morphologies are shown with topographically-related duneforms (climbing, falling, and echo dunes) located among spur-and-gully wall, landslide, and chaotic terrains, contrasting most martian dunes found in more topographically benign locations (e.g., craters, basins). VM dune fields superposed on Late Amazonian landslides are constrained to have formed and/or migrated over >10s of kilometers in the last 50 My to 1 Gy. Diversity of detected dune sand compositions, including unaltered ultramafic minerals and glasses (e.g., high and low-calcium pyroxene, olivine, Fe-bearing glass), and alteration products (hydrated sulfates, weathered Fe-bearing glass), is more pronounced in VM. Observations show heterogeneous sand compositions exist at the regional-, basinal-, dune field-, and dune-scales. Although not substantially greater than elsewhere, unambiguous evidence for recent dune activity in VM is indicated from pairs of high-resolution images that include: dune deflation, dune migration, slip face modification (e.g., alcoves), and ripple modification or migration, at varying scales (10s–100s m²). We conclude that VM dune fields are qualitatively and quantitatively distinct from other low- and mid-latitude dune fields, most readily attributable to the rift's unusual setting. Moreover, results imply dune field properties and aeolian processes on Mars can be largely influenced by regional environment, which may have their own distinctive set of boundary conditions, rather than a globally homogenous collection of aeolian sediment and bedforms.

1.0 Introduction and Motivation

Sand dunes on Mars offer an opportunity to investigate the nature of eroded sediments and their interactions with the atmosphere. As valuable recorders of atmosphere-surface interactions, planetary dune properties relate to differences in wind regime, sediment supply, the presence or absence of topography, and climate [Bagnold, 1941; Pye and Tsoar, 1990; Kocurek and Lancaster, 1999; Bourke *et al.*, 2010].

Although aeolian dune-forming sediments are widespread across Mars (Fig. 4.1b), they have the greatest areal extent as coherent fields of dunes in three distinct topographic settings: impact craters, high-latitude basins, and linear troughs or valleys, the largest being the Valles Marineris rift system (Fig. 4.1b). Craters are the most geographically wide-spread locations for dune fields [Greeley *et al.*, 1992]. Indeed, the majority ($\sim 78\%$ or $\sim 50,000 \text{ km}^2$) of low- to mid-latitude (65°N to 65°S) dune fields mapped in a near global study are found in impact craters [Hayward *et al.*, 2007a, 2007b]. The most extensive dune fields occur within the extremely flat circum-polar polar basins, as nearly continuous ergs or sand seas (e.g., Olympia Undae) [Lancaster and Greeley, 1990; Hayward *et al.*, 2010b]. The third common topographic setting for dune fields, linear troughs or valleys, includes the great structural rift system of Valles Marineris (VM). As one of the present day primary martian sediment sinks, VM holds nearly a third of the non-polar dune area ($\sim 22,000 \text{ km}^2$) on Mars [Chojnacki *et al.*, 2012a]. The rift contains large landslides [e.g., Lucchitta *et al.*, 1992], layered-deposits partially composed of water-altered minerals [Christensen *et al.*, 2001a; Gendrin *et al.*, 2005] and aeolian bedforms [Chojnacki and Moersch, 2009]. Early studies of VM interpreted dark mantles found on the chasm floors as extensive dune fields that were likely locally derived [Lucchitta *et al.*, 1992]. Our initial studies have described a diverse VM dune population with evidence of unusual dune

morphologies, contemporary activity, and locally-sourced dune sediment [*Chojnacki and Moersch, 2009; Chojnacki et al., 2010, 2011b, 2012a, 2012b*] that motivates more investigation.

Multiple factors suggest that VM dune fields are different from other terrestrial and martian dune fields. The atmospheric pressure at the canyon bottom (>8 mb), which is higher than at many locations on Mars, would result in a lower threshold friction velocity [*Greeley and Iversen, 1985*], making wind a potentially more effective geomorphic agent in VM [*Chojnacki et al., 2006*]. Variable chasmata widths and orientations may also play a role in the topographic focusing of winds and contrast with the lower relief of intracrater topography. Previous atmospheric modeling found substantially greater near-surface modeled winds in VM as compared with more topographically subdued locations [*Rafkin and Michaels, 2003; Toigo and Richardson, 2003*]. Finally, the variety of possible sediment sources found within VM (e.g., interior layered deposits (ILDs), spur-and-gully walls, landslides, mantling units) may contribute to a greater degree of dune sediment composition heterogeneity [*Chojnacki et al., 2012a, 2012b*].

Based on these observations and previous work in the region, we hypothesize that VM dune fields are compositionally, morphologically, and thermophysically distinct from other low- and mid-latitude dune fields due to their unusual geologic, topographic, and atmospheric setting. Testing this hypothesis not only yields a direct answer to the question “are VM dunes different from other low- and mid-latitude dunes?” but also provides insight into the importance of topography, geologic context, and possible source lithology for dune formation and distribution. We limited our survey of non-VM dune fields to the area between 50°N – 50°S latitude in order to exclude occurrences of high latitudinal processes (e.g., surface CO_2 ice, ice-dust mantles) that likely do not affect equatorial dunes during periods of low to moderate obliquities. This choice

is supported by the work of *Fenton and Hayward* [2009], who found increasing degradation of dune morphology and a decrease in inferred dune activity poleward of $\sim 50^{\circ}\text{S}$.

The central questions related to our hypothesis are:

1. Is there a geographic and/or genetic relationship between VM dune fields and spur-and-gully wall material, landslides, or interior layered deposits? Do these geologic features, not typically associated with craters, lead to differences in dune populations?
2. Does regional and/or local topography influence the morphology or geographic location of dune fields? Is there a similar correlation in VM with dune slip faces and chasmata orientation or ILD location?
3. Are particle sizes of VM dunes larger than those found in dunes elsewhere on Mars, perhaps because of the lower elevation (and the associated higher atmospheric pressure) and/or topographically-enhanced wind speeds?
4. Low-albedo soil measurements at landing sites appear globally homogeneous with respect to elemental and mineralogical composition [*Yen et al.*, 2005, 2013; *Brückner et al.*, 2008; *McSween et al.*, 2009, 2010]. Is there also compositional homogeneity in dune sands at low and mid-latitudes, which have a greater potential than soil for mobility and therefore for aeolian fractionation?
5. When did the sand dunes form? Is dune formation an ancient or geologically recent process?
6. If there is evidence for contemporary dune activity in VM, do the unique characteristics of VM's setting (e.g., topographic and albedo variations) produce greater activity than found in non-VM dune field locations?

We evaluate the hypothesis above by examining the morphology, composition, and thermophysical properties of Valles Marineris dunes in comparison with those of low- and mid-latitude martian dune field populations. After first giving background information on Martian dunes, we provide a summary of the data sets and methodology used herein. Next we present a survey of low- and mid-latitude non-VM dune fields and of VM dune fields investigated in terms of their contextual classification, topography, thermophysical properties, composition, and evidence for contemporary activity. As part of this overview, we investigate the degree of correlation between dune emplacement directions as indicated by dune morphology and wind directions as given by atmospheric modeling results. We then discuss these findings in the context of the hypothesis and associated questions, reflecting on the greater implications for other martian dune fields and questions about the scale and timing of aeolian processes.

2.0 Background: Dune Field Properties

The physical nature of martian aeolian sediments can be investigated using remote thermal infrared and/or *in situ* measurements. Estimated grain sizes derived from thermal inertia data of aeolian bedforms range from very fine to coarse sand ($\sim 100\text{--}1000\ \mu\text{m}$) [Fenton *et al.*, 2003; Fergason *et al.*, 2006a, 2006b; Chojnacki *et al.*, 2011a]. Many of these derived sand sizes are generally coarser than those of terrestrial dune sand, which is usually fine to medium sand ($125\text{--}250\ \mu\text{m}$) [Pye and Tsoar, 1990]. *In situ* results from the Spirit rover have identified poorly-sorted, fine to medium sand as the major component of Gusev crater ripples, which are likely to have formed primarily by saltation [Sullivan *et al.*, 2008; McGlynn *et al.*, 2012]. Observations of cloddy soils and sand dunes with elevated thermal inertia values, consistent with very-coarse sand ($>1\ \text{mm}$), leave open the possibility of partially consolidated sand surfaces

formed by some surficial induration process [Sullivan *et al.*, 2008; Chojnacki and Moersch, 2009; Chojnacki *et al.*, 2010].

In the last decade, a greater understating of Mars dune composition has developed. Studies using data from the Thermal Emission Spectrometer (TES) [Christensen *et al.*, 1992] onboard the Mars Global Surveyor (MGS) have revealed that low-albedo dunes have a primarily basaltic composition (e.g., plagioclase, pyroxene, olivine) with some occurrences of more intermediate compositions [Fenton *et al.*, 2003; Stockstill-Cahill *et al.*, 2008; Aben, 2003]. Studies using data from the Mars Express (MEx) mission's Observatoire pour la Minéralogie, l'Eau, les Glaces, et l'Activité (OMEGA) experiment [Bibring *et al.*, 2004] found evidence of ultramafic minerals associated with dark-toned duneforms in wide-spread Noachian-aged mid-latitude craters [Mustard *et al.*, 2005; Bibring and Langevin, 2008; Poulet *et al.*, 2008a; Tirsch *et al.*, 2011]. In contrast to the dominantly mafic composition of most mid-latitude dune fields, the high-latitude northern polar basins contain geographically contiguous sand seas partially composed of alteration products (e.g., gypsum, weathered glass) [Langevin, 2005; Horgan *et al.*, 2009; Horgan and Bell, 2012a]. A clear pattern of dune field compositions, extent of aeolian bedform alteration, or degree of sand compositional heterogeneity has yet to emerge.

Previous authors have correlated atmospheric modeling, both at the global- (e.g., NASA Ames Global Circulation Model (GCM) [Haberle *et al.*, 1993]) and mesoscales (e.g., Mars Regional Atmospheric Modeling System (MRAMS) [Rafkin *et al.*, 2001]), with dune slip face orientations and/or surface changes with varied success [Fenton *et al.*, 2005; Hayward *et al.*, 2007a, 2009; Chojnacki *et al.*, 2011a; Gardin *et al.*, 2012]. Certain locations in impact craters are thought to have zones of weak atmospheric flow, favoring dune field stability [Hayward *et al.*, 2007a, 2009]. Based on this observation, Hayward *et al.* [2007a] devised a parameter called

the dune centroid azimuth or DCA, which is defined as a geodesic line that extends from the crater's center toward the dune field's centroid. Those authors found correlations between DCAs and both slip face orientations and downwind directions derived from atmospheric models for small diameter (<25 km) smooth-floored craters in the southern latitudes (~65% agreement within 45°). This finding suggests that DCAs in topographically simple small craters are good proxies for long-term regional sediment movement and wind pattern stability. More complicated topographic settings (e.g., troughs, mountains, complex craters) influence atmospheric flow in less direct ways [Bourke *et al.*, 2004; Chojnacki *et al.*, 2010; Cardinale *et al.*, 2012; Silvestro *et al.*, 2012a] and may contribute to enhanced activity or the construction of geospatially and/or morphologically diverse dune populations.

Along with other important parameters influencing planetary dune properties (i.e., wind regime, sediment supply, topography), changes in climate can profoundly affect aeolian systems [Pye and Tsoar, 1990]. Although some martian dune fields may have been the products of an ancient climate [Edgett and Malin, 2000a; Fenton and Hayward, 2010], evidence is mounting that some bedforms have been active in the recent past. Studies using Mariner 9 and Viking orbiter observations showed that aeolian deposits are common on Mars and possibly active, with albedo variations occurring over the lifetimes of the individual missions [Sagan *et al.*, 1973; Greeley *et al.*, 1992]. The steady improvements in spatial resolution, spatial coverage, and temporal baseline of the data have led to a growing number of detections of contemporary bedform movement. The Mars Exploration Rovers (MER) have detected a number of wind events via observations of ripple migration, dust-cleaning events of the solar arrays, rover track modification, and dark streak activity [Greeley *et al.*, 2006; Geissler *et al.*, 2008; Sullivan *et al.*, 2008]. Orbital observations have been used to document unambiguous bedform modification,

and in some cases dune deflation and migration [Bourke *et al.*, 2008; Silvestro *et al.*, 2010b; Chojnacki *et al.*, 2011a; Geissler *et al.*, 2011; Bridges *et al.*, 2012b]. However, questions remain regarding the spatial extent, variety of topographic environment, and relations to other factors (i.e., elevation, albedo variability) in which this dune activity is taking place.

Although this study is focused on dark-toned dunes, a hierarchy of martian aeolian bedforms is recognized herein, analogous to the hierarchy of Wilson [1972] for terrestrial bedforms, consisting of four components (in order of increasing wavelengths): ripples, transverse aeolian ridges (TARs), dunes, and draas. Ripples are the smallest bedforms (<20 m) and can only be observed in high-resolution images, superposed on many surfaces, but primarily atop dunes or smaller sand patches [Bridges *et al.*, 2007; Zimbelman, 2010]. TAR is a general term for martian bedforms (wavelengths ~20–70 m) that may have formed as the result of either creep or saltation, the primary formation mechanisms for ripples or dunes, respectively [Bourke *et al.*, 2003; Zimbelman, 2010]. TARs are found without slip faces, and are often light in tone relative to their surroundings, symmetric in profile, and usually overridden by larger dark-toned dunes [Zimbelman, 2010]. Dunes, as defined above, are intermediate bedforms between TARs and draas with respect to wavelength. In terrestrial literature, the draa is the highest-order bedform with largest wavelengths (>1 km), and it includes lower-order superposed bedforms, such as dunes [Wilson, 1972, 1973]. All orders of aeolian bedforms have been observed in the study area (see below).

3.0 Data Sets and Methods

Detailed geomorphic, spectroscopic, and thermophysical mapping of VM and non-VM dune fields was performed to facilitate an understanding of their geologic history. The study area for VM dune fields was performed within the rift which is broadly encompassed by 0–18°S,

270–320°E longitude (Fig. 4.1b). The study area for the non-VM dune fields includes 50°N–50°S latitude, all longitudes, excluding those defined as VM dunes (Fig. 4.1b).

3.1 Morphology: Visible-Wavelength Images

To identify dune morphologies and provide geologic context, this study utilized four different visible-wavelength image data sets. The Mars Orbiter Camera (MOC) onboard MGS provided both narrow-angle (MOC NA) images (~3 m/pixel) and wide-angle images (~232 m/pixel) [Malin *et al.*, 1992]. The Context Camera (CTX) [Malin *et al.*, 2007] at 5–6 m/pixel and the High Resolution Imaging Science Experiment (HiRISE) at ~25 cm/pixel [McEwen *et al.*, 2007], both onboard the Mars Reconnaissance Orbiter (MRO), provide more recent image data. These visible-wavelength images were processed using U.S. Geological Survey (USGS) Integrated Software for Imagers and Spectrometers (ISIS) software [Gaddis, 1997; Anderson *et al.*, 2004] for radiometric calibration. Images were spatially registered to each other, and to other data sets, using ArcGIS and ENVI software.

3.2 Topography, Gradient, and Roughness: Digital Elevation Models

To quantitatively determine absolute elevation, investigate dune field topography, and estimate underlying slopes, digital elevation models (DEMs) were used. The Mars Orbiter Laser Altimeter (MOLA) [Smith *et al.*, 2001] instrument onboard MGS provided global gridded maps (128 pixels per degree) of elevations and slopes (0.5 km baseline). To quantitatively investigate relative surface relief and slopes at 50–200 m/pixel, Mars Express’s High Resolution Stereo Camera (HRSC) DEMs [Neukum *et al.*, 2004] were obtained from the HRSC data explorer at <http://hrscview.fu-berlin.de> and then reprojected and reformatted using ISIS software. At these higher resolutions, sub-kilometer topography of landforms (e.g., dune slip faces) can be analyzed.

To investigate dune field surface roughness, maps of MOLA roughness, derived from shot-to-shot topographic profiles (8 pixels per degree), were used at three baseline lengths: 0.6-km, 2.4-km, and 9.2-km (see description of these data sets by M. Kreslavsky at <http://www.planetary.brown.edu/planetary/rough/-TechInfo>). This data set is similar to previous roughness maps of Mars [Kreslavsky and Head, 1999, 2000, 2002], but with minor modifications. This parameter is a measure of roughness in dimensions of inverse length (e.g., m^{-1}). Because of the coarse scale of these maps, only 49 dune fields were large enough ($>200 \text{ km}^2$) and with sufficient spatial extent (2–3 pixels in the along-track direction) to be incorporated into this study.

3.3 Geologic Context: Geologic Maps

We determined the correlation of dune fields with geologic units of different estimated terrain ages (Noachian, Hesperian, Amazonian) whose proportion of coverage in our study area (equatorward of 50°) is similar [Scott and Tanaka, 1986; Greeley and Guest, 1987]. Dune field locations were overlaid on geologic maps digitized in ArcGIS by Skinner *et al.* [2006] at the 1:15,000,000-scale. The estimated ages of geologic units likely do not equate to dune field ages (either in start of activity or production of sediment), but set an upper bound for dune field ages.

3.4 Thermophysical Properties: Albedo and Thermal Inertia

We used data from the TES instrument [Christensen *et al.*, 1992] for estimating dune field albedos and thermal inertias. Three Mars Years (MY) of TES bolometric albedo measurements (MY 24, MY 25, and MY 26) were used, as calculated and resampled to $\sim 3 \text{ km/pixel}$ by Putzig and Mellon [2007a].

Also modeled by those authors, TES bolometric thermal inertia, at the same resolution, was used. Thermal inertia is the measure of a material's resistance to temperature change, and is

defined as the square root of the product of thermal conductivity (k), bulk density (ρ), and heat capacity (C):

$$I \equiv (k\rho C)^{1/2} \quad [\text{J m}^{-2} \text{K}^{-1} \text{s}^{-1/2}] \quad (1)$$

Thermal inertia is the dominant surface property controlling diurnal and seasonal temperature variations and yields insight into the physical properties (e.g., particle size, degree of induration) of a surface [Mellon *et al.*, 2000; Putzig and Mellon, 2007a]. For materials of the same composition, thermal inertia increases with grain size, denser packing, greater induration, and/or reduced pore space; therefore, interpretations of thermal inertia values are generally non-unique [e.g., Christensen, 1986]. Global TES-derived bolometric thermal inertia (modeled from both dayside and nightside data) ranges from 5–5000 $\text{Jm}^{-2}\text{K}^{-1}\text{s}^{-1/2}$ (units hereafter referred to as “tiu” for thermal inertia units) as modeled by Putzig and Mellon [2007a]. For our analyses using TES thermal inertia, pixels of relatively homogenous, bedrock-free dune area (as viewed in co-registered THEMIS and CTX images) were collected for large dune fields. Without this precaution, thermal inertia values extracted from the bulk dune field polygons (see section 3.7) were often inflated (>400 tiu) due to bedrock contamination and did not reflect the physical properties of dune surfaces. This method excluded smaller fields and those containing well-spaced barchans.

Modeled thermal inertia values from dayside measurements are generally lower than those derived from nightside data for most non-polar surfaces [Mellon *et al.*, 2000; Putzig and Mellon, 2007a, 2007b]. In the case of an idealized, non-layered, homogenous surface, dayside-derived and nightside-derived thermal inertia should approximately match [Mellon *et al.*, 2008]. Discrepancies in modeled thermal inertia are related to the fine components within the pixels dominating the signal in the day as compared with the rock or coarser grained components

dominating at night [Jakosky, 1986]. In addition, dayside-derived thermal inertia results are more prone to albedo and slope errors [Mellon *et al.*, 2000; Putzig and Mellon, 2007b]. However, differences between day and night modeled thermal inertia can be useful in the qualitative interpretation of surfaces composed of mixed particle sizes (e.g., dust, sand, cobbles) in addition to dune sand [Putzig and Mellon, 2007a, 2007b]. To explore diurnal variability in modeled thermophysical properties, we subtracted TES dayside-derived thermal inertia from nightside results to produce delta thermal inertia (ΔTI) measurements. Given the absence of large slopes or surface ice, the larger (positive) ΔTI value derived for a surface, the more substantial the intra-pixel particle size difference and the greater degree of thermophysical heterogeneity. ΔTI values within 25 tiu of zero are within the empirically-derived model uncertainty (for surfaces with thermal inertias ~ 280 tiu) [Putzig and Mellon, 2007a] and are considered null here.

We also derived thermal inertia maps (~ 100 m/pixel) from Mars Odyssey's Thermal Emission Imaging System (THEMIS) infrared images [Christensen *et al.*, 2004] using the thermal model of Putzig and Mellon [2007a], as implemented in the "jENVi" software suite at <http://arsia.gg.utk.edu/~utmars/jenvi/>. THEMIS nightside images were selected based on a qualitative 1–7 quality rating scale implemented by the THEMIS team and by comparing with dayside THEMIS observation (within a 15° by 15° area) of infrared dust opacity (both rating and dust opacity values available on Java Mission-planning and Analysis for Remote Sensing (JMARS) [Gorelick *et al.*, 2003]), in order to estimate visible dust opacity values and minimize atmospheric effects. To derive best-fit thermal inertia values, THEMIS predawn band 9 ($12.57 \mu\text{m}$) brightness temperature images were used in coordination with a seven-dimensional lookup table using thermal inertia, solar longitude, time of day, latitude, elevation, albedo, and visible

dust opacity. As an additional quality check, TES- and THEMIS-derived values of thermal inertia were compared for qualitatively homogenous areas (usually 3×3 TES pixels or greater). THEMIS images that produced thermal inertia values inconsistent with TES nightside bolometric values (>20% difference) were rejected or remodeled with a different dust opacity estimate to provide a better fit.

The thermal inertia values of unconsolidated surfaces are largely controlled by particle size. For a mono-sized distribution of particles, the effective particle size is equivalent to the mean particle size [Presley and Christensen, 1997]. In nature, particle size mixing, grain cementation, porosity, and sub-surface layering complicate this relationship [Ferguson *et al.*, 2006a; Putzig and Mellon, 2007a; Piqueux and Christensen, 2009]. Moreover, when dealing with a heterogeneous particle size distribution, the largest particle sizes are known to substantially influence thermal conductivity, the primary variable in deriving thermal inertia [Presley and Craddock, 2006; Piqueux and Christensen, 2009; Presley *et al.*, 2009]. However, some of these complications diminish when dealing with deposits of well-sorted sediments (e.g., aeolian bedforms) [Ferguson *et al.*, 2006b; Chojnacki *et al.*, 2011a]. When dune surfaces are thicker than several diurnal thermal skin depths (2–4 cm), heterogeneous subsurface materials (e.g., sand over bedrock) will not affect bulk diurnal thermal inertia [Putzig and Mellon, 2007b].

Based on sub-kilometer- to meter-scale surface observations and the assumption of a homogenous surface, effective particle sizes were derived from THEMIS thermal inertia values using the relationships determined by Piqueux and Christensen [2009]. Only locations atop the dune surfaces (lee face and stoss-side) that appeared homogenous and devoid of rock outcrops or mega-ripples observable at HiRISE or CTX resolutions were used for particle size derivations.

Depending on the spatial extent of the dune field in the image, thermal inertia values from 100s to 1000s of THEMIS pixels were collected for further analysis.

3.5 Composition: Spectra

To investigate dune sediment compositions, spectral data and compositional maps from two spectrometers were used.

3.5.1 Thermal Infrared Wavelengths: TES

We used three previously published TES-derived maps (at 16 pixels per degree) to investigate patterns in dune field dust coverage and composition. The TES spectrometer operates in the thermal infrared wavelength region at $1667\text{--}200\text{ cm}^{-1}$ ($6\text{--}50\text{ }\mu\text{m}$) [Christensen *et al.*, 1992, 2004]. The TES-derived dust cover index (DCI) [Ruff and Christensen, 2002] was used to determine the presence of silicate dust that may inhibit further spectroscopic study. This index utilizes the spectral behavior of surfaces at high wavenumbers ($>1300\text{ cm}^{-1}$) [Ruff and Christensen, 2002]. Additional TES indices, developed by Ruff and Christensen [2007] to investigate the origins of the global-scale surface type 2 (ST2) [Bandfield *et al.*, 2000], were here used for evaluating dune compositions. In the original works, the 465 and 530 indices are spectral ratios, which were devised to emphasize the presence of spectral features attributed to phyllosilicate minerals or other weathered materials. In practice, the 465 and 530 indices were found to correlate well with TES ST2 and primary silicate minerals, respectively [Ruff and Christensen, 2007].

3.5.2 Visible/Near Infrared Wavelengths: CRISM

To assess dune compositions and evidence of chemical alteration, we used near-infrared reflectance (VNIR) data from the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) onboard MRO [Murchie *et al.*, 2007, 2009c]. CRISM may be targeted with a high

spatial resolution of 18 or 36 m/pixel with 544 wavelengths (hyperspectral mode) or in a survey mode at 100 or 200 m/pixel with 72 wavelengths (multispectral mode), at a spectral sampling interval of 6.55 nm. Multispectral images were used in this study where hyperspectral data were not available. CRISM is composed of two spectrometers, a VNIR instrument covering wavelengths 362–1053 nm and an infrared (IR) instrument from 1002–3920 nm, both of which have specific processing requirements (see below) [Murchie *et al.*, 2007, 2009c]. CRISM data were processed using the CRISM Analysis Toolkit (CAT) interface and were corrected via the methods described by the CRISM team [Murchie *et al.*, 2009c; Seelos *et al.*, 2011]. This method includes calibration to radiance, division by the cosine of the solar incidence angle, and (for the IR) atmospheric correction by a scaled transmission spectrum derived from observations over Olympus Mons [McGuire *et al.*, 2009]. An additional algorithm was then applied to IR spectra to remove noisy spikes due to propagated detector noise [Parente, 2008]. Spectral summary products (Table 4.1) described by Pelkey *et al.* [2007] were created using CAT to assist in determining relative strengths of characteristic spectral features for further analysis.

In addition to examining these spectral summary maps, individual spectra were also assessed to better determine surface mineralogy. After processing of CRISM data as described above, individual spectra were derived by two different methods depending on the location of the most suitable denominator for rationing with the target: (1) several pixels (usually 5×5 or greater) from the target of interest were averaged, then divided by an average from a spectrally “bland” or unremarkable region in the same column of the unprojected image [Murchie *et al.*, 2009c] or (2) average spectra of irregularly shaped but equivalently sized regions of interest and bland regions were collected for the numerator and denominator, respectively, not necessarily (but usually) in the same column, and ratioed [Ehlmann *et al.*, 2009]. Spectral ratios help reduce

residual atmospheric features and systematic detector errors. In all cases, tens to hundreds of pixels were averaged together to improve the signal-to-noise ratio. Pixels used for mineral identification were collected from low-albedo dune surfaces that were clearly devoid of bedrock contamination in visible-wavelength images.

Here, CRISM spectra are presented from 0.45 to 2.6 μm , but the VNIR and near-IR images were processed and analyzed separately. Spectra from the two detectors were then joined by manually offsetting VNIR data to match the IR data in the $\sim 1\text{-}\mu\text{m}$ crossover region, and the bad bands of the VNIR ($>1.0102\text{ }\mu\text{m}$) to IR ($<1.0210\text{ }\mu\text{m}$) were excluded [Bishop *et al.*, 2009; McKeown *et al.*, 2009]. The $\sim 1.65\text{-}\mu\text{m}$ region may contain uncorrected atmospheric effects, and $\sim 2\text{-}\mu\text{m}$ region may have errors in correction for atmospheric CO_2 [Murchie *et al.*, 2009c]. Additionally, VNIR detector artifacts between 0.644 to 0.684 μm and the poor correction of light scattering by atmospheric dust at wavelengths less than $\sim 0.8\text{ }\mu\text{m}$ can occur [Murchie *et al.*, 2009c]. CRISM scene identification numbers used in this study are summarized in Tables 4.2 and 4.3 for low- and mid-latitude non-VM and VM dune field locations, respectively.

3.6 Dune Activity: Change Detection

Several data sets were employed to search for evidence of aeolian bedform changes. This study primarily utilized HiRISE visible-wavelength images acquired from 2006 to the present. All image pairs were calibrated as described in section 3.1 above, and spatially registered to fixed, static tie points (e.g., craters, boulders) using ArcGIS and ENVI software, allowing dune morphometry to be measured. For the one MOC-HiRISE pair presented in this study, change detection utilized the methods of Chojnacki *et al.* [2011a] (Chapter II) for image co-registration and change in area measurements.

We also employed three Mars Years of TES bolometric albedo data [Putzig and Mellon, 2007a] to search for changes in dune albedos. Increases in albedo (i.e., brightening) may be due to dust deposition following regional/global dust storms [Christensen, 1988a, 1988b; Sullivan *et al.*, 2008]. Conversely, decreases in albedo (i.e., darkening) can result from dust removal caused by horizontal wind shear events. TES bolometer data have an accuracy of 1–2% [Christensen *et al.*, 2001b], allowing albedo changes as low as a few percent to be detected. A higher value of $\pm 5\%$ of the original albedo was chosen as the minimum threshold for change events; for instance, a surface with an albedo of 0.12 in one year and 0.11 the next ($\sim 8\%$ change) would qualify as a “darkened” surface.

3.7 Dune and Dune Field Identification, Classification, and Mapping

We define dune fields as areas of relatively contiguous high-order bedforms possessing variable relief, composed primarily of sand-sized particles, and constructed by aeolian processes (i.e., saltation, creep). By contrast, we consider ergs to be larger dune fields of contiguous dark sand $>1000 \text{ km}^2$ in area, with multiple dune morphologies, and possessing a several-order hierarchy of bedforms including smaller superposed dunes [Wilson, 1973; Parsons and Abrahams, 2009].

For the majority of dune morphologies, we used the classification criteria defined by McKee [1979] based on gross bedform morphology. Recognized classes include barchan, barchanoid, linear, transverse, star, reversing, sand sheet, and dome dunes. In addition, three classes of topographically-related dunes (i.e., falling, climbing, and echo dunes) are identified by use of the morphological and topographic criteria defined by Chojnacki *et al.* [2010] (Chapter III). Additional occurrences of dunes possessing unusual morphologies that were not readily classified were also found [Hayward *et al.*, 2007a]. Overlapping dunes (compound) and/or

combinations of dune morphologies (complex) were commonly observed in large dune fields or ergs.

In order to place VM dune fields in a broader context, we drew from several previous works that examined other dune populations [Hayward *et al.*, 2007a, 2007b; Tirsch, 2009; Bridges *et al.*, 2012b]. Tirsch [2009] conducted a near-global study of 70 dune fields and sand sheets using MGS and MEx data, primarily OMEGA and HRSC, to investigate the composition and origin of intracrater dune materials. We restricted data from that study to the 44 deposits that were within $\pm 50^\circ$ latitude and were not isolated sand sheets without distinct duneforms (see below). In addition to these studies, we examined 184 low- and mid-latitude dune fields from the Mars Global Digital Dune Database (hereafter referred to as MGD³), provided as dune field polygons available from the USGS [Hayward *et al.*, 2007a, 2007b] (see Table 4.4).

Valles Marineris dune fields were mapped using a combination of co-registered CTX visible-wavelength images and THEMIS-derived thermal inertia images. Analysis of thermal inertia images assisted in interpreting the edges of dune fields where more diagnostic features (e.g., slip faces, wind scours) may not be present. Similarly, this method helped in excluding areas of bedrock and rubble that could spuriously elevate thermophysical measurements. See Tables 4.4–4.5 for naming convention of the MGD³ and VM dune fields.

Geomorphic mapping of dunes classified as unidirectional (i.e., forming under one dominant wind direction) was performed to provide paleowind direction indicators for the time of major dune formation or most recent modification. Slip face orientations were determined by annotating images with arrows perpendicular to the slip face (approximately in the middle of the feature) and pointing in the downwind direction, similar to previous studies [Fenton *et al.*, 2003; Hayward *et al.*, 2007a]. In cases where slip faces were large enough and presented flexures or

changes in the slip face curvature, multiple orientation arrows were used. The mean direction of all slip face arrows for a given dune field was calculated using the ArcGIS linear directional mean tool. This tool identifies the average direction for a set of lines or arrows and, for our usage, provides a “dune field mean slip face direction“. Finally, all quantitative properties (e.g., albedo) amassed from the MGD³ and VM dune field polygons were gathered from the same data sets and by the same method, to ensure that discrepancies in data processing or collection did not influence our results.

4.0 Results

4.1 Low- and Mid-latitude Non-Valles Marineris Dune Fields

4.1.1 Dune Field Occurrences, Morphology and Topography

The geographic locations of the 184 MGD³ low- and mid-latitude non-VM dune fields [Hayward *et al.*, 2007b] and their areal distributions are shown in Fig. 4.1b and Fig. 4.2a, respectively. The largest concentrations of dune fields occur in the Meridiani Planum, Syrtis Major, Hellas, and Argyre regions, with the majority of them south of the highland-lowland dichotomy [Christensen, 1988a]. No dune fields were reported in the 1:5 million quadrangles of Diacria, Arcadia, Cebrenia, Tharsis, Amenthes, Memnonia, or Phoenicis Lacus, which are primarily located in the Northern Hemisphere (Fig. 4.1b). Frequency and areal sum histograms of the non-VM dune fields are provided in Fig. 4.3a. Dune field area frequency peaks at 16–62 km² (~155 km² mean). As noted by Hayward *et al.* [2007], smaller dune fields are more abundant than larger ones. The total area of dune fields is ~28,300 km².

Craters host ~72% of dune fields in our non-VM study area, but the dune distribution and morphology for a given system can vary widely depending on the size and topographic complexity of the crater [Hayward *et al.*, 2007b]. Tirsch [2009] divided potentially mobile,

intracrater dark deposits into four useful and commonly observed categories: isolated sand sheets, small collections of individual dunes (simple dune fields), dunes arranged adjacent to crater topography (e.g., a central peak), and expansive complex and/or compound dune fields. This increasing level of morphological complexity is generally correlated with crater size. In particular, simple craters less than ~5 km in diameter often host unidirectional crescentic dunes (i.e., barchan, barchanoid), whereas larger complex craters, especially those that possess substantial central topography (e.g., central peak), contain dune morphologies that are less easy to interpret (as in, e.g., Moreux and Gale craters) [Hayward *et al.*, 2007a, 2009; Silvestro *et al.*, 2012b]. Fenton [2005] found intracrater dune morphologies in Hellas Planitia to be similar to those located in moderate-sized to large craters (e.g., Proctor crater), suggesting crater shape may not significantly determine morphology. The small percentage of intercrater plain dune fields are primarily adjacent to the Argyre or and Hellas basins [Hayward *et al.*, 2007a].

Martian dune morphology has been well categorized by several authors [Edgett and Malin, 2000b; Silvestro *et al.*, 2010a; Edgett, 1994; Tirsch, 2009; Hayward *et al.*, 2007a; Fenton, 2005; Fenton and Hayward, 2010]. Simple crescentic dunes, such as barchanoid, transverse, and barchan dunes, are recognized by the presence of a single, primary slip face and are generally considered to have been constructed under a unidirectional wind regime. Most (~75%) dune fields in the study area appear to be dominated by dunes classified as unidirectional [Hayward *et al.*, 2007a]. The non-VM dune morphological distribution has barchanoid (39%), barchan (22%), sand sheets (19%), and transverse (14%) morphologies (Fig. 4.2c). More complicated dune morphologies include star, reversing, and linear dunes, which are widely held to have formed under multiple and/or variable wind regimes and are less frequently reported (~5%). In the absence of adjacent topographic elements, dune morphologies are generally homogenous at

scales of 2–5 km, whereas dune morphologies on opposite sides of a given field are more likely to vary. Topographically-related dune fields, such as climbing and falling dunes (section 4.2.1) were not categorized in the MGD³, but are rarely reported [*Fenton et al.*, 2003; *Bourke et al.*, 2004].

Martian aeolian bedforms are found at nearly all elevations, but the distribution of dark-toned dunes is more restricted. Elevations for martian dune fields in our study area ranges from 1.7 km (highlands west of Argyre) to -7.5 km (bottom of Hellas). Mirroring global and regional topography, northern dune fields are located at lower elevations (<-3 km) than those in the Southern Highlands (>-2 km) with exceptions located in the low-lying (<-5 km) Hellas and Argyre basins. The distribution of dune field topographic reliefs and slopes can be viewed in Fig. 4.4. Note that often parameters of MOLA relief and slope are not due solely to dune morphology (i.e., trough-to-crest), but instead to the bulk terrain on which the dune field is superposed; exceptions to this pattern can be found with more massive dune fields in the Southern Highlands (e.g., the Proctor crater dune field), which are known to possess tall (100–300-m) dunes [*Fenton et al.*, 2003]. Reliefs are typically below 1 km (~400 m mean, 2 km maximum), indicating many relatively topographically gentle dune field environments (i.e., craters; Fig. 4.4a). The 12 examples with larger reliefs of 1–2 km are all for dune fields >200 km². Dune field slopes at MOLA resolution are similarly small (<5°, 1.4° mean; Fig. 4.4b). The largest relief and slope values are associated with dune fields located on the rims of Argyre and Hellas basins, or with large intracrater deposits (e.g., Kaiser, Rabe), whereas flat-lying dune fields (<1° mean, relief <1 km) are located in all regions (Fig. 4.4).

MOLA roughness values of the 36 dune fields are generally greatest at the 0.6-km baseline (i.e., high sub-km roughness) and then decreased at the 2.4- and 9.2-km scales (Fig.

4.5). This trend is interpreted as dune crest wavelengths being less than the 0.6-km baselines but lacking substantial topography and spacing (>1 km in size) at the longer baselines. These results are similar to trends found in the Olympia Planitia erg dunes using similar MOLA roughness measurements [Kreslavsky and Head, 2000]. Intercrater dune fields, often adjacent to large basins such as Hellas and Argyre, are typically rougher than crater-related dune fields at all baselines. Higher roughness values are generally related to more rugged intradune and adjacent terrain elements. For instance, many west Argyre locations with high MOLA relief, slope, and roughness values have dune fields superposed on irregular terrains with topographic elements at the 2–12-km scale.

4.1.2 Geologic Relationships

Of the 184 low- and mid-latitude non-VM dune fields examined, 113 (~62%, occupying a total of $\sim 13,200$ km² of dune area) are superposed on Noachian aged terrains: northwestern Argyre Basin, west of Hellas Basin, Meridiani Planum, and other scattered equatorial locations (Npl1; cratered unit, Nple; Etched unit, Npl2; subdued cratered unit, Npld; dissected plateau unit; see [Scott and Tanaka, 1986; Greeley and Guest, 1987]). Fifty-four dune fields (~30%, occupying $\sim 9,900$ km² of area) are located on Hesperian aged units: mostly north of Argyre Basin or Hellas Planitia (Hpl3; smooth unit, HNu; undivided material, Hs; Syrtis Major Formation, Hr; ridged plains material, Hplm; mottled smooth plain unit, Hh3; Hellas assemblage, Hch; older channel and flood-plains). Finally, 15 dune fields (~7%) are found straddling Noachian-Hesperian contacts; the undivided HNu unit straddles the Noachian-Hesperian boundary [Scott and Tanaka, 1986] but was counted as Noachian. Only two dune fields (~1%) are superposed on Amazonian terrains (Ah7; Hellas assemblage, Ael1; Elysium Formation).

These percentages as they relate to the proportions of terrains in the study area are discussed in section 5.5.

4.1.3 Thermophysical Properties

Albedo

TES global maps of bolometric albedo and thermal inertia were examined for trends in dune field thermophysical properties (Fig. 4.6). Mars Year 26 albedo values for non-VM dune fields generally range from 0.11–0.15 (0.14 mean; Fig. 4.6a); for comparison, mean values for all surfaces in this latitude range are closer to ~0.20. A few dune fields with outlying higher albedo values (>0.16; Arabia, north Syrtis, Hellas) likely indicate elevated dust coverage [Christensen, 1988b] or a greater areal fraction of light-toned interdune area or TARs enclosed in the MGD³ dune field polygons [Aben, 2003]. The darkest dune fields (<0.11) are located throughout the study area, but in greater abundances in Noachis and Argyre, with Proctor and Kaiser craters hosting some of the lowest albedo dunes. Indeed, minimal values show some dune surfaces reflecting less than ~7% of incoming light (albedo of ~0.07) and, along with lava flows and other bedrock areas, are some of the darkest surfaces on the planet.

Thermal Inertia

TES nightside-derived thermal inertia values for 32 of the largest non-VM dune fields are provided in Fig. 4.6b. All values occur between 180–390 tiu (280 tiu mean), consistent with silt- to sand-sized particles and are similar to previous TES and THEMIS thermal inertia measurements of dune fields [Aben, 2003; Fergason *et al.*, 2006a; Tirsch, 2009; Chojnacki *et al.*, 2011a]. TES dayside-derived thermal inertia values (not shown) of the dune fields are similar, but produce overall lower mean values. Only minor variations of ΔTI are detected for dune

surfaces (Fig. 4.6c); mean values range from -24 +55 tiu (+30 tiu mean). These low ΔTI values suggest homogeneity of grain sizes at the subpixel level, consistent with well-sorted dune surfaces. Dune sand with large variations in grain size would have greater nightside measurements relative to dayside. Dunes with elevated ΔTI values ($>+50$ tiu), located east of Argyre and Hellas, may result from high slopes, particle size heterogeneity, or dunes that are thin relative to the diurnal thermal skin depth (Fig. 4.6c).

4.1.4 Dune Field Composition

TES

TES global maps provide a useful way of examining global trends in dune field dust coverage and composition. The TES dust cover index values range from 0.85 to 0.99, with high values (>0.962) indicating “dust free” surfaces and lower values (<0.94) suggesting “dust covered” surfaces [Ruff and Christensen, 2002]. The DCI values for non-VM dune fields (Fig. 4.7a) are dominantly ($\sim 82\%$) consistent with dust free surfaces (~ 0.97 mean); most of the remaining dune fields, many in Hellas and Lunae Palus-Arabia, exhibited moderate values that indicate variable dust coverage. A single example in Meridiani Planum possessed values indicative of dust-mantled surfaces.

The 465 and 530 TES index values for dune fields are shown in Figs. 4.7b and 4.7c, respectively; index values above >1.006 and >1.005 (inset histograms, dashed lines) indicate 3-sigma detection thresholds for the 465 and 530 indices, respectively [Ruff and Christensen, 2007]. Dune fields with 465 index values above threshold (Fig. 4.7b) appear in Syrtis Major, east of Argyre and Hellas, and other scattered locations. Dune surfaces with elevated 465 index values are inferred to have sand compositions consistent with a ST2 surface, possibly

intermediate volcanic glass-rich basaltic andesite [Bandfield *et al.*, 2000] or an altered basaltic composition (high-silica amorphous or poorly crystalline phase) [Wyatt and McSween, 2002]. Dune fields with 530 index values above threshold (Fig. 4.7c) are primarily located in the Southern Highlands (e.g., Argyre, Noachis, Tyrrhenum, Aeolis, and Eridania) and Syrtis Major. Surfaces with elevated 530 index values are inferred to have sand compositions partially composed of olivine [Ruff and Christensen, 2007]. Locations with high 465 index values combined with low 530 index values (Fig. 4.7b, green stars) have been shown to be more indicative of ST2 surfaces than the 465-feature alone [Ruff and Christensen, 2007].

OMEGA

Data from the OMEGA instrument have shown dark surfaces to be highly correlated with the presence of mafic minerals [Bibring *et al.*, 2005; Mustard *et al.*, 2005]. In that instrument's VNIR spectral range, pyroxenes are the most common martian surface mineral group [Poulet *et al.*, 2007; Bibring and Langevin, 2008]. In general, low-calcium pyroxene (LCP) is enriched in Noachian terrains (e.g., Noachis, Hellas), whereas high-calcium pyroxenes (HCP) are more prominent in younger terrains associated with lava flows (e.g., Syrtis) and dark sand deposits [Mustard *et al.*, 2005]. Olivine detections, often correlated with pyroxenes, are also spatially diverse, but less spatially extensive. Tirsch [2009] examined 37 dune fields in the study area with OMEGA spectral data (Fig. 4.7d). That study found detections of pyroxene were the most common, either alone (59%) or with olivine (41%; Fig. 4.7d, inset). Dunes within the latter class possess spectra with broader 1- μm features attributed to olivine and are frequently located in the tropical latitudes [Tirsch *et al.*, 2011]. Tirsch *et al.* [2011] noted that absorption features

attributed to these minerals tend to be deeper in spectra of prominent dark-toned, barchanoid dunes than in spectra of adjoining sand sheets.

CRISM

In this study, CRISM spectral data of 30 non-VM dune fields were examined to identify major mineralogic trends (Fig. 4.7d, triangles; Table 4.2). Seven sites did not produce suitable spectra, possibly due to surface dust contamination. The remaining 23 spectra were divided into four classes after spectral analysis: HCP (38%), LCP (15%), olivine (23%), and a “basaltic” class that included similar spectral contributions from both pyroxene and olivine (23%). Pyroxenes (particularly HCP) were the most commonly detected and wide-spread sand mineralogy. Olivine detections were generally located in the tropical latitudes. No evidence of alteration products associated with dune sand was detected in our survey. CRISM results compare well with previous OMEGA studies [*Tirsch et al.*, 2011] and show the dominance of detected basaltic mineralogies in association with dune sand, but with subtle variations in the spectrally dominant phase.

4.1.5 Aeolian Activity

The results of several previous studies were integrated into this research [*Chojnacki et al.*, 2011a; *Geissler et al.*, 2011; *Silvestro et al.*, 2011; *Bridges et al.*, 2012b] in order to examine near-global trends in recent morphologic dune activity (Fig. 4.8a). Results from those authors are given in three categories: dune migration, surface change (e.g., slip face avalanches, deflation), and no detection. Examples of confirmed surface changes caused by aeolian processes include dune edge or slip face modifications, ripple advancements (at rates up to a few meters per MY), or bulk bedform deflation [*Chojnacki et al.*, 2011a; *Bridges et al.*, 2012b;

Silvestro et al., 2011]. Detected dune changes occurred at elevations ranging from +1.8 to -5 km. Most (64%) morphologic dune changes occurred in locations with TES DCI values consistent with dust free surfaces (e.g., Noachis and Syrtis Major; Fig. 4.7a), with a few exceptions in Arabia where changes occurred in locations of “moderate” dust coverage. In addition, the sites of dune activity were slightly lower in both elevation (~1 km) and albedo (~10%) in comparison to locations where no change was detected.

Many dune fields, like other Martian surfaces [*Putzig and Mellon*, 2007a], exhibited notable positive and negative albedo change over the three Mars Years of TES coverage (Figs. 4.8b–4.8c). Between MY 24 and 26, over 80% of the dune fields showed changes above the threshold value (5%). These locations exhibited similar albedo changes across several orbital tracks encompassing 10s to 1000s of km² of area, arguing against instrumental or atmospheric causes. Between MY 24 and 25 (Fig. 4.8b), ~30% of dune fields became darker (e.g., Meridiani, Tyrrhenum, Eridania), ~10% became brighter, and the remainder fell below threshold (Southern Highlands). Larger changes were detected between MY 25 and 26 (Fig. 4.8c), where dune fields either decreased (~50%) or increased (~25%) in albedo (e.g., Meridiani, Syrtis). As an example, a Syrtis Major dune field with a +53% increase in albedo (from 0.12 to 0.18) from MY 25 to MY 26 appeared to possess some dust coverage, indicated by abundant dust devil tracks on mid-toned dune surfaces (e.g., MOC R1102407 taken at the end of MY 26). These albedo changes of dark intracrater deposits are analogous, although greater in magnitude, to those described in previous studies that used coarser resolution data sets and longer (~10 MY) temporal baselines (i.e., Viking Orbiter data as compared with TES data) [*Geissler*, 2005; *Fenton et al.*, 2007].

Comparison can also be made between small-scale (1–20 m) changes in dune morphology (Fig. 4.8a) and dune field-sized (10s to 1000s km² of area) changes in albedo (Figs. 4.8b–4.8c). Of the previously documented dune surface changes within the study area (e.g., Meridiani, Noachis, Argyre), ~67% were found on surfaces that had mean darkening in one or more of the TES time steps (i.e., MY 24–25 or MY 25–26). Although many of these different data sets were not acquired contemporaneously, the moderate correlation between the two indicates that TES albedo changes may be reasonable proxies for aeolian sand movement. *Chojnacki et al.* [2011a] documented morphologic changes on dunes in Endeavour crater that occurred concurrently with TES-detected albedo changes. This location was shown to have experienced substantial dune activity between MY 24 and 29 (based on MOC-HiRISE-CTX analyses) and darkened by ~12% in albedo between MY 24 and 26 (from 0.14 to 0.12).

4.1.6 Aeolian Morphology Related to Topography and Modeled Winds

Many authors have mapped slip face orientations of unidirectional dune types (i.e., barchans, barchanoid, or transverse) or dark streak orientations in an attempt to infer local/regional wind regimes and sand pathways [*Greeley et al.*, 1993; *Fenton et al.*, 2003; *Hayward et al.*, 2007b; *Silvestro et al.*, 2010a]. These maps have then been compared to global and mesoscale wind models with various degrees of success in term of matching directionality, seasonality, and wind magnitude [*Hayward et al.*, 2007a; *Gardin et al.*, 2012]. In many cases, GCM models have limitations in resolving the effects of micro-topography, katabatic winds, and boundary layer turbulence, and do not always correlate well with locations of contemporary enhanced dune activity [*Bridges et al.*, 2012b]. In a broader context, GCMs do capture the larger-scale atmospheric patterns and may better reflect migration patterns that are longer than the durations of spacecraft missions [*Hayward et al.*, 2007a; *Gardin et al.*, 2012].

Mixed results have also been found in comparing mesoscale atmospheric models to aeolian surface markers. For instance, *Fenton et al.* [2005] compared wind directions from a mesoscale atmospheric model to the dunes in Proctor crater and found the model agreed with two of three slip face orientations and dust devil track orientations, but failed to produce wind speeds exceeding the threshold friction velocity. Similarly, the wind directions of season-specific MRAMS models aligned with direction of dune change in Endeavour crater, but failed to predict near-surface (~ 1.3 m) wind speeds necessary to generate surface shear stresses above threshold for basaltic sand [*Chojnacki et al.*, 2011a]. However, *Michaels* [2011] did find MRAMS predicted surface winds exceeded the threshold friction velocity of basaltic sand grains in Nili Patera where studies have detected unambiguous dune migration [*Silvestro et al.*, 2010b; *Bridges et al.*, 2012b].

Commonly, a regional cluster of craters is observed hosting separate dune fields that have dunes with a similar dominant slip face orientation, implying that they formed under a similar wind regime [*Hayward et al.*, 2007b; *Silvestro et al.*, 2010a; *Chojnacki et al.*, 2011a]. For example, all intracrater dunes within ~ 100 km of Endeavour crater, in Meridiani Planum, possess similar slip face orientations [*Chojnacki et al.*, 2011a]. As noted above, a significant ($\sim 65\%$) correlation exists between DCA, slip face orientation, and GCM wind directions for these smooth floored, moderate-sized craters [*Hayward et al.*, 2007a]. This correlation breaks down for dune fields found on larger, more irregular craters (e.g., central peaks, superposed craters) and in rugged intracrater terrain (as in, e.g., Moreux and Gale craters). In those instances of more substantial topography, individual dune slip face orientations are commonly not aligned with one another, DCAs, or flow direction predictions from atmospheric models.

4.2 Valles Marineris Dune Fields

Sixty dune fields were identified in seven Valles Marineris chasmata (Figs. 4.1b, 4.9–4.10; Tables 4.5). Dune fields are located in Ius, Melas, Hebes, Coprates, Juventae, Capri, and Ganges chasmata (Fig. 4.2b). According to available data, no substantial dune fields were found in Ophir, Candor, and Echus chasmata, despite the presence of dark-toned, isolated sand patches and individual duneforms in the locations. Dune field frequency and areal sum histograms (Fig. 4.3b) show frequency peaks at 8–125 km² (~400 km² mean), but the vast majority (~80%) of dune area is contained in four ergs (>1000 km²).

4.2.1 Dune Field Occurrences, Morphology and Topography

Barchanoid dunes are the most common morphology in VM (~32%), followed by transverse (18%), sand sheet (15%), and climbing or falling dunes (14%) (Fig. 4.2d; Table 4.5). Barchan, linear, dome, and echo dunes are less common and occur in characteristic settings of dune fields. Dome and barchan dunes commonly occur on the edges of the dune fields, where aeolian sediment is less abundant; echo dunes are usually found in chaotic floor material in the eastern chasmata. Dune morphologies are generally heterogeneous on scales of less than a few kilometers and the dune field shapes tend to be irregular at broader scales. In addition to their spatial size, dune fields can be broadly classified based on their topographic and geographic context. The four dune field classes defined in this study are floor, landslide, wall, and erg dune fields (Fig. 4.1b).

Floor Dune Fields

The broadest and most commonly occurring class (41 in total or ~67%) of VM dune fields is the floor dune field classification. Most of these dune fields are located on chasmata floors, adjacent to wall termini, mensae, chaotic materials, craters, and landslides. In some

cases, floor dune fields are partially or totally on top of ILDs (Figs. 4.11a, 4.11c). Fields in this class range in area from 10s to 100s of square kilometers ($\sim 100 \text{ km}^2$ mean). Floor dune fields are present in all chasmata analyzed in this study, but appear to be most common in Ius, Melas, Coprates, and Capri chasmata (Figs. 4.1b, 4.9–4.10).

Among the four dune field classes, floor dune fields display the most variability in dune morphology, having barchanoid, barchan, transverse, sand sheet, dome, echo, and linear dunes, including both compound and complex variations (Fig. 4.11). In the simplest case, unidirectional dune morphologies on central floors where they are largely removed from other canyon topography conform to a single dominant direction (Fig. 4.11b). More commonly, topographic elements, such as chaotic materials, disrupt simple dunes, instead forming echo and linear duneforms (Fig. 4.11c). Melas Chasma floor dune fields are often geographically associated with sulfate-bearing ILDs [Gendrin *et al.*, 2005] where dunes congregate on the flanks of the deposits (Fig. 4.11a). In this case, dunes extend southward, eventually grading into climbing dunes [Chojnacki *et al.*, 2010]. Floor dune fields are similar in scale and appearance to many non-VM dune fields found in moderate-sized craters ($\sim 25\text{--}40 \text{ km}$ in diameter), but with greater ($>1 \text{ km}$) relief and slopes ($>3^\circ$) (Fig. 4.4).

Landslide Dune Fields

The relatively rare landslide class ($\sim 9\%$) is designated for dune fields that are superposed on landslides. These dune fields can be found near the landslide scarps (heads) or the slide termini (toes) (Fig. 4.12; Chapter V). Some instances of this class are draped on landslide-floor contacts and are adjacent to neighboring floor dune fields. Two examples of this relationship are located in Coprates and Ius chasmata where dune slip face patterns make it unclear whether dune

fields migrated on to or off of landslides (Figs. 4.12a, 4.12c). Seven landslide dune fields are found superposed on landslides emerging from the north and south walls in Ius, Melas, Coprates, and Hebes chasmata (Fig. 4.9, 4.10a). Landslide dune fields are primarily composed of barchanoid or transverse dunes. These dune fields are mostly limited in spatial extent (10–130 km², ~75 km² mean) and typically consist of several small dune groups spread out across a landslide (Figs. 4.12a–4.12b). This class is exclusively superposed on some of the youngest martian surfaces (~1 Gy to ~50 My; Late Amazonian) where dune fields are found (section 4.2.2).

Wall Dune Fields

Dune fields located among chasmata wall slopes are referred to as “wall dune fields” and further sub-classified as either climbing or falling dune fields (Fig. 4.13). Based on qualitative descriptions in the terrestrial literature [Greeley and Iversen, 1985; Pye and Tsoar, 1990], previous work by Chojnacki *et al.* [2010] has defined martian *climbing dunes* as bedforms with lee faces on the uphill side and *falling dunes* as bedforms with lee faces on the downhill side, both lying on a topographic slope >8°. On Earth, these types of dunes are largely controlled by micro-topography; positive relief obstacles can accelerate and decelerate airflow, create turbulence, potentially enhancing erosion, deposition, and/or transport of sediment [Pye and Tsoar, 1990; Tsoar *et al.*, 1996; Lancaster and Tchakerian, 1996; Wiggs *et al.*, 2002]. Occurrences of these climbing and falling dune fields are located among the spur-and-gully walls in Melas, Coprates, and Hebes chasmata (Figs. 4.9b–4.9c, 4.10a), dominantly on north-facing slopes. This class is relatively rare across the rift (~16%) and Mars [Hayward *et al.*, 2007a; Tirsch, 2009; Chojnacki *et al.*, 2010].

The morphology of climbing dunes is very similar to that of barchanoids, but with sporadic slip faces that form on both spurs as well as gullies, primarily on the lower third of the wall slope. On confined steep slopes, a “sand ramp” morphology [Lancaster and Tchakerian, 1996; Chojnacki *et al.*, 2010; Bateman *et al.*, 2012] or “barchan trains” may be observed. An excellent example of the latter case is found in Hebes Chasma, where several barchans appear to have migrated southward up a linear gully with a slope of $\sim 10^\circ$ (Fig. 4.13a).

Many wall locations in the eastern VM show contiguous bedforms in gullies that are 1–3 km above the canyon floor (Fig. 4.13c) and are identified as falling dunes. Although they are variable in appearance, falling dunes are most commonly found with a symmetric “lobate” morphology, in which repeating, arcuate slip faces are arranged with horns pointing up hill (i.e., slip faces on the outside of crescent) or the upwind direction [Chojnacki *et al.* 2010]. The largest concentration of falling dunes is located on the walls within eastern Coprates Chasma. Most of these occurrences are found positioned precariously, midway up a horst [Schultz and Lin, 2001], informally referred to here as the “east Coprates massif” (Fig. 4.13c) [Beyer and McEwen, 2005]. Along with dark-toned dunes, inspection of HiRISE images in these areas (not shown) reveals repetitive rippled textures that cover smoother talus slopes ($30\text{--}40^\circ$ in HRSC topography) several kilometers up the wall, whereas low-order bedforms are located on lower slopes. Outside of VM, climbing and falling dunes have been rarely reported, with minor occurrences along crater rims pits or troughs [Fenton *et al.*, 2003; Bourke *et al.*, 2004].

Ergs

Four major ergs are located in Juventae and Ganges chasmata (Figs. 4.1b, 4.10b, 4.10d). In addition to size $>1000\text{ km}^2$, several other factors, including many orders of superposed

bedforms, distinguish this VM class from others. These expansive ergs cover thousands of square kilometers of area and typically stretch from the base of wall slopes across the chasm to the base ILDs, landslides, or adjacent walls, dominating a significant portion of the total canyon bottom surface area. Although few in number, these ergs account for over three-quarters of the total dune area mapped in VM, and three of them comprise the largest dune fields (3000–8000 km²) outside of the north polar region (Fig. 4.1; Tables 4.4, 4.5).

Valles Marineris erg morphologies are characterized by expansive interior sand sheets that grade into barchanoid or transverse dunes on erg margins (Fig. 4.14). Sharp slip faces tend to develop adjacent to positive relief topography, whereas many interior portions may be devoid of prominent aeolian bedforms (Fig. 4.14b). High sediment supply and persistent winds patterns form compound-complex dunes many 100s of meters tall (Fig. 4.14a). Echo dunes or smaller crescentic sand scours [Bishop, 2011] can also be found in erg interiors, where winds have interacted with chaotic material or other positive relief topography (Fig. 4.14b). Often, a series of barchanoid and transverse dunes will grade into peripheral sand sheets, whereas barchan and dome dunes may develop on the perimeter of the field where the sediment supply is more limited (Fig. 4.14c). Additional distinctions between the four classes of VM dune fields in thermophysical properties, composition, and topographic regimes are discussed below.

Dune Field Topography

Elevations of VM dune fields range from -1 to -5 km, but are commonly around -3.5 km (Table 4.5). The reliefs of the dune field polygons, which largely relate to the underlying topography, range from 100s of meters to ~4 kilometers (Fig. 4.4a), with an average relief of ~1 kilometer. Also, MOLA slopes of the dune fields are ~1–17°, with mean slopes of ~4° (Fig.

4.4b). Dune field topography was also examined using HRSC data, which can resolve individual duneforms. These measurements produce more extreme values for dune and dune field slopes ($\sim 0\text{--}45^\circ$, $\sim 9^\circ$ mean; Figs. 4.11–4.14, insets). Because the upper limit of sand dune surfaces is expected to be at the angle of repose ($32\text{--}34^\circ$) [Pye and Tsoar, 1990], extreme slope values are attributed to outcrop topography, such as wall spurs and mensae. The wall dune class is found with the highest mean MOLA elevations (-1.7 km), the greatest mean reliefs (~ 1.7 km), and the highest mean HRSC slopes ($\sim 15^\circ$; Fig. 4.13). The erg and the landslide dune fields have the lowest mean elevations, -3.5 and -4.0 km, respectively. Erg surfaces frequently show HRSC slopes below 8° ($\sim 5^\circ$ mean) excluding where major slip faces or infrequent non-dune materials yield values up to $\sim 30^\circ$ (Fig. 4.14). In contrast, hummocky landslides with superposed dune fields show HRSC slope measurements over both dune and non-dune materials from $10^\circ\text{--}33^\circ$ ($\sim 7^\circ$ mean; Fig. 4.12). Floor dune fields on the flanks of ILDs show some of the greatest relief ($2.1\text{--}3.7$ km; Fig. 4.11); however, all classes have at least one dune field, which is greater than 2 km in relief.

The 13 VM dune fields large enough (>200 km²) to be assessed with the MOLA roughness data set possess some of the largest roughness values (at all baselines) examined for this study (Fig. 4.5). Floor dune fields with large barchanoid dunes are considerably rougher at both the 0.6-km and 9.2-km baselines as compared with the other classes and non-VM dune fields. Floor dune fields had higher roughness values, whereas the ergs are consistently smoother at all baselines. Coprates Chasma possess the roughest dune fields, and those in Melas Chasma are the smoothest. Morphologic inspection of dune field terrains with these endmember roughness values corroborates these trends (e.g., Figs. 4.11a, 4.12a, 4.13c).

4.2.2 Geologic Relationships

Of the 60 Valles Marineris dune fields, 35 (~51%), occupying a total of ~20,000 km², are superposed on Hesperian terrains in Ius, Hebes, Juventae, half of Melas, half of Coprates, half of Capri, half of Ganges chasmata (HNu; undivided material, Hvl; layered material, Hcht; older chaotic material, Hch; older channel material). Another 29 dune fields (~49%) for a total of ~18,200 km² are superposed on Amazonian terrains in half of Ius, half of Hebes, half of Melas, half of Coprates, half of Capri, half of Ganges and most of Juventae (Avf; floor material, As; slide material). Seventeen dune fields are found straddling Hesperian-Amazonian contacts. No VM dune fields are detected on Noachian terrains (~37% of VM interior units); however, many wall dunes are adjacent to these ancient units (e.g., eastern Coprates Chasma).

Several dune fields, particularly the landslide class, are located on some of the youngest (1–200 My) geologic units on Mars [Scott and Tanaka, 1986; Quantin *et al.*, 2004]. Quantin *et al.* [2004a] used THEMIS and MOC data to estimate crater size-frequency ages for VM landslides (~3.5 Gy to ~150 My), some of which possess superposed dune fields. The two youngest examples of these dune fields, located in Ius and Hebes chasmata, are superposed on massive 150–200 million-year-old landslides (Figs. 4.12b–4.12c).

4.2.3 Thermophysical Properties

Albedo

The thermophysical properties of Valles Marineris dune fields are summarized in Table 4.7. These dune fields, like other Martian dune populations, are usually low in TES MY 26 albedo. Most of these albedos range between 0.10 and 0.16, with a mean value of 0.13 (Fig. 4.6a). As a class, wall dune fields show the lowest albedo (~0.12), whereas the landslide dune

fields have the brightest surfaces (~ 0.14). Dune fields of the western chasmata are higher and more variable in albedos than those in the eastern portion, notably Coprates (Fig. 4.6a).

Thermal Inertia

Nightside-derived thermal inertias for the 20 VM dune fields large enough for use with TES range from 200–390 tiu (295 tiu mean) and are consistent with particle sizes of fine to very coarse sand sizes (160–2000 μm) for unconsolidated sediments (Fig. 4.6b; Table 4.7). Dune fields superposed on landslides in Melas Chasma or adjacent walls in Coprates Chasma possess the highest mean values (>330 tiu), suggestive of large particle sizes, thermally thin dune surfaces, and/or induration. Diurnal variations in modeled thermal inertia (i.e., ΔTI) are $\sim 2\times$ as variable for VM dunes ($+25$ – 170 tiu, 55 tiu mean; Fig. 4.6c; Table 4.7) as for non-VM dunes. Relatively low values (<35 tiu) are associated with ergs in Juventae and Ganges chasmata. The low mean ΔTI values of the ergs imply uniform, non-layered dune surfaces [Putzig and Mellon, 2007b], consistent with the vast and uninterrupted areas of sand sheets. In contrast to these more uniform dune surfaces, floor dune fields in eastern Coprates and northern Melas, and those in the landslide class show higher ΔTI values (>70 tiu). These higher values are related to particle size heterogeneity (e.g., lag deposits), induration, and/or thermoclinometric effects (i.e., dayside slope-related illumination variations).

The higher spatial resolution of THEMIS thermal inertia data, compared with data derived from TES, reveals the heterogeneity of the martian dunes in greater detail (Figs. 4.15–4.16). For example, most dune lee and stoss sides are clearly resolvable at the 100-m scale, preventing ambiguities encountered using TES data. THEMIS thermal inertia values of dune surfaces range from 220–440 tiu (280 tiu mean), consistent with fine to very coarse sand (Table

4.8). Without evidence for bedrock contamination, excessively high values (>350 tiu) may be evidence for surface induration or particle sizes greater than sand (see below). The thermal inertias of dune surfaces often contrast with values of the immediately surrounding terrains and interdune areas, which may be moderately lower (<100 tiu; dust mantled surfaces) or greater (>800 tiu; canyon floor basement rock) (Figs. 4.15–4.16).

A compound-complex floor dune field in eastern Ius Chasma contains barchanoids, sand sheets, and one unusually immense duneform or draa (Figs. 4.9a, 4.15a). This draa has a long south-facing slip face (Fig. 4.15c) that measures ~ 280 -m-tall in HRSC topography (Fig. 4.15a, A–A' profile) and is one of the largest VM aeolian bedforms identified in this study. Several locations near the stoss side and crest possess thermal inertia values (~ 340 tiu) that translate into unconsolidated grain sizes (~ 3.5 mm) that are generally considered too high for mobile sediment (Table 4.8), possibly indicating induration (see sections 5.3, 5.5).

In many of the dune fields, barchanoid surfaces, particularly their stoss sides near the dune crests, displayed elevated thermal inertias relative to neighboring sand sheets (Figs. 4.15a, 4.15d, 4.16e; Table 4.8). A thermal inertia profile of Ganges Chasma floor dunes shows one example of this trend (Fig. 4.15d, B–B' profile). Here, barchanoid dunes (Fig. 4.15e) exhibit thermal inertia values of 260–320 tiu consistent with 700–1500 μm sediment (coarse to very coarse sand), whereas sand sheet values between 260–320 tiu are consistent with ~ 500 μm sediment (medium sand). Stoss sides with elevated thermal inertias are present at all orientations. For nightside thermal measurements, thermoclinometric effects would produce anomalously higher derived thermal inertia values only on west facing slopes.

THEMIS measurements in Coprates Chasma produce some of the higher thermal inertia values for all VM dunes (Table 4.8). In the cases of dunes on or near walls, high values (>300

tiu) might be caused by thermal adjacency effects from other walls [Bandfield, 2009], and/or mass-wasting of high thermal inertia material into the dune fields. An example of this second scenario can be found in collapse pits south of Coprates Chasma (Figs. 4.9c, 4.16a–4.16d), where fans of talus with rock sizes down to the limit of HiRISE detectability are observed adjacent to dune surfaces. HiRISE images show countless boulder tracks trailing into the dune field from the north and southwest walls (which have 18–25° slopes in HRSC topography), where they dissipate into dune sand. A small population of boulders can be found within the dune field superposed on rippled surfaces (e.g., Fig. 4.16d). THEMIS thermal inertia and IRA albedo profiles show a clear thermophysical contact between the wall rock and talus, but an indistinguishable transition between the wall talus and dunes (Fig. 4.16a, A–A' profile). Dune surfaces far from any potential contamination from wall debris (Figs. 4.16b–4.16d) also show elevated thermal inertia values (~325 tiu), consistent with very coarse sand. Thermal adjacency and slope effects are less likely to cause these elevated values because this trend occurs relatively consistently, regardless of slope orientation.

Typical thermophysical properties of VM ergs can be viewed in Fig. 4.16e where Juventae Chasma interior sand sheet surfaces appear relatively homogenous at the 100-m scale. Barchanoid and transverse dunes at the edges of the erg exhibit elevated thermal inertia values relative to most of the interior sand sheets (Fig. 4.16f–4.16g; Table 4.8). A well-defined thermophysical contact separates the dune field from the higher thermal inertia ILD and wall units to the west. However, several thermal anomalies are apparent within the erg interior. Barchanoid and transverse dune surfaces near the northern wall possess thermal inertia values in excess of 400 tiu that are associated with dune stoss slopes (Fig. 4.16g), perhaps caused by

induration or large particle sizes. HiRISE images reveal brighter-tone, lower-order bedforms located in dune troughs, and no evidence for mass-wasting (e.g., boulders).

A more complicated explanation is required for the high thermal inertia large landform slightly to the east (Fig. 4.16e). Based on morphological evidence, *Chojnacki and Hynek* [2008] hypothesized such interior one-kilometer-tall mounds are the remnants of a much larger layered deposit of a similar origin as other Juventae ILD mounds, but now partially buried by erg sand. Higher resolution images clearly show a ~12-km-long series of SW-NE trending slip faces on the brink of the 700–1000-m-tall mound (Fig. 4.16e, B–B' profile, white arrows). A draa approaching 1 km in height would be the largest duneform ever reported on Mars [*Bourke et al.*, 2006], but would require further study. In addition, ubiquitous lower-order superposed aeolian bedforms and a series of prominent slip faces suggest a complicated aeolian system has developed in these locations.

4.2.4 Dune Field Composition

TES

The TES-derived DCI [*Ruff and Christensen*, 2002] values of VM dune fields (~0.964 mean) indicate dust coverage is relatively minimal (Fig. 4.7a). Wall and floor dunes are found with “dust free” values (>0.962), except for a few floor dune fields in the western chasmata with moderate dust coverage. The erg and landslide dune field classes possess lower DCI values (0.965–0.966 mean) indicating greater dust coverage than other classes. Mean values of the TES 530 index of VM dune fields are not above the threshold set by *Ruff and Christensen* [2007] (Fig. 4.7c, inset; Table 4.6). However, many western VM dune fields possess mean 465 index values above threshold, in addition to having low 530 index values (Fig. 4.7b, green stars),

making them candidates for altered ST2-type terrains and altered basaltic compositions. These locations in Ius and Melas chasmata are also sites where OMEGA [Horgan *et al.*, 2012b, 2012c] and CRISM (see below; Fig. 4.17) observations have identified evidence of weathered glass-rich basalt.

CRISM

CRISM mafic spectral index values of VM dunes, such as olivine (OLINDEX), LCP (LCPINDEX), and HCP (HCPINDEX) (Table 4.1; [Pelkey *et al.*, 2007]), are typically moderate to high relative to surrounding terrain, whereas IRA albedo and ferric phases (BD530) are low; these trends are consistent with mafic (ferrous) compositions. Less commonly, spectral evidence of alteration products, such as weathered glass and hydrated minerals, is found (Fig. 4.17; Chapter V). Commonly in erg interiors or with landslide surfaces, no suitable spectrum was available from a given CRISM observation, presumably due to surficial dust. Highlighting the compositional diversity, Figs. 4.18–4.19 provide views of VM dune surfaces with a variety of CRISM indices. These images reveal variable index values for different dune surfaces, indicating variations in mineral compositions and/or particle sizes [King and Ridley, 1987b; Mustard *et al.*, 2005].

The distribution of CRISM mineralogic detections of VM dune surfaces is shown in Fig. 4.17a. A characteristic basaltic spectrum is the most typical shape (~39%; 4.17a, inset), with some dune surface spectra showing evidence of a more pronounced pyroxene or olivine spectral shape (Figs. 4.17c, 4.18–4.20). These basaltic spectra show positive slopes from 1.0 to 2.6 μm , deep ~1- μm features, and shallow absorptions between 1.8 and 2.4 μm (e.g., Figs. 4.17c, 4.18a, 4.19c, 4.20a, 4.20c, 4.20g). Laboratory studies of pyroxene and olivine mixtures (i.e., basaltic

compositions) produce spectra [*Clenet et al.*, 2011] comparable to some of these examples. A subtle shift of the $\sim 1\text{-}\mu\text{m}$ feature to longer wavelengths (from 1.10 to 1.18 μm) indicates the presence of unweathered Fe-bearing basaltic glass (Fig. 4.18a) [*Horgan and Bell*, 2012a; *Tirsch et al.*, 2012]. Spectra that displayed a $\sim 1.05\text{-}\mu\text{m}$ feature coupled with a prominent $\sim 2.3\text{-}\mu\text{m}$ feature, are identified as HCP (17%) (e.g., Figs. 4.17c, 4.18b-4.18c, 4.19c, 4.20a) [*Burns*, 1993]. Less commonly (5%), a similar shaped spectra displayed shorter wavelength band centers at $\sim 0.9\text{ }\mu\text{m}$ and $\sim 2.1\text{ }\mu\text{m}$, and are consistent with LCP (e.g., Figs. 4.17c, 4.19b). Olivine-bearing sand was identified based on a broad $\sim 1\text{-}\mu\text{m}$ feature several 100s of nm-wide (17%; e.g., Figs. 4.17c, 4.19a, 4.20e) [*Burns*, 1993]. In some cases, VNIR spectra of dunes reveal broad features shortward of $\sim 0.7\text{ }\mu\text{m}$ (Fig. 4.17c, “olivine” spectrum), which likely indicate the presence of ferric iron and alteration [*Cloutis and Bell*, 2004]. In all likelihood, these dunes are composed of multiple minerals and glasses in varying proportions, some of which are not readily detectable in the VNIR. As an important example, most plagioclase phases, known to be abundant on the martian surface from thermal infrared studies [*Bandfield et al.*, 2000; *Bandfield*, 2002a], do not have prominent absorption features in CRISM’s spectral range.

Observations in western VM display a variety of spectral shapes with subtle variations that we attribute to different mafic minerals and glasses (Figs. 4.18a–4.18b, 4.20a). Spectra of one northwest Ius Chasma dune field exhibit broad absorption minima centered on $\sim 1.07\text{ }\mu\text{m}$, attributed to the presence of basaltic glass (Fig. 4.18a). Another dune field in southeastern Ius Chasma shows evidence of compositional heterogeneity across the CRISM observation. An HCP spectrum is associated with interior smooth sand sheets, whereas a more generic “basaltic” (olivine and HCP) spectrum is found along slip faces and dune field edges (Fig. 4.18b). This

variation may be due to more pronounced aeolian mixing of sediment along dunes edges. One Hebes Chasma sand sheet superposed on hydrated-sulfate-bearing ILDs also displays spectral features consistent with HCP (Fig. 4.18c).

CRISM multispectral observations of many Melas and Coprates chasmata dunes produce spectra that differ from many martian dune spectral shapes. Reflectance spectra of these dunes possess overall negative slopes from 0.7–2.6 μm , are concave-up from ~ 0.7 to ~ 1.4 μm , have minima near 1.1 μm , and often show absorptions longward of 2 μm (Fig. 4.17b). When ratioed with regional plateau dust spectra, these spectra maintain a negative slope, some concavity is usually preserved, and the absorption minimum occurs between 1.1 and 1.15 μm (Fig. 4.17c). These spectral attributes are inconsistent with laboratory measurements of olivine or pyroxene, whose spectra would possess positive slopes and show the 1- μm absorption at a shorter wavelength (e.g., Figs. 4.18–4.19). Previous work by *Horgan and Bell* [2012a] attributed these spectral features to the presence of weathered Fe-bearing glass with thin (several μm) silica-enriched glassy rinds [*Minitti et al.*, 2007; *Horgan et al.*, 2011; *Salvatore et al.*, 2010]. Dunes whose spectra are categorized as weathered glass are also detected in Hebes Chasma and they collectively account for $\sim 15\%$ of spectral detections (Fig. 4.17; Chapter V).

CRISM spectra of eastern VM dune fields suggest even greater compositional diversity (Figs. 4.17a, 4.20). Previous CRISM studies [*Bishop et al.*, 2009; *Parente et al.*, 2011] of sulfate-bearing ILDs in western Juventae Chasma (Fig. 4.10b), informally designated mounds A–C [*Catling et al.*, 2006], identified HCP and olivine as common constituents of surrounding wall and floor outcrops. Spectra of the sand comprising barchanoid dunes atop mound C (Fig. 4.11c) display features consistent with olivine, whereas dunes on the western ILD flank possessed

spectra consistent with HCP (Chapter V) [Chojnacki *et al.*, 2012b]. In addition, some spectra of dunes adjacent to mound C show evidence of absorption features at ~ 1.96 and ~ 2.1 μm attributable to hydration (Chapter V) and most readily associated with monohydrated sulfates (i.e., kieserite), a common constituent of ILDs [Bishop *et al.*, 2009; Parente *et al.*, 2011]. East of these locations (Fig. 4.7d), spectra of dunes possess broad and deep absorption features from 1.1 to 1.3 μm that are attributed to Fe-enriched olivine (Fig. 4.19a). Spectra of the erg interior sand sheets are more consistent with a basaltic composition (e.g., pyroxene and olivine; Fig. 4.17a). The relatively rare detection of LCP-bearing dune sand within VM is found with small barchan dunes on the eastern margin of the erg (Figs. 4.10b, 4.17a; Chapter V).

Spectra of Ganges Chasma sand dunes also suggest compositional heterogeneity with evidence of olivine, LCP, HCP, glass, various combinations (i.e., basalt), and hydrated minerals (Figs. 4.19b-4.19c, 4.20e, 4.20g; Chapter V). Intra-dune field variations in sand composition are also identified in several locations. For instance, notable high values in pyroxene (VAR) and olivine (OLINDEX) indices occur in different locations within the dune field (Fig. 4.19b); the corresponding spectra match well with laboratory measurements of intermediate pyroxene and olivine- and glass-bearing basalt spectra. Spectra of other sand sheets in north central Ganges Chasma show features consistent with basaltic Fe-bearing glass (Figs. 4.17c, 4.19c). However, several small barchan dunes can be found ~ 800 m in the inferred upwind direction with a different composition; here, a prominent absorption band in spectra at ~ 2.3 μm suggests a more HCP-rich component to these dunes (Fig. 4.19c).

Finally, Capri Chasma hosts seven small (~ 10 -80 km^2) floor dune fields, primarily in the central portion of the chasm (Figs. 4.1b, 4.10d). Most of these dunes are located on the north flanks of Capri Mensa, previously identified to be partially composed of water-altered minerals

(e.g., sulfates, hematite, and opaline silicates) [Roach *et al.*, 2010a]. These dunes display diagnostic spectral absorption features at ~ 1.9 , ~ 2.1 , and $2.4\ \mu\text{m}$, attributed to monohydrated sulfates, particularly kieserite (Fig. 4.17c; Chapter V). Eight percent of the dunes in VM give spectral evidence of sulfates.

4.2.5 Aeolian Activity

Dunes and dune fields in VM displayed evidence for six types of aeolian surface change (Figs. 4.9–4.10, yellow stars, Table 4.9). In rough order of increasing spatial scale and of presentation in the following text, these changes are: ripple migration/re-alignment, dune perimeter modifications, slip face modifications (e.g., development of alcoves), dune migration, dune deflation, and dune field bolometric albedo change (darkening and lighting). Repeated coverage with HiRISE images offers the opportunity for meter-scale change detection. Surface changes associated with several Ius Chasma dune and ripple surfaces are shown in Fig. 4.21a. Dark-toned ripples showed local realignment (Fig. 4.21b, top left bar in the two images; also see Attachment 4.1), sometimes by as much as $\sim 90^\circ$, over the course of two martian seasons (Table 4.9); other ripples showed no change in orientation (bottom right). Additional small ($10\text{--}20\ \text{m}^2$) occurrences of sand movement (both deposition and erosion) can be found located along the dune perimeter edge of the sand sheet where darker-toned sand overrides lighter-tone sedimentary deposits that provided adequate contrast (Figs. 4.21a–4.21b, black arrows; Attachment 4.2).

High-resolution monitoring of dune fields in Ius, Capri, Ganges, and Juventae chasmata shows indications for slip face movement or adjustment. As an example, a slip face edge of a Juventae Chasma dune atop mound C in 2006 and 2010 is shown in Figs. 4.21c–4.20d. The second time-step shows three prominent $\sim 20\ \text{m}$ -wide alcoves (Fig. 4.21d) paired with three

slightly wider fans near the base of the slip face. This observation is interpreted to be evidence for relatively large grain flow events that occurred following slip face avalanches, similar to features described elsewhere on Mars [*Horgan et al.*, 2012a]. Assuming a mean thickness of 2 meters, measurements of these alcoves suggest $\sim 200 \text{ m}^3$ of sand was displaced. In addition, there is evidence for dune slip face retreat and grain flow events in Ius and Capri chasmata, respectively (see Table 4.9). These changes are similar in length scale and time interval to those observed on aeolian bedforms in other mid- to low-latitude (often intracrater) locations [*Silvestro et al.*, 2010b, 2011; *Bridges et al.*, 2012b].

At a larger scale, barchan dune deflation was inferred in Ganges Chasma between 2005 and 2008 using MOC-HIRISE images (Figs. 4.22d–4.22f; Table 4.9) [*Chojnacki et al.*, 2011b]. That work estimated that one small ($\sim 5000 \text{ m}^2$) barchan decreased in area $\sim 20\%$ over the ~ 1.6 Mars Year gap between images. This dune field was examined for more evidence of surface change using repeat HiRISE observations. Several occurrences (Figs. 4.22a, white triangles) of superposed dark-toned ripples show movement and migration relative to canyon floor immobile elements (Figs. 4.22b–4.22c; Attachment 4.3). One of the smaller barchans showed advancement up to one full TAR wavelength ($\sim 11 \text{ m}$) over the ~ 2 Mars Years between images, which is comparable with other migrating dunes detected on Mars [*Bridges et al.*, 2012b]. Here, as with other studies, TARs do not show evidence for surface movement [*Bridges et al.*, 2012b]; this lack of detected TAR movement, along with their common stratigraphic position below dunes, contributes to the widely-held view that they are immobile for time periods far exceeding the lifetimes of spacecraft missions [*Golombek et al.*, 2010; *Bridges et al.*, 2012b]. Previously unexposed TARs in the first image are found in the second image (Figs. 4.22f–4.22g; Attachment 4.4); other downwind TARs become covered in dark-toned dune sand.

Dune fields are also found with notable changes in TES albedo over larger areas and longer time interval (Figs. 4.8b–4.8c; Attachment 4.5; Table 4.7). Typically (~45%) VM dune fields decreased in albedo between MY 24 and 25 (e.g., from 0.13 to 0.11, -25%), possibly linked to dust removal events. Although visible-wavelength image coverage during this time frame was not adequate for dune change detection analysis, active dust devils and their tracks were common [Malin and Edgett, 2001]. At a larger scale, annual recurring cloud trails form in multiple locations across the rift, and were interpreted as high-velocity thermally driven updrafts formed along VM canyon walls composed of dust and water ice particles [Clancy *et al.*, 2009]. This unique phenomenon may be responsible for major dust lifting events and large albedo changes in VM. The TES albedo of VM dune fields typically (~55%) increased (e.g., from 0.11 to 0.14, +40%) between MY 25 and 26 (Figs. 4.8c; Attachment 4.5; Table 4.7), a time frame of “brightening” that coincides with the 2001 global dust storm.

4.2.6 Aeolian Morphology Related to Topography and Modeled Winds

Slip face mapping of Valles Marineris dunes provides insight into paleowinds regimes, which can be compared with local topography and wind vectors predicted by atmospheric models. In the simplest cases, slip faces of many dunes within a given field tend to conform to a dominant trending direction. For instance, a floor dune field in south Coprates Chasma (Figs. 4.9c, 4.11b) is found with all dune slip faces oriented in a similar eastward direction (parallel to the canyon orientation). Climbing and falling dunes also possess morphologies that suggest unidirectional wind regimes (Fig. 4.13). Exceptions to these trends can be found with dune fields in more irregularly shaped chasmata with multiple canyon reentrants and large central mensae. For instance, east Ganges Chasma sand streak and dune slip face orientations imply primary easterly winds are overprinted by a secondary northerly wind regime, the latter being

consistent with the direction of dune migration (Figs. 4.10d, 4.22). This example might be evidence of the types of converging flow regimes that have dictated individual dune orientations elsewhere [*Fenton and Beyer, 2012; Fenton et al., 2012*].

Regional trends can be investigated by examining all the dune field mean slip face orientations in an individual chasm (Fig. 4.9–4.10, black arrows). For both of these chasmata with long, linear canyons, slip face orientations are dominantly orthogonal with the chasmata' orientation. Ius and Coprates chasmata have dune field mean slip face directions that generally imply longitudinal (up- or down-canyon) flow regimes, primarily to the west (Figs. 4.9a, 4.9c). In chasmata with tall central mensae (Hebes and Ganges chasmata), dune slip face directions are primarily diagonally away from the mensa (Figs. 4.9b, 4.10a, 4.10d). Taken together, the mean slip face orientations of all the dunes in Melas Chasma suggest a general southward migration of dunes. Capri Chasma appears to have more complex wind patterns, as individual dune slip faces are variable depending on proximity to ILDs or walls (Fig. 4.10c). Dune slip faces in Juventae Chasma also appear to be affected by their proximity to ILDs and chaotic material, but possess an overall southward-facing trend, toward the reentrant portion of the chasm (Fig. 4.10b).

Global circulation models do not resolve VM or its topographic influence on regional winds [*Haberle et al., 1993; Hayward et al., 2007a*]. Dune field DCA and dune slip face orientations to the north and south of VM are all suggestive of a northerly wind regime in agreement with GCM results [*Hayward et al., 2007a; Gardin et al., 2012*], but have very little correlation with those paleowind indicators within the more topographically complicated VM. However, as Melas Chasma was one of final Mars Exploration Rover (MER) candidate landing sites, two independent teams have performed mesoscale modeling of VM winds. *Rafkin and Michaels* [2003] and *Toigo and Richardson* [2003] describe complex wind regimes driven by the

Tharsis Plateau's effect on global tidal circulation combined with the production of extensive slope winds. In these models, katabatic winds drain into the chasmata, particularly the western half [Rafkin and Michaels, 2003; Toigo and Richardson, 2003]. Compounding the channeling of modeled atmospheric flow, the rift's west-east orientation amplified the tidal forcing of winds [Toigo and Richardson, 2003]. Model results indicate that topographic forcing of near-surface winds is stronger in VM than at the other MER candidate landing sites (e.g., Gusev crater, Meridiani). As an example of this topographic forcing, moderate morning drainage winds (~20 m/s) enter the chasmata perpendicular to canyon rims, controlled by spur and gully topography by following the local downhill gradient. These types of drainage winds were suggested to be responsible for the channeling of the dark-toned sands that form falling dunes, which are generally restricted to gullies [Chojnacki *et al.*, 2010]. In addition, the areas of barchan deflation and migration in east Ganges Chasma (Fig. 4.22; [Chojnacki *et al.*, 2011b]) are spatially coincident with zones of enhanced downslope wind flow and sediment flux (to the south), as estimated by mesoscale models [Fenton *et al.*, 2012]. Model predict that by afternoon or early evening moderate northerly winds (~30 m/s) will flow up southern walls in western VM [Rafkin and Michaels, 2003; Toigo and Richardson, 2003]. Some of these predicted upslope winds are spatially coincident with climbing dunes along the southern walls of Hebes and Melas chasmata (Figs. 4.11a, 4.13a). Mesoscale models also resolve well-defined topographic focusing of canyon floor winds caused by Geryon Montes in Ius Chasma (Fig. 4.9a, white arrows). This massif divides west-east canyons into two narrower canyons and produces a prominent Venturi effect as modeled atmospheric flow accelerates through the constrictions [Rafkin and Michaels, 2003] (section 5.2).

4.3 Statistical Analysis of Dune Field Populations

To test the hypothesis that the non-VM and VM dune fields were drawn from different populations, we performed some basic statistical tests on the following dune field variables: relief, slope, roughness (at all 3 baselines), albedo, nightside thermal inertia, ΔTI , DCI, TES 465, TES 530, and TES albedo changes (both MY 24–25 and MY 25–26 (Table 4.10)). For this task, the Mann–Whitney U test was employed, as it examines whether two independent samples have similar distributions. It is the non-parametric alternative to the two-sample T-test for the equality of two sample means. Prior to this test, it was determined that the variables listed above did not follow a normal distribution using the Kolmogorov-Smirnov test; thus, the T-test was not appropriate. The Mann–Whitney test is commonly used to examine temporal trends in terrestrial climate data and its use is well understood [Yue and Wang, 2002; Xu *et al.*, 2003; Mauget *et al.*, 2011]. This test determines the variables' data rankings, uses those ranking sets to calculate a Mann-Whitney unbiased (U) statistic, and then normalizes the U statistic into a standardized score (Z) statistic. In other words, it examines the differences between mean ranks and variances of the two samples and reports a probability (p-value) that the two samples were derived from the same population. At a given alpha level, one may reject the null-hypothesis that the samples were derived from the same population. Using this test with our chosen alpha level of 5%, it was determined that there is a 95% confidence level, usually by 4 to 17 orders of magnitude, for 11 of the 13 variables that the two samples were not drawn from the same population (Table 4.10). The TES ΔTI and 530 index variables were statistically separable, but at lower confidence levels and might have failed given a higher confidence level (e.g., $\alpha = 1\%$) or different pixel selection. Variables that failed this test were TES nightside thermal inertia and the TES 465

index (Figs. 4.6b, 4.7b), for which the VM and non-VM dune field values were not statistically separable using this method.

5.0 Discussion: Synthesis of Results and Implications

5.1 Valles Marineris Dune Fields and Their Geologic Settings

Geologic settings of VM dune fields and low- and mid-latitude non-VM dune fields were compared to address how geologic features not typically associated with craters (e.g., landslides, spur-and-gully wall material) may influence different dune populations. Depending on the chasmata, VM dune fields are spatially coincident with floors, layered deposits (Hebes, Melas, Juventae, Capri, Ganges), landslides (Hebes, Melas, Coprates), and/or walls (Coprates, Melas) (Fig. 4.9–4.14). Other than erg sand sheets, dune fields are rarely found on central floors without adjacent landforms (e.g., layered deposits, chaotic material). Dune fields in close association with layered deposits are not unique to VM – previous authors have reported intracrater dunes adjacent to layered deposits in other locations (e.g., Gale crater, Proctor crater) [Fenton *et al.*, 2003; Tirsch, 2009; Hobbs *et al.*, 2010]. However, dune fields superposed on the summits of large layered mounds (e.g., in Juventae, Melas chasmata) do appear to be unique to VM (e.g., Figs. 4.10b, 4.11c; Chapter V).

The landslide dune field class is contextually unusual. There are >500,000 km² of landslide area on Mars, primarily (~93%) outside of VM [Scott and Tanaka, 1986; Greeley and Guest, 1987]. Although many of these landslides are much larger than those in VM (e.g., Lycus Sulci, Arsia Sulci), no previous studies [Edgett and Malin, 2000b; Malin and Edgett, 2001; Hayward *et al.*, 2007a] have reported superposed dark-toned dune fields, nor did our analyses of MGD³ dune locations find them. Landslides are locations of high roughness [Kreslavsky and Head, 1999], where the irregular terrain is a likely factor in producing scattered, instead of

continuous, clusters of dunes [Greeley and Iversen, 1985; Pye and Tsoar, 1990]. Often these widely-spaced duneforms may develop around isolated roughness elements forming echo dunes or more irregular duneforms (Fig. 4.12). This clustering of duneforms can also be seen in some intercrater dunes (e.g., east of Hellas) [Fenton, 2005; Silvestro *et al.*, 2010a], but are not typical in major intracrater dune fields (e.g., Proctor, Rabe). It is likely that small, dark-toned bedforms are present on other martian landslides, but their presence is more pronounced in VM. Valles Marineris landslide dune fields are unlike most martian dune fields previously described, with regard to their context, frequency, and spatial extent, and broaden the context within which dune fields may be found elsewhere on Mars.

Climbing and falling dune fields of VM also contrast with dunes in more topographically subdued crater locations. The frequent “lobate” falling dune morphology, where horns are oriented uphill and in the inferred upwind direction [Chojnacki *et al.*, 2010], is unusual for dunes, as crescentic dunes of similar scale possess horns that point downwind [Bagnold, 1941]. Thus, when considering the inferred prevailing wind direction, the falling dune morphology is distinctive compared to simple crescentic dune morphology of most martian dunes. Climbing dunes within VM can also possess moderate slopes and reliefs (Figs. 4.11a, 4.13a–4.13b), but typically have less distinctive barchanoid morphologies. Minor occurrences of such duneforms are found superposed on trough, pits, or crater walls [Fenton *et al.*, 2003; Bourke *et al.*, 2004; Silvestro *et al.*, 2010a], so this dune class is not unique to VM, although more frequently reported there (Figs. 4.2c–4.2d). However, non-VM wall dunes are without the substantial slope and relief values that accompany VM wall dunes; the extreme topographic nature of VM wall dunes is a distinguishing factor from those found outside the rift. In addition to topographic

relief and wind regime (section 5.2), availability of sediment may be a key factor in making wall dunes more abundant in VM than elsewhere on Mars (Chapter V) [*Chojnacki et al.*, 2010].

The abundant co-occurrences of debris flow, fan, landslide, and spur-and-gully wall morphologies with dune fields, as described above for VM, suggest the interleaving of mass-wasting and aeolian processes. Sites in east Coprates, Melas, Ius, and Hebes chasmata show numerous locations where gravity-driven and aeolian processes have occurred in the same locations (Figs. 4.11a, 4.12c, 4.13). The orientations of low-order bedforms, crescentic scours (2–5-m-wide), and bright streaks (5–15-m-long) on the wall slopes above falling and floor dunes are consistent with downslope winds, parallel to the talus slide direction. The boulder tracks and occasional boulders on dune surfaces (e.g., Fig. 4.16) imply mass-wasting has occurred after bulk dune construction. Reports of any geologic materials other than airfall dust superposed on martian dune surfaces (e.g., craters, boulders) are exceedingly rare [*Malin and Edgett*, 2001; *Bourke et al.*, 2010]. Additional sites (e.g., Fig. 4.13c; Chapter V) [*Chojnacki et al.*, 2010] show similar morphologic evidence of dunes and lower-order aeolian bedforms juxtaposed with surfaces that were created by mass-wasting, suggesting these processes are interleaved to construct relatively distinct and variable morphologies (e.g., crescentic scours, ripples, dunes, talus, scree) by gravity and wind.

Terrestrial occurrences of sedimentary deposits called “sand ramps” formed from the combination of hill slope, aeolian, and fluvial processes have been described in the Mojave Desert as part of sand transport corridors [*Lancaster and Tchakerian*, 1996; *Bateman et al.*, 2012]. These Holocene-aged wedged-shaped sand ramps are found within gullies, on steep slopes (10–30°), and have a complicated stratigraphy of cross-bedded sand, fluvial, and talus horizons [*Lancaster and Tchakerian*, 1996; *Bateman et al.*, 2012]. Rarely recognized in the

terrestrial record, sand ramps require an active aeolian system to exist within high relief environments that are tectonically stable, and possess additional influxes of mass-wasting and/or fluvial runoff material. Wall dunes in VM (Figs. 4.11a, 4.13) appear analogous (morphologically, topographically, contextually) to these sand ramp deposits of the western United States, except without a known fluvial component. These observations support the hypothesis of broadly contemporaneous mass-wasting and aeolian processes have occurred in these locations.

5.2 Topographic Effects

The regional and/or local topographic settings may profoundly affect the morphology and geospatial distribution of martian dune fields. The topographic relief, slope, and roughness values of the two dune field populations and the terrains they are superposed on are clearly distinguishable from one another (Figs. 4.4–4.5; sections 4.1.1, 4.2.1, 4.3; Tables 4.6, 4.10). MOLA roughness values at the 0.6-km baseline, attributed to dune crest-to-crest topography [Kreslavsky and Head, 2000], are similar for both populations (Fig. 4.5). However, VM dune roughness data are considerably greater at the longer baselines and, based on morphologic inspection, relate to intradune rugged terrains (e.g., wall debris, chaotic material). The most important differences between the Valles Marineris rift vs. crater or basin topographic regimes are the deep, linear troughs vs. shallow, bowl-shaped structures. The sedimentary effect of the interaction of a flowing atmosphere with craters is dependent on crater diameter and depth [Greeley and Iversen, 1985], whereas for canyon-like structures it is dependent on the canyon depth, canyon floor width, and spacing of consecutive canyons (if more than one exists) [Wiggs *et al.*, 2002; Vardoulakis *et al.*, 2003; Ahmad *et al.*, 2005]. These factors related to canyon topography influence the geographic positions of the aeolian bedforms and their sand pathways.

A large factor when considering wind flow encountering linear canyon systems, is its angle of incidence, the angle at which the winds contact the rim. Terrestrial observations show that as wind flow over a plateau passes over a canyon rim orthogonally, it briefly decelerates downslope of the cliff edge (flow separation), reaches a minimum flow velocity in the center of the canyon, and accelerates momentarily on the opposite side of the valley when heading up the escarpment (streamline compression) [Greeley and Iversen, 1985; Pye and Tsoar, 1990; Wiggs *et al.*, 2002; Bourke *et al.*, 2004; Al-Enezi *et al.*, 2008]. These terrestrial observations are relevant for VM wall dunes; falling and climbing dunes, whose slip face orientations are dominantly orthogonal to the gradient of the canyon wall (Fig. 4.13), are located in the zones of predicted atmospheric acceleration. Modeled drainage winds (~20 m/s) [Rafkin and Michaels, 2003; Toigo and Richardson, 2003] entering Coprates Chasma from the south are channelized within gullies following the local downhill gradient, consistent with locations of falling dunes [Chojnacki *et al.*, 2010]. Local narrowing of topography (i.e., gullies) can cause streamline convergence [Spirn, 1986; Pye and Tsoar, 1990; Al-Enezi *et al.*, 2008] and lead to enhanced aeolian sediment transport. In addition, saltation path lengths are longer when winds align with the downhill gradient [Iversen and Rasmussen, 1994, 1999] resulting in enhanced transport. For northerly canyon bottom winds, which encounter the southern walls of Hebes, Melas, and Coprates chasmata, modeled atmospheric flow tends to momentarily increase [Rafkin and Michaels, 2003; Toigo and Richardson, 2003]. Climbing dunes located in these chasmata (Figs. 4.11a, 4.13a–4.13b) are located in the bottom-third of wall slopes and coincident with these predicted zones of enhanced atmospheric flow. There continued transport farther up walls is likely prohibited where topography steepens, gullies terminate, and/or saltation path lengths are reduced.

When near-surface atmospheric flow encounters linear topography at a parallel to sub-parallel orientation, topographic focusing of winds often follows [*Vardoulakis et al.*, 2003; *Ahmad et al.*, 2005]. Early aeolian studies recognized the common formation of sand drifts where atmospheric flow exits and decelerates from a topographic constriction with the resulting reduction in sediment carrying capacity [*Bagnold*, 1941]. The “Windy Gap” part of Ferris dune field, Wyoming, provides a larger scale example where airflow streamline convergence occurs because of the topographic constriction of a mountain pass; this constriction results in a two- to three-fold acceleration of airflow within in the gap and subsequent dune deposition on the downwind side following the decrease in airflow [*Pye and Tsoar*, 1990]. Massifs in Algeria and California have been found to divide (i.e., diverge) and/or focus (i.e., converge) airflow and resulting sand flow paths upwind and downwind (respectively), of massifs [*Greeley and Iversen*, 1985; *Pye and Tsoar*, 1990]. Evidence of this streamline convergence occurring near VM wall massifs or other canyon constrictions is observed in mesoscale modeling results [*Rafkin and Michaels*, 2003; *Toigo and Richardson*, 2003]. These types of constrictions include the low elevations around Geryon Montes and the east Coprates massif and are common locations for nearby floor dune fields (Figs. 4.9a, 4.9c, white arrows, 4.13c). Dunes in wider canyons without massifs are either associated with topographic elements, such as mensae or landslides (east Ius, Melas, and Capri chasmata) and chaotic material (north Juventae and east Ganges chasmata), or are part of ergs (south Juventae and west/central Ganges chasmata). These correlations between canyon massifs and dune field proximity may mimic the dune centroid azimuth relationship with crater-related dune fields [*Hayward et al.*, 2007a, 2009].

In contrast to widely-spaced dune fields adjacent to wall massifs, the topographically abrupt and deep (~8 km) Juventae and Ganges chasmata show continuous expanses of erg sand

sheets and dunes (Fig. 4.1b). Morphologic and thermophysical measurements in the chasmata indicate a significant volume of sand has filled these basins, more than is observed in craters (e.g., Proctor, Gale) [Fenton *et al.*, 2003; Hobbs *et al.*, 2010]. For instance, the ergs and dune fields of Ganges Chasma (Figs. 4.14b–4.14c) occupy nearly 70% of the chasms' interior area, not including Ganges Mensa, whereas dune fields in craters of comparable size (diameter ~150 km) typically only occupy 30% of the floor area [Hayward *et al.*, 2007b]. In addition, ergs $\geq 1000 \text{ km}^2$ occur more often within VM, whereas the terrain within our study latitudes did not have any occurrences in the 2000 km^2 or greater bin sizes (Fig. 4.3; Tables 4.4, 4.6). This type of extreme topographic forcing may contribute to the greater areal distribution and occurrence of large VM ergs as compared with low- and mid-latitude non-VM dune fields. To summarize, topographic data, terrestrial field studies, mesoscale modeling results, and geospatial relationships suggest Valles Marineris to be fundamentally different topographic and aeolian regime as compared with craters or basins, including polar-basins.

5.3 Dune Sand Grain Sizes

Inferred particle sizes of the different dune field populations were compared to assess if dunes were constructed from measurably different sand sizes. Thermal infrared measurements from TES indicate VM dune fields have greater thermal inertia and ΔTI values, on average, than other martian dune fields (Figs. 4.6b–4.6c; Tables 4.6, 4.10; sections 4.1.3, 4.2.3, 4.3). Except for four non-VM dunes measurements, TES thermal inertia values of dune areas correspond with very fine to coarse sand sizes ($\sim 500 \text{ }\mu\text{m}$ mean), assuming no surface induration. In comparison, TES thermal inertia values for dune fields within VM correspond with medium to very coarse sand ($\sim 900 \text{ }\mu\text{m}$ mean) using the same assumption. For the larger ΔTI values, particularly in VM, we interpreted these results to indicate enhanced thermophysical heterogeneity attributed to

mixing of variable sand sizes and, in some cases, proximity to the dune field's suspected sand source(s) (Chapter V).

For VM dunes with THEMIS thermal inertia values $\sim 270\text{--}330$ tiu we favor the interpretation of loose coarse grained sand grains, instead of induration because of evidence for adjacent sand sources (Chapter V) and the overall geologic context of these sites (e.g., close proximity of wall talus). For example, Coprates olivine-bearing dunes (Figs. 4.16a, 4.17) have high thermal inertias (~ 325 tiu or very-coarse sand), which are interpreted to be caused by the dunes' close proximity to their sediment source (i.e., wall spurs and talus) of the same composition (Chapter V) [Chojnacki *et al.*, 2012b]. Indurated dunes rendered inactive in this scenario would eventually become buried by wall debris. Results from mesoscale atmospheric models of VM [Rafkin and Michaels, 2003; Toigo and Richardson, 2003; Fenton *et al.*, 2012] imply near-surface winds capable of saltation for coarse sand sized particles (~ 25 m/s). For example, that near-surface wind velocity value would produce the wind threshold friction velocity ($u_{*t} \approx 2.8$ m/s) required for entraining very coarse sand sizes (~ 2000 μm) [Greeley *et al.*, 1980]. Creep of even larger particles is plausible if coarse sand grains are saltating. More mobile fine-sand sizes (~ 200 μm) would require threshold friction and surface winds nearly half of those values (i.e., ~ 1.5 m/s and 13 m/s, respectively). Those calculations assume basaltic sand grains (for density) at VM atmospheric pressure (~ 8 mb) and an aerodynamic roughness value of 3 cm [Sullivan *et al.*, 2005]. It is worth noting the detected coarse-grained population is not necessarily representative of the bulk of the dune volumetrically. Frequently in terrestrial and martian examples, the coarsest grains are concentrated in ripples on the stoss sides in the upper portions of a dune profile, whereas finer sand makes up the bulk of bedform [Bagnold, 1941; Pye and Tsoar, 1990; Sullivan *et al.*, 2008; Kocurek *et al.*, 2013].

Previous investigations using GCM results found evidence for enhanced aeolian erosion at lower elevations and higher pressures [Armstrong and Leovy, 2005]. To assess if the lower VM elevation (and the associated higher atmospheric pressure) might make a substantial difference in the winds' potential to saltate coarser sand, predicted relationships between particle diameter and friction velocity at various pressures are investigated [Iversen and White, 1982]. We calculate for the difference in atmospheric pressure (~ 4 mb), 25 m/s winds at the bottom of VM (at ~ 8 mb) have the same potential for initiating saltation of 2 mm grains as 35 m/s winds would on the plateau above the rift (at ~ 4 mb). Whereas 25 m/s winds are capable of moving 2 mm grains at the bottom of VM, an equivalent wind could only move grains up to 800 μm on the plateau. On the other hand, high-elevation surfaces ($>+10$ km, e.g., Tharsis Montes) with drastically lower pressures (~ 0.5 mb) than locations discussed above have occurrences of “reticulate bedforms” that are hypothesized to form from saltating dust aggregates as depositional features [Bridges *et al.*, 2010; Geissler *et al.*, 2013]. For the Tharsis Montes “bedforms”, where pressures are as low as ~ 0.5 mb, near surface-winds of 75 m/s ($u_{*t} \approx 10$ m/s) are required to move particles ~ 10 μm in diameter, which may be feasible based on atmospheric modeling [Bridges *et al.*, 2010]. However, these morphologies' basic formation mechanism (i.e., depositional vs. erosional) remains controversial.

In relation to those calculated results, we also found a moderate inverse-correlation between dune change detections and elevation within our study area latitudes; in other words, detections of dune change were more frequent at lower elevations and therefore higher pressures (section 4.1.5). In support of this finding, observations of dune activity in the low-elevation northern polar basins [Bridges *et al.*, 2012b; Geissler *et al.*, 2012]) suggest enhanced dune activity at lower elevations (higher pressures) and a greater relationship

5.4 Dune Sediment Compositional Heterogeneity

Sand, dust, and lithic fragments are the major components of low-albedo soils at five widely-separated landing sites [Herkenhoff *et al.*, 2008]. This study adopts the planetary use of *soil*, as “the uppermost, unconsolidated, weathered portion of the regolith, in contrast to the terrestrial usage that also specifies that soil must have an organic component” [McSween and Keil, 2000]. Low-albedo soils are the products of physical weathering of igneous minerals (e.g., pyroxene, plagioclase) with additional components formed from diagenesis and chemical weathering (e.g., sulfate, nanophase oxides, clays); impact comminution and cosmic ray irradiation also play a minor but persistent role in soil composition [McSween *et al.*, 2010; McGlynn, 2012]. Soil measurements at landing sites appear remarkably homogenized with respect to elemental chemistry and mineralogy [McSween *et al.*, 2009; McGlynn, 2012]. Moreover, soils have been widely regarded as the products of a global mechanism in which bolide impacts and aeolian processes (e.g., transport, sorting) are partially responsible for their homogenization [Yen *et al.*, 2005; Brückner *et al.*, 2008; McGlynn *et al.*, 2011; McSween *et al.*, 2010]. For example, crater rays identified in thermal infrared data show ejecta can extend over 1000 kilometers from the impact site [Tornabene *et al.*, 2006]. Dust particles (<60 μm) are a small component of most martian soils and are known to be near uniform in composition [McSween and Keil, 2000; Ruff and Christensen, 2002; Christensen *et al.*, 2003; Wolff *et al.*, 2009]. The mechanism by which the dust is homogenized is likely global-encircling dust storms occurring every few Mars Years [Geissler, 2005; Cantor *et al.*, 2006].

Martian sand has been suggested to be relatively homogenous, with respect to particle size and composition, by eons of saltation and derivation from a uniform crustal component [Soderblom and Bell, 2008]. But density and particle size fractionation can arise from variations

in settling velocities, grain entrainment thresholds, and grain transport velocities [Hughes *et al.*, 2000], and can profoundly affect compositional variability of sediment. Additionally, local (10s of kilometers) derivation of sediment may contribute to sand compositional heterogeneity (Chapter V). Emerging results concerning dune sand compositions and thermal inertias (see section 5.4) suggest not all dune sand is homogenous. As in earlier studies [Mustard *et al.*, 2005; Poulet *et al.*, 2008b; Tirsch *et al.*, 2011], our CRISM results indicate pyroxene (particularly HCP) is the most commonly detected sand mineralogy, but did resolve subtle geographic variations in mafic compositions over several scales. For example, results in Meridiani Planum (Fig. 4.7d) display variability over the regional scale (100s of kilometers), whereas Gale crater dune sand compositions (olivine and HCP) vary over shorter local distances (10s of kilometers). This latter example of a moderate size aeolian system with ample data coverage may suggest observational bias (i.e., data availability) to be a factor in detection of sand mineralogical diversity.

Studies of high-latitude dunes, known to be partially composed of alteration products (e.g., gypsum, weathered glasses), clearly contradict any notion that martian sand has been globally homogenized [Langevin *et al.*, 2005; Horgan *et al.*, 2009; Horgan and Bell, 2012b]. Although no clear evidence of alteration products was found in our analysis of non-VM dune fields, preliminary results using OMEGA and TES data suggest dunes partially composed of weathered glass extend into the lower latitudes [Horgan *et al.*, 2012b, 2012c], suggesting greater compositional variability than is implied by Fig. 4.7.

Whereas our results described above show subtle spatial variations in basaltic sand compositions of intracrater dunes, analysis of VM dune spectra reveals substantial compositional heterogeneity at several scales. At the basinal scale, four mineral classes were detected across

different dune surfaces in Juventae and Ganges chasmata (Fig. 4.17a; olivine, HCP, LCP, sulfate). Sometimes the spectrally dominant mineral detected in one dune field was different from the dominant mineral in another area a few kilometers away. For example, two western Juventae Chasma dune fields in proximity to one another (~10 km) possess different spectral signatures (i.e., olivine and HCP), whereas erg sand spectra on the eastern margin show evidence of LCP (Fig. 4.17a, Chapter V). At smaller scales, adjacent duneforms (~1 km) in Ganges or Coprates chasmata show different spectrally dominant phases (Figs. 4.17a, 4.19c) and provide evidence for aeolian compositional fractionation, attributed to variable mineral densities and the sorting capacity of the wind (Chapter V). These examples of spectrally distinct dune fields and duneforms in proximity to one another argue for discrete, relatively unmixed sediment sources, counter to the suggestion that martian soils and (by extension) dark sands have been globally homogenized.

5.5 Timing of Dune Field Formation

Motivating questions for this work include: when did the VM dune fields form and is dune formation a geologically recent or ancient process? *Absolute* age dating of dune formation is not possible using orbital measurements. Although crater counting has been used to estimate many planetary surfaces, unconsolidated martian aeolian bedforms do not lend themselves to this technique because of their inability to retain craters [Bourke *et al.*, 2010]. However, some age constraints can be made based on relative age relationships using crater counts on adjacent terrain, geologic context, thermophysical measurements, and possible origins of regional layered deposits. Valles Marineris dune fields are located on relatively young geologic terrains. The high proportion of VM dune fields superposed on Amazonian units (49%) is not solely due to a greater proportion of these younger units within VM, since only ~25% of the below-the-rim units

are Amazonian [*Scott and Tanaka*, 1986]. Conversely, most (99%) of low- and mid-latitude non-VM dunes are located on Noachian or Hesperian units, despite those units occupying only about two-thirds of the area between 50°N to 50°S (Table 4.6). Although the Amazonian Epoch spans ~3 billion years, VM landslides with superposed dune fields are dated at 150 Myr–1 Gyr [*Quantin et al.*, 2004]. These age constraints on underlying landslide surfaces suggest these dune fields (1) were formed and subsequently migrated onto these landslides and/or (2) were derived from local landslide-related rock and sediment. Either of these options could have occurred any time after slide emplacement. Evidence of an off-slide sand source is found in Ius, Coprates, and Melas chasmata, where dunes extending up to 30 kilometers are found straddling landslide-floor contacts (e.g., Figs. 4.12a, 4.12c). In these cases, dune field migration must have occurred on to or off of the slide, but these dunes' formation period(s) are poorly constrained to some time after the date of the landslide they are now superposed on. As a potential example of sand derivation from the slide itself, one east Ius Chasma dune field has a similar CRISM-derived composition as adjacent eroding landslide scarps 1 to 20 kilometers upslope (i.e., HCP; see Chapter V); the landslide on which these dunes are now superposed was dated at 150 My [*Quantin et al.*, 2004]. This evidence, along with topographic and morphologic analyses, supports the local derivation of dune sediment from adjacent landslide scarps and by extension their formation in the last 150 My. From these lines of reasoning, we suggest at least some VM sand dunes formed and were actively migrating >10s of kilometers during the Late Amazonian period (150 Myr–1 Gyr).

Induration of martian surfaces provides a means of bedform stabilization, which might preserve duneforms that formed under a different climate. Based on photogeologic evidence, several authors have identified martian examples of indurated duneforms that range in degree of

inferred induration [Edgett and Malin, 2000a; Malin and Edgett, 2001; Fenton and Hayward, 2010]. *In situ* evidence for bedform induration was detected at the El Dorado ripple field in Gusev crater, where surface crusts were found along the sides of trenches dug by the MER Spirit rover [Sullivan *et al.*, 2008]. Surface induration of regolith, including sand, is thought to be a global process possibly linked to slow chemical weathering and/or salt formation [Moore *et al.*, 1999; Jakosky *et al.*, 2006; Sullivan *et al.*, 2008]. The thermophysical properties of dunes also offer some insight into their possible generations; for example, certain dunes and draas in Ius and Juventae chasmata (Figs. 4.15a, 4.16e, 4.16g; sections 4.2.3, 5.4; Table 4.8) possess THEMIS thermal inertia values (350–400 tiu) consistent with surface induration and/or particles larger than very coarse sand (>2 mm). Photogeologic inspection using HiRISE indicates elevated values are not due to intradune objects (e.g., outcrops, boulders). Ius Chasma (Fig. 4.15a, ~ 350 tiu) may offer the best candidate for an indurated, giant draa that has stopped undergoing bulk horizontal migration and remains as an example of a past generation of bedform. In this and other examples in Juventae Chasma (Figs. 4.16e, 4.16g, ~ 400 tiu), proposed indurated dune surfaces are far from wall or other possible sources of contamination of particles larger than sand by mass wasting. Thus, bedform induration appears more plausible than dunes that are constructed of pebble-size particles (2–64 mm), which would require unrealistic wind speeds [Greeley *et al.*, 1980]. Assuming the conditions presented in section 5.3, a threshold friction velocity of ~ 4 m/s capable of entraining particle sizes ~ 3 mm in diameter [Iversen and White, 1982; Greeley and Iversen, 1985] would require near-surface (~ 1 m) wind velocities in excess of 35 m/s. Results from mesoscale atmospheric models of VM suggest such winds are present but at several 100s of meters above canyon walls, whereas the weakest winds occur at lower altitudes and central canyon floors where these draas are located [Rafkin and Michaels, 2003;

Toigo and Richardson, 2003]. Thus, these high thermal inertia values observed of VM dunes may be indicative of indurated (fossilized) dune surfaces and may be examples of bedforms that are in an intermediate state of induration prior to large-scale lithification, burial, or deflation, as inferred elsewhere on Mars [*Edgett and Malin, 2000a; Malin and Edgett, 2001; Fenton and Hayward, 2010*].

Aeolian processes have persisted for the majority of Mars' history. As on Earth, generations of dune fields likely formed, persisted for a time, were subsequently buried and locked in the geologic record (e.g., as sandstone), exhumed by erosional processes, and potentially served as sediment sources for other bedforms [*Kocurek and Lancaster, 1999; Muhs, 2004*]. Hesperian-aged ILDs are widespread across the rift, often with adjacent duneforms (e.g., Fig. 4.9–4.10, 4.11a, 4.12b, 14). Layered deposit genesis is largely debated, but may have involved aeolian deposition followed by cementation and diagenesis by aqueous-related processes [*Nedell et al., 1987; Malin and Edgett, 2000; Lucchitta et al., 1992; Roach et al., 2010a; Murchie et al., 2009a, 2009b; Chojnacki and Hynek, 2008*]. An aeolian-related sedimentary petrogenesis of layered deposits is supported by MER observations at Meridiani Planum where the stratigraphy at Endurance and Victoria craters is consistent with thick stacks of complex dune morphologies (i.e., sand sheets, crescentic dunes) that were constructed under multiple wind regimes and changing climates [*Grotzinger et al., 2005; Hayes et al., 2011*]. These Hesperian-aged sulfate-bearing sandstones (i.e., layered deposits) are largely interpreted to be ancient aeolian sand sheets and dunes that were cemented by repeated fluctuations in the groundwater table [*Squyres et al., 2004; Grotzinger et al., 2005; Arvidson et al., 2006; Andrews-Hanna et al., 2010*]. An aeolian origin of the VM layered deposits is consistent with the observed thinly-bedded, mid-toned, flat-lying strata consisting of moderate thermal inertias,

friable materials, some of which show cross-bedding [Malin and Edgett, 2000; Catling *et al.*, 2006; Chojnacki and Hynek, 2008; Fergason *et al.*, 2006a; Edgett and Malin, 2000a]. Dunes in Capri, Juventae, and Ganges chasmata are partially composed of monohydrated-sulfates (Fig. 4.17), which appear to be sourced from adjacent ILD-derived talus of the same composition (Chapter V). This evidence supports the hypothesis that these ILDs are partially composed of ancient aeolian deposits acting today as stored sand supplies (Chapter V). Though these observations of layered deposits are far from diagnostic, it seems plausible, if not probable, that at some time in VM's early history aeolian dunes formed and were eventually entombed as sedimentary rock resulting from changes in diagenetic alteration, wind regime, climate, and/or from burial by mass-wasting. Thus, in response to the question as to whether dune formation in VM is a geologically recent (e.g., landslide-related dunes) or ancient (e.g., layered deposits) process, we suggest there is evidence for both.

5.6 Present-Day Activity of VM and Non-VM dunes

To assess if the rift setting shows a greater propensity for observable dune change than others less topographically extreme environments, we compared our results with other studies (sections 4.1.5, 4.2.5). Our findings show VM dune morphologies and dune field surface albedos are actively changing at the ~1-m and ~3–6-km scales, which correspond to the spatial resolutions of the HiRISE and TES instruments (Figs. 4.8, 4.21–4.22; Table 4.9). That the TES bolometric albedos of VM dune fields vary significantly from Mars Year to Mars Year (–25% – +40%) over large areas (10s–1000s km²) suggests changes in dust coverage (Figs. 4.8b–4.8c; Table 4.7). The TES DCI values of the majority of dune fields within our wider study area (50° N to 50° S) suggest they are not substantially mantled in dust and, in view of likely dust deposition via episodic dust storms [Kieffer *et al.*, 1992; Cantor, 2007], indicate that dust-

cleaning events are occurring. Dunes that substantially decreased in albedo may be linked to dust devils or larger thermally driven updrafts [Greeley *et al.*, 2006; Clancy *et al.*, 2009]. For example, frequent occurrences of dust devils have been observed on canyon floors by past authors [Malin and Edgett, 2001; Cantor *et al.*, 2006] and in unpublished images (e.g., MOC M0000030, MOC2230). In addition, dust may be put into suspension from impacts with salting sand grains because of wind shear forces or the vibrations caused by particle saltation. Low-albedo regions, such as Syrtis Major, have been observed to be brighter following global dust storms only to return to pre-storm albedo values later, supporting the notion of dust deposition and subsequent lifting/removing at large scales [Christensen, 1988a]. On the other hand, dune fields in central Capri Chasma that decreased in TES albedo and showed morphologic evidence for surface change (Tables 4.7, 4.9) are spatially coincident with several repeated cloud trails [Clancy *et al.*, 2009], a aeolian phenomenon not described to occur elsewhere on Mars.

The time period between MY 25 and 26, during which VM and many non-VM dune fields increased in albedo (Figs. 4.8b–4.8c; Table TT7), includes the 2001 global dust storm (at the middle of MY 25). This event consisted of a series of causally-linked regional and local storms, some of which were observed to have interacted substantially with VM [Smith *et al.*, 2002; Cantor, 2007]. During this storm, multiple regional dust storm tracks intersected into VM (e.g., Syria-Claritas storm system) where variable altitude dust clouds were observed to obscure VM entirely [Cantor, 2007]. We suggest this storm to be the cause for the increased albedos of many low-latitude dune fields between MY 25 and 26.

Although mean values and ranges of albedo change were similar for the VM and non-VM dunes (Table 4.6), the two populations were statically separable (Table 4.10). Additionally, modal values for the VM dunes were more extreme and qualitatively more uniform than for

many other regions' dunes (Figs. 4.8b–4.8c, insets). In comparison, several non-VM dune fields in proximity to one another (<100 km), which are superposed on rugged terrain west of Argyre and Hellas, showed opposite albedo changes from one another (>+20% and <-20%). This evidence suggests that factors like the local micro-topography (i.e., sub-kilometer) and wind regime, rather than the settling of a regional-wide blanket of dust, can affect how a dune field's albedo evolves. The net albedo change for all dunes over the three time periods (MY 24–25, MY 25–26, and MY 24–26) is ~0%; these trends suggest equilibrium is reached where dust removed from one location is balanced by dust deposition on another.

Our analysis of geomorphologic aeolian activity reveals changes comparable to those found in other dune change studies (e.g., Fig. 4.8a, Table 4.6) [*Bridges et al.*, 2012b; and references therein]. Inferred modifications from aeolian activity include: dune deflation, dune migration, slip face modification (e.g., alcoves), and ripple modification or migration, at varying scales (10s–100s m²). Consistent with earlier findings [*Silvestro et al.*, 2010b; *Chojnacki et al.*, 2011a], most bedform modifications occurred on the dune field's inferred upwind direction and in the mean slip face directions, suggesting current effective wind patterns are similar to past dune-building winds. Barchan dune migration rates in Ganges Chasma (Fig. 4.22) averaged to ~5 m per Mars Year, and thus are slightly higher than average when compared with global studies of migrating dunes (~3 m per Mars Year) [*Bridges et al.*, 2012]. In addition, our observations are consistent with mesoscale wind modeling results [*Rafkin and Michaels*, 2003; *Toigo and Richardson*, 2003; *Fenton et al.*, 2012], which produce winds capable of saltating sand-sized particles and entraining dust particles [*Greeley et al.*, 1980].

Of the 12 repeat observations examined, six pairs showed evidence for VM dune modification; *Bridges et al.* [2012] also had two VM dune change null detections (totaling to

~43% of the number of repeat images examined collectively). That study also found 20 of 57 pairs outside of the rift with definitive sand movement (~35%). However, the statistics of small numbers do not allow definitive conclusions to be drawn. Results suggest Valles Marineris dune morphologic activity is not substantially different from intracrater locations in the study area.

6.0 Assessment of Hypothesis and Conclusions

We have examined all dune fields within 50°N–50°S latitude, in order to compare regional variations of dune field properties and characterize the distinct aeolian system found in VM. Detailed geomorphic, spectroscopic, and thermophysical mapping reveal subtle to drastic differences in the two environments and dune field properties. Our hypothesis is that VM dune fields and their properties are distinct from other low- and mid-latitude dune fields. Qualitative and quantitative characterizations of the two dune populations (summarized in Table 4.6) support this hypothesis. Most parameters were statistically separable using the Mann–Whitney test and inconsistent with the two populations being drawn from the same distribution. Thus we suggest conditions in the rift were favorable for yielding morphologically, compositionally, and thermophysically distinct dune fields from other regions.

We view these differences in dune field characteristics and properties to be directly related to the unique geologic, topographic, and atmospheric intra-rift setting. Moreover, these distinctive dune population characteristics (e.g., geologic contexts, composition, topographic and thermophysical properties) imply dune field properties and aeolian processes on Mars can be largely influenced by regional environment, which might have their own distinctive set of boundary conditions, rather than a globally homogenous collection of aeolian sediment and bedforms. Aside from the broader perspective these examples provide to planetary aeolian studies, other notable results include:

- Ergs in Juventae and Ganges chasmata constitute the largest non-polar dune fields on Mars, and indicate these systems had/have a persistent and plentiful sediment supply.
- A previously unreported class of dune fields is found superposed on 50 Myr–1 Gyr old VM landslides. These occurrences provide constraints on dune field-formation timing and may imply local sourcing.
- Many locations within Coprates Chasma show evidence for coupled mass-wasting and aeolian processes resulting in morphologies that are analogous to sedimentary sand ramps of the Mojave Desert.
- Diversity of detected dune sand compositions, including unaltered ultramafic minerals and alteration products, is more pronounced in VM. Previously unreported occurrences of dunes bearing monohydrated sulfate and weathered Fe-glass are described outside of the polar regions and broaden the mineralogical context of aeolian sediment.
- Compositional heterogeneity of duneform sand at basin-scales suggests local sourcing, whereas heterogeneity at dune-scales implies aeolian fractionation.
- High thermal inertia values (>350 tiu) of large draa and smaller dune surfaces are consistent with surface induration, which have likely rendered these bedforms inactive.

Thus, these points suggest great diversity in martian aeolian features as well as the intra-rift environments of Valles Marineris. Continued work in this region will investigate possible dune sediment sources, including ILDs, spur-and-gully walls, landslides, and extra-rift sources (Chapter V).

References

- Aben, L. K. (2003), Compositional and thermophysical analysis of Martian aeolian dunes, M.S. Thesis, Ariz. State Univ, Tempe, AZ.
- Ahmad, K., M. Khare, and K. K. Chaudhry (2005), Wind tunnel simulation studies on dispersion at urban street canyons and intersections—a review, *Journal of Wind Engineering and Industrial Aerodynamics*, 93(9), 697–717, doi:10.1016/j.jweia.2005.04.002.
- Anderson, J. A., S. C. Sides, T. L. Sucharski, and K. J. Becker (2004), Modernization of the Integrated Software for Imagers and Spectrometers, p. Abstract #2039.
- Andrews-Hanna, J., M. Zuber, R. Arvidson, and S. Wiseman (2010), Early Mars hydrology: Meridiani playa deposits and the sedimentary record of Arabia Terra, *J. Geophys. Res.*, 115, doi:10.1029/2009JE003485.
- Andrews-Hanna, J. C., R. J. Phillips, and M. T. Zuber (2007), Meridiani Planum and the global hydrology of Mars, *Nature*, 446(7132), 163–166, doi:10.1038/nature05594.
- Armstrong, J. C., and C. B. Leovy (2005), Long term wind erosion on Mars, *Icarus*, 176(1), 57–74, doi:10.1016/j.icarus.2005.01.005.
- Arvidson, R. E. (1974), Wind-blown streaks, splotches, and associated craters on Mars: Statistical analysis of Mariner 9 photographs, *Icarus*, 21(1), 12–27, doi:10.1016/0019-1035(74)90086-4.
- Arvidson, R. E. et al. (2006), Nature and origin of the hematite-bearing plains of Terra Meridiani based on analyses of orbital and Mars Exploration rover data sets, *J. Geophys. Res.*, 111(E12), doi:10.1029/2006JE002728.
- Bagnold, R. A. (1941), *The Physics of Blown Sand and Desert Dunes*, Methuen, London.
- Bandfield, J. L. (2002a), Global mineral distributions on Mars, *J. Geophys. Res.*, 107(E6), doi:10.1029/2001JE001510.
- Bandfield, J. L. (2002b), Spectroscopic study of the Moses Lake dune field, Washington: Determination of compositional distributions and source lithologies, *J. Geophys. Res.*, 107(E11), doi:10.1029/2000JE001469.
- Bandfield, J. L. (2009), Effects of surface roughness and graybody emissivity on martian thermal infrared spectra, *Icarus*, 202(2), 414–428, doi:10.1016/j.icarus.2009.03.031.
- Bandfield, J. L., V. E. Hamilton, and P. R. Christensen (2000), A Global View of Martian Surface Compositions from MGS-TES, *Science*, 287(5458), 1626–1630, doi:10.1126/science.287.5458.1626.
- Bandfield, J. L., A. Deanne Rogers, and C. S. Edwards (2011), The role of aqueous alteration in the formation of martian soils, *Icarus*, 211(1), 157–171, doi:10.1016/j.icarus.2010.08.028.
- Bandfield, J. L., C. S. Edwards, D. R. Montgomery, and B. D. Brand (2013), The dual nature of the martian crust: Young lavas and old clastic materials, *Icarus*, 222(1), 188–199, doi:10.1016/j.icarus.2012.10.023.
- Baratoux, D., N. Mangold, O. Arnauds, J.-M. Bardintzeff, B. Platevoët, M. Grégoire, and P. Pinet (2011), Volcanic sands of Iceland - Diverse origins of aeolian sand deposits revealed at Dyngjúsandur and Lambahraun, *Earth Surface Processes and Landforms*, 36(13), 1789–1808, doi:10.1002/esp.2201.
- Bateman, M. D., R. G. Bryant, I. D. L. Foster, I. Livingstone, and A. J. Parsons (2012), On the formation of sand ramps: A case study from the Mojave Desert, *Geomorphology*, 161–162, 93–109, doi:10.1016/j.geomorph.2012.04.004.

- Beyer, R. A., and A. S. McEwen (2005), Layering stratigraphy of eastern Coprates and northern Capri Chasmata, Mars, *Icarus*, 179(1), 1–23, doi:10.1016/j.icarus.2005.06.014.
- Bibring, J. P., and Y. Langevin (2008), Mineralogy of the Martian surface from Mars Express OMEGA observations, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. F. I. Bell, pp. 153–168, Cambridge University Press, Cambridge.
- Bibring, J. P., A. Soufflot, and J. Mustard (2004), OMEGA: Observatoire pour la Mineralogie, l'Eau, les Glaces, et l'Activite, ESA Special Publication, 1240, 37–49.
- Bibring, J. P. et al. (2005), Mars surface diversity as revealed by the OMEGA/Mars Express observations, *Science*, 307(5715), 1576–1581, doi:10.1126/Science.1108806.
- Bishop, J. L. et al. (2009), Mineralogy of Juventae Chasma: Sulfates in the light-toned mounds, mafic minerals in the bedrock, and hydrated silica and hydroxylated ferric sulfate on the plateau, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003352.
- Bishop, M. A. (2011), Aeolian scours as putative signatures of wind erosion and sediment transport direction on Mars, *Geomorphology*, 125(4), 569–574, doi:10.1016/j.geomorph.2010.10.029.
- Blasius, K. R., J. A. Cutts, J. E. Guest, and H. Masursky (1977), Geology of the Valles Marineris: First Analysis of Imaging From the Viking 1 Orbiter Primary Mission, *J. Geophys. Res.*, 82(28), 4067–4091, doi:10.1029/JS082i028p04067.
- Bourke, M., S. A. Wilson, and J. R. Zimbelman (2003), The variability of transverse aeolian ridges in troughs on Mars, vol. XXXIV, p. Abstract 2090.
- Bourke, M., M. Balme, R. Beyer, K. Williams, and J. Zimbelman (2006), A comparison of methods used to estimate the height of sand dunes on Mars, *Geomorphology*, 81(3-4), 440–452, doi:10.1016/j.geomorph.2006.04.023.
- Bourke, M., K. Edgett, and B. Cantor (2008), Recent aeolian dune change on Mars, *Geomorphology*, 94(1-2), 247–255, doi:10.1016/j.geomorph.2007.05.012.
- Bourke, M. C., J. E. Bullard, and O. S. Barnouin-Jha (2004), Aeolian sediment transport pathways and aerodynamics at troughs on Mars, *J. Geophys. Res.*, 109(E7), doi:10.1029/2003JE002155.
- Bourke, M. C., N. Lancaster, L. K. Fenton, E. J. R. Parteli, J. R. Zimbelman, and J. Radebaugh (2010), Extraterrestrial dunes: An introduction to the special issue on planetary dune systems, *Geomorphology*, 121, 1–14, doi:10.1016/j.geomorph.2010.04.007.
- Breed, C. S., M. J. Grolier, and J. F. McCauley (1979), Morphology and Distribution of Common “Sand” Dunes on Mars: Comparison With the Earth, *J. Geophys. Res.*, 84(B14), 8183–8204, doi:10.1029/JB084iB14p08183.
- Bridges, N. T., P. E. Geissler, A. S. McEwen, B. J. Thomson, F. C. Chuang, K. E. Herkenhoff, L. P. Keszthelyi, and S. Martinez-Alonso (2007), Windy Mars: A dynamic planet as seen by the HiRISE camera, *Geophys. Res. Lett.*, 34(23), doi:10.1029/2007GL031445.
- Bridges, N. T. et al. (2010), Aeolian bedforms, yardangs, and indurated surfaces in the Tharsis Montes as seen by the HiRISE Camera: Evidence for dust aggregates, *Icarus*, 205(1), 165–182, doi:10.1016/j.icarus.2009.05.017.
- Bridges, N. T., F. Ayoub, J.-P. Avouac, S. Leprince, A. Lucas, and S. Mattson (2012a), Earth-like sand fluxes on Mars, *Nature*, advance online publication, doi:10.1038/nature11022.
- Bridges, N. T. et al. (2012b), Planet-wide sand motion on Mars, *Geology*, 40(1), 31–34, doi:10.1130/G32373.1.

- Brückner, J., G. Dreibus, R. Gellert, S. W. Squyres, H. Wänke, A. Yen, and J. Zipfel (2008), Mars Exploration Rovers: chemical composition by the APXS, in *The Martian Surface*, edited by J. Bell and J. Bell, pp. 58–102, Cambridge University Press, Cambridge.
- Burns, R. G. (1993), Origin of electronic spectra of minerals in the visible to near-infrared region, in *Remote Geochemical Analysis: Elemental and Mineralogical Composition*, edited by C. M. Pieters and P. A. J. Englert, Cambridge University Press.
- Cantor, B. (2007), MOC observations of the 2001 Mars planet-encircling dust storm, *Icarus*, 186(1), 60–96, doi:10.1016/j.icarus.2006.08.019.
- Cantor, B. A., K. M. Kanak, and K. S. Edgett (2006), Mars Orbiter Camera observations of Martian dust devils and their tracks (September 1997 to January 2006) and evaluation of theoretical vortex models, *J. Geophys. Res.*, 111(E12), E12002, doi:10.1029/2006JE002700.
- Cardinale, M., G. Komatsu, S. Silvestro, and D. Tirsch (2012), The influence of local topography for wind direction on Mars: two examples of dune fields in crater basins, *Earth Surf. Process. Landforms*, n/a, doi:10.1002/esp.3289.
- Carr, M. (2006), *The surface of Mars*, [Rev. ed.], Cambridge University Press, Cambridge; New York.
- Catling, D., S. Wood, C. Leovy, D. Montgomery, H. Greenberg, C. Glein, and J. Moore (2006), Light-toned layered deposits in Juventae Chasma, Mars, *Icarus*, 181(1), 26–51, doi:10.1016/j.icarus.2005.10.020.
- Chapman, M. G., and K. L. Tanaka (2001), Interior trough deposits on Mars: Subice volcanoes?, *J. Geophys. Res.-Planets*, 106(E5), 10087–10100.
- Chapman, M. G., M. T. Gudmundsson, A. J. Russell, and T. M. Hare (2003), Possible Juventae Chasma subice volcanic eruptions and Maja Valles ice outburst floods on Mars: Implications of Mars Global Surveyor crater densities, geomorphology, and topography, *J. Geophys. Res.*, 108(E10), doi:10.1029/2002JE002009.
- Chojnacki, M., and B. M. Hynek (2008), Geological context of water-altered minerals in Valles Marineris, Mars, *J. Geophys. Res.*, 113(E12), doi:10.1029/2007JE003070.
- Chojnacki, M., and J. E. Moersch (2009), Valles Marineris Dune Fields: Thermophysical Properties, Morphology, and Provenance, p. Abstract #2486, Lunar and Planetary Institute, Houston.
- Chojnacki, M., B. M. Jakosky, and B. M. Hynek (2006), Surficial properties of landslides and surrounding units in Ophir Chasma, Mars, *J. Geophys. Res.*, 111(E4), doi:10.1029/2005JE002601.
- Chojnacki, M., J. Moersch, and D. Burr (2010), Climbing and falling dunes in Valles Marineris, Mars, *Geophys. Res. Lett.*, 37, doi:10.1029/2009GL042263.
- Chojnacki, M., D. M. Burr, J. E. Moersch, and T. I. Michaels (2011a), Orbital observations of contemporary dune activity in Endeavor crater, Meridiani Planum, Mars, *J. Geophys. Res.*, 116, doi:10.1029/2010JE003675.
- Chojnacki, M., J. E. Moersch, and D. M. Burr (2011b), Variable scale surface change of Valles Marineris dune fields and adjacent terrains, in *EPSC-DPS*, vol. 6, p. 1529.
- Chojnacki, M., J. E. Moersch, D. M. Burr, and J. J. Wray (2012a), Potential Sediment Sources and Pathways in Valles Marineris Dune Fields: Implications for Martian Aeolian Systems, in *Third International Planetary Dunes Workshop: Remote Sensing and Image Analysis of Planetary Dunes*, p. Abstract #7040, Lunar and Planetary Institute, Flagstaff, Arizona.

- Chojnacki, M., J. E. Moersch, D. M. Burr, and J. J. Wray (2012b), Valles Marineris Dune Fields: Sediment Pathways and Provenance, p. Abstract #2444, Lunar and Planetary Institute, Houston.
- Christensen, P. (1986), The spatial distribution of rocks on Mars, *Icarus*, 68(2), 217–238, doi:10.1016/0019-1035(86)90020-5.
- Christensen, P. (1988a), Eolian intracrater deposits on Mars: Physical properties and global distribution, *Icarus*, 56(3), 496–518, doi:10.1016/0019-1035(83)90169-0.
- Christensen, P. R. (1988b), Global Albedo Variations on Mars: Implications for Active Aeolian Transport, Deposition, and Erosion, *J. Geophys. Res.*, 93(B7), 7611–7624.
- Christensen, P. R. et al. (1992), Thermal Emission Spectrometer Experiment: Mars Observer Mission, *J. Geophys. Res.*, 97(E5), 7719, doi:10.1029/92JE00453.
- Christensen, P. R., R. V. Morris, M. D. Lane, J. L. Bandfield, and M. C. Malin (2001a), Global mapping of Martian hematite mineral deposits: Remnants of water-driven processes on early Mars, *J. Geophys. Res.-Planets*, 106(E10), 23873–23885.
- Christensen, P. R. et al. (2001b), Mars Global Surveyor Thermal Emission Spectrometer experiment: Investigation description and surface science results, *J. Geophys. Res.*, 106(E10), 23823–23871, doi:10.1029/2000JE001370.
- Christensen, P. R. et al. (2003), Morphology and composition of the surface of Mars: Mars Odyssey THEMIS results, *Science*, 10.1126/science.1080885.
- Christensen, P. R. et al. (2004), The Thermal Emission Imaging System (THEMIS) for the Mars 2001 Odyssey Mission, *Space Science Reviews*, 110(1/2), 85–130, doi:10.1023/B:SPAC.0000021008.16305.94.
- Christensen, P. R., J. L. Bandfield, A. D. Rogers, T. D. Glotch, V. E. Hamilton, S. W. Ruff, and M. B. Wyatt (2008), Global Mineralogy Mapped from the Mars Global Surveyor Thermal Emission Spectrometer, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell, pp. 195–220, Cambridge University Press, Cambridge.
- Clancy, R. T., M. J. Wolff, B. A. Cantor, M. C. Malin, and T. I. Michaels (2009), Valles Marineris cloud trails, *J. Geophys. Res.*, 114, doi:10.1029/2008JE003323.
- Clark, R. N., T. V. V. King, M. Klejwa, G. A. Swayze, and N. Vergo (1990), High Spectral Resolution Reflectance Spectroscopy of Minerals, *J. Geophys. Res.*, 95(B8), 12653–12680, doi:10.1029/JB095iB08p12653.
- Clark, R. N., G. A. Swayze, A. J. Gallagher, T. V. V. King, and W. M. Calvin (1993), The U. S. Geological Survey, Digital Spectral Library: Version 1: 0.2 to 3.0 microns, U.S. Geological Survey Open File Report, 93-592, 1340 pages.
- Clenet, H., J. Flahaut, C. Quantin, P. Pinet, Y. Daydou, and P. Allemand (2011), Compositional Diversity of Mafic Rocks in the Vicinity of Valles Marineris, Mars, Using Modified Gaussian Model, vol. 38th Lunar and Planetary Science Conference, p. abstract 1674.
- Cloutis, E., and J. F. I. Bell (2004), Mafic silicate mapping on Mars: effects of palagonitic material, multiple mafic silicates, and spectral resolution, *Icarus*, 172(1), 233–254, doi:10.1016/j.icarus.2004.06.002.
- Cloutis, E. A., and M. J. Gaffey (1991), Spectral-compositional variations in the constituent minerals of mafic and ultramafic assemblages and remote sensing implications, *Earth, Moon, and Planets*, 53(1), 11–53, doi:10.1007/BF00116217.
- Cloutis, E. A., M. J. Gaffey, T. L. Jackowski, and K. L. Reed (1986), Calibrations of Phase Abundance, Composition, and Particle Size Distribution for Olivine-Orthopyroxene Mixtures

- From Reflectance Spectra, *Journal of Geophysical Research*, 91(B11), 11641–11653, doi:10.1029/JB091iB11p11641.
- Cloutis, E. A. et al. (2006), Detection and discrimination of sulfate minerals using reflectance spectroscopy, *Icarus*, 184(1), 121–157, doi:10.1016/j.icarus.2006.04.003.
- Dalton, J. B., O. Prieto-Ballesteros, J. S. Kargel, C. S. Jamieson, J. Jolivet, and R. Quinn (2005), Spectral comparison of heavily hydrated salts with disrupted terrains on Europa, *Icarus*, 177(2), 472–490, doi:10.1016/j.icarus.2005.02.023.
- Delamere, W. A., L. L. Tornabene, A. S. McEwen, K. Becker, J. W. Bergstrom, N. T. Bridges, E. M. Eliason, D. Gallagher, K. E. Herkenhoff, and L. Keszthelyi (2010), Color imaging of Mars by the High Resolution Imaging Science Experiment (HiRISE), *Icarus*, 205(1), 38–52, doi:10.1016/j.icarus.2009.03.012.
- Dohm, J. M., J.-P. Williams, R. C. Anderson, J. Ruiz, P. C. McGuire, G. Komatsu, A. F. Davila, J. C. Ferris, D. Schulze-Makuch, and V. R. Baker (2009), New evidence for a magmatic influence on the origin of Valles Marineris, Mars, *Journal of Volcanology and Geothermal Research*, 185(1-2), 12–27, doi:10.1016/j.jvolgeores.2008.11.029.
- Edgett, K. S. (1994), The Sand Component of the Modern Martian Aeolian Sedimentary System, Ph.D. Dissertation, Ariz. State Univ., Tempe, AZ.
- Edgett, K. S. (2002), Low-albedo surfaces and eolian sediment: Mars Orbiter Camera views of western Arabia Terra craters and wind streaks, *J. Geophys. Res.*, 107(E6), doi:10.1029/2001JE001587.
- Edgett, K. S., and N. Lancaster (1993), Volcaniclastic aeolian dunes: terrestrial examples and application to martian sands, *Journal of Arid Environments*, 25(3), 271–297, doi:10.1006/jare.1993.1061.
- Edgett, K. S., and M. C. Malin (2000a), Examples of Martian Sandstone: Indurated, Lithified, and Cratered Eolian Dunes in MGS MOC Images, p. Abstract #1071, Lunar and Planetary Institute, Houston.
- Edgett, K. S., and M. C. Malin (2000b), New views of Mars eolian activity, materials, and surface properties: Three vignettes from the Mars Global Surveyor Mars Orbiter Camera, *J. Geophys. Res.*, 105(E1), 1623–1650, doi:10.1029/1999JE001152.
- Edwards, C. S., P. R. Christensen, and V. E. Hamilton (2008), Evidence for extensive olivine-rich basalt bedrock outcrops in Ganges and Eos chasmas, Mars, *J. Geophys. Res.*, 113(E11), doi:10.1029/2008JE003091.
- Edwards, C. S., J. L. Bandfield, P. R. Christensen, and R. L. Fergason (2009), Global distribution of bedrock exposures on Mars using THEMIS high-resolution thermal inertia, *J. Geophys. Res.*, 114(E11), doi:10.1029/2009JE003363.
- Ehlmann, B. L. et al. (2009), Identification of hydrated silicate minerals on Mars using MRO-CRISM: Geologic context near Nili Fossae and implications for aqueous alteration, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003339.
- Ehlmann, B. L., J. F. Mustard, S. L. Murchie, J.-P. Bibring, A. Meunier, A. A. Fraeman, and Y. Langevin (2011), Subsurface water and clay mineral formation during the early history of Mars, *Nature*, 479(7371), 53–60, doi:10.1038/nature10582.
- Al-Enezi, A., K. Pye, R. Misak, and S. Al-Hajraf (2008), Morphologic characteristics and development of falling dunes, northeast Kuwait, *Journal of Arid Environments*, 72(4), 423–439, doi:10.1016/j.jaridenv.2007.06.017.

- Fenton, L. K. (2005), Potential sand sources for the dune fields in Noachis Terra, Mars, *J. Geophys. Res.*, 110(E11), doi:10.1029/2005JE002436.
- Fenton, L. K. (2006), Dune migration and slip face advancement in the Rabe Crater dune field, Mars, *Geophysical Research Letters*, 33, doi:10.1029/2006GL027133.
- Fenton, L. K., and R. A. Beyer (2012), Reconstructing the Formative Winds of a Dune Field in Ganges Chasma, Mars, by Bootstrapping with the Rule of Maximum Gross Bedform-Normal Transport, p. Abstract #7028, Lunar and Planetary Institute, Houston.
- Fenton, L. K., and R. K. Hayward (2010), Southern high latitude dune fields on Mars: Morphology, aeolian inactivity, and climate change, *Geomorphology*, 121(1-2), 98–121, doi:10.1016/j.geomorph.2009.11.006.
- Fenton, L. K., J. L. Bandfield, and A. W. Ward (2003), Aeolian processes in Proctor Crater on Mars: Sedimentary history as analyzed from multiple data sets, *J. Geophys. Res.*, 108(E12), doi:10.1029/2002JE002015.
- Fenton, L. K., A. D. Toigo, and M. I. Richardson (2005), Aeolian processes in Proctor Crater on Mars: Mesoscale modeling of dune-forming winds, *J. Geophys. Res.*, 110(E6), doi:10.1029/2004JE002309.
- Fenton, L. K., P. E. Geissler, and R. M. Haberle (2007), Global warming and climate forcing by recent albedo changes on Mars, *Nature*, 446(7136), 646–649, doi:10.1038/nature05718.
- Fenton, L. K., T. I. Michaels, and R. A. Beyer (2012), Aeolian Sediment Sources and Transport in Ganges Chasma, Mars: Morphology and Atmospheric Modeling, vol. XLIII, p. Abstract 2441.
- Ferguson, R. L., P. R. Christensen, and H. H. Kieffer (2006a), High-resolution thermal inertia derived from the Thermal Emission Imaging System (THEMIS): Thermal model and applications, *J. Geophys. Res.*, 111(E12), doi:10.1029/2006JE002735.
- Ferguson, R. L., P. R. Christensen, J. F. Bell, M. P. Golombek, K. E. Herkenhoff, and H. H. Kieffer (2006b), Physical properties of the Mars Exploration Rover landing sites as inferred from Mini-TES-derived thermal inertia, *J. Geophys. Res.*, 111(E2), doi:10.1029/2005JE002583.
- Flahaut, J., C. Quantin, P. Allemand, P. Thomas, and L. Le Deit (2010), Identification, distribution and possible origins of sulfates in Capri Chasma (Mars), inferred from CRISM data, *J. Geophys. Res.*, 115(E11), doi:10.1029/2009JE003566.
- Flahaut, J., J. F. Mustard, C. Quantin, H. Clenet, P. Allemand, and P. Thomas (2011), Dikes of distinct composition intruded into Noachian-aged crust exposed in the walls of Valles Marineris, *Geophysical Research Letters*, 38(15), doi:10.1029/2011GL048109.
- Flahaut, J., C. Quantin, H. Clenet, P. Allemand, J. F. Mustard, and P. Thomas (2012), Pristine Noachian crust and key geologic transitions in the lower walls of Valles Marineris: Insights into early igneous processes on Mars, *Icarus*, doi:10.1016/j.icarus.2011.12.027.
- Freeman, J. J., A. Wang, and B. L. Jolliff (2007), Pathways to Form Kieserite from Epsomite at Mid to Low Temperatures, with Relevance to Mars, p. Abstract #1298, Lunar and Planetary Institute, Houston.
- Frey, H. (1979), Martian canyons and African rifts: Structural comparisons and implications, *Icarus*, 37(1), 142–155, doi:10.1016/0019-1035(79)90122-2.
- Fryberger, S. G. (1979), Dune forms and wind regime, vol. 1042, pp. 137–170, U.S. Geol. Surv. Prof. Pap.

- Gaddis, L. (1997), An overview of the Integrated Software for Imaging Spectrometers (ISIS), *Lunar Planet. Sci.*, XXVIII, abstract 1226.
- Gardin, E., P. Allemand, C. Quantin, S. Silvestro, and C. Delacourt (2012), Dune fields on Mars: Records of a climate change?, *Planetary and Space Science*, 60(1), 314–321, doi:10.1016/j.pss.2011.10.004.
- Gay, S. P. (1999), Observations regarding the movement of barchan sand dunes in the Nazca to Tanaca area of southern Peru, *Geomorphology*, 27, 279–293.
- Geissler, P. E. (2005), Three decades of Martian surface changes, *J. Geophys. Res.*, 110(E2), doi:10.1029/2004JE002345.
- Geissler, P. E. et al. (2008), First in situ investigation of a dark wind streak on Mars, *J. Geophys. Res.*, 113(E12), doi:10.1029/2008JE003102.
- Geissler, P. E., N. W. Stanzos, N. T. Bridges, and the HiRISE Science Team (2011), Shifting sands on Mars: 3 case studies, vol. XLII, p. Abstract 2537.
- Geissler, P. E., M. E. Banks, N. T. Bridges, S. Silvestro, and HiRISE Science Team (2012), HiRISE Observations of Sand Dune Motion on Mars: Emerging Global Trends, p. Abstract #7053, Lunar and Planetary Institute, Houston.
- Geissler, P. E., L. K. Fenton, and N. T. Bridges (2013), Dust “Drifts” on Mars, p. Abstract #2573, Lunar and Planetary Institute, Houston.
- Gendrin, A. et al. (2005), Sulfates in martian layered terrains: the OMEGA/Mars Express view, *Science*, 307(5715), 1587–1591, doi:10.1126/Science.1109087.
- Ghrefat, H. A., P. C. Goodell, B. E. Hubbard, R. P. Langford, and R. E. Aldouri (2007), Modeling grain size variations of aeolian gypsum deposits at White Sands, New Mexico, using AVIRIS imagery, *Geomorphology*, 88(1-2), 57–68, doi:10.1016/j.geomorph.2006.10.013.
- Golombek, M., K. Robinson, A. McEwen, N. Bridges, B. Ivanov, L. Tornabene, and R. Sullivan (2010), Constraints on ripple migration at Meridiani Planum from Opportunity and HiRISE observations of fresh craters, *J. Geophys. Res.*, 115, doi:10.1029/2010JE003628.
- Golombek, M. P. et al. (2006), Erosion rates at the Mars Exploration Rover landing sites and long-term climate change on Mars, *J. Geophys. Res.*, 111(E12), doi:10.1029/2006JE002754.
- Gorelick, N. S., M. Weiss-Malik, B. Steinberg, and S. Anwar (2003), JMARS: A Multimission Data Fusion Application, in 34th Lunar and Planetary Science Conference, p. Abstract #2057, Lunar and Planetary Institute, Houston, Texas.
- Greeley, R. (2002), Terrestrial analogs to wind-related features at the Viking and Pathfinder landing sites on Mars, *J. Geophys. Res.*, 107(E1), doi:10.1029/2000JE001481.
- Greeley, R., and J. E. Guest (1987), Geologic map of the eastern equatorial region of Mars (USGS I-1802-B, 1987, 1:15M scale).
- Greeley, R., and J. D. Iversen (1985), Wind as a geological process on Earth, Mars, Venus and Titan, edited by R. Greeley, Cambridge Univ. Press, New York.
- Greeley, R., and M. D. Kraft (2001), Survivability of Aggregate Sands on Mars, p. Abstract #1839, Lunar and Planetary Institute, Houston.
- Greeley, R., R. Leach, B. White, J. Iversen, and J. Pollack (1980), Threshold windspeeds for sand on Mars: Wind tunnel simulations, *Geophys. Res. Lett.*, 7(2), 121, doi:10.1029/GL007i002p00121.
- Greeley, R., N. Lancaster, S. W. Lee, and P. C. Thomas (1992), Martian Aeolian Processes, Sediments and Features, in Mars, edited by H. H. Kieffer, B. M. Jakosky, C. W. Snyder, and M. S. Matthews, pp. 730–766, University of Arizona Press, Tucson.

- Greeley, R., A. Skyeck, and J. B. Pollack (1993), Martian Aeolian Features and Deposits: Comparisons With General Circulation Model Results, *J. Geophys. Res.*, 98(E2), 3183–3196, doi:10.1029/92JE02580.
- Greeley, R. et al. (2006), Gusev crater: Wind-related features and processes observed by the Mars Exploration Rover Spirit, *J. Geophys. Res.*, 111(E2), doi:10.1029/2005JE002491.
- Grotzinger, J. P. et al. (2005), Stratigraphy and sedimentology of a dry to wet eolian depositional system, Burns formation, Meridiani Planum, Mars, *Earth and Planetary Science Letters*, 240(1), 11–72, doi:10.1016/j.epsl.2005.09.039.
- Haberle, R. M. (1986), The climate of Mars, *Scientific American*, 254(5).
- Haberle, R. M., J. B. Pollack, J. R. Barnes, R. W. Zurek, C. B. Leovy, J. R. Murphy, H. Lee, and J. Schaeffer (1993), Mars Atmospheric Dynamics as Simulated by the NASA Ames General Circulation Model 1. The Zonal-Mean Circulation, *J. Geophys. Res.*, 98(E2), 3093–3123, doi:10.1029/92JE02946.
- Hall, K. (1999), The role of thermal stress fatigue in the breakdown of rock in cold regions, *Geomorphology*, 31(1–4), 47–63, doi:10.1016/S0169-555X(99)00072-0.
- Hall, K., and M.-F. André (2001), New insights into rock weathering from high-frequency rock temperature data: an Antarctic study of weathering by thermal stress, *Geomorphology*, 41(1), 23–35, doi:10.1016/S0169-555X(01)00101-5.
- Hartmann, W. K., and G. Neukum (2001), Cratering chronology and the evolution of Mars, *Space Science Reviews*, 96(1–4), 165–194.
- Hartmann, W. K., J. Anguita, M. A. de la Casa, D. C. Berman, and E. V. Ryan (2001), Martian Cratering 7: The Role of Impact Gardening, *Icarus*, 149(1), 37–53, doi:10.1006/icar.2000.6532.
- Hayes, A. G., J. P. Grotzinger, L. A. Edgar, S. W. Squyres, W. A. Watters, and J. Sohl-Dickstein (2011), Reconstruction of eolian bed forms and paleocurrents from cross-bedded strata at Victoria Crater, Meridiani Planum, Mars, *J. Geophys. Res.*, 116, doi:10.1029/2010JE003688.
- Hayward, R. K., K. F. Mullins, L. K. Fenton, T. M. Hare, T. N. Titus, M. C. Bourke, A. Colaprete, and P. R. Christensen (2007a), Mars Global Digital Dune Database and initial science results, *J. Geophys. Res.*, 112(E11), doi:10.1029/2007JE002943.
- Hayward, R. K., K. F. Mullins, L. K. Fenton, T. M. Hare, T. N. Titus, M. C. Bourke, A. Colaprete, and P. R. Christensen (2007b), Mars Global Digital Dune Database: MC2 – MC29, USGS Open File Report.
- Hayward, R. K., T. N. Titus, T. I. Michaels, L. K. Fenton, A. Colaprete, and P. R. Christensen (2009), Aeolian dunes as ground truth for atmospheric modeling on Mars, *J. Geophys. Res.*, 114(E11), doi:10.1029/2009JE003428.
- Hayward, R. K., J. R. Zimbelman, L. K. Fenton, T. N. Titus, and G. E. Cushing (2010a), Mars Analog: Grand Falls Dune Field, Arizona, in *Second International Planetary Dunes Workshop: Planetary Analogs _ Integrating Models, Remote Sensing, and Field Data*, p. Abstract #2004, Lunar and Planetary Institute, Alamosa, Colorado.
- Hayward, R. K., L. K. Fenton, K. L. Tanaka, T. N. Titus, and P. R. Christensen (2010b), Mars global digital dune database: Dune volume estimates in the north polar region, vol. XLI, p. Abstract 1109.
- Hazen, R. M., D. Papineau, W. Bleeker, R. T. Downs, J. M. Ferry, T. J. McCoy, D. A. Sverjensky, and H. Yang (2008), Mineral evolution, *American Mineralogist*, 93(11–12), 1693–1720, doi:10.2138/am.2008.2955.

- Herkenhoff, K. E., M. P. Golombek, E. A. Guinness, J. B. Johnson, A. Kusack, L. Richter, R. J. Sullivan, and S. Gorevan (2008), In situ observations of the physical properties of the Martian surface, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell and J. Bell, pp. 451–467, Cambridge University Press, Cambridge.
- Hiroi, T., and C. M. Pieters (1994), Estimation of grain sizes and mixing ratios of fine powder mixtures of common geologic minerals, *J. Geophys. Res.*, 99(E5), 10867–10879, doi:10.1029/94JE00841.
- Hobbs, S. W., D. J. Paull, and M. C. Bourke (2010), Aeolian processes and dune morphology in Gale Crater, *Icarus*, 210(1), 102–115, doi:10.1016/j.icarus.2010.06.006.
- Horgan, B., and J. F. Bell (2012a), Widespread weathered glass on the surface of Mars, *Geology*, doi:10.1130/G32755.1.
- Horgan, B., P. Mann, J. Stromberg, and E. A. Cloutis (2011), Acid Alteration of Basalts, Andesites, and Anorthosites: Near-IR Spectra and Implications for Martian Soil Formation, in 42nd Lunar and Planetary Science Conference, p. Abstract #2415, Lunar and Planetary Institute, Houston.
- Horgan, B., L. K. Fenton, and P. R. Christensen (2012a), Comparing Active Modes of Mass Movement on Martian Dunes, p. Abstract #7052, Lunar and Planetary Institute, Houston.
- Horgan, B., D. Clarke, and J. F. I. Bell (2012b), Evidence for Glass-Rich Pyroclastics in Martian Dunes, in *Third International Planetary Dunes Workshop: Remote Sensing and Image Analysis of Planetary Dunes*, p. Abstract #7050, Lunar and Planetary Institute, Flagstaff, Arizona.
- Horgan, B., M. Chojnacki, J. Iai, D. Clarke, J. Joseph, and J. F. Bell III (2012c), Widespread Explosive Volcanism on Mars Inferred from the Global Distribution of Glass-Rich Sediments, in *Geological Society of America*, p. Paper No. 60–4, Geological Society of America, Charlotte, N.C.
- Horgan, B. H., J. F. Bell, E. Z. Noe Dobrea, E. A. Cloutis, D. T. Bailey, M. A. Craig, L. H. Roach, and J. F. Mustard (2009), Distribution of hydrated minerals in the north polar region of Mars, *J. Geophys. Res.*, 114(E1), doi:10.1029/2008JE003187.
- Horgan, B. H. N., and J. F. Bell (2012b), Seasonally active slipface avalanches in the north polar sand sea of Mars: Evidence for a wind-related origin, *Geophys. Res. Lett.*, 39(9), L09201, doi:10.1029/2012GL051329.
- Hughes, M. G., J. B. Keene, and R. G. Joseph (2000), Hydraulic Sorting of Heavy-Mineral Grains by Swash on a Medium-Sand Beach, *Journal of Sedimentary Research*, 70(5), 994–1004, doi:10.1306/112599700994.
- Hurowitz, J. A., S. M. McLennan, N. J. Tosca, R. E. Arvidson, J. R. Michalski, D. W. Ming, C. Schroder, and S. W. Squyres (2006), In situ and experimental evidence for acidic weathering of rocks and soils on Mars, 111(E2).
- Hynek, B. M., R. J. Phillips, and R. E. Arvidson (2003), Explosive volcanism in the Tharsis region: Global evidence in the Martian geologic record, *J. Geophys. Res.*, 108(E9), doi:10.1029/2003JE002062.
- Iversen, J. D., and K. R. Rasmussen (1994), The effect of surface slope on saltation threshold, *Sedimentology*, 41(4), 721–728, doi:10.1111/j.1365-3091.1994.tb01419.x.
- Iversen, J. D., and K. R. Rasmussen (1999), The effect of wind speed and bed slope on sand transport, *Sedimentology*, 46(4), 723–731, doi:10.1046/j.1365-3091.1999.00245.x.

- Iversen, J. D., and B. R. White (1982), Saltation threshold on Earth, Mars and Venus, *Sedimentology*, 29(1), 111–119, doi:10.1111/j.1365-3091.1982.tb01713.x.
- Jakosky, B. M. (1986), On the thermal properties of Martian fines, *Icarus*, 66, 117–124.
- Jakosky, B. M., B. M. Hynek, S. M. Pelkey, M. T. Mellon, S. Martinez-Alonso, N. E. Putzig, N. Murphy, and P. R. Christensen (2006), Thermophysical properties of the MER and Beagle II landing site regions on Mars, *J. Geophys. Res.*, 111(E8), E08008.
- Johnsson, M. J. (1993), The system controlling the composition of clastic sediments, *Geological Society of America Special Paper*, 284, 1–19.
- Kieffer, H. H., B. M. Jakosky, C. W. Snyder, and M. S. Matthews (Eds.) (1992), *Mars, Space science series.*, University of Arizona Press, Tucson.
- King, T. V. V., and W. I. Ridley (1987a), Relation of the Spectroscopic Reflectance of Olivine to Mineral Chemistry and Some Remote Sensing Implications, *J. Geophys. Res.*, 92(B11), 11457–11469, doi:10.1029/JB092iB11p11457.
- King, T. V. V., and W. I. Ridley (1987b), Relation of the Spectroscopic Reflectance of Olivine to Mineral Chemistry and Some Remote-Sensing Implications, *Journal of Geophysical Research - Solid Earth and Planets*, 92(B11), 11457–11469.
- Kirk, R. L. et al. (2008), Ultrahigh resolution topographic mapping of Mars with MRO HiRISE stereo images: Meter-scale slopes of candidate Phoenix landing sites, *J. Geophys. Res.*, 113, doi:10.1029/2007JE003000.
- Kirk, R. L., E. Howington-Kraus, and M. R. Rosiek (2009), Build Your Own Topographic Model: A Photogrammetry Guest Facility for Planetary Researchers, in 40th Lunar and Planetary Science Conference, p. Abstract #1414, Lunar and Planetary Institute, Houston.
- Kocurek, G., and N. Lancaster (1999), Aeolian system sediment state: theory and Mojave Desert Kelso dune field example, *Sedimentology*, 46, 505–515, doi:10.1046/j.1365-3091.1999.00227.x.
- Kocurek, G., M. Carr, R. Ewing, K. G. Havholm, Y. C. Nagar, and A. K. Singhvi (2007), White Sands Dune Field, New Mexico: Age, dune dynamics and recent accumulations, *Sedimentary Geology*, 197(3-4), 313–331, doi:10.1016/j.sedgeo.2006.10.006.
- Kocurek, G., N. T. Bridges, K. S. Edgett, W. Goetz, K. W. Lewis, M. B. Madsen, D. M. Rubin, R. J. Sullivan, and MSL Science Team (2013), Rocknest Sand Shadow at the Curiosity Field Site: Morphology, Origin and Stabilization, p. Abstract #1375, Lunar and Planetary Institute, Houston.
- Kreslavsky, M. A., and J. W. Head (1999), Kilometer-scale slopes on Mars and their correlation with geologic units: Initial results from Mars Orbiter Laser Altimeter (MOLA) data, *J. Geophys. Res.*, 104(E9), 21911–21924, doi:10.1029/1999JE001051.
- Kreslavsky, M. A., and J. W. Head (2000), Kilometer-scale roughness of Mars: Results from MOLA data analysis, *J. Geophys. Res.*, 105(E11), 26695–26711, doi:10.1029/2000JE001259.
- Kreslavsky, M. A., and J. W. Head (2002), Mars: Nature and evolution of young latitude-dependent water-ice-rich mantle, *Geophysical Research Letters*, 29, doi:10.1029/2002GL015392.
- Lancaster, N. (2009), *Geomorphology of Desert Environments*, edited by A. J. Parsons and A. D. Abrahams, pp. 557–595, Springer Netherlands.
- Lancaster, N., and R. Greeley (1990), Sediment Volume in the North Polar Sand Seas of Mars, *J. Geophys. Res.*, 95(B7), 10921–10927, doi:10.1029/JB095iB07p10921.

- Lancaster, N., and V. P. Tchakerian (1996), Geomorphology and sediments of sand ramps in the Mojave desert, *Geomorphology*, 17(1-3), 151–165, doi:10.1016/0169-555X(95)00101-A.
- Langevin, Y. (2005), Summer Evolution of the North Polar Cap of Mars as Observed by OMEGA/Mars Express, *Science*, 307(5715), 1581–1584, doi:10.1126/science.1109438.
- Langevin, Y., F. Poulet, J.-P. Bibring, and B. Gondet (2005), Sulfates in the North Polar Region of Mars Detected by OMEGA/Mars Express, *Science*, 307(5715), 1584–1586, doi:10.1126/science.1109091.
- Lewis, K. W., O. Aharonson, J. P. Grotzinger, S. W. Squyres, J. F. Bell, L. S. Crumpler, and M. E. Schmidt (2008), Structure and stratigraphy of Home Plate from the Spirit Mars Exploration Rover, *Journal of Geophysical Research*, 113(E12), doi:10.1029/2007JE003025.
- Lichtenberg, K. A. et al. (2010), Stratigraphy of hydrated sulfates in the sedimentary deposits of Aram Chaos, Mars, *J. Geophys. Res.*, 115, doi:10.1029/2009JE003353.
- Lucchitta, B. (1999), Geologic map of Ophir and central Candor Chasmata (MTM -05072) of Mars, U.S. Geol. Surv. Misc. Invest. Ser. Map I-2568, scale 1:500,000.
- Lucchitta, B. K. (2010), 5 - Lakes in Valles Marineris, in *Lakes on Mars*, pp. 111–161, Elsevier, Amsterdam.
- Lucchitta, B. K., A. McEwen, G. D. Clow, P. Geissler, R. B. Singer, R. A. Schulz, and S. W. Squyres (1992), The Canyon System on Mars, in *Mars*, edited by H. H. Kieffer, B. M. Jakosky, C. W. Snyder, and M. S. Matthews, pp. 453–492, University of Arizona Press, Tucson.
- Malin, M. C., and K. S. Edgett (2000), Sedimentary rocks of early Mars, *Science*, 290(5498), 1927–1937.
- Malin, M. C., and K. S. Edgett (2001), Mars Global Surveyor Mars Orbiter Camera: Interplanetary cruise through primary mission, *J. Geophys. Res.*, 106(E10), 23429–23570, doi:10.1029/2000JE001455.
- Malin, M. C., K. S. Edgett, L. V. Posiolova, S. M. McColley, and E. Z. N. Dobreá (2006), Present-Day Impact Cratering Rate and Contemporary Gully Activity on Mars, *Science*, 314(5805), 1573–1577, doi:10.1126/science.1135156.
- Malin, M. C. et al. (2007), Context Camera Investigation on board the Mars Reconnaissance Orbiter, *J. Geophys. Res.*, 112(E5), doi:10.1029/2006JE002808.
- Mangold, N., A. Gendrin, B. Gondet, S. Lemouelic, C. Quantin, V. Ansan, J. Bibring, Y. Langevin, P. Masson, and G. Neukum (2008), Spectral and geological study of the sulfate-rich region of West Candor Chasma, Mars, *Icarus*, 194(2), 519–543, doi:10.1016/j.icarus.2007.10.021.
- Mangold, N., D. Baratoux, O. Arnalds, J.-M. Bardintzeff, B. Platevoet, M. Grégoire, and P. Pinet (2011), Segregation of olivine grains in volcanic sands in Iceland and implications for Mars, *Earth and Planetary Science Letters*, 310(3-4), 233–243, doi:10.1016/j.epsl.2011.07.025.
- Mauget, S. A., E. C. Cordero, and P. T. Brown (2011), Evaluating Modeled Intra- to Multidecadal Climate Variability Using Running Mann–Whitney Z Statistics, *J. Climate*, 25(5), 1570–1586, doi:10.1175/JCLI-D-11-00211.1.
- McEwen, A. et al. (2010), The High Resolution Imaging Science Experiment (HiRISE) during MRO's Primary Science Phase (PSP), *ICARUS*, 205(1), 2–37, doi:10.1016/j.icarus.2009.04.023.
- McEwen, A. S., M. C. Malin, M. H. Carr, and W. K. Hartmann (1999), Voluminous volcanism on early Mars revealed in Valles Marineris, *Nature*, 397(6720), 584–586, doi:10.1038/17539.

- McEwen, A. S. et al. (2007), Mars Reconnaissance Orbiter's High Resolution Imaging Science Experiment (HiRISE), *J. Geophys. Res.*, 112(E5), doi:10.1029/2005JE002605.
- McGlynn, I. O. (2012), Formation and Alteration of Basaltic Soils on Mars, PhD Dissertation, Univ. of Tenn., Knoxville, TN.
- McGlynn, I. O., C. M. Fedo, and H. Y. McSween (2011), Origin of basaltic soils at Gusev crater, Mars, by aeolian modification of impact-generated sediment, *J. Geophys. Res.*, 116, doi:10.1029/2010JE003712.
- McGlynn, I. O., C. M. Fedo, and H. Y. McSween (2012), Soil mineralogy at the Mars Exploration Rover landing sites: An assessment of the competing roles of physical sorting and chemical weathering, *J. Geophys. Res.*, 117(E1), E01006, doi:10.1029/2011JE003861.
- McGuire, P. C., J. L. Bishop, A. J. Brown, A. A. Fraeman, G. A. Marzo, M. Frank Morgan, S. L. Murchie, J. F. Mustard, M. Parente, and S. M. Pelkey (2009), An improvement to the volcano-scan algorithm for atmospheric correction of CRISM and OMEGA spectral data, *Planetary and Space Science*, 57(7), 809–815, doi:10.1016/j.pss.2009.03.007.
- McKee, E. D. (1979), Introduction to a study of global sand seas, in *A Study of Global Sand Seas*, U.S. Geol. Surv. Prof. Pap., 1052, 3–17.
- McKeown, N. K., J. L. Bishop, E. Z. Noe Dobrea, B. L. Ehlmann, M. Parente, J. F. Mustard, S. L. Murchie, G. A. Swayze, J.-P. Bibring, and E. A. Silver (2009), Characterization of phyllosilicates observed in the central Mawrth Vallis region, Mars, their potential formational processes, and implications for past climate, *J. Geophys. Res.*, 114, doi:10.1029/2008JE003301.
- McSween, H. Y., and K. Keil (2000), Mixing relationships in the Martian regolith and the composition of globally homogeneous dust, *Geochimica et Cosmochimica Acta*, 64(12), 2155–2166, doi:10.1016/S0016-7037(99)00401-9.
- McSween, H. Y. et al. (2004), Basaltic Rocks Analyzed by the Spirit Rover in Gusev Crater, *Science*, 305(5685), 842–845, doi:10.1126/science.3050842.
- McSween, H. Y. et al. (2006), Characterization and petrologic interpretation of olivine-rich basalts at Gusev Crater, Mars, *J. Geophys. Res.*, 111(E2), E02S10.
- McSween, H. Y., G. J. Taylor, and M. B. Wyatt (2009), Elemental Composition of the Martian Crust, *Science*, 324(5928), 736–739, doi:10.1126/science.1165871.
- McSween, H. Y., I. O. McGlynn, and A. D. Rogers (2010), Determining the modal mineralogy of Martian soils, *J. Geophys. Res.*, 115, doi:10.1029/2010JE003582.
- Mellon, M. T., B. M. Jakosky, H. H. Kieffer, and P. R. Christensen (2000), High-resolution thermal inertia mapping from the Mars Global Surveyor Thermal Emission Spectrometer, *Icarus*, 148(2), 437–455.
- Mellon, M. T., R. L. Fergason, and N. E. Putzig (2008), The thermal inertia of the surface of Mars, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell, pp. 399–427, Cambridge University Press, Cambridge.
- Michaels, T. I. (2011), Aeolian Phenomena at Nili and Meroe Paterae, vol. XLII, p. Abstract 2697.
- Michalski, J. R., M. D. Kraft, T. G. Sharp, and P. R. Christensen (2006), Effects of chemical weathering on infrared spectra of Columbia River Basalt and spectral interpretations of martian alteration, *Earth and Planetary Science Letters*, 248(3–4), 822–829, doi:10.1016/j.epsl.2006.06.034.

- Minitti, M. E., C. M. Weitz, M. D. Lane, and J. L. Bishop (2007), Morphology, chemistry, and spectral properties of Hawaiian rock coatings and implications for Mars, *J. Geophys. Res.*, 112(E5), E05015, doi:10.1029/2006JE002839.
- Modi, A. L. (2011), Processes controlling the composition of first-cycle sediments deposited in an arid-climate, with implications for provenance reconstruction studies, M.S. Thesis, Univ. of Tenn, Knoxville, TN.
- Moore, H. J., D. B. Bickler, J. A. Crisp, H. J. Eisen, J. A. Gensler, A. F. C. Haldemann, J. R. Matijevic, L. K. Reid, and F. Pavlics (1999), Soil-like deposits observed by Sojourner, the Pathfinder rover, *Journal of Geophysical Research*, 104(E4), 8729, doi:10.1029/1998JE900005.
- Moroz, L. V. et al. (2009), Spectral properties of simulated impact glasses produced from martian soil analogue JSC Mars-1, *Icarus*, 202(1), 336–353, doi:10.1016/j.icarus.2009.02.007.
- Muhs, D. R. (2004), Mineralogical maturity in dunefields of North America, Africa and Australia, *Geomorphology*, 59(1-4), 247–269, doi:10.1016/j.geomorph.2003.07.020.
- Muhs, D. R., R. L. Reynolds, J. Been, and G. Skipp (2003), Eolian sand transport pathways in the southwestern United States: importance of the Colorado River and local sources, *Quaternary International*, 104(1), 3–18, doi:10.1016/S1040-6182(02)00131-3.
- Murchie, S. et al. (2007), Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) on Mars Reconnaissance Orbiter (MRO), *J. Geophys. Res.*, 112(E5), doi:10.1029/2006JE002682.
- Murchie, S. et al. (2009a), Evidence for the origin of layered deposits in Candor Chasma, Mars, from mineral composition and hydrologic modeling, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003343.
- Murchie, S. L. et al. (2009b), A synthesis of Martian aqueous mineralogy after 1 Mars year of observations from the Mars Reconnaissance Orbiter, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003342.
- Murchie, S. L. et al. (2009c), Compact Reconnaissance Imaging Spectrometer for Mars investigation and data set from the Mars Reconnaissance Orbiter's primary science phase, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003344.
- Mustard, J. F., F. Poulet, A. Gendrin, J. P. Bibring, Y. Langevin, B. Gondet, N. Mangold, G. Bellucci, and F. Altieri (2005), Olivine and pyroxene, diversity in the crust of Mars, *Science*, 307(5715), 1594–1597.
- Mustard, J. F. et al. (2007), CRISM-OMEGA Observations of Phyllosilicate-Olivine Stratigraphy in Nili Fossae, Mars, in 38th Lunar and Planetary Science Conference, p. Abstract #2071, Lunar and Planetary Institute, League City, Texas.
- Nedell, S., S. W. Squyers, and D. W. Anderson (1987), Origin and evolution of the layered deposits in the Valles Marineris, Mars, *Icarus*, 70(3), 409–414, doi:10.1016/0019-1035(87)90086-8.
- Nesse, W. D. (2012), *Introduction to mineralogy*, Oxford University Press, New York.
- Neukum, G., and et. al. (2004), HRSC: The High Resolution Stereo Camera of Mars Express, *Eur. Space Agency Spec. Publ.*, 1240, 17–35.
- Orofino, V., R. Politi, A. Blanco, and S. Fonti (2006), Diffuse reflectance of altered olivine grains: Remote sensing detection and implications for Mars studies, *Planetary and Space Science*, in press.
- Parente, M. (2008), A New Approach to Denoising CRISM Images, p. Abstract #2528, Lunar and Planetary Institute, Houston.

- Parente, M., H. D. Makarewicz, and J. L. Bishop (2011), Decomposition of mineral absorption bands using nonlinear least squares curve fitting: Application to Martian meteorites and CRISM data, *Planetary and Space Science*, 59(5-6), 423–442, doi:10.1016/j.pss.2011.01.009.
- Parsons, A. J., and A. D. Abrahams (Eds.) (2009), *Geomorphology of Desert Environments*, Springer Netherlands, Dordrecht.
- Pelkey, S. M. et al. (2007), CRISM multispectral summary products: Parameterizing mineral diversity on Mars from reflectance, *J. Geophys. Res.*, 112(E8), doi:10.1029/2006JE002831.
- Pelletier, J. D., and C. Rasmussen (2009), Quantifying the climatic and tectonic controls on hillslope steepness and erosion rate, *Lithosphere*, 1(2), 73–80, doi:10.1130/L3.1.
- Piqueux, S., and P. R. Christensen (2009), A model of thermal conductivity for planetary soils: 1. Theory for unconsolidated soils, *J. Geophys. Res.*, 114(E9), doi:10.1029/2008JE003308.
- Poulet, F., and S. Erard (2004), Nonlinear spectral mixing: Quantitative analysis of laboratory mineral mixtures, *J. Geophys. Res.-Planets*, 109(E2).
- Poulet, F., N. Mangold, D. Loizeau, J. P. Bibring, Y. Langevin, J. Michalski, and B. Gondet (2008a), Abundance of minerals in the phyllosilicate-rich units on Mars, *Astronomy & Astrophysics*, 487(2), L41–U193, doi:10.1051/0004-6361:200810150.
- Poulet, F., R. E. Arvidson, C. Gomez, R. V. Morris, J.-P. Bibring, Y. Langevin, B. Gondet, and J. Griffes (2008b), Mineralogy of Terra Meridiani and western Arabia Terra from OMEGA/MEx and implications for their formation, *Icarus*, 195(1), 106–130, doi:10.1016/j.icarus.2007.11.031.
- Poulet, F., J.-P. Bibring, Y. Langevin, J. F. Mustard, N. Mangold, M. Vincendon, B. Gondet, P. Pinet, J.-M. Bardintzeff, and B. Platevoet (2009), Quantitative compositional analysis of martian mafic regions using the MEx/OMEGA reflectance data 1. Methodology, uncertainties and examples of application, *Icarus*, 201(1), 69–83, doi:10.1016/j.icarus.2008.12.025.
- Presley, M. A., and P. R. Christensen (1997), Thermal conductivity measurements of particulate materials 2. Results, *J. Geophys. Res.*, 102(E3), 6551–6566, doi:10.1029/96JE03303.
- Presley, M. A., and R. A. Craddock (2006), Thermal conductivity measurements of particulate materials: 3. Natural samples and mixtures of particle sizes, *J. Geophys. Res.*, 111(E9), doi:10.1029/2006JE002706.
- Presley, M. A., R. A. Craddock, and N. Zolotova (2009), The effect of salt crust on the thermal conductivity of one sample of fluvial particulate materials under Martian atmospheric pressures, *J. Geophys. Res.*, 114(E11), doi:10.1029/2009JE003355.
- Putzig, N., and M. Mellon (2007a), Apparent thermal inertia and the surface heterogeneity of Mars, *Icarus*, 191(1), 68–94, doi:10.1016/j.icarus.2007.05.013.
- Putzig, N., and M. Mellon (2007b), Thermal behavior of horizontally mixed surfaces on Mars, *Icarus*, 191, 52–67, doi:10.1016/j.icarus.2007.03.022.
- Pye, K., and H. Tsoar (1990), *Aeolian sand and sand dunes*, Unwin Hyman, London.
- Quantin, C., P. Allemand, N. Mangold, and C. Delacourt (2004), Ages of Valles Marineris (Mars) landslides and implications for canyon history, *Icarus*, 172(2), 555–572, doi:10.1016/j.icarus.2004.06.013.
- Quantin, C., J. Flahaut, H. Clenet, P. Allemand, and P. Thomas (2012), Composition and structures of the subsurface in the vicinity of Valles Marineris as revealed by central uplifts of impact craters, *Icarus*, 221(1), 436–452, doi:10.1016/j.icarus.2012.07.031.

- Rafkin, S., R. M. Haberle, and T. I. Michaels (2001), The Mars Regional Atmospheric Modeling System: Model Description and Selected Simulations, *Icarus*, 151(2), 228–256, doi:10.1006/icar.2001.6605.
- Rafkin, S. C. R., and T. I. Michaels (2003), Meteorological predictions for 2003 Mars Exploration Rover high-priority landing sites, *J. Geophys. Res.*, 108(E12), doi:10.1029/2002JE002027.
- Ramsey, M. S., P. R. Christensen, N. Lancaster, and D. A. Howard (1999), Identification of sand sources and transport pathways at the Kelso Dunes, California, using thermal infrared remote sensing, *Geological Society of America Bulletin*, 111(5), 646–662, doi:10.1130/0016-7606(1999)111<0646:IOSSAT>2.3.CO;2.
- Raupach, M. R., and J. J. Finnigan (1997), The influence of topography on meteorological variables and surface-atmosphere interactions, *Journal of Hydrology*, 190, 182–213.
- Roach, L. H. et al. (2007), CRISM Spectral Signatures of the North Polar Gypsum Dunes, p. Abstract #1970, Lunar and Planetary Institute, Houston.
- Roach, L. H., J. F. Mustard, S. L. Murchie, J. -P. Bibring, F. Forget, K. W. Lewis, O. Aharonson, M. Vincendon, and J. L. Bishop (2009), Testing evidence of recent hydration state change in sulfates on Mars, *J. Geophys. Res.*, 114, doi:10.1029/2008JE003245.
- Roach, L. H., J. F. Mustard, M. D. Lane, J. L. Bishop, and S. L. Murchie (2010a), Diagenetic haematite and sulfate assemblages in Valles Marineris, *Icarus*, 207(2), 659–674, doi:10.1016/j.icarus.2009.11.029.
- Roach, L. H., J. F. Mustard, G. Swayze, R. E. Milliken, J. L. Bishop, S. L. Murchie, and K. Lichtenberg (2010b), Hydrated mineral stratigraphy of Ius Chasma, Valles Marineris, *Icarus*, 206(1), 253–268, doi:10.1016/j.icarus.2009.09.003.
- Rogers, D., and P. R. Christensen (2003), Age relationship of basaltic and andesitic surface compositions on Mars: Analysis of high-resolution TES observations of the northern hemisphere, *J. Geophys. Res.*, 108(E4), doi:10.1029/2002JE001913.
- Rossi, A. P., G. Neukum, M. Pondrelli, S. van Gasselt, T. Zegers, E. Hauber, A. Chicarro, and B. Foing (2008), Large-scale spring deposits on Mars?, *J. Geophys. Res.*, 113(E8), doi:10.1029/2007JE003062.
- Ruff, S. W., and P. R. Christensen (2002), Bright and dark regions on Mars: Particle size and mineralogical characteristics based on Thermal Emission Spectrometer data, *J. Geophys. Res.*, 107, doi:10.1029/2001JE001580.
- Ruff, S. W., and P. R. Christensen (2007), Basaltic andesite, altered basalt, and a TES-based search for smectite clay minerals on Mars, *Geophysical Research Letters*, 34(10), doi:10.1029/2007GL029602.
- Russell, P. et al. (2008), Seasonally active frost-dust avalanches on a north polar scarp of Mars captured by HiRISE, *Geophys. Res. Lett.*, 35(23), L23204, doi:10.1029/2008GL035790.
- Sagan, C. et al. (1973), Variable Features on Mars, 2, Mariner 9 Global Results, *J. Geophys. Res.*, 78(20), 4163–4196, doi:10.1029/JB078i020p04163.
- Salvatore, M. R., J. F. Mustard, M. B. Wyatt, and S. L. Murchie (2010), Definitive evidence of Hesperian basalt in Acidalia and Chryse planitiae, *J. Geophys. Res.*, 115(E7), doi:10.1029/2009JE003519.
- Schultz, R. A., and J. Lin (2001), Three-dimensional normal faulting models of the Valles Marineris, Mars, and geodynamic implications, *J. Geophys. Res.*, 106(B8), 16549–16566, doi:10.1029/2001JB000378.

- Scott, D. H., and K. L. Tanaka (1986), Geology map of the western equatorial region of Mars, U.S. Geol. Surv. Misc. Invest. Map, I-1802-A, scale 1:15,000,000.,
- Seelos, F. P., S. L. Murchie, D. C. Humm, O. S. Barnouin, F. Morgan, H. W. Taylor, C. Hash, and CRISM Team (2011), CRISM Data Processing and Analysis Products Update: Calibration, Correction, and Visualization, p. Abstract #1438, Lunar and Planetary Institute, Houston.
- Silvestro, S., G. Di Achille, and G. G. Ori (2010a), Dune morphology, sand transport pathways and possible source areas in east Thaumasia Region (Mars), *Geomorphology*, 121(1-2), 84–97, doi:10.1016/j.geomorph.2009.07.019.
- Silvestro, S., L. K. Fenton, D. A. Vaz, N. T. Bridges, and G. G. Ori (2010b), Ripple migration and dune activity on Mars: Evidence for dynamic wind processes, *Geophys. Res. Lett.*, 37(20), L20203.
- Silvestro, S., D. A. Vaz, L. K. Fenton, and P. E. Geissler (2011), Active aeolian processes on Mars: A regional study in Arabia and Meridiani Terrae, *Geophysical Research Letters*, 38, doi:10.1029/2011GL048955.
- Silvestro, S., L. K. Fenton, T. I. Michaels, A. Valdez, and G. G. Ori (2012a), Interpretation of the complex dune morphology on Mars: dune activity, modelling and a terrestrial analogue, *Earth Surface Processes and Landforms*, 37(13), 1424–1436, doi:10.1002/esp.3286.
- Silvestro, S., L. K. Fenton, T. I. Michaels, A. Valdez, and G. G. Ori (2012b), Interpretation of the complex dune morphology on Mars: dune activity, modelling and a terrestrial analogue, *Earth Surf. Process. Landforms*, n/a, doi:10.1002/esp.3286.
- Skinner, J. A., T. M. Hare, and K. L. Tanaka (2006), Digital Renovation of the Atlas of Mars 1:15,000,000-Scale Global Geologic Series Maps, p. Abstract #2331.
- Smith, D. E. et al. (2001), Mars Orbiter Laser Altimeter: Experiment summary after the first year of global mapping of Mars, *J. Geophys. Res.*, 106(E10), 23689–23722, doi:10.1029/2000JE001364.
- Smith, M. D., B. J. Conrath, J. C. Pearl, and P. R. Christensen (2002), Thermal Emission Spectrometer Observations of Martian Planet-Encircling Dust Storm 2001A, *Icarus*, 157(1), 259–263, doi:10.1006/icar.2001.6797.
- Soderblom, L., and J. Bell (2008), Introduction and Historical Perspective: 1. Exploration of the Martian surface: 1992–2007, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell, pp. 3–20, Cambridge University Press, Cambridge.
- Solomon, S. C. et al. (2005), New Perspectives on Ancient Mars, *Science*, 307(5713), 1214–1220, doi:10.1126/science.1101812.
- Spirn, A. W. (1986), *Air Quality at the Street-Level: Strategies for Urban Design*, Cambridge: Harvard Graduate School of Design.
- Squyres, S. W. et al. (2004), The Opportunity Rover's Athena science investigation at Meridiani Planum, Mars, *Science*, 306(5702), 1698–1703.
- Stockstill-Cahill, K. R., F. S. Anderson, and V. E. Hamilton (2008), A study of low-albedo deposits within Amazonis Planitia craters: Evidence for locally derived ultramafic to mafic materials, *J. Geophys. Res.*, 113(E7), doi:10.1029/2007JE003036.
- Stopar, J. D., G. Jeffrey Taylor, V. E. Hamilton, and L. Browning (2006), Kinetic model of olivine dissolution and extent of aqueous alteration on mars, *Geochimica et Cosmochimica Acta*, 70(24), 6136–6152, doi:10.1016/j.gca.2006.07.039.
- Sullivan, R. et al. (2005), Aeolian processes at the Mars Exploration Rover Meridiani Planum landing site, *Nature*, 436(7047), 58–61, doi:10.1038/nature03641.

- Sullivan, R. et al. (2008), Wind-driven particle mobility on Mars: Insights from Mars Exploration Rover observations at “El Dorado” and surroundings at Gusev Crater, *J. Geophys. Res.*, 113(E6), doi:10.1029/2008JE003101.
- Thomas, P., and C. Weitz (1989), Sand dune materials and polar layered deposits on Mars, *Icarus*, 81(1), 185–215, doi:10.1016/0019-1035(89)90133-4.
- Tirsch, D. (2009), Dark aeolian sediments in Martian craters: Composition and sources, Ph.D. dissertation, Freie Univ., Berlin.
- Tirsch, D., R. Jaumann, A. Pacifici, and F. Poulet (2011), Dark aeolian sediments in Martian craters: Composition and sources, *J. Geophys. Res.*, 116(E3), doi:10.1029/2009JE003562.
- Tirsch, D., R. A. Craddock, T. Platz, A. Maturilli, J. Helbert, and R. Jaumann (2012), Spectral and petrologic analyses of basaltic sands in Ka’u Desert (Hawaii) - implications for the dark dunes on Mars, *Earth Surface Processes and Landforms*, 37(4), 434–448, doi:10.1002/esp.2266.
- Toigo, A. D., and M. I. Richardson (2003), Meteorology of proposed Mars Exploration Rover landing sites, *J. Geophys. Res.*, 108, doi:10.1029/2003JE002064.
- Tornabene, L. L., J. E. Moersch, H. Y. McSween, A. S. McEwen, J. L. Piatek, K. A. Milam, and P. R. Christensen (2006), Identification of large (2–10 km) rayed craters on Mars in THEMIS thermal infrared images: Implications for possible Martian meteorite source regions, *Journal of Geophysical Research*, 111(E10), doi:10.1029/2005JE002600.
- Tsoar, H., B. White, and E. Berman (1996), The effect of slopes on sand transport — numerical modelling, *Landscape and Urban Planning*, 34(3–4), 171–181, doi:10.1016/0169-2046(95)00235-9.
- Vaniman, D. T., D. L. Bish, S. J. Chipera, C. I. Fialips, J. William Carey, and W. C. Feldman (2004), Magnesium sulphate salts and the history of water on Mars, *Nature*, 431(7009), 663–665, doi:10.1038/nature02973.
- Vardoulakis, S., B. E. . Fisher, K. Pericleous, and N. Gonzalez-Flesca (2003), Modelling air quality in street canyons: a review, *Atmospheric Environment*, 37(2), 155–182, doi:10.1016/S1352-2310(02)00857-9.
- Wiggs, G., J. Bullard, B. Garvey, and I. Castro (2002), Interactions Between Airflow and Valley Topography with Implications for Aeolian Sediment Transport, *Physical Geography*, 23(5), 366–380, doi:10.2747/0272-3646.23.5.366.
- Williams, J.-P., D. A. Paige, and C. E. Manning (2003), Layering in the wall rock of Valles Marineris: intrusive and extrusive magmatism, *Geophys. Res. Lett.*, 30(12), doi:10.1029/2003GL017662.
- Wilson, I. G. (1972), Universal discontinuities in bedforms produced by the wind, *Journal of Sedimentary Research*, 42(3), 667–669, doi:10.1306/74D725EF-2B21-11D7-8648000102C1865D.
- Wilson, I. G. (1973), *Ergs*, *Sedimentary Geology*, 10(2), 77–106, doi:10.1016/0037-0738(73)90001-8.
- Witbeck, N. E., K. L. Tanaka, and D. H. Scott (1991), *Geologic Map of Valles Marineris Region, Mars*, MAP I-2010, U.S. Geol. Surv., Reston, Va.,
- Wolff, M. J., M. D. Smith, R. T. Clancy, R. Arvidson, M. Kahre, F. Seelos, S. Murchie, and H. Savijärvi (2009), Wavelength dependence of dust aerosol single scattering albedo as observed by the Compact Reconnaissance Imaging Spectrometer, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003350.

- Wyatt, M. B., and H. Y. McSween (2002), Spectral evidence for weathered basalt as an alternative to andesite in the northern lowlands of Mars, *Nature*, 417(6886), 263–266, doi:10.1038/417263a.
- Xu, Z. X., K. Takeuchi, and H. Ishidaira (2003), Monotonic trend and step changes in Japanese precipitation, *Journal of Hydrology*, 279(1–4), 144–150, doi:10.1016/S0022-1694(03)00178-1.
- Yen, A. S. et al. (2005), An integrated view of the chemistry and mineralogy of martian soils, *Nature*, 436(7047), 49–54, doi:10.1038/nature03637.
- Yen, A. S. et al. (2013), Evidence for a Global Martian Soil Composition Extends to Gale Crater, p. Abstract #2495, Lunar and Planetary Institute, Houston.
- Yue, S., and C. Wang (2002), The influence of serial correlation on the Mann–Whitney test for detecting a shift in median, *Advances in Water Resources*, 25(3), 325–333, doi:10.1016/S0309-1708(01)00049-5.
- Zimbelman, J. R. (2010), Transverse Aeolian Ridges on Mars: First results from HiRISE images, *Geomorphology*, 121(1-2), 22–29, doi:10.1016/j.geomorph.2009.05.012.

Appendix IV

Table 4.1 CRISM Spectral Parameters Used in This Study^a.

Name	Parameter	Rationale
RBR	red/blue ratio	rock/dust ratio
BD530	0.53 μm band depth	crystalline ferric minerals
RPEAK1	reflectance peak at 1 μm	Fe mineralogy
BDI1000VIS	1 μm integrated band depth; VIS wavelengths	crystalline Fe^{+2} or Fe^{+3} minerals
IRA	1.3 μm reflectance	IR albedo
OLINDEX	olivine index	olivine will be strongly positive; favors fayalite
OLINDEX2 ^b	olivine index 2	olivine will be strongly positive; less influenced by albedo
LCPINDEX	pyroxene index	pyroxene will be strongly positive; favors LCP
HCPXINDEX	pyroxene index	pyroxene will be strongly positive; favors HCP
VAR	spectral variance between 1 μm to 2.3 μm	olivine and pyroxene will have high values
BD1900R ^b	1.9 μm band depth	H ₂ O, chemically bound or adsorbed
BDI2000	2 μm integrated band depth	pyroxene abundance and particle size
BD2100	2.1 μm band depth	monohydrated minerals
D2300	2.32 μm relative to 2.17 μm	hydrated min; particularly Fe/Mg phyllosilicates
SINDEX	drop in reflectance from ~2.3 μm to ~2.4 μm	indicative of hydrated minerals, particularly sulfates

^aAll spectral parameters are from *Pelkey et al.* [2007] unless otherwise indicated.

^b*Salvatore et al.* [2010].

Table 4.2 Non-VM^a CRISM Observations^b Used in This Study.

CRISM Targeted	CRISM Multispectral
FRT128F3	MSP117BD_01
FRT1380F	MSP3717_07
FRT147A8	MSP3FAF_05
FRT148DB	MSP49EA_07
FRT1646D	MSP55CD_03
FRT16A9E	MSP5C9A_05
FRT45F2	MSP624D_01
FRT7F0E	MSP8D60_01
FRT8236	MSPB325_01
FRT8541	MSW3125_01
FRT8FE6	MSW5161_01
FRT93BE	
FRT9620	
FRT98AB	
FRT9905	
FRT99C4	
FRTA492	
FRTA5A6	
FRTCA5B	
FRTCF49	
HRL10815	
HRL18642	
HRL1D9EA	
HRL3551	
HRL434E	
HRL5C69	
HRLC1CE	
HRS1BC38	
HRS48EB	

^aNon-Valles Marineris dune fields between 50°N–50°S.

^bObservation type and resolution for Tables 4.2–4.3 are as follows: full resolution targeted (FRT) 18 m/pixel, half resolution targeted (HRL) 36 m/pixel, half resolution short targeted (HRS) 36 m/pixel, multispectral window (MSW) 100 m/pixel, and multispectral survey (MSP) 200 m/pixel observation.

Table 4.3 Valles Marineris CRISM Observations Used in This Study.

CRISM Targeted	CRISM Multispectral
FRT125AD	MSP1AA73_01
FRT126D4	MSP10600_01
FRT14969	MSP13D02_01
FRT163F5	MSP14646_01
FRT1693A	MSP1495E_01
FRT16E3D	MSP14C8E_01
FRT16E9C	MSP14D58_01
FRT181D0	MSP157CE_03
FRT1FD85	MSP1AAF1_05
FRT21CB0	MSP1AC4A_01
FRT68A0	MSP33F3_05
FRT7FAA	MSP361C_03
FRT85A4	MSP3DEC_01
FRT93E3	MSP3F10_03
FRT9445	MSP42B7_07
FRTBFE4	MSP4477_05
HRL12E34	MSP44E2_07
HRL177C5	MSP46F2_05
HRL28A6	MSP4824_07
HRL444C	MSP5042_01
HRL633F	MSP5D0E_05
HRL81EF	MSP79F6_01
HRL843E	MSPC156_01
HRL89A2	MSPC3CF_01
HRL9B5	MSPCF0F_01
HRLC038	MSW2FFA_01
HRLD770	MSW42B3_01
HRS18E7B	MSW43F5_01
HRS1E2FB	MSW517D_01
HRSB8EB	MSW5B28_01
	MSW6393_01
	MSW6671_01
	MSW9DF5_01

Table 4.4 Non-VM Dune Fields^a Used in This Study.

Dune Field ID ^b	Quadrangle ^c	Area (km ²)	Dune Field ID ^b	Quadrangle ^c	Area (km ²)
0020-494	Noachis	56	0455-415	Noachis	22
0045+031	Arabia	28	0460-411	Noachis	7
0067+071	Arabia	115	0461-382	Noachis	21
0081+097	Arabia	156	0461-402	Noachis	32
0090+130	Arabia	35	0462-408	Noachis	214
0101+066	Arabia	82	0465-417	Noachis	501
0113-497	Noachis	119	0472-322	Noachis	1
0129+159	Arabia	176	0565-454	Noachis	39
0152-425	Noachis	8	0570-021	Iapygia	912
0176-432	Noachis	149	0597+369	Ismenius Lacus	13
0193-486	Noachis	73	0611-481	Hellas	61
0194-468	Noachis	1583	0628+266	Syrtis Major	316
0288-450	Noachis	257	0649+342	Casius	15
0289+489	Ismenius Lacus	424	0656+228	Syrtis Major	829
0289+500	Ismenius Lacus	20	0671+088	Syrtis Major	528
0290+491	Ismenius Lacus	11	0675-490	Hellas	183
0295+489	Ismenius Lacus	16	0680+068	Syrtis Major	906
0304-475	Noachis	1924	0682-448	Hellas	22
0326-500	Noachis	117	0688-479	Hellas	5
0338-472	Noachis	395	0700-429	Hellas	20
0339-454	Noachis	18	0739-425	Hellas	11
0339-488	Noachis	257	0740-426	Hellas	8
0345-407	Noachis	77	0753-434	Hellas	77
0346-494	Noachis	211	0754-356	Hellas	9
0347-437	Noachis	1310	0755+207	Syrtis Major	100
0348-463	Noachis	80	0791-449	Hellas	76
0366-456	Noachis	309	0791+202	Syrtis Major	186
0372-471	Noachis	62	0793+198	Syrtis Major	92
0374-411	Noachis	18	0795+196	Syrtis Major	10
0376-486	Noachis	47	0887-488	Hellas	48
0380-426	Noachis	67	0890-478	Hellas	124
0380-447	Noachis	7	0894-475	Hellas	10
0386-448	Noachis	117	0894-476	Hellas	17
0388-454	Noachis	8	0937-089	Tyrrhenum	222
0389-485	Noachis	63	0939-069	Tyrrhenum	49
0395-432	Noachis	114	1249-136	Tyrrhenum	477
0419-449	Noachis	24	1279-148	Tyrrhenum	601
0420-423	Noachis	8	1283-141	Tyrrhenum	186
0443+417	Ismenius Lacus	678	1284-162	Tyrrhenum	201
0444-267	Sinus Sabaeus	7	1284-163	Tyrrhenum	22
1286-143	Tyrrhenum	155	3072-482	Argyre	147
1335-347	Eridania	4	3073-462	Argyre	9
1335-348	Eridania	5	3074-478	Argyre	23
1370-050	Aeolis	425	3075-458	Argyre	20
1375-177	Aeolis	69	3076-032	Coprates	40
1376-059	Aeolis	414	3076-447	Argyre	81
1384-341	Eridania	34	3078-439	Argyre	35
1411-362	Eridania	10	3079-445	Argyre	47
1480-435	Eridania	103	3080-466	Argyre	5
1539-491	Eridania	52	3082-392	Argyre	203
1541-381	Eridania	60	3082-431	Argyre	6

Table 4.4 Continued

Dune Field ID ^b	Quadrangle ^c	Area (km ²)	Dune Field ID ^b	Quadrangle ^c	Area (km ²)
1544-475	Eridania	44	3082-494	Argyre	55
1553-490	Eridania	16	3085-489	Argyre	110
1557-486	Eridania	43	3088-432	Argyre	28
1586-459	Eridania	12	3089-430	Argyre	9
1593+111	Elysium	122	3090-369	Argyre	48
1627-489	Eridania	13	3094-399	Argyre	22
1858+118	Amazonis	47	3096-436	Argyre	4
1967-459	Phaethontis	24	3097-395	Argyre	183
2019-421	Phaethontis	37	3098-404	Argyre	52
2021-415	Phaethontis	38	3102-486	Argyre	807
2427-491	Thaumasia	6	3105-459	Argyre	213
2428-465	Thaumasia	8	3108-468	Argyre	65
2938-497	Thaumasia	1904	3109-466	Argyre	23
2975-411	Thaumasia	104	3114-403	Argyre	69
2976-437	Thaumasia	63	3122-395	Argyre	9
2995-434	Thaumasia	38	3122-397	Argyre	79
3010-480	Argyre	118	3128-440	Argyre	54
3034-437	Argyre	67	3134-004	Coprates	105
3037-474	Argyre	75	3153-446	Argyre	629
3037-480	Argyre	4	3158-451	Argyre	823
3046-476	Argyre	176	3203-217	Margaritifer Sinus	168
3048-436	Argyre	22	3233-354	Argyre	8
3049+082	Lunae Palus	9	3237-439	Argyre	32
3050-441	Argyre	11	3241-353	Argyre	341
3052-421	Argyre	110	3264-256	Margaritifer Sinus	233
3057-450	Argyre	58	3313-381	Argyre	27
3057-468	Argyre	230	3347-481	Argyre	223
3059-453	Argyre	20	3352-407	Argyre	313
3060+347	Mare Acidalium	30	3423+137	Oxia Palus	30
3069-411	Argyre	148	3438+040	Oxia Palus	42
3070-468	Argyre	20	3443+084	Oxia Palus	245
3072-459	Argyre	29	3450+133	Oxia Palus	10
3468+151	Oxia Palus	68			
3469+180	Oxia Palus	17			
3470-331	Argyre	32			
3471+158	Oxia Palus	30			
3487+005	Oxia Palus	83			
3492+185	Oxia Palus	126			
3493+053	Oxia Palus	83			
3499+053	Oxia Palus	19			
3502+047	Oxia Palus	57			
3518+013	Oxia Palus	6			
3521+072	Oxia Palus	22			
3525+102	Oxia Palus	21			
3527+107	Oxia Palus	308			
3527+126	Oxia Palus	94			
3528+080	Oxia Palus	145			
3572+107	Oxia Palus	23			
3573-492	Argyre	79			
Total		28,338 km ²			

Table 4.4 Continued

^aMars Global Digital Dune Database polygons [*Hayward et al.*, 2007b
<http://pubs.usgs.gov/of/2007/1158/>] between 50°N–50°S.

^bDune field IDs where the first four digits are the dune field's centroid East longitude, last three digits are the dune field's latitude (each coordinate given to the first decimal place without the decimal), and the separating + or – sign indicating which hemisphere.

^cThe 1:5 M scale Mars Chart quadrangle region the dune field is located in.

Table 4.5 Valles Marineris Dune Field Morphologic Data.

Dune Field ID	Location ^a	Area ^b (km ²)	SF-AzMean ^c	Dune Type ^d	Class ^e	Geographic Association ^f	Mean Elevation ^g (km)	Relief ^h (km)	Slope ⁱ (°)
2753-078	NW Ius	22	291	SS, B, U	F	Floor, LS	-3.1	0.6	2.9+/-1.4
2763-078	SW Ius	135	110	Bd, T, U	F	Floor, Wall	-2.9	2.2	4.3+/-1.9
2765-712	NW Ius	4	95	SS, Bd, D, E	F	Floor	-4.2	0.1	1.8+/-1.0
2779-805	SW Ius	10	277	Bd, T, B	F	Floor, Wall	-3.7	0.5	4.9+/-0.7
2811-086	SE Ius	226	272	Bd, SS, B, U	F	Floor, Wall	-4.1	0.7	1.9+/-1.2
2816-871_LS	SE Ius	127	294	Bd, U	LS?	LS	-3.5	0.5	2.8+/-1.6
2821-089	SE Ius	125	99	Bd, T, SS	F	LS	-3.6	1.3	3.6+/-1.7
2822-092	W Hebes	217	296	SS, T, U	F	ILD	-2.7	3.7	7+/-2.0
2823-078_LS	NE Ius	29	145	Bd, T, D	LS	LS	-3.5	0.9	3.2+/-1.6
2836-005	N Hebes	39	264	Bd, SS, T	F?	ILD, Wall	-2.8	0.7	3.8+/-2.0
2845-155_CD	SE Hebes	1	180	Cd, B, L	CD	Wall	-1.8	1.0	6.4+/-1.3
2846-126	E Hebes	54	182	Bd, T, D	F?	Floor, ILD	-3.6	0.8	3.1+/-1.3
2849-116_CD	SW Melas	9	204	Cd, Bd	CD	Wall	-1.0	1.2	9+/-3.9
2850-120_LS	N Hebes	43	53	Bd, T	LS	ILD, LS	-3.3	2.0	3.2+/-1.7
2856-113	SW Melas	271	205	Bd, SS, U	F	Floor, ILD	-1.0	0.6	1.2+/-0.6
2886-131	SE Melas	219	251	Bd, L, U	F	ILD	-2.5	2.1	2.3+/-1.1
2886-133_CD	SE Melas	82	177	Cd, Bd	CD	Wall	-2.6	2.1	8.1+/-3.2
2890-122	S Melas	13	149	Bd	F	Floor, ILD	-2.9	0.3	1.8+/-0.4
2894-099	NE Melas	101	116	Bd, T, D	F	Wall, LS, ILD	-4.5	0.4	1.8+/-0.6
2901-101_LS	NE Melas	177	105	Bd, T, Fd?	LS	Wall	-4.3	1.8	3.9+/-2.1
2901-119	SE Melas	133	127	T, SS	F	Floor	-5.0	0.2	3.9+/-2.1
2908-105	NE Melas	52	99	T, Bd	F	ILD, Wall	-4.6	1.2	2.2+/-1.1
2910-107_LS	NE Melas	8	294	B, Bd	LS	LS, Wall	-4.8	0.1	3.9+/-2.1
2946-126_CD	SW Coprates	23	159	Cd, B	CD	Wall	-3.6	2.4	6.2+/-3.1
2968-140	SW Coprates	56	164	B, Bd	F	Floor, LS	-2.8	1.0	4.5+/-3.8
2971-046_ERG	Juventae	4634	132	Bd, SS, T, E	ERG	ILD, Floor,	-3.0	2.9	2.3+/-1.6

Table 4.5 (continued)

Dune Field ID	Location ^a	Area ^b (km ²)	SF-AzMean ^c	Dune Type ^d	Class ^e	Geographic Association ^f	Mean Elevation ^g (km)	Relief ^h (km)	Slope ⁱ (°)
2979-331	N Juventae	8	130	T, Bd, B, D, E	F	ILD	-1.3	0.3	3.4+/-1.4
2982-136_CD	W Coprates	34	175	Cd, Bd	CD	Wall	-1.3	2.7	7.4+/-3.1
2984-347	N Juventae	109	131	Bd, SS, E, B	F	Chaos	-3.2	0.5	2.4+/-1.6
2985-264	N Juventae	6	149	Bd, T	F	Chaos	-2.4	0.2	3.1+/-1.6
3018-142	Coprates	240	199	Bd, T, SS	F	Wall	-4.6	2.4	3.2+/-2.5
3019-144_FD	Coprates	20	9	Fd, Bd, Cd	FD	Wall	-2.6	2.6	10.5+/-2.7
3023-153	S of Coprates	159	106	Bd, T, Fd?	F	Wall	-2.8	2.0	6.8+/-3.5
3023-155	S of Coprates	22	136	Bd, T, SS	F	Wall	-1.2	0.5	2.7+/-1.2
3025-144	Coprates	45	260	Bd, T, B	F	Floor, Wall	-5.0	0.5	1.9+/-2.1
3038-143_FD	W Coprates massif	10	271	Fd	FD	Wall	-3.4	1.1	5.6+/-2.6
3043-150	Coprates	392	291	B, Bd, T	F	Floor, Wall	-4.9	1.0	2.5+/-3.1
3047-152_LS	Coprates	62	258	Bd	LS	LS, Floor, Wall	-4.7	0.4	1.7+/-0.9
3049-148	SE Coprates	10	324	Bd, T	F	Wall	-3.9	1.0	16.8+/-1.8
3057-145_FD	N Coprates massif	11	15	Fd, L	FD	Wall	-2.6	0.7	8.7+/-1.5
3060-155_FD	S Coprates massif	26	16	Fd, L	FD	Wall	-2.0	1.2	6.7+/-3.2
3062-148_FD	S Coprates massif	11	144	Fd, Bd, L	FD	Wall	-2.3	2.1	6.9+/-2.1
3064-143	NE Coprates	216	301	Bd, B, D	F	Floor, Wall	-4.7	0.9	1.9+/-2.9
3067-155	SE Coprates	25	263	T, Bd, Fd?	F	Floor, Wall	-4.5	1.3	3.7+/-2.4
3075-149	E Coprates	17	143	L, T	F	Floor, Wall	-4.8	0.7	3.4+/-3.4
3089-130	N Capri	10	352	SS, Bd	F	Floor, Wall	-4.3	0.8	6.3+/-4.3
3089-142	W Capri	72	286	B, SS, U	F	Floor	-4.7	0.5	1+/-0.8
3090-147	W Capri	20	357	Bd, L, SS	F	Floor	-4.3	0.5	3.4+/-1.8
3107-068	N Ganges	214	302	Bd, SS, L	F	ILD, LS	-3.3	2.4	3.8+/-2.3
3113-068	N Ganges	105	0	Bd, SS	F	ILD, Chaos	-3.5	1.0	2.1+/-2.2
3116-131	Capri	4	126	T, Bd	F	ILD	-2.9	0.7	6.2+/-2.0
3120-070	N Ganges	122	83	SS, Bd, E	F	ILD, Chaos	-3.7	0.3	1.1+/-1.2
3123-132	Capri	33	208	Bd, SS	F	Crater, ILD	-3.0	0.3	1.9+/-1.1

Table 4.5 (continued)

Dune Field ID	Location ^a	Area ^b (km ²)	SF-AzMean ^c	Dune Type ^d	Class ^e	Geographic Association ^f	Mean Elevation ^g (km)	Relief ^h (km)	Slope ⁱ (°)
3086-798_ERG	W Ganges	3155	(W 127) & (E 247)	SS, Bd, E	ERG	Floor, ILD, Chaos, Wall	-3.7	1.6	2.3+/-1.9
3115-820_ERG	C Ganges	8036	316	SS, Bd, E	ERG	Floor, ILD, Chaos, Wall	-3.7	1.4	1.6+/-1.5
3138-869_ERG	E Ganges	1050	232	SS, Bd, T, E	ERG	Floor, ILD, Chaos, Wall	-3.6	0.7	1+/-0.9
3135-127	Capri	64	248	Bd, T, D	F	ILD	-2.6	1.5	2+/-1.1
3145-131	S Capri	16	71	Bd, T	F	Chaos	-3.6	0.3	2.7+/-2.1
3147-076	Ganges	355	178	Bd, SS, T, L, B, D	F	LS, Wall	-3.7	0.1	1.1+/-0.6
3154-078	E Ganges	141	171	Bd,T, SS, L, B, D	F	LS	-3.6	0.2	0.9+/-0.7

^aDune field IDs where the first four digits are the dune field's centroid East longitude, last three digits are the dune field's latitude (each coordinate given to the first decimal place without the decimal), and the separating – sign indicates the hemisphere. Dune field class identifier follows the ID for landslide (LS), falling (FD), climbing (CD), and erg (ERG), where the remainder are floor dune fields.

^bDune field location in VM.

^cAverage slipface azimuth as calculated using the ArcGIS linear directional mean tool from all resolved slipface measurements.

^dDune morphologies identified within the dune field: barchan (B), barchanoid (Bd), linear (L), transverse (T), sand sheet (SS), dome (D), falling dune (Fd), climbing dune (Cd), and echo dune (E). “?” indicates some ambiguity in interpretation.

^eDune field class: floor (F), landslide (LS), falling (FD), climbing (CD) and erg (ERG).

^fAdjacent geographic units: Floor, ILDs, Landslides (LS), Wall, and Chaos.

^gMean MOLA elevation for dune field polygons.

^hMOLA-derived relief as calculated from the minimum elevation subtracted from maximum values within the dune field polygon.

ⁱMean MOLA-derived slope and standard deviation measurements for dune field polygons.

Table 4.6 Comparison of Measured and Derived Properties Between Non-VM and Valles Marineris Dune Fields.

		Mid-Latitude Dune Fields (+/-50° latitude)	Valles Marineris	Similar or Different?	Statistically Separable ^a
Area		frequency peaks at 16–62 km ² (~155 km ² mean), no ergs >2000 km ²	frequency peaks at 8–125 km ² (~400 km ² mean), three ergs >2000 km ²	different	N/A
Morphology		barchanoid, barchan, sand sheet, transverse, linear, dome and star	barchanoid, transverse, sand sheet, falling, barchan, linear, dome, climbing, echo	different	N/A
Major Landforms		craters, troughs, intracrater plains	linear chasmata, hummocky landslides, spur-and-gullies, ILDs, collapse pits, craters	more variable in VM	N/A
Elevation ^b		-7.6 km – +1.7 km (-1.7 km mean)	-5.1 km – +0.4 km (-3.6 km mean)	similar, slightly lower in VM	N/A
Relief		100 m – 1900 m (~400 m mean)	100 m – 3700 m (~1200 m mean)	substantially greater in VM	yes
Underlying Slope		0.2° – 4.4° (1.4° mean)	0.9° – 16.8° (4.3° mean)	substantially greater in VM	yes
Roughness	0.6 km	2.48E-05 m ⁻¹ – 2.97E-04 m ⁻¹ (1.17E-04 m ⁻¹ mean)	7.54E-05 m ⁻¹ – 3.07E-04 m ⁻¹ (1.62E-04 m ⁻¹ mean)	similar at shortest baseline, but greater for VM at larger baselines	yes
	2.4 km	5.82E-06 m ⁻¹ – 8.08E-05 m ⁻¹ (3.02E-05 m ⁻¹ mean)	1.82E-05 m ⁻¹ – 1.35E-04 m ⁻¹ (5.80E-05 m ⁻¹ mean)		
	9.2 km	8.25E-07 m ⁻¹ – 2.53E-05 m ⁻¹ (7.06E-06 m ⁻¹ mean)	3.87E-06 m ⁻¹ – 3.18E-05 m ⁻¹ (1.72E-05 m ⁻¹ mean)		
Terrain Age		~62% Noachian, ~30% Hesperian and ~1% Amazonian	~57% Hesperian and ~49% Amazonian	different	N/A
Albedo	MY26	0.09 – 0.23 (0.14 mean).	0.09 – 0.25 (0.13 mean)	similar, slightly darker in VM	yes
Thermal Inertia	Nightside	180 – 390 tiu (280 tiu mean)	200 – 390 tiu (290 tiu mean)	similar	no
	ΔTI	+25 – +55 tiu (30 tiu mean)	+25 – 170 tiu (55 tiu mean)	slightly greater in VM	yes

Table 4.6 Continued

		Mid-Latitude Dune Fields (+/-50° latitude)	Valles Marineris	Similar or Different?	Statistically Separable ^a
Composition	VNIR	HCP, LCP, olivine, basalt	HCP, LCP, olivine, weathered-glass, kieserite	greater frequency of alteration products	N/A
	DCI	0.929 – 0.991 (0.970 mean; 82% "dust free")	0.931 – 0.975 (0.970 mean; 79% "dust free")	similar	yes
	465	0.995 – 1.017 (1.004 mean; 57% > 3-sigma)	0.994 – 1.015, 1.004 (27% > 3-sigma)	different, greater amount of Non-VM above threshold	no
	530	0.984 – 1.024 (1.002 mean; 58% > 3-sigma)	0.986 – 1.015, 1.000 (none > 3-sigma)	different, greater amount of Non-VM above threshold	yes
Surface Activity?	Photo-geologic	dune and ripple migration, dune perimeter modifications, slip face adjustment, dune deflation	ripple migration/re-alignment, dune perimeter modifications, slip face adjustment, dune deflation	similar	N/A
	Albedo	MY24–25: +20% – -30% (0% mean) MY25–26: +80% – -37% (0% mean)	MY24–25: +30% – -25% (-7% mean) MY25–26: +42% – -12% (+7% mean)	similar, but more spatially uniform in VM	yes

^aResults from the Mann-Whitney test (see Table 4.10). Here N/A, not applicable.

^bRange of values, along with the mean value, are provided for all quantitative parameters.

Table 4.7 Valles Marineris Dune Field TES Thermophysical Data.

Dune Field ID	Nightside Thermal Inertia ^a (tiu)	ΔT ^b (tiu)	MY 26 Albedo ^c	MY 24–25 Albedo Change ^d (%)	MY 25–26 Albedo Change ^d (%)
2753-078			0.147	no change ^d	25
2763-078			0.117	no change	no change
2765-712			0.153	no change	6
2779-805			0.143	no change	no change
2811-086	295+/-51	60	0.115	6	no change
2816-871_LS			0.135	-8	9
2821-089	220+/-30	34	0.131	no change	no change
2822-092			0.175	20	no change
2823-078_LS			0.160	no change	42
2836-005			0.162	-7	no change
2845-155_CD			0.130	no change	-6
2846-126			0.138	-13	7
2849-116_CD			0.110	30	no change
2850-120_LS			0.152	-8	6
2856-113	295+/-34	null	0.130	-16	7
2886-131	285+/-38	40	0.115	-6	15
2886-133_CD			0.127	-12	30
2890-122			0.135	-9	11
2894-099	298+/-86	81	0.133	-8	no change
2901-101_LS	346+/-51	170	0.143	-15	34
2901-119			0.138	-12	7
2908-105			0.138	-25	26
2910-107_LS			0.146	-16	19
2946-126_CD			0.139	-6	6
2968-140	366+/-37	124	0.117	-18	13
2971-046_ERG	231+/-34	null	0.129	-8	7
2979-331			0.132	no change	no change
2982-136_CD			0.105	no change	-12
2984-347			0.121	-9	no change
2985-264			0.154	-20	7
3018-142			0.125	-9	no change
3019-144_FD			0.119	no change	no change
3023-153			0.119	-12	26
3023-155			0.117	-9	no change
3025-144			0.123	-6	no change
3038-143_FD			0.132	no change	no change
3043-150	333+/-49	59	0.121	-6	16
3047-152_LS			0.113	no change	7

Table 4.7 Continued

Dune Field ID	Nightside Thermal Inertia ^a (tiu)	ΔTI^b (tiu)	MY 26 Albedo ^c	MY 24–25 Albedo Change ^d (%)	MY 25–26 Albedo Change ^d (%)
3049-148			0.116	-13	10
3057-145_FD			0.122	no change	no change
3060-155_FD			0.125	-9	12
3062-148_FD			0.126	no change	12
3064-143	332+/-62	null	0.125	-9	13
3067-155	338+/-107	92	0.129	-17	11
3075-149			0.135	-8	no change
3089-130			0.144	-10	23
3089-142			0.138	-6	8
3090-147			0.136	-6	12
3107-068	284+/-37	null	0.127	-12	no change
3113-068	325+/-29	102	0.130	-10	-6
3116-131			0.131	-12	no change
3120-070	284+/-22	69	0.149	no change	8
3123-132			0.131	-17	7
3086-798_ERG	211+/-26	26	0.136	-9	no change
3115-820_ERG	204+/-30	null	0.125	-13	no change
3138-869_ERG	249+/-27	26	0.134	-6	no change
3135-127	298+/-33	83	0.127	-8	no change
3145-131			0.118	no change	-10
3147-076	314+/-44	29	0.120	no change	no change
3154-078	307+/-11	35	0.121	-13	no change

^aData from *Putzig and Mellon, 2007a*. Only pixels of relatively homogenous, bedrock-free dune area were collected for large dune fields.

^bTES dayside-derived subtracted from nightside-derived thermal inertia [*Putzig and Mellon, 2007a*], where values <25 tiu (null) are within the expected model error [*Christensen et al., 2001b*].

^cTES bolometric albedo [*Putzig and Mellon, 2007a*] and the expected error is 1–2% [*Christensen et al., 2001b*].

^dTES albedo changes between -5% and +5% are reported as “no change”.

Table 4.8 Valles Marineris Dune Field THEMIS Thermal Inertia Data.

Dune Field ID	Morphology and Comment	Thermal Inertia ^a (tiu)	Estimated Particle Size ^b (mm)	THEMIS ID	Chasma	Dune Field Class ^c
2763-078	draa	340+/-40	3.5	I34691007	Ius	F
2763-078	eastern barchanoids	282+/-31	1.0	I34691007	Ius	F
2811-086	barchanoids	304+/-23	1.6	I09671005	Ius	F
2821-086	transverse dunes	264+/-25	0.7	I01345002	Ius	F
2823-078_LS	barchanoids near scarp	244+/-19	0.5	I01345002	Ius	LS
2836-005	barchanoids	226+/-28	0.4	I34204005	Hebes	F
2845-155_CD	barchan train	303+/-17	1.6	I07936014	Hebes	CD
2846-126	western barchanoids	263+/-15	0.6	I07936014	Hebes	F
2846-126	eastern sand sheet	260+/-11	0.6	I07936014	Hebes	F
2850-120_LS	barchanoids	254+/-15	0.5	I07936014	Hebes	LS
2849-116_CD	climbing dune in gully	265+/-5	0.9	I00933002	Melas	F
2856-113	dome dunes	295+/-11	1.4	I34204005	Melas	F
2886-131	northeastern transverse dunes	272+/-12	0.8	I16909010	Melas	F
2886-131	southeast sand sheets	225+/-21	0.3	I35377006	Melas	F
2886-133_CD	climbing dunes	240+/-11	0.4	I16909010	Melas	F
2890-122	barchanoids	286+/-13	1.2	I01894002	Melas	F
2894-099	dome dunes	307+/-17	1.7	I35040006	Melas	F
2901-101_LS	barchanoids atop landslide	320+/-17	2.1	I35040006	Melas	LS
2908-105	southern barchanoids	288+/-18	0.9	I34079005	Melas	F
2971-046_ERG	western barchanoids	326+/-35	2.7	I01669003	Juventae	ERG
2971-046_ERG	western sand sheet	253+/-22	0.6	I01669003	Juventae	ERG
2971-046_ERG	western barchanoids	283+/-16	1.1	I01669003	Juventae	ERG
2971-046_ERG	central sand sheet	220+/-15	0.3	I36899010	Juventae	ERG
2979-331	summit barchans	317+/-20	3.0	I36899010	Juventae	F
2968-140	barchans	315+/-9	1.9	I34266008	Coprates	F
3018-142	barchanoids at base of walls	348+/-21	4.9	I35414004	Coprates	F
3019-144_FD	falling dune in gully	371+/-19	8.4	I35414004	Coprates	FD
3019-144_FD	falling dune in gully	396+/-16	10.1	I35414004	Coprates	FD
3023-153	transverse dunes in collapse pit	229+/-20	2.8	I36250009	Coprates	F
3023-153	barchanoids in collapse pit	318+/-26	0.5	I36250009	Coprates	F
3023-155	barchanoids in collapse pit	273+/-19	1.2	I36250009	Coprates	F
3038-143_FD	falling dune in gully	359+/-17	21.3	I05264008	Coprates	FD
3062-148_FD	falling dune in gully	333+/-12	2.8	I08185014	Coprates	FD
3049-148	barchanoids at base of walls	342+/-15	2.8	I08185014	Coprates	F
3060-155_FD	falling dune in gully	321+/-19	2.2	I08185014	Coprates	FD
3064-143	eastern barchanoids	290+/-14	1.1	I01831004	Coprates	F
3064-143	western sand sheet	255+/-16	0.5	I01831004	Coprates	F

Table 4.8 Continued

Dune Field ID	Morphology and Comment	Thermal Inertia ^a (tiu)	Estimated Particle Size ^b (mm)	THEMIS ID	Chasma	Dune Field Class ^c
3089-142	western barchans	279+/-17	1.0	I35264006	Capri	F
3135-127	eastern barchanoids	300+/-33	1.7	I02742002	Capri	F
3135-127	eastern sand sheets	236+/-21	0.5	I02742002	Capri	F
3086-798_ERG	sand sheets	232+/-20	0.4	I35264006	Ganges	ERG
3107-068	southern barchanoids	261+/-19	0.6	I33941005	Ganges	F
3107-068	northern sand sheets	236+/-10	0.4	I33941005	Ganges	F
3113-068	northern barchanoids	260+/-15	0.7	I02405002	Ganges	F
3113-068	northern sand sheets	241+/-14	0.5	I02405002	Ganges	F
3115-820_ERG	sand sheets	229+/-19	0.4	I02405002	Ganges	ERG
3120-070	southern barchanoids	311+/-19	2.0	I35476012	Ganges	F
3120-070	northern sand sheets	289+/-21	1.2	I35476012	Ganges	F
3115-820_ERG	echo dunes	267+/-11	1.2	I35476012	Ganges	ERG
3138-869_ERG	eastern barchanoids	249+/-16	0.5	I35139005	Ganges	ERG
3138-869_ERG	sand sheets	232+/-14	0.4	I35139005	Ganges	ERG
3147-076	northern barchanoids	261+/-14	0.6	I35426006	Ganges	F
3147-076	southern sand sheets	239+/-13	0.4	I35426006	Ganges	F
3154-078	barchanoids	239+/-16	0.4	I36549007	Ganges	F

^aTHEMIS thermal inertia (see text for description).

^bEffective particle sizes were derived from THEMIS thermal inertia values, following the method of *Piqueux and Christensen* [2009] (see text for description). Absolute error is estimated to be ~10%.

^cDune field class: floor (F), landslide (LS), falling (FD), climbing (CD) and erg (ERG).

Table 4.9 Changes in Valles Marineris Dune/Bedform Morphologies.

Chasmata (Dune Field ID ^a)	Image ID (Relative Position)	DOA ^b	L _s and Season	Sub Solar Azimuth ^c	Incident Angle	Observation	Interpretation
Ius (2763-078)	PSP_009223_1720 (W)	7/15/08	98.8°, Southern Winter	153.2°	59.8°	new streaks, scars on slip faces	slip face avalanches / modification
	ESP_022607_1725 (E)	5/24/11	298.1°, Southern Summer	214.6°	36.9°		
Ius (2765-712)	PSP_003949_1725 (E)	5/31/07	111°, Southern Winter	155.4°	49.0°	changes in ripple orientation, changes in light-toned rock albedo	ripple re-alignment, aeolian reworking / sand deflation / sand deposition
	PSP_009579_1725 (W)	8/11/08	248°, Southern Spring	212.4°	59.0°		
Juventae (2979-331)	TRA_000875_1765 (W)	10/3/06	115.4°, Southern Winter	210.6	210.6	new scar and fan on slip face	slip face avalanche
	ESP_018624_1765 (E)	7/17/10	119.6°, Southern Winter	208.7°	208.7°		
Capri (3135-127)	ESP_020931_1675 (E)	1/13/11	216.5°, Southern Spring	357.0°	52.0°	new streaks, scars on slip faces	slip face avalanches / modification
	ESP_022276_1670 (W)	4/28/11	282.4°, Southern Summer	342.6°	39.0°		
Ganges (3147-076)	S0501628 (E)	4/26/05	200.16°, Southern Spring	358.9°	43.5°	reduction in dune area, dune change in location	dune deflation / dune migration
	PSP_008536_1725 (W)	5/22/08	75.3°, Southern Autumn	215.0°	57.0°		
	ESP_026100_1725 (S)	2/20/12	72.9°, Southern Autumn	215.0°	56.4°	ripple / dune change in location	ripple / dune migration

^aCompare with TES albedo changes in Table 4.7.^bImage date of acquisitions (DOA).^cAngle in degrees at the center of the image from the three o'clock position clockwise to the direction of the Sun.

Table 4.10 Mann-Whitney Results.

	Non-VM Population	VM Population	U(max) ^a	Z ^b	Prob> U ^c
Relief	184	60	2249	-6.93168	4.10E-13
Slope	184	60	1571	-8.42809	3.51E-17
Roughness (0.6 km)	170	40	1397	-5.79289	6.92E-09
Roughness (2.4 km)	170	40	1090	-6.67919	2.40E-11
Roughness (9.2 km)	170	40	1038	-6.83097	8.43E-12
Albedo	180	60	7559	4.39582	8.01E-6
Nightside Thermal Inertia	32	20	243.5	-1.42975	0.1527
Δ TI	32	20	206	-2.13608	0.0314
DCI	174	53	6155	3.68749	2.26E-04
465 Index	174	53	4480	-0.31496	0.7531
530 Index	174	53	5697	2.59325	0.0095
Albedo: MY24-25%	180	60	7783	4.92560	8.41E-07
Albedo: MY25-26%	180	60	2762	-5.79846	6.69E-09

^aMaximum unbiased (U) statistic calculated from the Mann–Whitney test.

^bStandardized score (Z) or normalized U statistic.

^cBelow the 0.05 level, the two distributions are statistically separable.

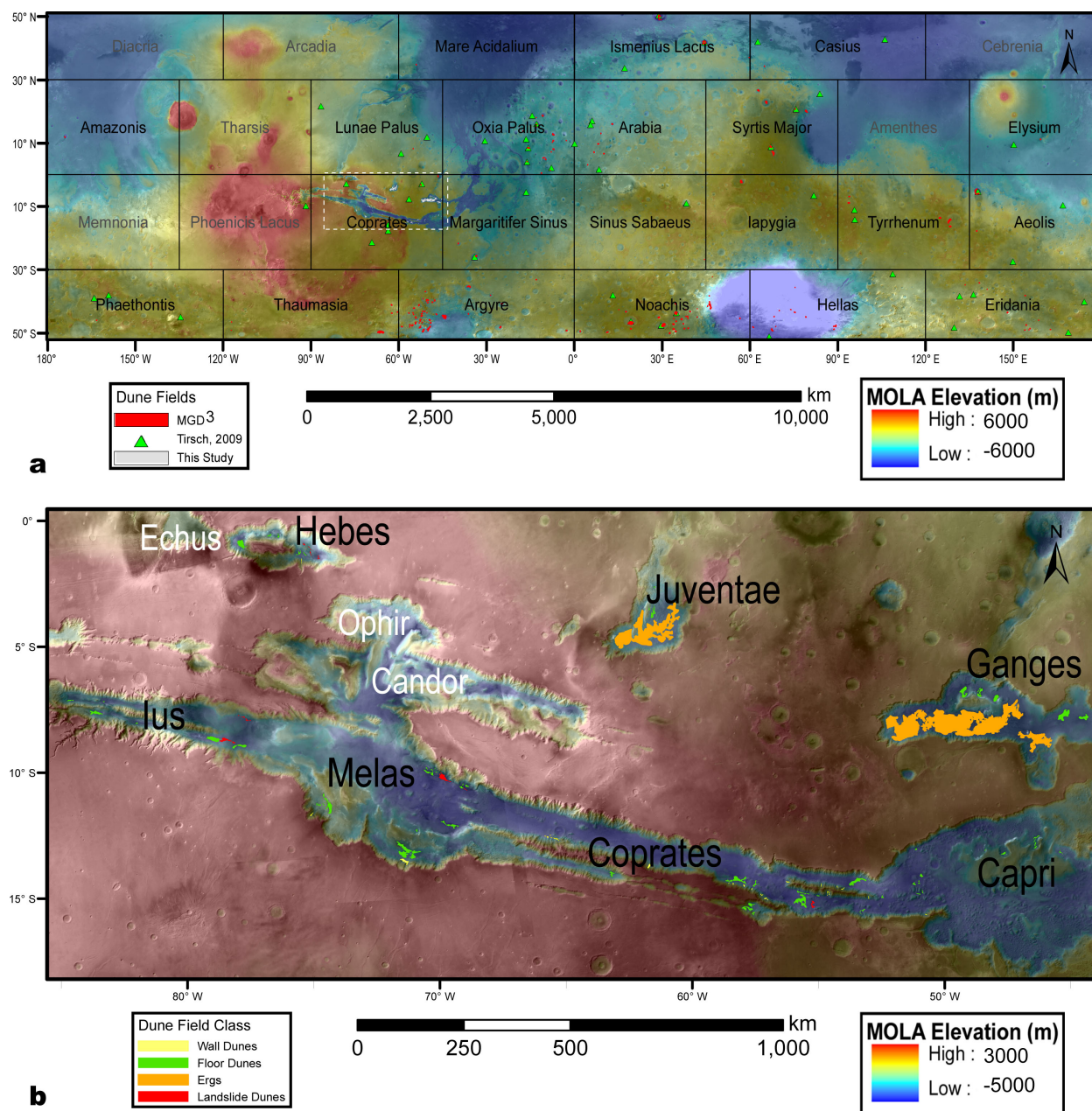


Figure 4.1 MOC wide-angle mosaic colorized with MOLA topography of the dune fields from 50°N–50°S latitude.

(a) The distribution of low- and mid-latitude dune fields, as mapped by *Tirsch* [2009] and *Hayward et al.* [2007]. Mars 1:5 million quadrangle grids and their names are given in black for grids that contain dune fields and in gray for those that do not. White dashed box indicates the Valles Marineris study area shown in Fig 4.1b. (b) Valles Marineris distribution of dune field classes mapped herein. Chasmata names are given in black for those that contain dune fields and in white for those that do not.

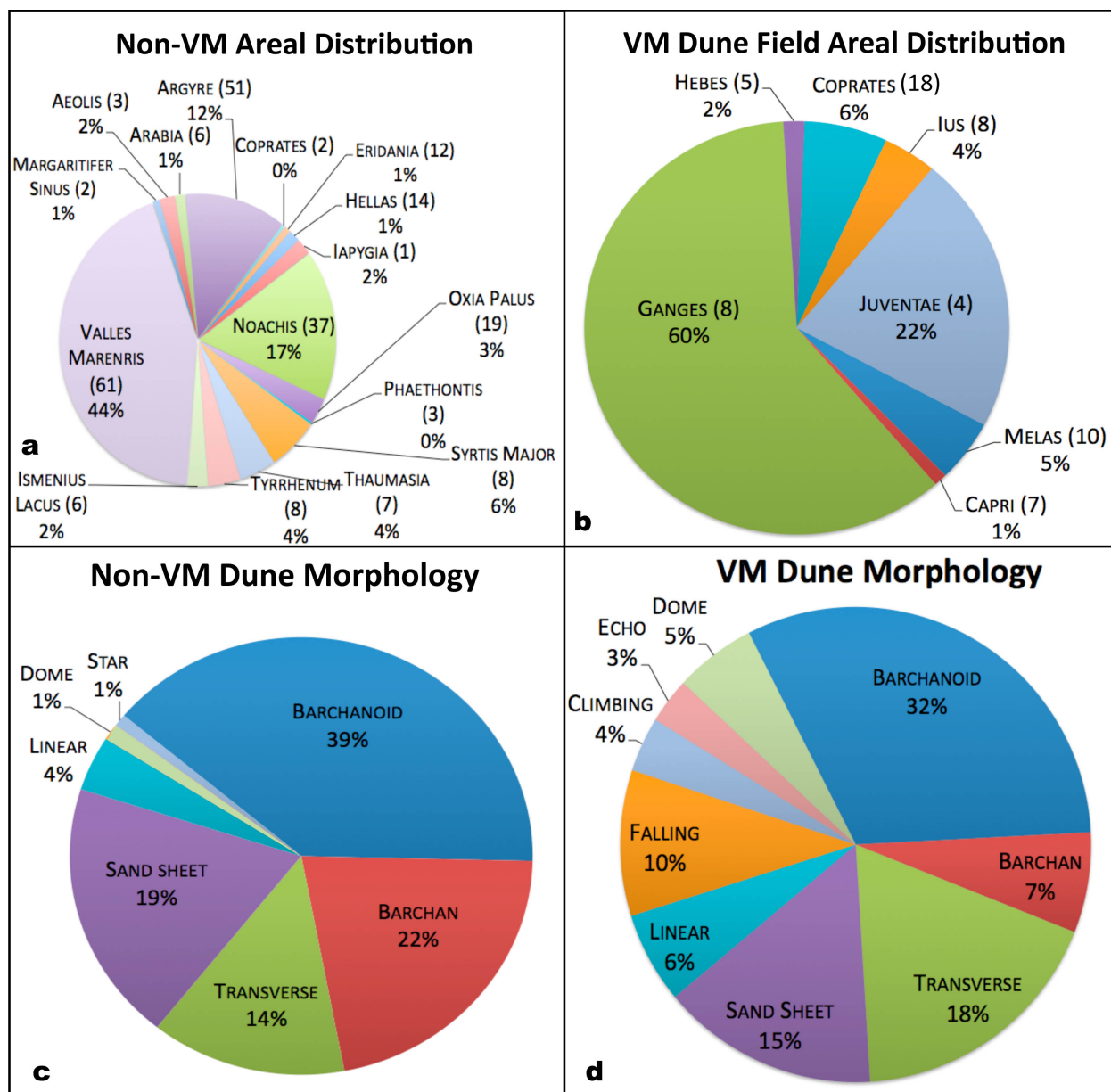


Figure 4.2 Geographic and morphologic distributions of dune fields studied herein. (a) Areal distribution of non-VM dune fields from 50°N to 50°S by quadrangle, as mapped by *Hayward et al.* [2007b]. The total area of the 184 dune fields is ~28,000 km² not including the VM dune fields mapped herein. The percentages next to each quadrangle name represent the percentage of total dune field coverage, and the numbers in parentheses indicate the number of individual dune fields. The Amazonis, Elysium, Casius, Lunae Palus, Mare Acidalium, Sinus, and Sabaeus quadrangles each contained one small (<50 km²) dune field but are not listed. (b) Valles Marineris dune field distribution by chasmata mapped herein. The total area of the 60 VM dune fields is ~22,000 km². Percentages and numbers have the same meaning as in Fig. 4.2a. Numeric distribution (percentage of occurrences per dune field population) of dune morphologies of (c) non-VM, as mapped by *Hayward et al.* [2007b], and (d) VM dune fields.

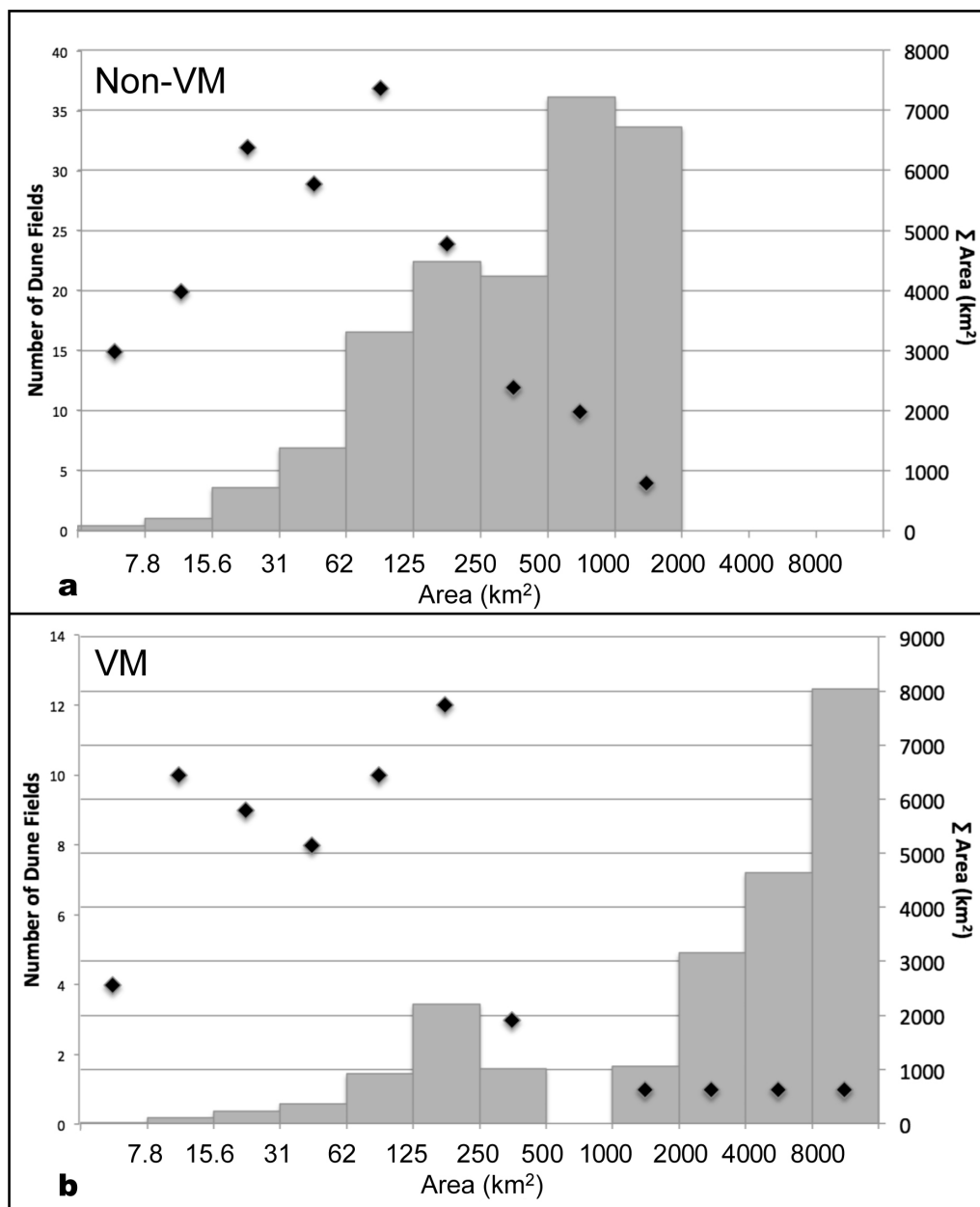


Figure 4.3 Frequency, an area histograms for dune fields studied herein. Combined frequency (black diamonds and left y-axis) and areal sum (gray boxes and right-y-axis) histograms of (a) dune fields from 50°N to 50°S and (b) dune fields in Valles Marineris. Dune field area frequency peaks at 16–62 km² (~155 km² mean) for non-VM and 8–125 km² (~400 km² mean) for VM dune fields. Both populations' area distributions are bimodal and the four VM ergs dominate the area for the VM populations due to their large areal sums (>1000 km²).

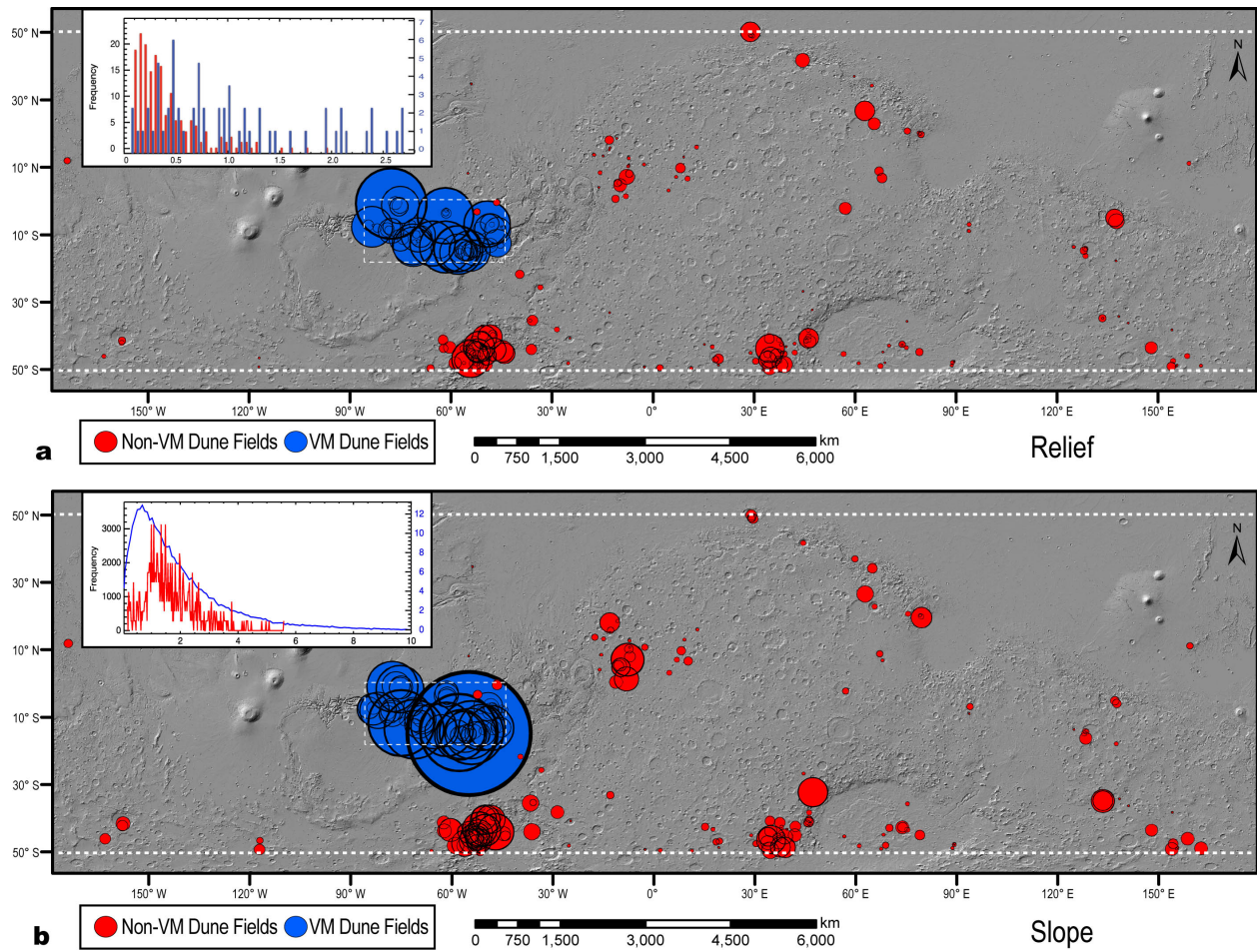


Figure 4.4 Topographic spatial and statistical trends of 184 non-VM (red) and 60 VM dune fields (blue).

Mean values for given variable are plotted on a MOLA shaded relief base map where circle diameter is linearly proportional to parameter value. Inset histograms are provided (high value tail of distribution not shown) where the non-VM frequency is shown on the left y-axis and the VM frequency is shown on the right y-axis. Values for both parameters were statistically separable for the two dune field populations. (a) MOLA dune field relief (dune field polygon minimum elevation subtracted from maximum elevation), showing the greater relief of VM dune fields. For reference, the diameter of the key symbol is equal to ~1 km of relief and the largest dune field relief is ~3.7 km in Hebes Chasma. (b) MOLA dune field slopes revealing much higher slopes for VM dune populations. For reference, the diameter of the key symbol is equal to ~3° of slope and the largest dune field mean slope value is found in Coprates Chasma (~17°).

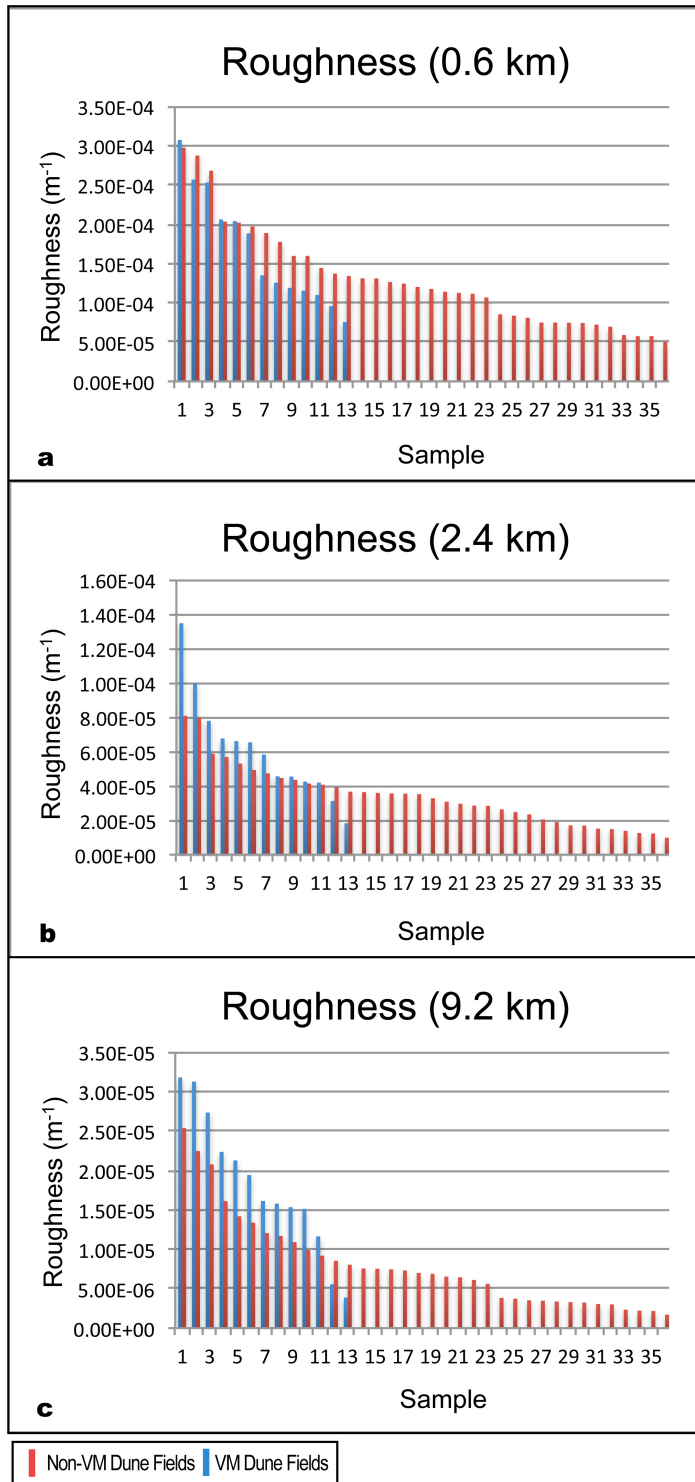


Figure 4.5 MOLA roughness values of 36 non-VM and 13 VM dune fields. Only dune fields that were large enough ($>200 \text{ km}^2$) and with sufficient spatial extent (2–3 pixels in the along-track direction) were used. Roughness plots are shown for the (a) 0.6-km, (b) 2.4-km, and (c) 9.2-km baselines.

Figure 4.6 Thermophysical spatial and statistical trends of non-VM (red) and VM (blue) dune fields studied herein.

Mean values for given variable are plotted on a MOLA shaded relief base map where circle diameter is linearly proportional to parameter value. Inset histograms are provided. Albedo and ΔTI values were statistically separable for the two dune field populations. (a) MY 26 TES albedo of 184 non-VM and 60 VM dune fields showing the moderately low-albedo of VM dunes. For reference, the diameter of the key symbol is equal to an albedo of ~ 0.12 albedo and the largest dune field albedo is ~ 0.24 in Oxia Palus. (b) TES nightside-derived thermal inertia of 32 non-VM and 20 VM dune fields showing variable dune thermophysical properties. For reference, the diameter of the key symbol is equal to ~ 180 tiu and the largest value is ~ 390 tiu in Coprates Chasma. (c) TES ΔTI of the same number of dune fields as in the preceding figure where large values may indicate greater thermophysical heterogeneity (e.g., greater abundance of intradune area, outcrops, dust mantles). For reference, the diameter of the key symbol is equal to a ~ 40 tiu change and the largest is ~ 170 tiu in Melas Chasma. ΔTI values within 25 tiu of zero are considered null here and are grayed out on the inset histograms.

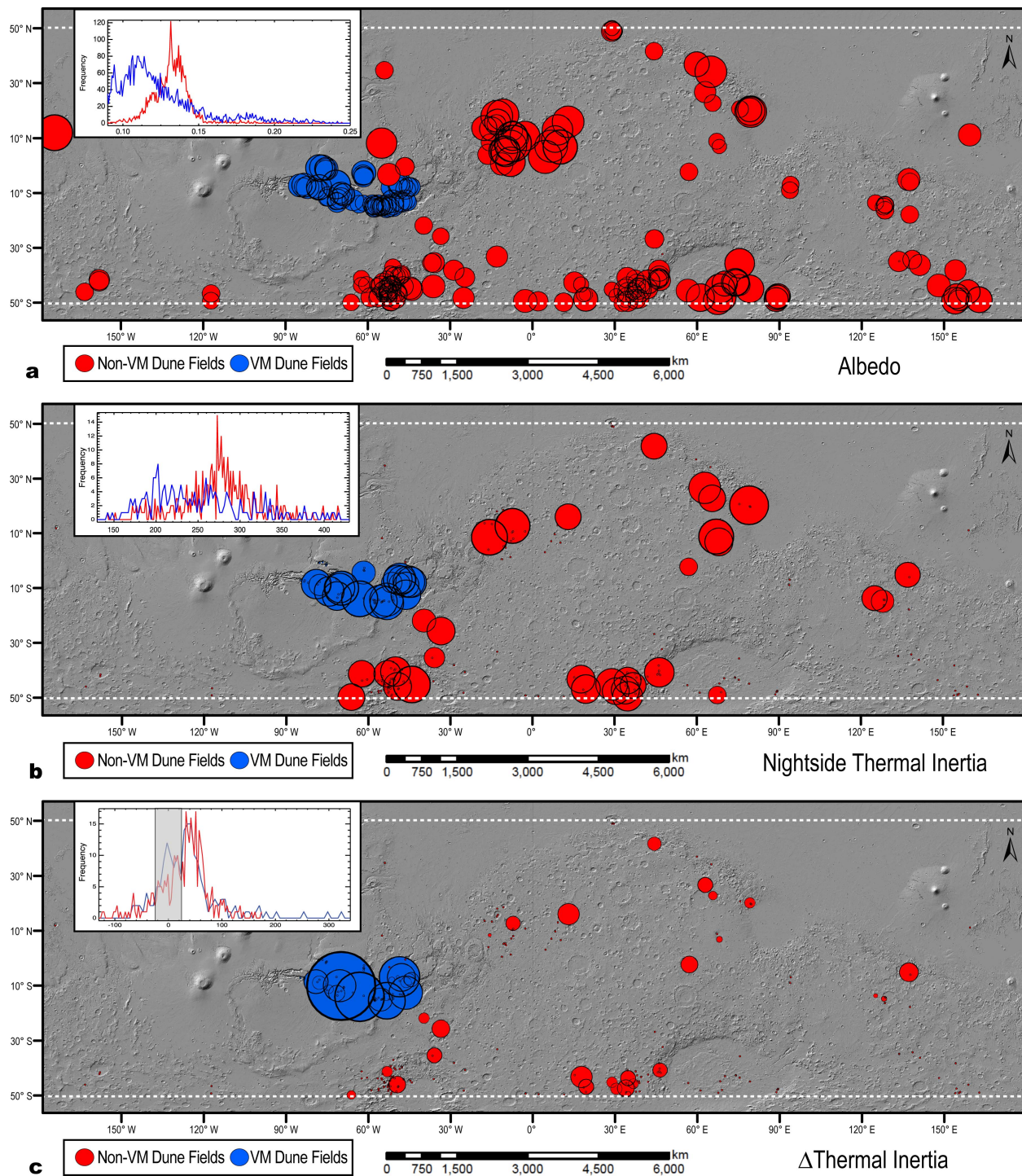


Figure 4.6

Figure 4.7 Spatial and statistical trends of dune field compositions.

The three TES spectral indices show trends in dust coverage and composition of 184 non-VM (red) and 60 VM (blue) dune fields. TES DCI and 530 values were statistically separable for the two dune field populations. (a) Mean TES dust cover index of dune field, with values consistent with dust-free surfaces (green circles) being the most common. Inset histogram is provided where threshold values below and above 0.962 (dashed line) indicate moderate dust coverage and dust free surfaces, respectively. (b) TES 465 index (mean values) of dune fields, where locations with elevated 465 index, but low 530 index (green stars) are also provided. Inset histogram is provided where values above the dashed line indicate detections to 3-sigma. (c) TES 530 index (mean values) of dune fields. Inset histogram is provided where values above the dashed line indicate detections to 3-sigma. (d) OMEGA-derived compositions of 37 non-VM dune fields (circles), as mapped by *Tirsch* [2009], in comparison to CRISM results of 23 dune fields (triangles) studied herein. Pyroxenes, either alone or with olivine, are the most common detections (insets).

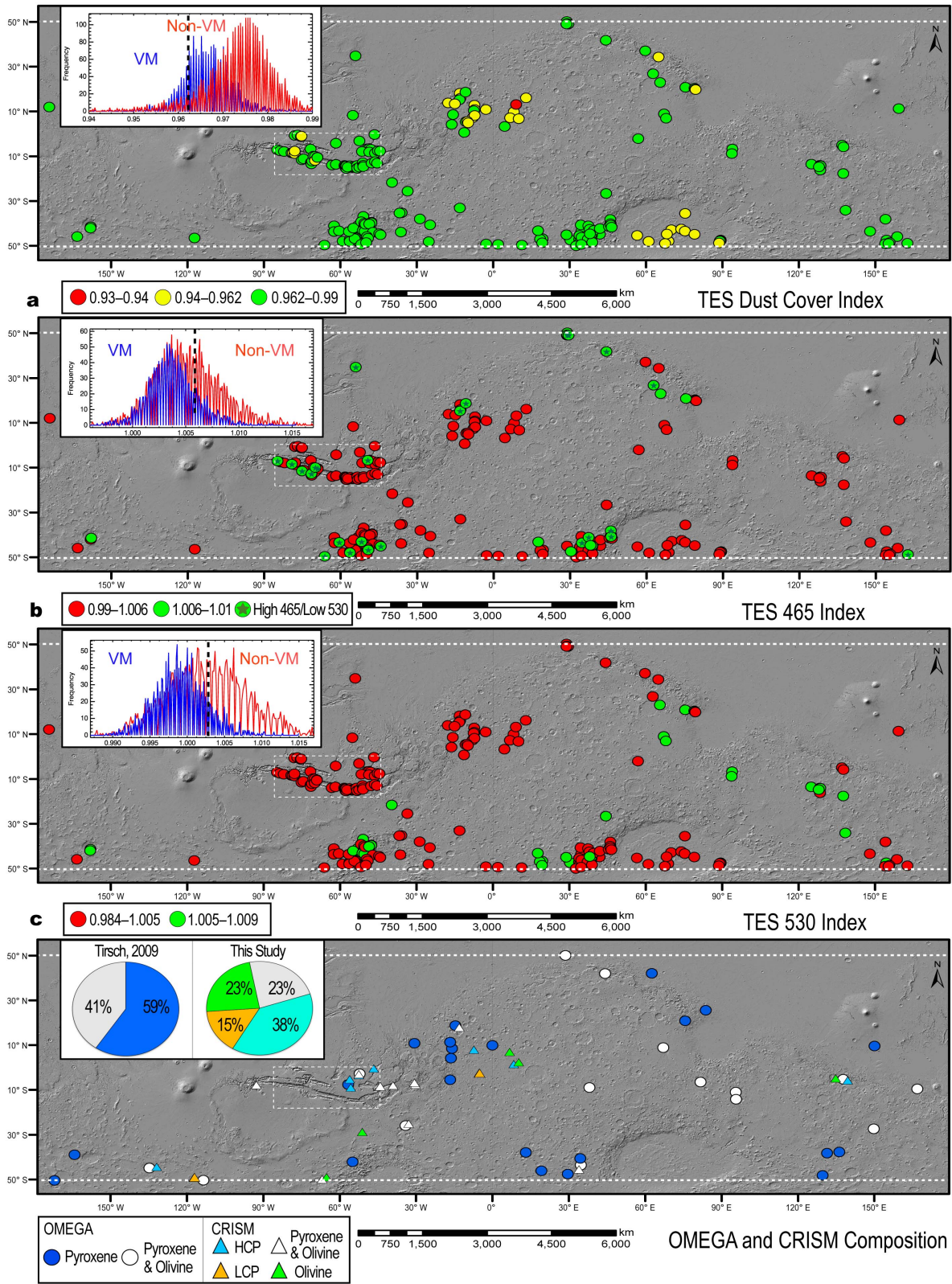


Figure 4.7

Figure 4.8 Dune morphologic change and dune field albedo change results.

(a) Wide-spread dune changes from *Bridges et al.* [2012] and *Silvestro et al.* [2011] in comparison to bedform activity in VM (this study). TES temporal changes in bolometric albedo (in percent) of 184 non-VM and 60 VM dune fields from Mars Year (b) 24–25 and (c) MY 25–26. Histograms are provided (insets) where albedo changes $\pm 5\%$ are considered null and are grayed out. Values for both parameters were statistically separable for the two dune field populations. Dune fields show a roughly even distribution of brightening and darkening with many showing drastic changes in albedo ($>+70\%$ or $<-35\%$). For reference, a surface that had a detected albedo change from 0.11 to 0.12 is equivalent to $\sim 8\%$.

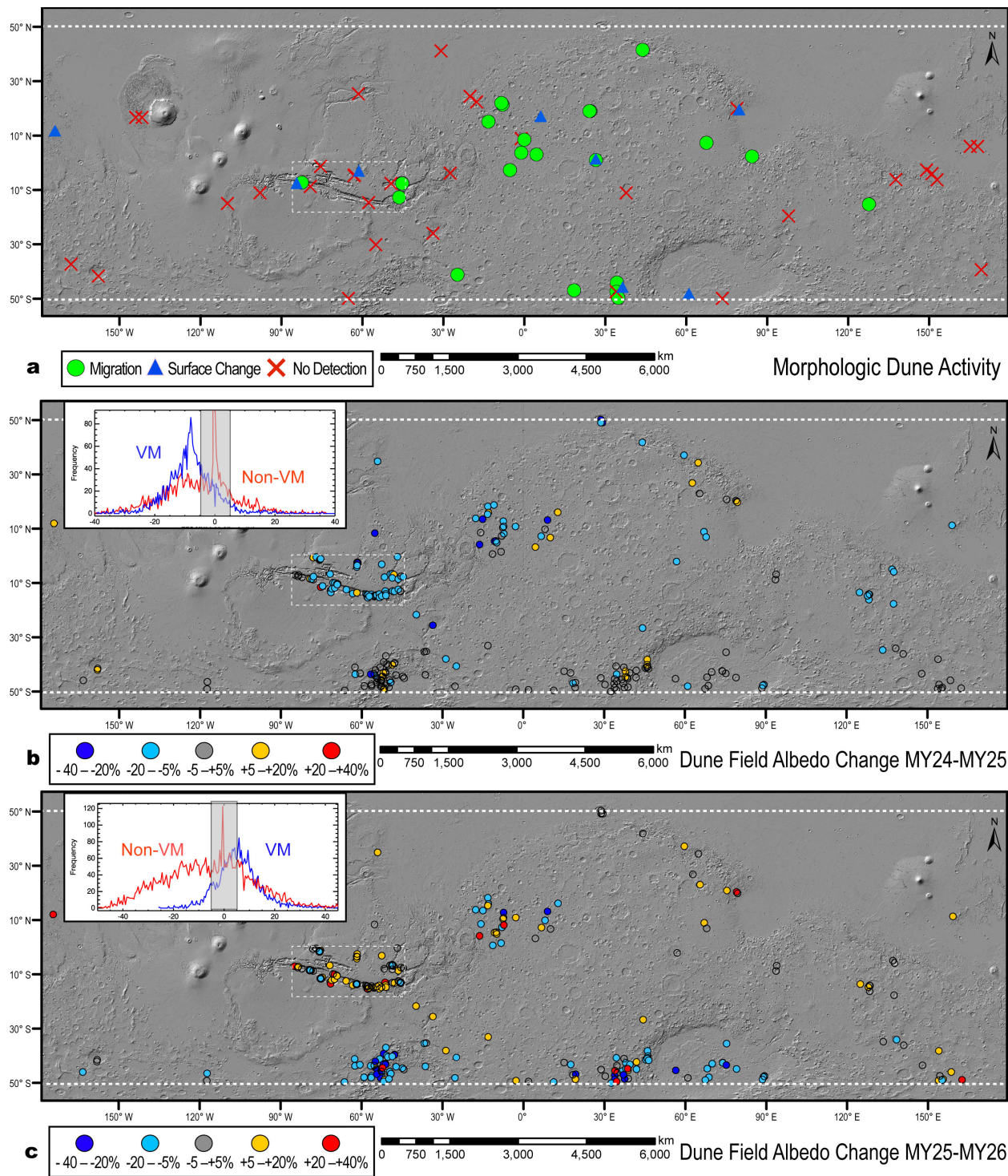


Figure 4.8

Figure 4.9 THEMIS daytime infrared mosaic of (a) Ius, (b) Melas, and (c) Coprates chasmata colorized with MOLA topography.

For Fig. 4.9–4.10 dune field mean slip face directions are shown in black. Also, dune field classifications are as follows: wall (yellow), floor (green), erg (orange), and landslide (red). Also see Fig. 4.1b. Dune field labels are shown in white and are consistent with Tables 4.5, and 4.7, and 4.8. Bold white arrows indicate ends of massifs that may be locations of enhanced atmospheric flow or sediment deposition, as inferred from mesoscale modeling. Yellow stars indicate locations with detected surface activity. (THEMIS image credit: *Gorelick and Christensen* [2005]).

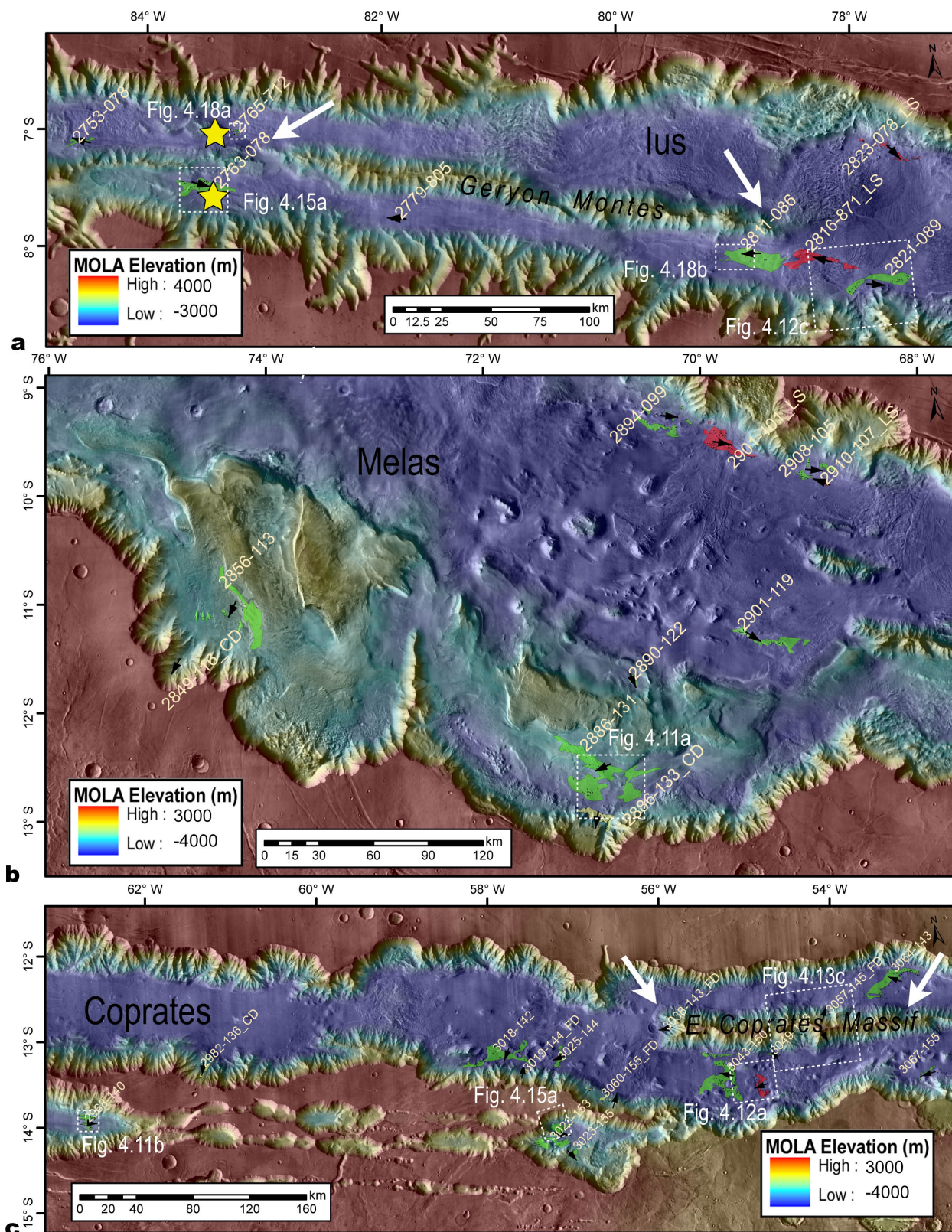


Figure 4.9

Figure 4.10 THEMIS daytime infrared mosaic of (a) Hebes, (b) Juventae, (c) Capri, and (d) Ganges chasmata colorized with MOLA topography. Prominent Juventae Chasma ILD mounds are labeled A–C after *Catling et al.* [2006]. Refer to Fig. 4.9 for details.

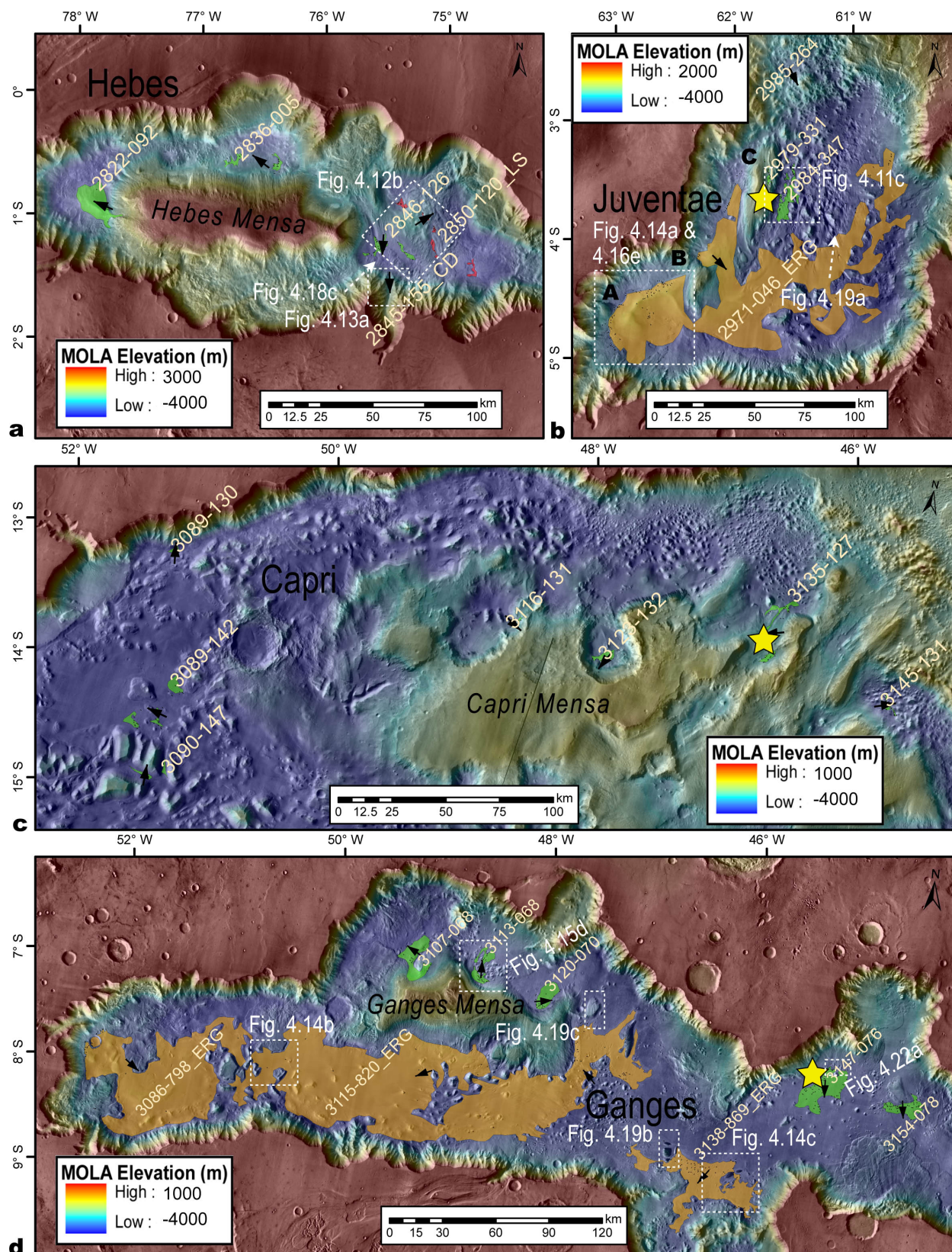


Figure 4.10

Figure 4.11 CTX views and HRSC topography of select VM floor dune fields (Figs. 4.1b, 4.9–4.10, green polygons).

Orientations perpendicular to individual slip faces are shown with green arrows and the mean direction for the dune field arrows in turquoise. In the inset plots, HRSC elevation (black, in units of meters) and slope (dashed red, in units of degrees) profiles are shown where dashed lines (when shown) indicate contacts between dune and non-dune surfaces. North is toward the top in all images unless otherwise indicated. Context boxes (white) are provided for inset figures showing un-annotated dune morphologies. The same conventions are used for Figs. 4.12–4.14.

(a) CTX mosaic of southeast Melas Chasma floor dunes (inset, P16_007245_1648_XN_15S071W) found atop of ILDs. Additional occurrences of climbing dunes on southern wall slopes are also seen. Profile A–A' using HRSC DEM (orbit 334, 100 m/pixel). (b) In south Coprates, barchanoids and barchans (inset, HiRISE PSP_006480_1660) are found with dominantly unidirectional (eastward) slip face orientations. CTX P14_006480_1667_XN_13S063W where profile B–B' uses HRSC DEM (orbit 471, 150 m/pixel). (c) Juventae Chasma dunes are located within chaotic terrain (inset, HiRISE ESP_024294_1765) and atop of mound C. Sand supply increases southward into the main erg (see Fig. 4.14a). Profile C–C' using HRSC DEM (orbit 1070, 50 m/pixel).

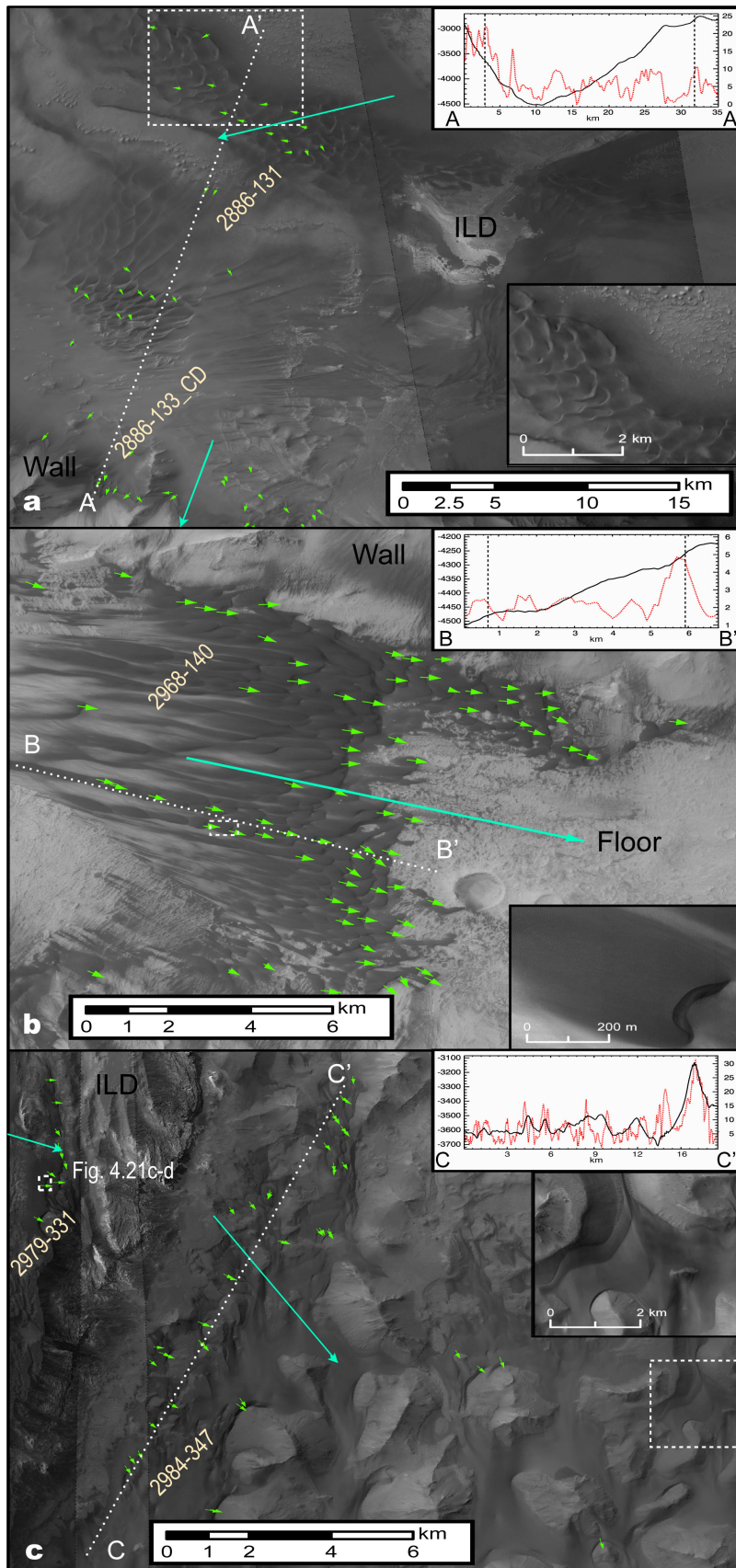


Figure 4.11

Figure 4.12 CTX views and HRSC topography of select VM landslide dune fields (Figs. 4.1b, 4.9–4.10, red polygons).

(a) Coprates Chasma dunes are found scattered on landslide and floor surfaces (inset, P18_007930_1649_XI_15S055W). Profile A–A' using HRSC DEM (orbit 1004, 75 m/pixel). (b) CTX mosaic of Hebes Chasma where a jumbled landslide originating off the north wall is found partially overriding ILDs. Dunes (inset, HiRISE ESP_017385_1785) are stretched out over ~30 km in the central portion of the slide. Profile B–B' using HRSC DEM (orbit 2116, 50 m/pixel). Also see Fig. 4.18c. (c) CTX mosaic of east Ius Chasma dunes (inset, HiRISE ESP_024532_1715) found superposed on dusty landslide surfaces and adjacent to slide-wall contacts. Profile C–C' using HRSC DEM (orbit 2149, 50 m/pixel).

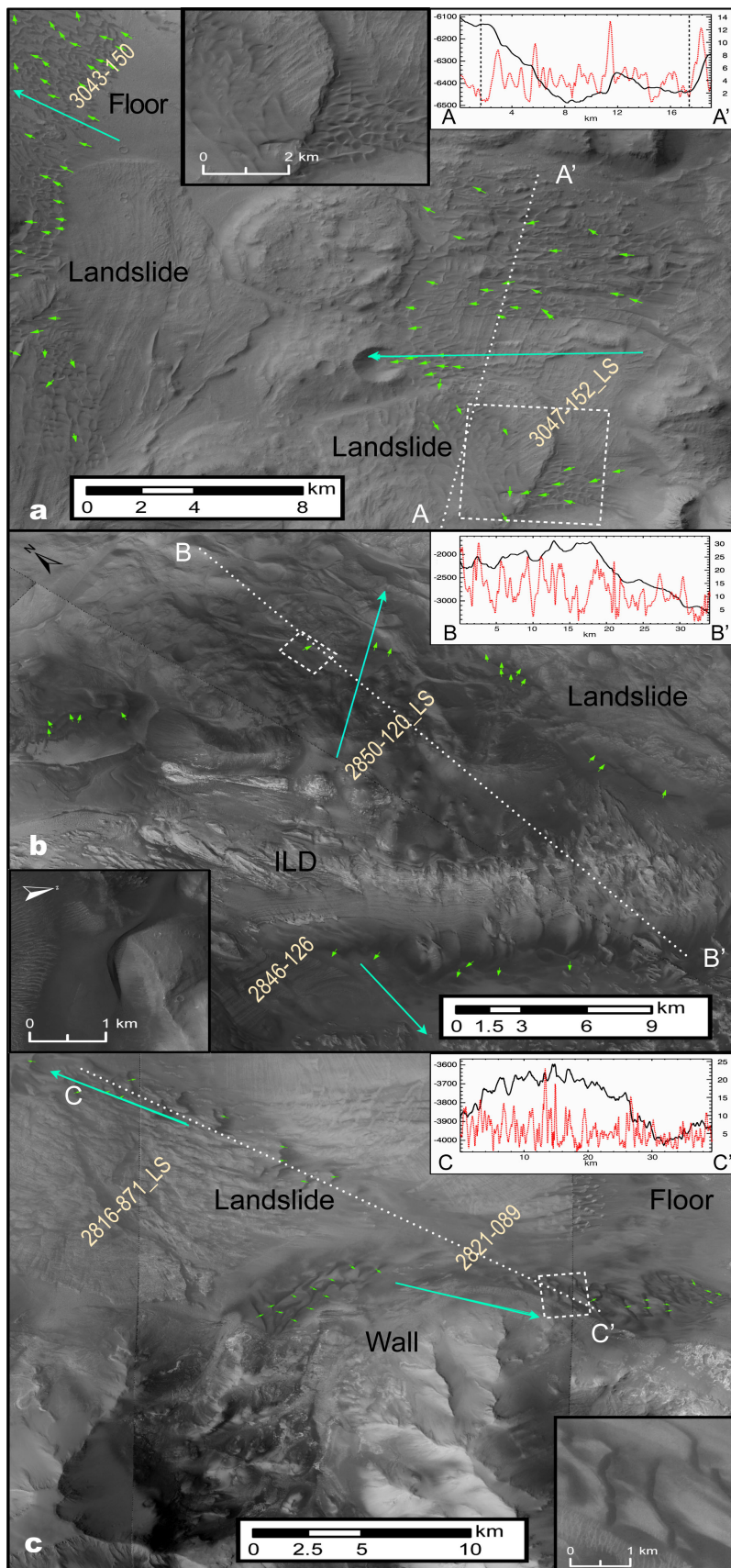


Figure 4.12

Figure 4.13 CTX views and HRSC topography of select VM (a–b) climbing and (c) falling dunes (Figs. 4.1b, 4.9–4.10, yellow polygons).

Images are generally oriented so lower elevations are toward the bottom of the images. (a) In south Hebes Chasma, a barchan train (inset) climbing a steep ($\sim 10^\circ$) and narrow gully southward. CTX P04_002485_1784_XN_01S075W and profile A–A' using HRSC DEM (orbit 2116, 50 m/pixel). (b) In west Coprates Chasma, climbing dunes (inset, HiRISE ESP_018189_1675) are spread out over ~ 20 km in a gully. CTX mosaic and profile B–B' using HRSC DEM (orbit 1995, 50 m/pixel). (c) Falling dunes (left inset, P13_006216_1663_XN_13S055W; right inset, P16_007429_1671_XN_12S054W) are abundant on the flanks of the east Coprates massif. CTX mosaic where profiles C–C' and D–D' uses HRSC DEM (orbit 982, 50 m/pixel).

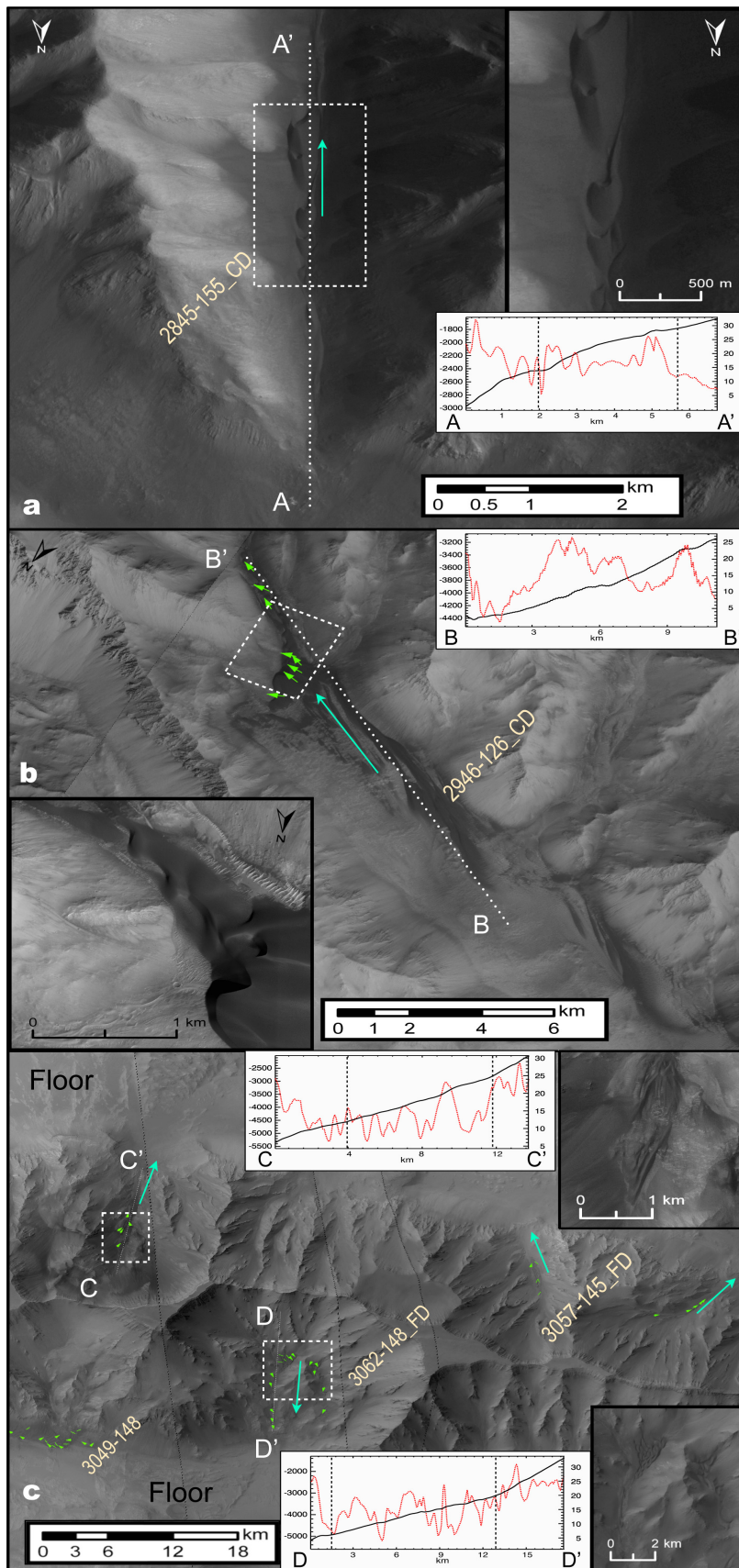


Figure 4.13

Figure 4.14 MOC CTX mosaics and HRSC topography of select VM ergs (Figs. 4.1b, 4.10, orange polygons).

(a) West Juventae Chasma erg with barchanoids (inset, HiRISE ESP_017490_1755) on the perimeter, whereas sand sheets are more commonly on the interior. Profile A–A' using HRSC DEM (orbit 243, 100 m/pixel). Compares with Fig. 4.17b. (b) In central Ganges Chasma, featureless mid-toned sand sheets, with low slopes, atop of floor and landslide surfaces dominant, whereas slip faces develop near topographic elements (inset, HiRISE PSP_007798_1720). Profile B–B' using HRSC DEM (orbit 2222, 75 m/pixel). (c) In east Ganges Chasma, erg sand is found in lesser abundance, but with a greater proportion of slip faces as compared with ergs to the west (inset, HiRISE ESP_017885_1715). Profile C–C' using HRSC DEM (orbit 2145, 50 m/pixel).

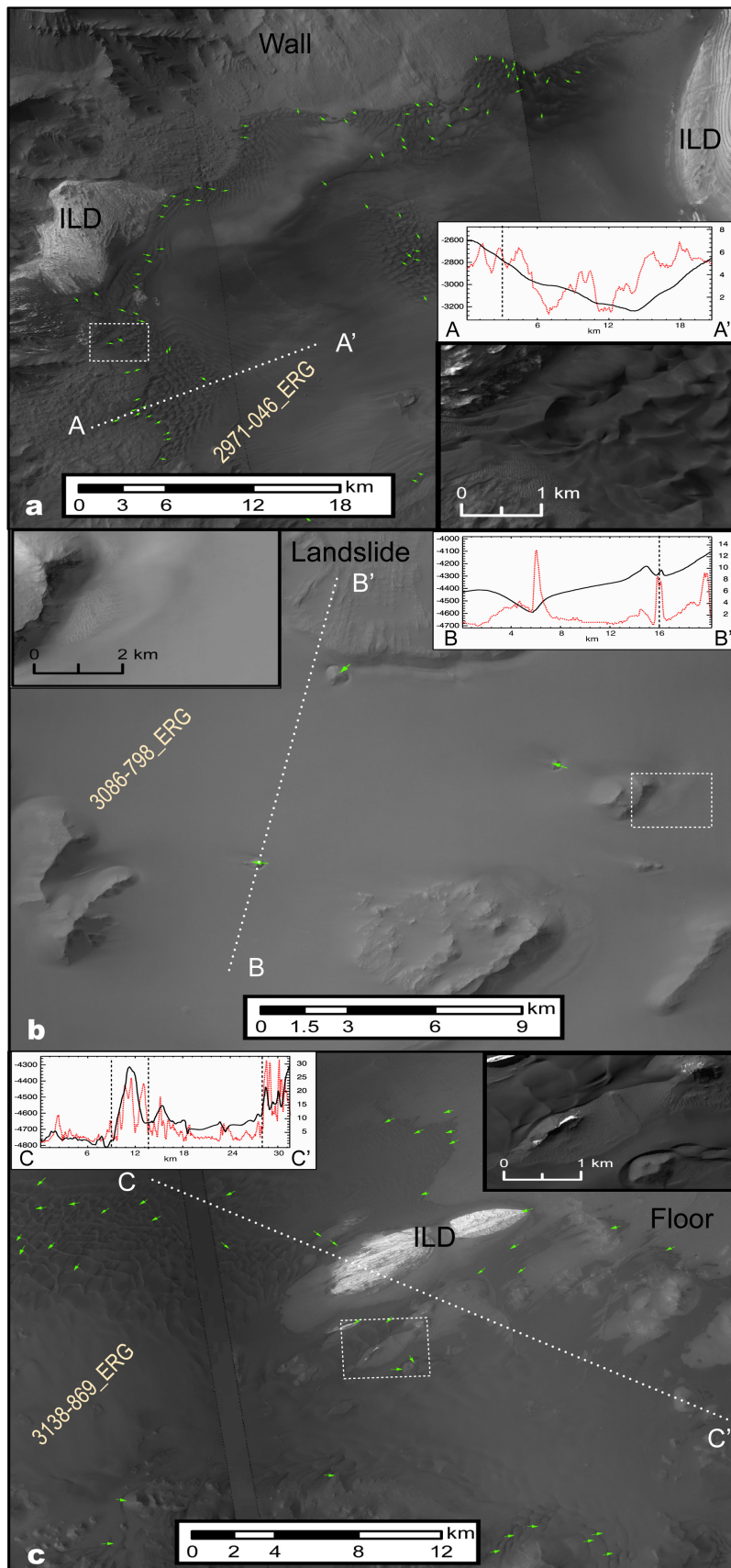


Figure 4.14

Figure 4.15 CTX mosaics colorized with THEMIS thermal inertias of VM dune fields. (a) An east Ius Chasma dune field. THEMIS I34691007 Thermal inertia profile A–A' values increase along (b) the draa stoss side and crests that may indicate coarser particles and/or locations of more frequent dust-cleaning events (HiRISE PSP_009223_1720). (c) The large draa (white arrows in (a)) is nearly 300-m-tall and may be indurated due to its high thermal inertia (see inset plot). HiRISE ESP_017293_1720 and HRSC DEM (orbit 2979, 75 m/pixel). (d) Northeast of Ganges Mensa thermal inertia/CRISM IRA albedo profiles B–B' show trends of the prominent (e) barchanoid dunes and sand sheets to the north (THEMIS I02405002). Vertical lines on that plot indicate locations of ILD-dune and dune-canyon floor contacts (diamonds). (f) Mass-wasting of ILD material adjacent to dunes in HiRISE IRB false-color.

Figure 4.16 CTX mosaics colorized with THEMIS thermal inertia of VM dune fields. (a) Dunes in a collapse pit south of Coprates Chasma where the thermal inertia/CRISM IRA albedo A–A' profiles show the inverse relationship between the two data sets as transverse dune troughs consist of brighter/finer sediment, whereas crest are darker/coarser (THEMIS I36250009). Vertical lines on that plot indicate locations of wall rock-talus and dune-canyon floor contacts (diamonds). (b–c) North wall mass-wasting and aeolian processes in HiRISE RED and IRB false-color. (d) Boulder tracks are found on steep talus slopes and less commonly boulders are found atop of dune field surfaces (black arrows, HiRISE ESP_023806_1645). (e) Juventae Chasma's erg dunes consist of (f) barchanoid dunes onlapping ILDs (HiRISE PSP_004291_1755), whereas southeastward slip faces diminish as the sand surfaces decrease in thermal inertia. Thermal inertia/HRSC topography B–B' profiles show high values (>350 tiu) are found in the northeast on stoss side and crest surfaces (g) and a giant draa ~700-m-tall (white arrows, see text). HRSC DEM (orbit 243, 75 m/pixel).

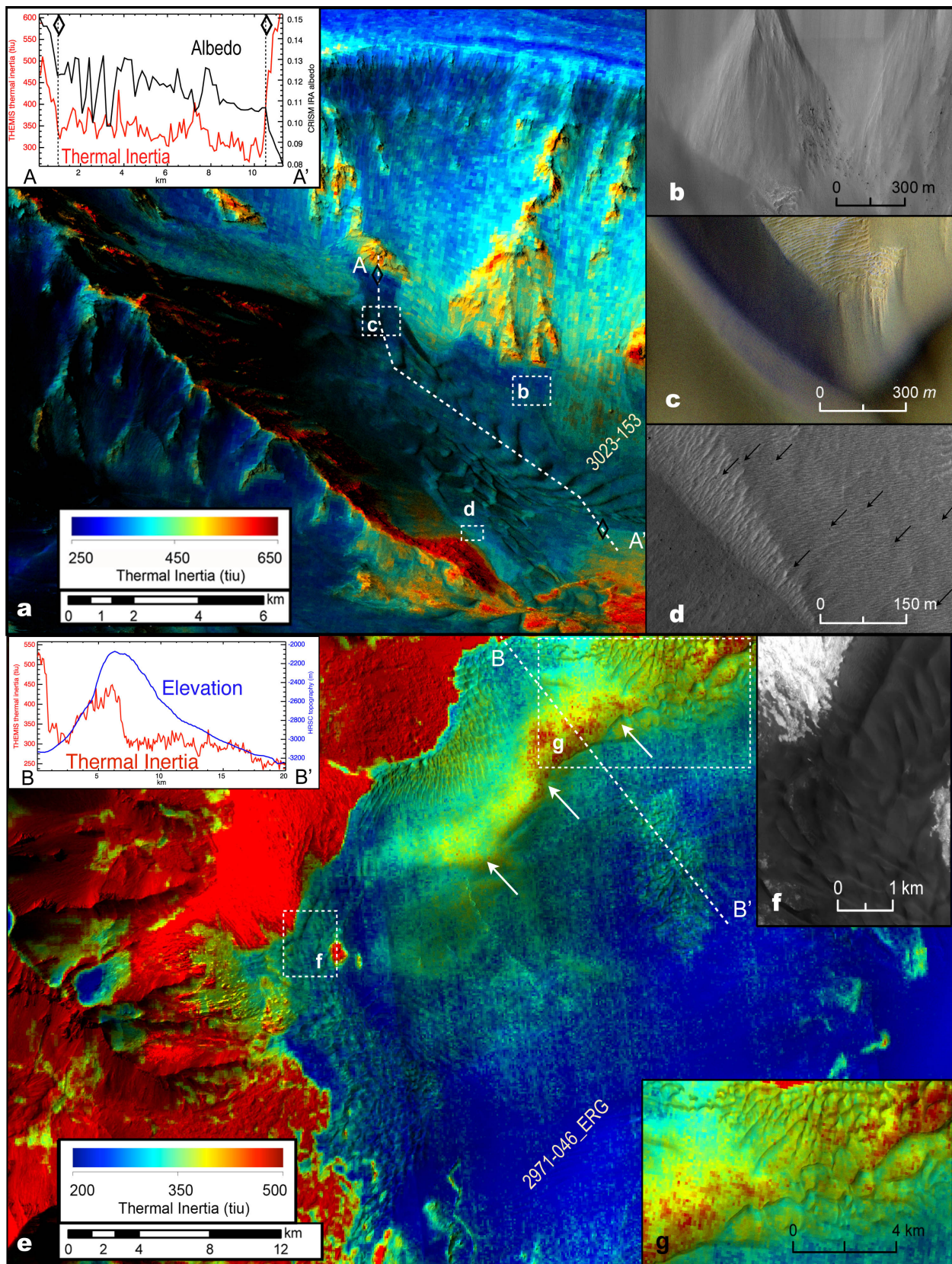


Figure 4.16

Figure 4.17 Valles Marineris dune field CRISM endmember compositions.

(a) A MOC wide-angle mosaic showing compositional classes across VM. Arrows indicate locations of sample spectra shown below. (b) Unratioed I/F CRISM spectra of endmember dune compositions. Dune field IDs are given followed by alphanumeric CRISM image labels where observation type and resolutions are listed in Table 4.3. (c) The same dune spectra shown in (b) but ratioed (solid lines) with dust spectra in comparison to candidate laboratory spectra (dashed lines). “BAS101 Acid A” is an artificially weathered glassy basalt sample with a grain size of 500–1000 μm , as described in *Horgan et al.* [2011].

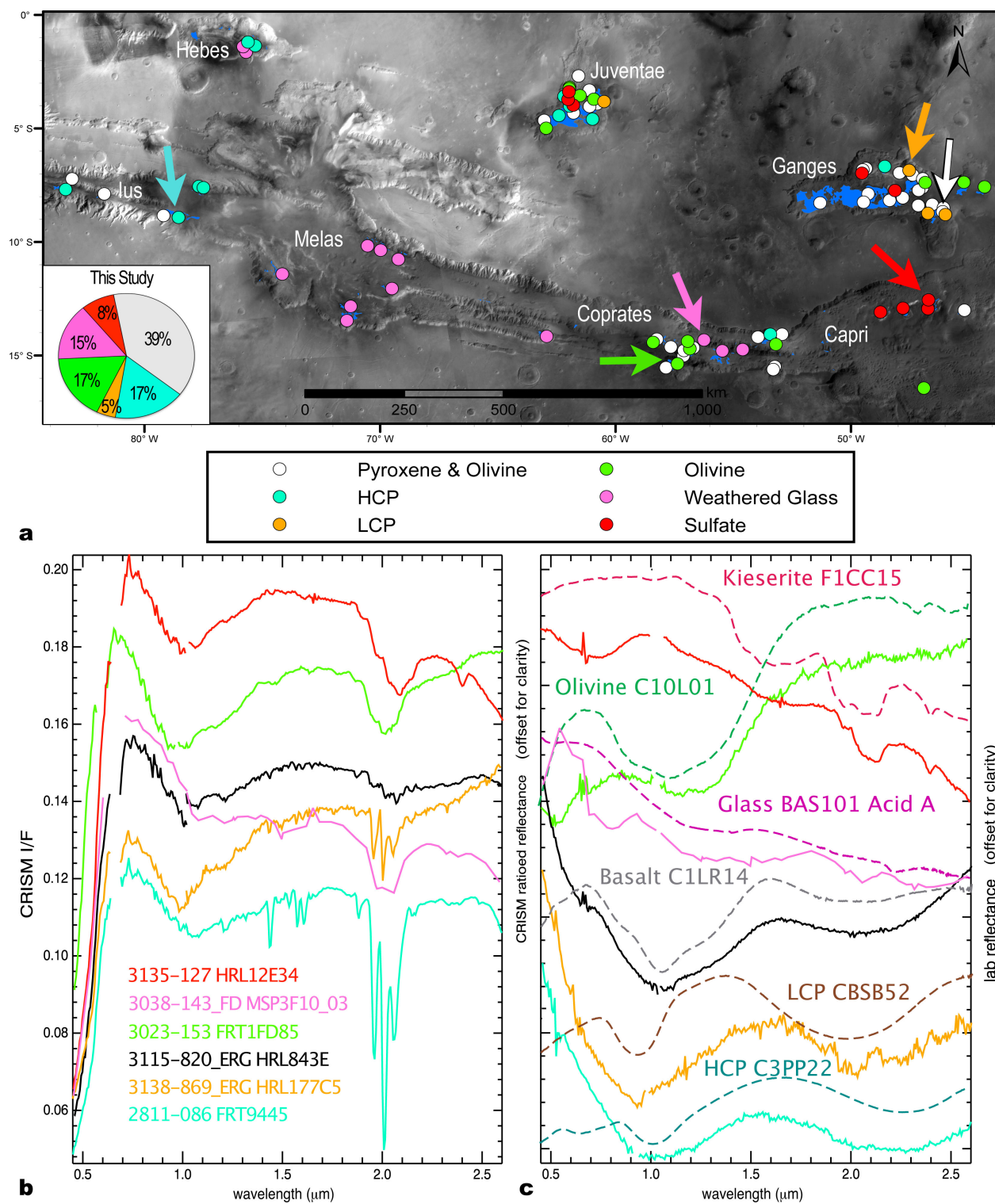


Figure 4.17

Figure 4.18 CRISM compositional views and spectra of western VM dunes.

For 4.18–4.19, color-coded arrows indicate locations of CRISM spectra (solid) and all lab spectra are shown with dashed lines. (a) CRISM on CTX B19_017082_1726_XN_07S083W of floor dunes partially composed of basaltic glass in northwest Ius Chasma, where colors are BD1900R (magenta), OLINDEX (green), and BDI2000 (blue). (b) CRISM on a CTX mosaic of southwest Ius Chasma floor dunes partially composed of basalt and HCP, where colors are BD2100 (magenta), OLINDEX2 (green), and HCPINDEX (blue). (inset, HiRISE ESP_013627_1710). (c) Oblique northward view of Hebes Chasma sand sheets and layered deposits using CTX B12_014418_1789_XN_01S075W with HRSC elevation (orbit 438 DEM, 100 m/pixel), where colors are SINDEX (magenta), LCPINDEX (green), and BDI2000 (blue). Dune sand spectra are consistent with HCP (inset, HiRISE ESP_014418_1790). Also see Fig. 4.12b.

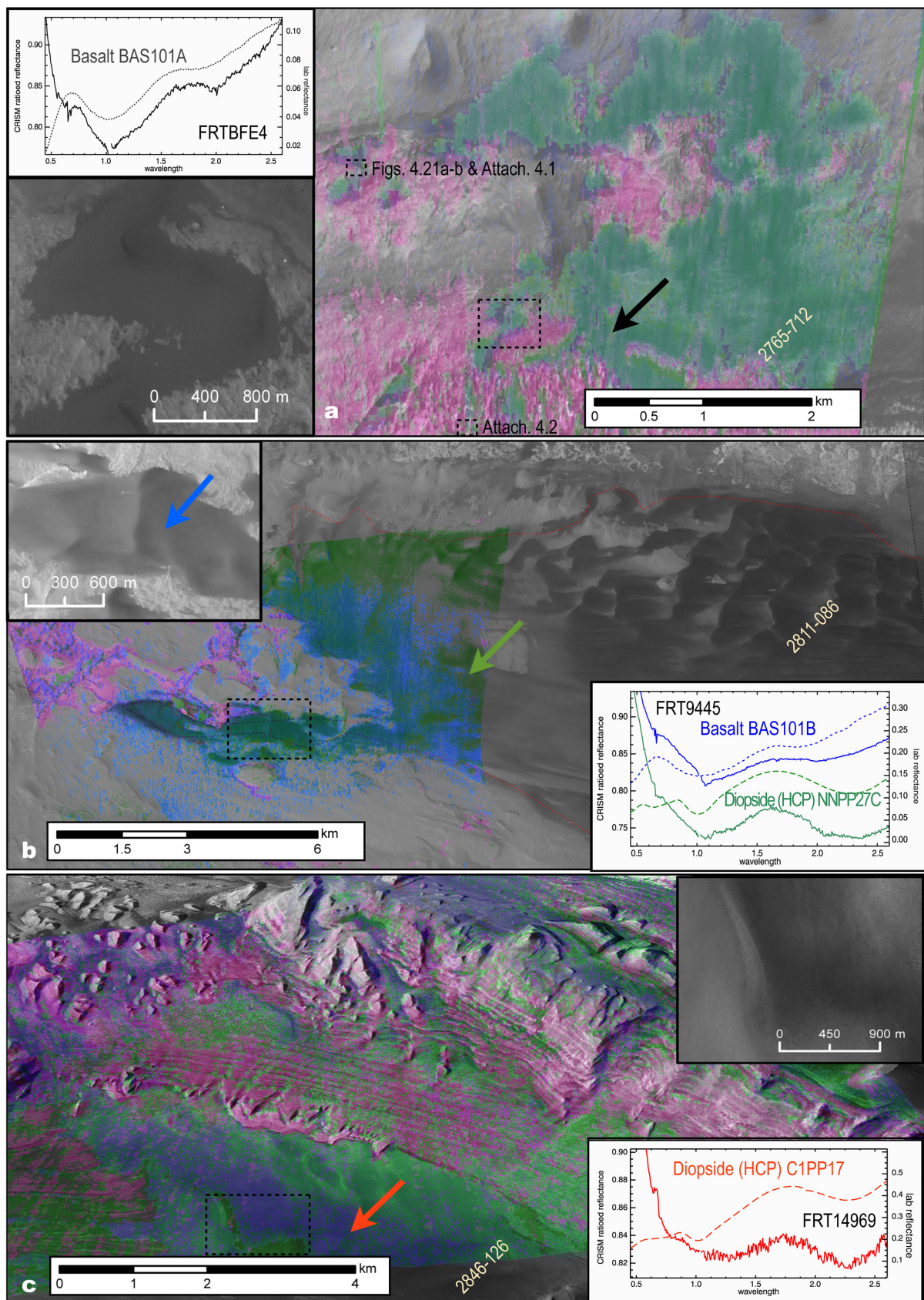


Figure 4.18

Figure 4.19 CRISM compositional views and spectra of eastern VM dunes.

(a) Oblique northward view of floor dunes among Juventae Chasma chaos terrain using CTX B09_012967_1770_XI_03S061W with HRSC elevation (orbit 449, 200 m/pixel), where colors are ISLOPE1 (yellow), OLINDEX (green), and LCPINDEX (blue). Dune sand spectra are consistent with olivine. (inset, HiRISE ESP_012967_1765). (b) CRISM on a CTX/HiRISE mosaic of erg dunes in eastern Ganges Chasma, where colors are SININDEX (magenta), OLINDEX (green), and VAR (yellow). Dune sand spectra are consistent with glassy basalts and intermediate- to low-calcium pyroxenes. (inset) Oblique southward view (dashed white arrow) of dunes and ILD surfaces where the FOV is ~3 km (HiRISE ESP_017173_1715). (c) CRISM on a CTX mosaic of central Ganges Chasma floor dunes, where colors are HCPINDEX (blue), OLINDEX (green), and LCPINDEX (yellow). Note the large difference in absorption depths and positions of the 2- μ m features between the two dunes in close proximity, indicating pyroxene variations (inset, HiRISE ESP_011516_1730).

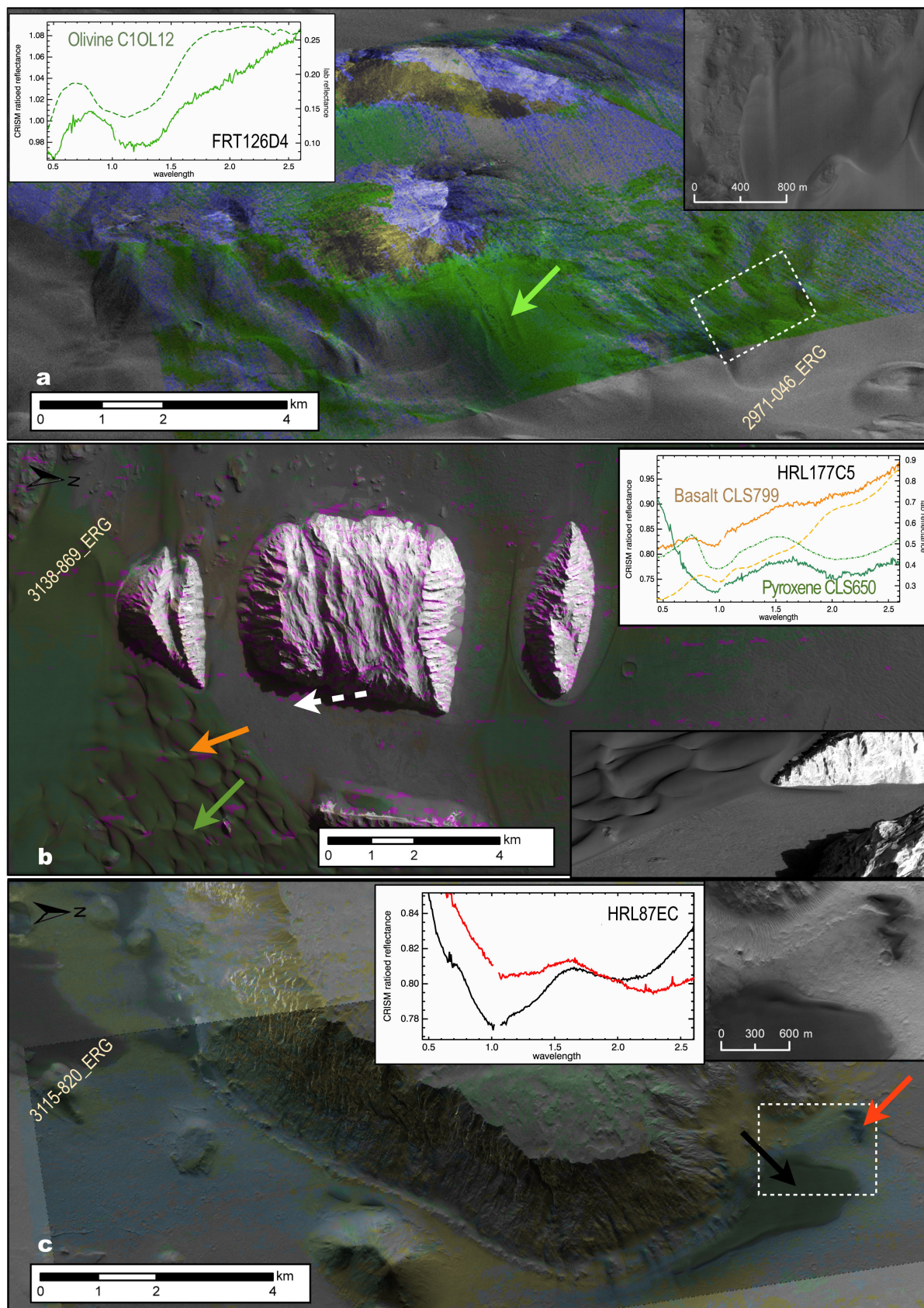


Figure 4.19

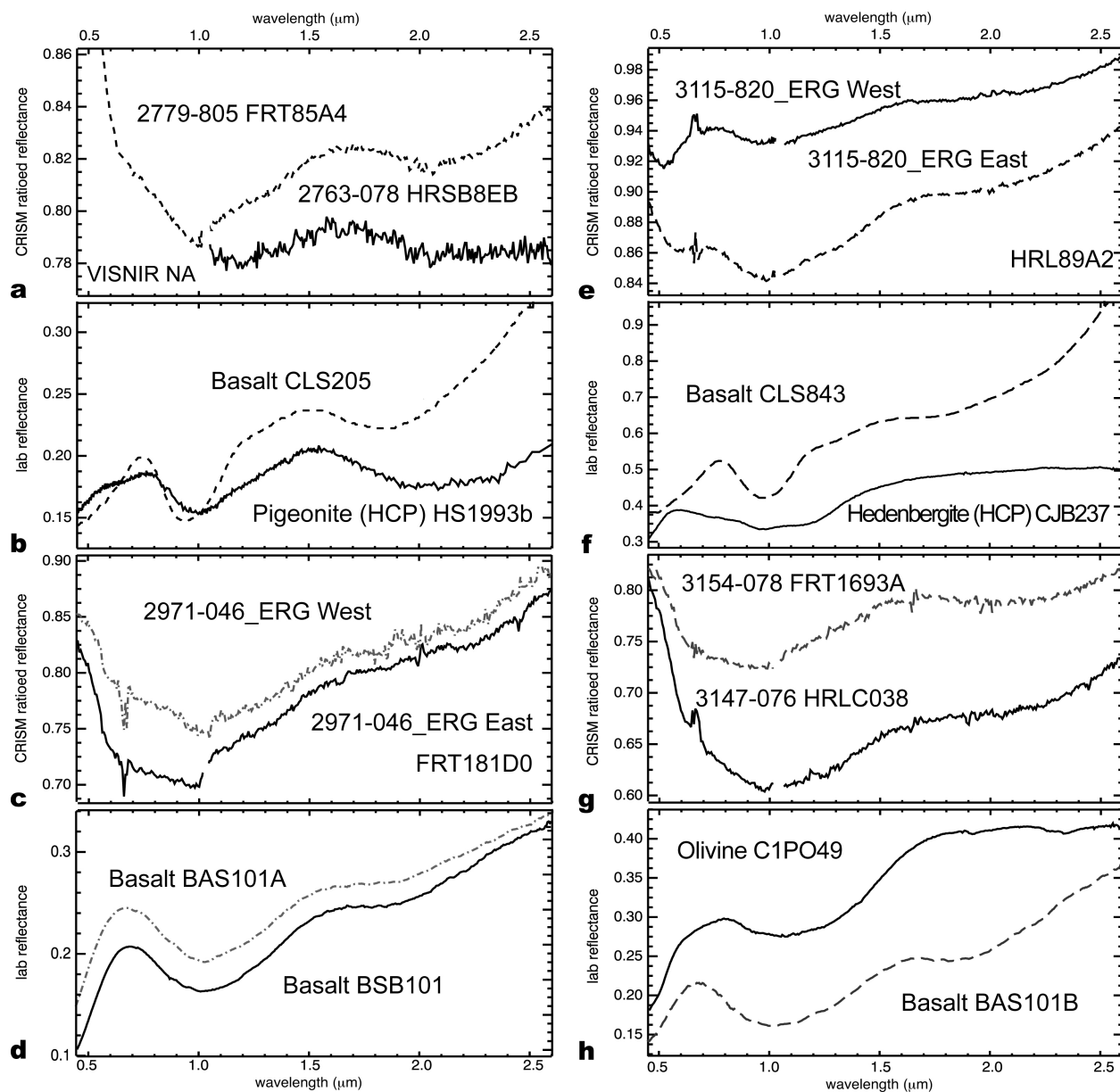


Figure 4.20 Interleaved CRISM spectra of VM dunes and corresponding laboratory spectra of candidate rocks and minerals.

See Figs. 4.9–4.10 for locations of corresponding VM dune fields. (a) Basalt- and HCP-bearing floor dunes in Ius Chasma. (c) Glassy-basaltic erg sand in south Juventae Chasma. (e) Glassy-basaltic and HCP-bearing erg sand in south central Ganges Chasma. (g) Olivine- and glass-bearing basalt in east Ganges Chasma. The interleaved Figs. 4.20b, 4.20d, 4.20f, 4.20h are candidate laboratory spectra and should be compared with the dune spectra in the preceding panel.

Figure 4.21 Morphologic dune and ripple changes in Ius and Juventae chasmata.

All time steps images are provided with the same FOV and all symbols are static. (a–b) Ripple realignment or adjustment is observed in some locations (left, white bars), whereas other rippled surfaces (right, white bars) show no change. Other locations show evidence for dune perimeter modifications (black arrows). Note that there is a change in solar azimuth; see Table 4.9 for image information. See Fig. 4.18a for location. (c–d) Changes in a Juventae Chasma dune slip face are observed over ~2 MY. Three slip face alcoves appear paired with fans of sand below (black arrows). See Fig. 4.11c for location.

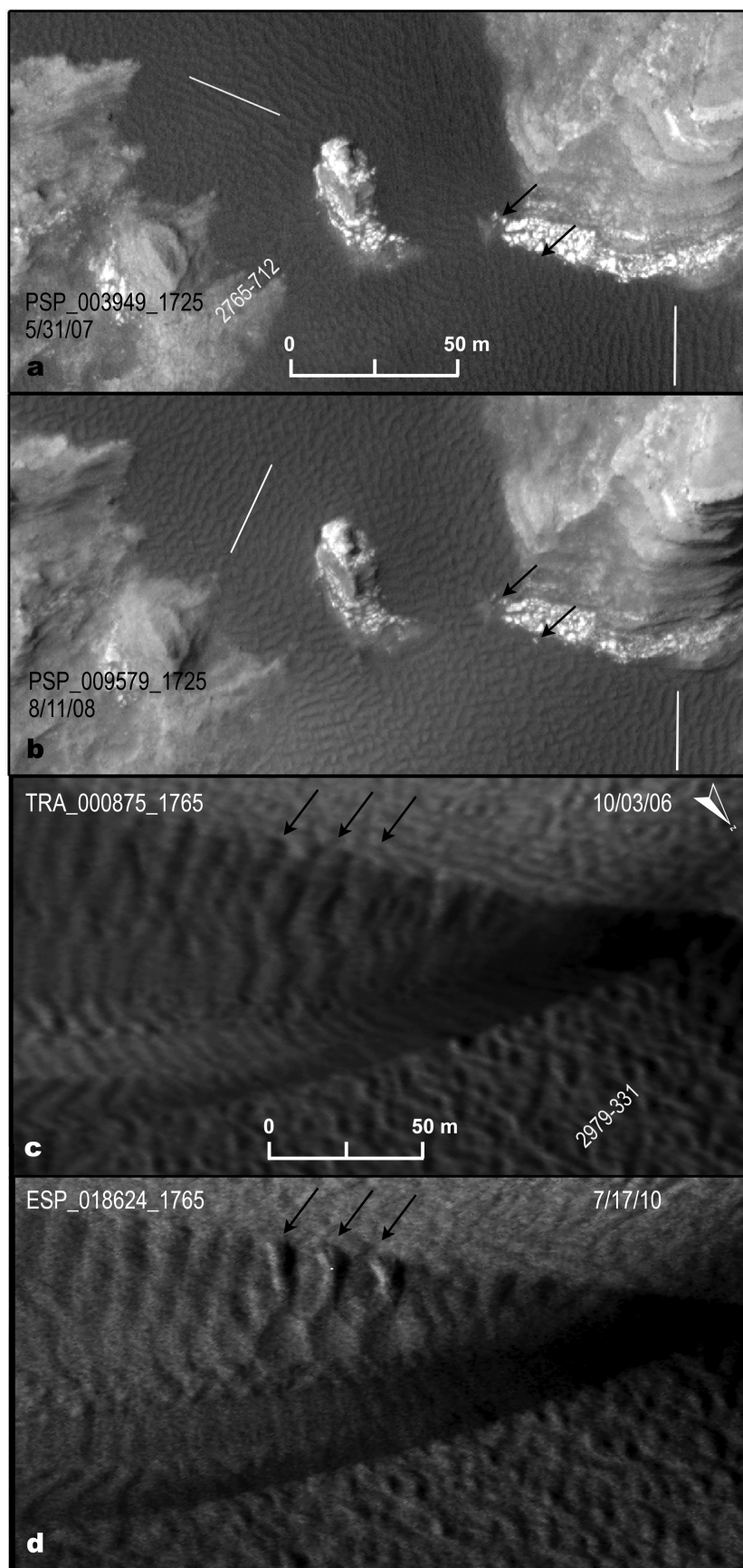


Figure 4.21

Figure 4.22 Ganges Chasma dune and ripple changes.

(a) Deflation, modification, and migration (white triangles) of dunes and ripples in HiRISE PSP_008536_1725. The same FOV is provided for 4.22b–4.22c and for 4.22d–4.22f. See Table 4.9 for image information and Fig. 4.10d for locations. (b–c) Changes in the stoss side (black arrows) and slip face position (white arrow) over the ~2 MY between images. Other examples of ripple migration can be found (e.g., lower black arrows). (d–f) A barchan dune in a 2005 MOC images (yellow polygon), HiRISE images in and 2008 (white polygon) and 2012 (pink polygons) showing bulk transport to the south. Insets in Figs. 4.22f–4.22g are ~20-m-wide.

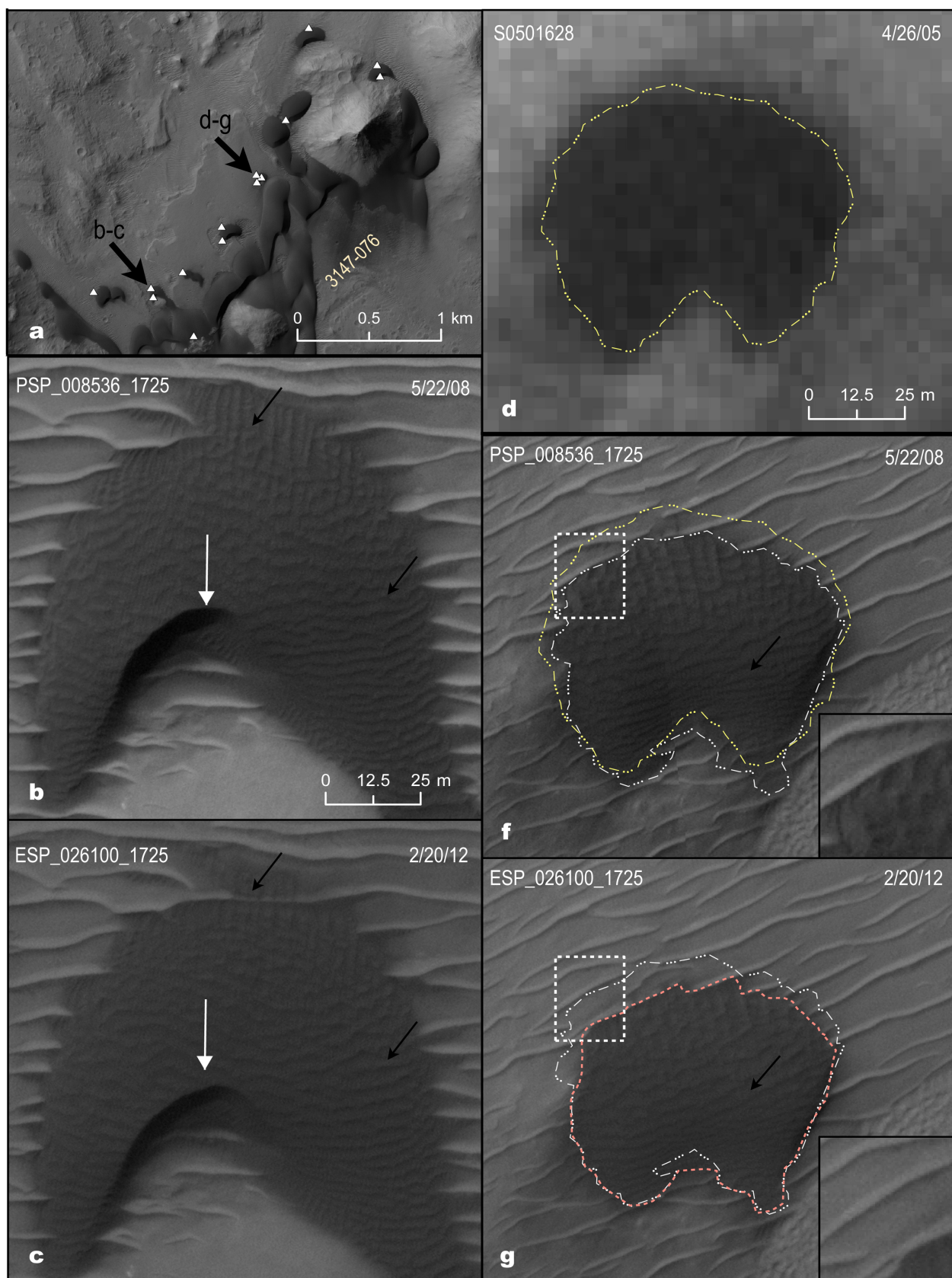


Figure 4.22

Attachment 4.1 An animated time-step sequence of ripple movement in Ius Chasma. Compare with Fig. 4.21a–4.21b and see Fig. 4.18a for location. For attachments 4.1–4.4 image information is listed on Table 4.9.

Attachment 4.2 An animated time-step sequence of sand deflation in Ius Chasma.
See Fig. 4.18a for location.

Attachment 4.3 An animated time-step sequence of a barchan change in Ganges Chasma. Compare with Figures 4.22b–4.22c and see Fig. 4.22a for location.

Attachment 4.4 An animated time-step sequence of a barchan change in Ganges Chasma. Note a small apparent change in static landforms due to image parallax. Also, the second time step has a slight offset between CCD pixels in the center of the image. Compare with Figures 4.22f–4.22g and see Fig. 4.22a for location.

Attachment 4.5 An animated time-step sequence of three Mars Years of TES bolometric albedo of the Valles Marineris region.

FOV is approximately 6000 km across. Typically dune fields get darker from MY 24 to 25 (*e.g.*, from 0.13 to 0.11), and then get brighter from MY 25 to 26 (*e.g.*, from 0.11 to 0.14), and the largest dune albedo changes were up to 35%. Compare with Figures 4.8b–4.8c.

CHAPTER V: VALLES MARINERIS DUNE SEDIMENT PROVENANCE AND PATHWAYS

Abstract

Although low-albedo sand is a prevalent component of the martian surface, sources and pathways of the sands are uncertain. As one of the principal present day martian sediment sinks, the Valles Marineris (VM) rift system hosts a diversity of dune field populations associated with a variety of landforms that serve as potential sediment sources (PSSs), including: spur-and-gully walls, interior layered deposits (ILDs), and landslides. Here, we test the hypothesis that VM dune fields are largely derived from a variety of local and regional (intra-rift) sediment sources. Results show several dune fields are superposed on ancient wall massifs and ILDs that are topographically isolated from extra-rift sand sources. Spectral analysis of dune sand reveals compositional heterogeneity at the basinal-, dune field-, and dune-scales, arguing for discrete, relatively unmixed sediment sources. In Coprates and Melas chasmata, mapping is consistent with the principle sand source for dunes being Noachian-aged upper and lower wall materials composed of primary (igneous) minerals and glasses, some of which show evidence for alteration. In contrast, examination of dune fields in Capri, Juventae, and Ganges chasmata show evidence for partial sediment derivation from adjacent Early Hesperian-aged hydrated sulfate-bearing ILD units as stored secondary sources. Dunes containing these “soft” secondary minerals are unlikely to have been derived from distant sources due to the physical weathering of sand grains during transport. Isolated extra-rift dune fields, sand sheets, and sand patches are located on the plateaus surrounding VM and the adjoining areas, but do not form interconnected networks of sand pathways into the rift in the current epoch. Nevertheless, if past wind regimes were consistent with more recent regimes inferred from morphological analysis (i.e., dune slip faces, wind streaks), and were sufficient in strength and duration, small dune populations within Aurorae Chaos and north of eastern VM might have been part of ancient extended sand pathways into VM. We favor local and regional (intra-rift) derivation of dune sand from a variety of these lithologic sources. Furthermore, we suggest these examples of local and regional sourcing are more easily detected and volumetrically significant than many other locations on Mars, due to the extreme topography, compositional diversity of local lithologies, and extent of source units.

1.0 Introduction and Hypothesis

Low-albedo sand is a prevalent component of the martian surface, as commonly observed at landing sites and inferred from orbit [*Soderblom and Bell*, 2008]. The wide-spread nature of dunes and other aeolian bedforms on Mars shows that near-surface winds have commonly interacted with plentiful mobile sediments. Aeolian sediment formation must have occurred over most of Mars' history, as bedform sand and sandstone are associated with a wide-range of terrain ages (Chapter IV) [*Grotzinger et al.*, 2005; *Edgett and Malin*, 2000a]. However, the sources, both in terms of lithology and geographic location, of the sand are uncertain. Similarly, definitive sand transport pathways that connect sources to sinks have been elusive [*Fenton*, 2005; *Fenton et al.*, 2005]. Thus, the sources and pathways of martian sands are an active area of investigation [*Bourke et al.*, 2004; *Fenton*, 2005; *Langevin et al.*, 2005; *Stockstill-Cahill et al.*, 2008; *Tirsch et al.*, 2011; *Silvestro et al.*, 2010a; *Horgan et al.*, 2009].

Whereas martian sand compositions are primarily basaltic [*Soderblom and Bell*, 2008], terrestrial sand compositions are dominated by felsic minerals (dominantly quartz), which are principally derived from the long-term chemical and physical weathering of felsic continental crust [*Pye and Tsoar*, 1990; *Muhs*, 2004]. Important primary sources of potential dune sediment include glacial erosion of bedrock and pyroclastic deposits. Following the weathering of country rock, sediment transport (including fluvial, aeolian, glaciogenic, and gravity-driven processes) distributes the sand and possibly alters it, both texturally and compositionally [*Greeley and Iversen*, 1985; *Pye and Tsoar*, 1990; *Muhs et al.*, 2003]. Secondary sources may involve sedimentary deposits (e.g., sandstone), river banks, or marine or lacustrine beaches [*Pye and Tsoar*, 1990; *Bandfield*, 2002b; *Lancaster*, 2009].

Besides a primary or secondary classification, these terrestrial sources may also be grouped according to size and proximity from the sand deposit. “Local” sources are small supplies of the sand that is produced by the erosion of crystalline or clastic bedrock that is frequently proximal to the sand deposits [Johnsson, 1993]. This *in situ* derivation of sand is most commonly reported from volcanoclastic or weakly-cemented sedimentary surfaces [Edgett and Lancaster, 1993; Muhs, 2004; Hayward et al., 2010a; Baratoux et al., 2011]. “Regional” sources are more expansive sediment supply sites, which may occur distal from the dunes fields, but are generally larger than the dunes they source [Kocurek and Lancaster, 1999]. A “distant” source may connect with extended sand pathways that stretch 100s to 1000s of kilometers to a sink with fluvial and/or aeolian transport being an integral part of the system [Fryberger, 1979; Gay, 1999; Muhs et al., 2003]. Combinations are likely to occur; for example, the Kelso Dunes of the Mojave Desert are primarily derived from Mojave River wash (~55 km; regional source), but partially sourced from Providence Mountain fans (~5 km; local source) [Kocurek and Lancaster, 1999; Ramsey et al., 1999]. However, distant sources for terrestrial dunes are often difficult to identify without active sand transport pathways, low rates of chemical weathering, and long-term tectonic stability.

Although similar processes for sediment derivation may occur on Mars, exact mechanisms and their relative contributions are poorly constrained. The production of low-albedo sands and soils throughout martian history has likely involved multiple processes, including fluvial, glacial, lacustrine, impact-related, aeolian, volcanoclastic, and aqueous chemical alteration, most of which are compounded by gravity-driven processes [Edgett, 1994; Hall, 1999; Greeley et al., 2006; Murchie et al., 2009b; McGlynn et al., 2011]. Based on geologic context gained from low-resolution images, several authors have pointed to correlations

between putative sediment sources (e.g., layered deposits, crater floors) and adjacent dark-toned aeolian bedforms [Thomas and Weitz, 1989; Lucchitta *et al.*, 1992; Edgett, 1994]. More recent studies with higher resolution data have shown evidence of more specific sand sources [Bourke *et al.*, 2004; Fenton, 2005; Stockstill-Cahill *et al.*, 2008; Tirsch *et al.*, 2011; Chojnacki *et al.*, 2010]. For instance, Tirsch *et al.* [2011] proposed local sourcing of aeolian sediment from buried, ancient ash deposits, based on spectral similarities between dark layers within crater walls and the adjacent bedforms. Other authors have described potential regional sources and sand pathways using photogeologic evidence (e.g., dark sediment emerging from outcrops) and paleowind indicators (e.g., dune slip faces) [Bourke *et al.*, 2004; Fenton, 2005]. Chojnacki *et al.* [2010] argued for local wall outcrop sources and discrete sand pathways for Valles Marineris falling dunes based on thermophysical and topographic relationships. Perhaps the most established case of a dune sediment source is that of the Olympia Undae gypsum-rich sand that is likely derived from the lower layers of the polar layered deposits or basal units as a regional source [Langevin *et al.*, 2005; Horgan *et al.*, 2009].

As one of the principal martian sediment sinks, Valles Marineris (VM) contains nearly a third of the non-polar dune area on Mars (Chapter IV). The ~22,000 km² of VM dune fields show an assortment of morphologies, thermophysical properties, compositions, and associated landforms. Intra-rift (within the chasmata) potential sand sources (PSSs) include: interior layered deposits (ILDs), spur-and-gully walls, and landslides [Chojnacki *et al.*, 2012a, 2012b]. The rift system exposes a ~10-km-thick section of martian crust from which dune material may have been derived. Valles Marineris also possess a diversity of primary and secondary lithologies that may serve as mineralogic “markers” in detecting dune sediment provenance and linking sources to sinks.

Based on these observations and implications, we hypothesize that *local and regional (intra-rift)* sources are major contributors to Valles Marineris dune fields. If the hypothesis is found to be true, this fact would provide insight into local and/or regional sand pathways and the degree to which VM sand-sized sediments have been homogenized or altered over relatively short distances. If the hypothesis is false and active sand pathways are observed leading into VM, this result would imply distant (extra-rift) sources exist and sand on Mars could travel considerable distances.

The central questions related to our hypothesis are:

1. What intra-rift lithologies appear to produce aeolian sediment and what evidence might constrain their relationship to sand dunes? Do these relationships vary with chasmata?
2. Are there extended sand pathways leading into VM? Evidence for pathways would include networks of dunes and dark streak on the plateaus surrounding the rift or adjoining canyon areas.
3. If sources can be inferred, then how do the mineral, chemical, and corresponding spectral signatures of sands differ from those of the inferred source material? Is there evidence for chemical alteration during transport?
4. If local/regional sources are found, is there evidence for sand fractionation, either compositionally or texturally? Does dune sand composition correlate with inferred grain sizes?
5. Assuming sand sources are found, what were the geologic formations mechanisms and ages for the source lithologies? Similarly, what were the mechanisms or processes responsible for the breakdown and mobilization of sediment from that source rock?

In this paper, we investigate the morphology, composition, and thermophysical properties of Valles Marineris dune fields in order to infer their provenance and sand pathways. First we provide an overview of the study area and possible source regions. A summary of the data sets and methodology used for our analysis follows. Next we evaluate possible local or regional (intra-rift) sources for VM dunes, in addition to possible extended sand pathways leading into VM. We then discuss the findings in the context of the hypothesis and related questions.

2.0 Study Region

Our study region consists of the equatorial Valles Marineris rift system (Fig. 5.1). At over 4,000 km in length, Valles Marineris is not only the most areally extensive graben complex on Mars, it has also been identified as one of the largest structural complexes in the Solar System [Blasius *et al.*, 1977; Frey, 1979]. This geologically rich region of sub-rectangular, interconnected troughs (Fig. 5.1b) hosts massive landslides [e.g., Quantin *et al.*, 2004a], thickly-bedded sedimentary layered deposits [e.g., Blasius *et al.*, 1977], and extensive erosional and depositional aeolian landforms [e.g., Lucchitta *et al.*, 1992; Chapter IV]. Evidence of massive physical weathering and sediment breakdown is found throughout the rift, including erosion and mass-wasting of wall slopes, ILDs, and landsides [Lucchitta *et al.*, 1992; Chojnacki *et al.*, 2006, 2010; Chojnacki and Hynek, 2008]. Previous mapping in Valles Marineris identified 60 dune fields among several canyon environments (Chapter IV, Table 4.5). Four classes of VM dune fields were previously described: floor, landslide, wall, and erg (Fig. 5.1b; Chapter IV, Figs. 4.11–4.14). Floor dunes, the most frequent and wide-spread class of dune fields, occur in all chasmata and are located adjacent to or among ILDs, spur-and-gully walls, chaotic materials, craters, and landslides. Landslide dune fields are superposed on Late Amazonian (<1 Ga) landslides in Ius, Melas, Coprates, and Hebes chasmata. Wall-related dune fields (falling and

climbing dunes [*Chojnacki et al.*, 2010]) are concentrated among Coprates, Melas, and Hebes chasmata spur-and-gully walls. Finally, massive ($>1000 \text{ km}^2$) ergs are located in Juventae and Ganges chasmata adjacent to ILDs, walls, chaotic materials, and landslides.

Extended sand pathways or corridors have likely delivered sand from extra-rift sources over VM's $>3 \text{ Ga}$. These pathways would exist on the higher-lying plateaus or adjoining lower-lying regions (Fig. 5.1a). This vast area ($>5,000,000 \text{ km}^2$) displayed in Fig. 5.1a includes the 1:5 million-quadrangle regions of Lunae Palus (north), Oxia Palus (northeast), Margaritifer Sinus (east), Coprates (central), Phoenicis Lacus (west), and Tharsis (northwest) (Fig. 5.1a). The Tharsis rise dominates the area to the northwest and is thought to have been the center of martian volcanic and tectonic activity for much of the planet's history [*Scott and Tanaka*, 1986; *Carr*, 2006; *Dohm et al.*, 2009]. Large-scale deformation is clearly linked to the Tharsis load, manifested as radial faults, circumferential wrinkle ridges, and transform faults about the rise [*Carr*, 2006] that might inhibit sand pathways. Potential sand corridors include Hesperian-age channels (e.g., Maja Valles) emerging from chasmata in the area north and east of VM, forming the western part of the circum-Chryse outflow channel systems [*Lucchitta et al.*, 1992; *Chapman et al.*, 2003]. Adjacent higher-lying terrains are heavily degraded by impact craters that might serve as sediment sinks or sources [*Christensen*, 1988b; *Tirsch et al.*, 2011]. Southeast margins show low-elevation chaotic terrains (e.g., Aurorae Chaos) and the Margaritifer Sinus valley network, including Holden crater, to the southeast. South of VM lies a broad expanse of highland cratered terrain including the Late Noachian tectonic province of Thaumasia Planum and the Solis Lacus classic albedo feature [*Scott and Tanaka*, 1986; *Dohm et al.*, 2009]. Finally, the collapse terrain of Noctis Labyrinthus is connected to the western edge of VM by deep pits and fossae. Previous studies described dark-toned dune fields (red triangles) and sand sheets

within the area shown in Fig. 5.1a, five from the Mars Global Digital Dune Database [Hayward *et al.*, 2007a], two dune fields from Tirsch [2009], in addition to the 15 identified by this study.

3.0 Data Sets and Methods

In order to address the hypothesis identified above we performed data collection, mapping, and analysis using multiple spacecraft data sets.

3.1 Assessing Sand Sources, Pathways, and Provenance

We adopt the convention of Fenton [2005] when characterizing martian sources as local or regional. Local sources are discrete areas, which are in spatial contact with the larger aeolian bedforms they source via short pathways. Regional sources are larger in area than the dune fields they provide sediment for and can occur in contact or several mean dune field widths away from that source with sand pathways generally 10s of kilometers in length. Regional sediment sourcing might include the cannibalization of sand from a larger dune field to smaller, more mobile barchan or climbing dunes [Kocurek and Lancaster, 1999]. In addition, these sources would need to have experienced wind regime(s) that are appropriate, with respect to strength, direction, and persistence, in order to deliver sand across long distances.

External (distant) sources are not explicitly sought, due to the vast extent of the surrounding area ($>5,000,000 \text{ km}^2$), relative non-uniqueness of typical basaltic compositions, and possibility of chemical weathering. Rather, we look for evidence for extended sand pathways that have the potential for delivering sediment to VM from the surrounding plateaus or adjoining canyon areas. Sand pathways are recognized by intermittent clusters of dunes and dark streaks with similar orientations, indicating they formed under a consistent wind regime [Pye and Tsoar, 1990; Lancaster and Tchakerian, 1996; Al-Enezi *et al.*, 2008]. Pathways to the rift must be viable; crater chains, chaotic materials, fossae, or catenae would likely prohibit extended routes

[Christensen, 1988b]. Alternatively, topographic depressions (e.g., chasmata, valles, troughs) connected with VM may serve as sand corridors and lead to enhanced aeolian sand transport [Pye and Tsoar, 1990; Edgett, 1994; Al-Enezi *et al.*, 2008].

Sediment provenance includes a given sediments' derivation location(s) and their compositional relationship to that source. Assessing dune sediment provenance without *in situ* geochemical data, wind measurements, and sediment textural characteristics is a challenging endeavor. However, the vast quantity of Mars remote sensing data and apparent lack of large-scale crustal recycling and intense chemical weathering [Johnsson, 1993; Solomon *et al.*, 2005; Hazen *et al.*, 2008; Soderblom and Bell, 2008] enable this type of study, despite the lack of field validation.

Based on terrestrial studies [Pye and Tsoar, 1990; Johnsson, 1993; Kocurek and Lancaster, 1999; Ramsey *et al.*, 1999] and the scale of available remote sensing data sets (sections 3.2–3.5), the following characteristics may be used to establish local or regional sand sources for martian dune fields:

- i. Morphologic evidence for eroded detrital material mass wasted from bedrock or lithified deposits (e.g., scree or talus) should be present. Any evidence of ongoing erosion of that bedrock or lithified deposit (e.g., boulder tracks) would indicate that the mass-wasting was geologically recent [Russell *et al.*, 2008; Bateman *et al.*, 2012] .
- ii. Transport direction of sand, as indicated by wind streak and dune slip face orientations, should be broadly away from the source. Also, although there are exceptions, dune sediment sources typically occur at higher elevations than the dunes they supply [Greeley and Iversen, 1985; Pye and Tsoar, 1990].

- iii. For underlying slopes at or near the angle of repose for sand-sizes ($\sim 32^\circ$), dune sediment pathways will be unidirectionally downslope [*Bagnold*, 1941; *Tsoar et al.*, 1996].
- iv. Inferred particle size/degree of induration of the putative source material should be greater than or equal to that of dune sediment. Dune sand particle sizes should be equal or smaller with increasing distance downwind of the PSS (i.e., decreasing thermal inertia), as a result of the preferential transport of finer material [*Wilson*, 1973; *Johnsson*, 1993; *Ramsey et al.*, 1999].
- v. The mean size of any talus boulders should decrease between the PSS and the associated dune field.
- vi. In the case of a clastic protolith partially comprised of light-toned alteration products (e.g., phyllosilicates, sulfates), reflectance or albedo of sediment downwind from the PSS should decrease due to deflation of weaker light-toned particles and greater durability of low-albedo mafic material [*Ghrefat et al.*, 2007; *Nesse*, 2012].
- vii. Given limited chemical weathering, dune sediment spectra should resemble the spectra of the putative source bedrock and talus [*Johnsson*, 1993; *Ramsey et al.*, 1999]. Additional spectral effects due to compositional mixing and/or particle size differences [*Hiroi and Pieters*, 1994; *Poulet and Erard*, 2004] or aeolian driven compositional fractionation [*McGlynn et al.*, 2012] should be considered.

Evaluation of the spectral and thermophysical characteristics described above (iv, iv-vii) is best accomplished using contemporaneous remote observations of both the dune field and the PSS (i.e., the same observation with near-identical atmospheric path lengths and emission angles), to reduce errors associated with atmospheric and temporal variations. These criteria assume some sediment sources still show evidence for sand production and are spatially

coincident with dunes. It is possible some sand sources persisted for a time, then ceased sediment production and are no longer detectable. The scale of the observation (i.e., image resolution) is an important factor when discussing provenance, as most terrestrial studies describe details impossible from orbital observations. In addition, exceptions to these criteria may occur; for example, climbing dunes located high on VM walls may be sourced from units and dunes at lower elevations (conflicting with prediction ii above) [Chojnacki *et al.*, 2010]. Finally, high-resolution (spectral and spatial) data sets with adequate spatial coverage are required to properly assess dune provenancial relationships [Edgett, 1994].

3.2 Dune and Outcrop Morphology: Visible-Wavelength Images

To identify photogeologic evidence of dune sediment provenancial relationships and pathways, medium- and high-resolution visible-wavelength images from three spacecraft instruments were used. To characterize fine-scale morphology, this study utilized images from the Context Camera (CTX) [Malin *et al.*, 2007] at 5–6 m/pixel and the High Resolution Imaging Science Experiment (HiRISE) at 25–100 cm/pixel [McEwen *et al.*, 2007], both onboard the Mars Reconnaissance Orbiter (MRO). Supplementary images from the Mars Express’s (MEx) High Resolution Stereo Camera (HRSC) [Neukum *et al.*, 2004] were examined (~13 m/pixel). In addition, wide-angle (~230 m/pixel) mosaics from the Mars Orbiter Camera (MOC) [Malin *et al.*, 1992] onboard the Mars Global Surveyor (MGS) were used for regional context views. All images were radiometrically calibrated using U.S. Geological Survey (USGS) Integrated Software for Imagers and Spectrometers (ISIS) software [Gaddis, 1997; Anderson *et al.*, 2004] and spatially co-registered to each other and to other data sets (sections 3.3–3.5), with ArcGIS and ENVI software.

3.3 Dune and Outcrop Topography: Digital Elevation Models

To measure topographic relationships between dunes and PSSs (e.g., maximum gradient, relative elevations), three types of digital elevation models (DEMs) were employed. Gridded regional topographic elevation data from the Mars Orbiter Laser Altimeter (MOLA) onboard MGS were used at 128 pixels per degree [Smith *et al.*, 2001]. To investigate relative surface topography and slopes at 50–150 m/pixel, HRSC DEMs were used [Neukum *et al.*, 2004]. Using HiRISE stereo pairs, high-resolution DEMs (1 m/pix) [Kirk *et al.*, 2008] were constructed with SOCET SET® BAE system photogrammetry software at the U.S.G.S. processing guest facility [Kirk *et al.*, 2009]. The resulting HiRISE DEMs were co-registered with MOLA data points and have elevation postings every 1 m with sub-meter vertical precision.

3.4 Dune and Outcrop Thermophysical Properties: Albedo and Thermal Inertia

In order to survey the region for dark-toned PSSs and sand pathways, this study examined albedo data from the Thermal Emission Spectrometer (TES) [Christensen *et al.*, 1992] onboard MGS. TES bolometric albedo measurements were used, as calculated and resampled to ~3 km/pixel (20 pixels per degree) by Putzig and Mellon [2007a].

To assess the thermophysical properties of dune and PSS surfaces, thermal inertia data were used. Thermal inertia is largely controlled by particle size with additional influence from other factors (e.g., degree of induration, reduced pore space) [Ferguson *et al.*, 2006a; Putzig and Mellon, 2007; Piqueux and Christensen, 2009]. However, for large sandy bedforms that are likely well-sorted and thicker than several diurnal thermal skin depths (2–4 cm), some of these complications diminish. Particle sizes were derived from thermal inertia values using the assumption of a homogenous, mono-modal sediment population [Piqueux and Christensen, 2009].

High-resolution thermal inertia maps at ~100 m/pixel from Mars Odyssey's (MO) Thermal Emission Imaging System (THEMIS) nighttime infrared images [*Christensen et al.*, 2004; *Ferguson et al.*, 2006a] were derived for investigation of intra-rift PSSs and dunes (see Chapter IV, section 3.3). Particle size analysis was performed for contiguous dune areas, using 100s to 1000s of THEMIS pixels that appeared qualitatively homogenous and devoid of rock outcrops in corresponding CTX and HiRISE images.

3.5 Dune and Outcrop Composition: Spectra

For investigation of PSS and dune sediment compositional relationships, spectral data from two instruments were used.

3.5.1 CRISM Hyperspectral and Multispectral Images

The Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) onboard MRO provided visible and near-infrared reflectance (VNIR) spectra, used to determine outcrop, talus, and dune sand compositions [*Murchie et al.*, 2007]. CRISM is comprised of two spectrometers, a VNIR instrument (362–1053 nm) and an infrared (IR) instrument (1002–3920 nm) and can be operated in two modes. Targeted observations provide a high spatial and spectral resolution of 18 or 36 m/pixel with 544 bands (hyperspectral), whereas survey mode observations are at 100 or 200 m/pixel with 72 bands (multispectral).

CRISM data were processed using the CRISM Analysis Toolkit (CAT) [*Seelos et al.*, 2011] interface, as implemented in the ENVI software package. Data reduction followed the method described by *Murchie et al.* [2009c], which includes calibration to radiance, photometric correction by division by the cosine of the solar incidence angle, and (for the IR) atmospheric correction by a scaled transmission spectrum derived from observations over Olympus Mons [*McGuire et al.*, 2009]. Spectral summary products described by *Pelkey et al.* [2007] were then

created using CAT to highlight mineralogic spectral features of interest for further analysis. The spectral summary products used in this study are listed in Chapter IV's Table 4.1.

In addition to examining maps of spectral summary products, individual spectra were also assessed. Most of the spectra presented herein are presented as ratioed spectra, in which target spectra were averaged and then divided by the average of a spectrally unremarkable (compositionally “bland”) region of similar area, usually (but not necessarily) in the same column as the target pixels in the unprojected image. This method is commonly used in CRISM spectral analyses to accentuate absorption features and reduce systematic errors (detector spikes) and residual atmospheric features [e.g., *Ehlmann et al.*, 2009; *Murchie et al.*, 2009c]. The averaging of tens to hundreds of pixels in the target area and the bland normalization area helps improve signal-to-noise ratio. Here, CRISM spectra are presented from 0.45 to 2.6 μm , but the VNIR and IR images were processed and analyzed separately. Spectra from the two detectors were then joined by manually offsetting VNIR data to match the IR data in the $\sim 1\text{-}\mu\text{m}$ crossover region, and the bad bands of the VNIR ($>1.0102\text{ }\mu\text{m}$) to IR ($<1.0210\text{ }\mu\text{m}$) were excluded [*Bishop et al.*, 2009]. Additional caution should be employed for interpretations of VNIR spectra at wavelengths less than $\sim 0.8\text{ }\mu\text{m}$, which can suffer due to poor corrections for atmospheric dust [*Murchie et al.*, 2009c]. To assess the presence of ferric or ferrous iron, VNIR spectra were analyzed and presented using continuum removal, following the methods described by *Roach et al.* [2010a]. CRISM data products used in this study are summarized in Table 5.1 and Chapter IV's Table 4.3 for extra-rift and VM locations, respectively.

Because determination of the compositions of dunes and PSSs was a key goal of this study, our methods differed in several respects from those employed in past CRISM studies. When possible, dune and PSS spectra were collected from the same columns in the unprojected

CRISM images, to avoid systematic detector errors. These target spectra were ratioed by the same spectral denominator, not necessarily collected from the same column as either target. This method eliminates differences due solely to different denominators, but may create target spectra with variable detector artifacts found in two target spectra. However, comparison between dune and PSS unratioed spectra can avoid issues from potential variable “noisy columns”. Where no suitable spectrally bland areas could be used as denominators, the original unratioed spectra of dune and PSS were analyzed. This method typically produces spectra with contributions from the atmosphere, instrument artifacts, and surface dust leading to difficulties in mineral detection [Murchie *et al.*, 2009c], but can be useful when comparing relative compositional differences of adjacent surfaces. CRISM multispectral images were often employed in this study (in addition to full spectral resolution data, where available), as they provide superior spatial coverage of dune fields and PSSs in a single observation. The long north-south survey mode observations allow multiple targets to be divided by the same denominator of a dusty plateau surface. Qualitative comparisons between plateau “dust” spectral denominators show relatively homogenous spectra from east to west VM. Finally, CRISM infrared albedo (IRA, reflectance depth at 1.3 μm ; Chapter IV, Table 4.1) was used to examine VNIR reflectance trends between dunes and PSSs.

3.5.2 HiRISE Color Images

In coordination with CRISM-mapped mineralogy, HiRISE multi-band color images were used to assess meter-scale variations in color and inferred composition [McEwen *et al.*, 2007, 2010]. Three broad-bands (IR at 874 nm, RED at 694 nm, and BG at 536 nm) cover the center ~1 km swath of the 6-km-wide HiRISE images and are at wavelength positions that maximize the signal-to-noise and reduce the effects of atmospheric hazes [Delamere *et al.*, 2010]. The

three bands were independently scaled to I/F values and shadowed areas subtracted or offset following the methods of *Delamere et al.* [2010]. This technique removes most atmospheric and topographic contributions so color variations are related to surface composition. Using corrected I/F values band ratios were created (IR/RED, IR/BG, and BG/RED). Generally, this approach allows the analyst to distinguish between mafic and altered compositions minerals due to the relative absorption features of ferrous and ferric iron at HiRISE band positions and can be useful when mapping at the scale of outcrops [*Delamere et al.*, 2010; *Chojnacki et al.*, 2011a]. In addition to HiRISE grayscale RED images, enhanced false-color (IRB) and enhanced 3-color (RGB) images are used in figures; *Delamere et al.* [2010] provide details on creation of those products.

3.6 Dune Field Mapping and Paleowind Directional Analysis

Dune field identification and mapping followed the methodology described in Chapter IV, section 3.7. Possible sand pathways and intra-rift PSSs were identified using TES albedo followed by examination of CTX and THEMIS images. Dune slip face and dark streak orientations were inspected to provide insight into major wind regime(s), sediment transport direction(s), and spatial relationships with PSS. As in previous studies [*Hayward et al.*, 2007a; *Tirsch et al.*, 2011; *Gardin et al.*, 2012], only those dunes formed under a unidirectional wind regime (e.g., barchan, transverse) were used for slip face analysis. All usable slip faces in a given field were averaged to provide a “mean slip face direction” that was used for paleowind analysis (sections 4.1, 4.2.2). For crater-related dunes, the dune fields’ position within the hosting crater, or dune centroid azimuth (DCA) [*Hayward et al.*, 2007a], was also examined to approximate long-term migration direction (see Chapter IV, section 1). See Table 5.2 for naming convention of dune fields.

4.0 Results: Valles Marineris Dune Field Potential Sediment Sources and Extra-Rift Pathways

Valles Marineris dune fields are found adjacent to a variety of landforms that may serve as intra-rift sources. Extended sand pathways on the surrounding plateau and adjoining areas are also possible source for VM dune sand and are investigated.

4.1 Valles Marineris Intra-Rift PSSs

4.1.1 Spur-and-Gully Wall PSSs

In eastern Coprates Chasma (Fig. 5.2a), dune fields are commonly below or among spur-and-gully walls [*Chojnacki and Moersch, 2009; Chojnacki et al., 2010*]. One example of this relationship (Fig. 5.3) shows a compound dune field of transverse and barchanoid dunes in collapse pits south of the main chasm and in close association with a series of eroding spurs on three sides. Fans of talus, with rock sizes down to the limit of HiRISE resolution (i.e., ~1 m), can be observed downslope grading into the dune field (Figs. 5.3b–5.3d). Dune slip faces are generally oriented eastward (away from the PSS) and follow the topographic gradient. HRSC topographic profiles between spurs and dunes were evaluated for potential sand pathways (Fig. 5.3a, inset) to test the possibility that sand blew upslope onto unrelated outcrops. Slopes of these PSS talus deposits range from ~30–40°, above the angle of repose for most dry sand-sized materials. THEMIS thermal inertia of these dunes show relatively high values (>300 tiu), and for unconsolidated sediments, are consistent with coarse to very coarse sand (Chapter IV). Thermal inertia and CRISM IRA albedo values decrease monotonically downslope from the surrounding walls toward the dune field's eastern edge along the A to A' profile (Chapter IV, Fig. 4.16a). Because this trend occurred relatively consistently, regardless of slope orientation, it does not appear to be solely due to illumination or thermoclinometric (slope-related) effects.

CRISM observations of lower spurs, talus, and dunes reveal the same broad spectral absorption feature, from 1.0 to 1.3 μm , due to olivine (Fig. 5.3a, lower inset). Unratioed spectra of lower wall talus and dunes are nearly identical except for overall offsets in reflectance values (Fig. 5.4a). Spectra of east and west spurs higher in the sequence (Fig. 5.3a, upper inset) show narrow absorption features at 1.4, 1.9, and 2.2 or 2.3 μm , likely due to the presence of Al- (e.g., kaolinite) or Fe/Mg-phyllosilicates (e.g., smectite), respectively, and are unlike spectra of dunes. In addition, spectra of the collapsed terrain just to the south of this location indicate the presence of low-calcium pyroxenes (LCP) [Poulet *et al.*, 2009], which is not evident in local sand spectra.

An ~800-m-tall wall scarp and associated talus above a large floor dune field (~220 km^2 , Figs. 5.2a, 5.5–5.6) is located eastward in the main chasm. Crescentic dune slip faces close to the wall are broadly oriented away from the wall scarp (Fig. 5.6a, red arrows). In contrast, the dark streaks, barchanoids, and barchans in this field are oriented toward the west where they grade into relatively thin sand sheets (Fig. 5.5). HiRISE observations of these outcrops resolve meter-scale boulders and 100-m-scale talus-fans leading into the dune field (Fig. 5.6b, inset). Profile A–A' (Fig. 5.6a, inset) represents a proposed potential sand pathway, where THEMIS thermal inertias and CRISM IRA albedos generally decrease downslope and away from the scarp southwestward. Barchanoid dunes adjacent to talus have thermal inertia values (320–350 tiu) consistent with coarse sand, whereas dunes paleo-downwind . Also on that profile, HRSC-derived slopes of the scarp above dunes are relatively steep (Fig. 5.6a, inset, $>30^\circ$). CRISM spectra of certain dunes and scarp (Fig. 5.6b, inset) are most consistent with high-calcium pyroxene (HCP), based on the broad ~1- and ~2.3- μm features. However, other adjacent dunes and lower talus possess spectra with narrower ~1- μm absorptions and more subdued ~2- μm features, which are consistent with spectra of a mix of olivine and pyroxene (basalt). In contrast

to these dune, scarp, and talus spectral compositions, other mid-wall spurs in that figure have spectral characteristics that match LCP, Fe/Mg-smectite [Flahaut *et al.*, 2012], and other basaltic compositions (Fig. 5.4b).

Just west of this location, mapping has shown several relationships between different dune populations and local outcrops (Figs. 5.2, 5.7). Dark-toned dune fields are superposed on southern chasm wall slopes and both sides of a ~200-km-long horst or massif (Fig. 5.2a; Chapter IV, Fig. 4.13c) [Schultz and Lin, 2001], hereafter referred to as the “east Coprates massif” [Beyer and McEwen, 2005]. Falling dunes, a type of topographically constrained duneform, are located within gullies adjacent to upper wall dark-toned outcrops and 2–3 km above the canyon floor (Figs. 5.7b–5.7c). Underlying slopes of falling dunes are 8–18°, according to HIRISE and HRSC DEMs, and steep (>30°) slip faces are arranged on the downslope side, consistent with downslope transport [Chojnacki *et al.*, 2010]. Slopes of the higher- and lower-lying spurs, gullies, and talus above and below these falling dunes commonly exceed the angle of repose for sand-size particles (>32°; Chapter IV, Fig. 4.13). Falling dunes possess relatively high THEMIS thermal inertia values (300–350 tiu), consistent with large grain sizes (~1–2 mm) [Chojnacki *et al.*, 2010]. Spectra of falling dunes and adjacent upper wall outcrops possess similar spectral features and slopes, both on the south wall and east Coprates massif (Figs. 5.4c, 5.7b–5.7c). Below these occurrences, dark-toned lower wall outcrops and associated talus of the massif possess similar spectra to those of local floor dunes in CRISM observations (Fig. 5.7d). Spectra of falling dunes and adjacent floor dunes were also compared, and found, in many cases, to match (Figs. 5.4c, 5.7e). Spectral pairing, along with photogeologic and topographic evidence of dark-toned sediment mass-wasting from one dune field to another, indicates some floor and falling dune sands are formed by similar materials (Fig. 5.2a; Table 5.3).

Northeast Melas Chasma also hosts several large dune fields (50–177 km²) adjacent to dark-toned lower walls (Fig. 5.8). Barchanoid, transverse, and dome dunes are located along a ~80-km-stretch of the northern wall. In general, slip faces are oriented downslope and away (south to southeast) from the eroding lower walls (e.g., Figs. 5.9a, 5.10a). Morphologic evidence of mass-wasting of lower wall elements is evident, with relatively steep (~30°) fans of detrital material contacting canyon floors and dunes (Figs. 5.8b–5.8d, 5.10b). The thermophysical properties of these surfaces show a wide range of values (Fig. 5.9d), but thermal inertias generally decrease downslope and southeastward (away from the spurs). Analysis of CRISM multispectral observations for several dune and outcrop-talus locations reveals nearly identical spectra, most of which have negative slopes (Fig. 5.9a, inset). Unratioed reflectance spectra of lower walls and dunes possess overall negative slopes from 0.7 to 2.6 μm , concave-up slopes between 0.7 and 1.4 μm , and minima near 1.1 μm (Chapter IV); when ratioed with dust spectra, these spectra maintain a negative slope, concavity is preserved, and often show subtle absorptions longward of 2 μm (Fig. 5.9a, inset), similar to spectra of coatings on Hawaiian glass-rich basalts [Minitti *et al.*, 2007]. Horgan and Bell [2012a] proposed that these spectral attributes are the result of weathered Fe-bearing glass and/or silica-enriched leached rinds on glass as has been observed in spectra of high-northern latitude erg locations. Additional examples of this weathered-glass dune composition are detected in Coprates and Hebes chasmata (Figs. 5.2, 5.4g–5.4h, 5.7c–5.7d; Table 5.3; Chapter IV) as well as other mid- to low-latitude locations [Horgan *et al.*, 2012b, 2012c]. Furthermore, these spectra are distinct from examples of unweathered Fe-bearing basaltic glass sand spectra [Salvatore *et al.*, 2010; Tirsch *et al.*, 2012].

HiRISE images of the Melas Chasma PSSs and dunes reveal discrete talus fans of mid- and dark-toned sediment derived from spurs and cliffs, continuing downslope, where they contact dune surfaces (Fig. 5.10). Boulders tracks, boulders, and larger lobes of talus appear superposed on some dune surfaces and show interleaved mass-wasting and aeolian processes (Figs. 5.10c–5.10d). Variations in HiRISE color relate to differences in composition; calibrated I/F band values for both surfaces (mid- and dark-toned) do show notable discrepancies when ratioed with one another (Fig. 5.10f). Mid-toned tan surfaces (spurs, cliffs, lighter-toned fans, TARs) have color ratios consistent with ferric minerals (i.e., high IR/RED and IR/BG ratios), whereas darker purple surfaces (darker-toned fans, boulders, dunes) have color ratios consistent with ferrous iron (i.e., high BG/RED and low IR/BG ratios; Fig. 5.10f). Color ratios of the light- and dark-toned talus surfaces match TARs and dunes, respectively. Additional morphologic and thermophysical evidence for sediment derivation from wall units in other chasmata (Juventae, Ius, Hebes, Ganges) was found (Fig. 5.2a; Table 5.3), but further analysis was limited due to a lack of high-resolution spectral data.

4.1.2 Interior Layered Deposit PSSs

Juventae Chasma hosts an expansive erg ($\sim 4600 \text{ km}^2$; Figs. 5.1b, 5.2a) that dominates the southern half of the chasm, whereas smaller floor dunes are located to the north near sulfate-bearing ILDs and chaotic material [Chapman *et al.*, 2003; Catling *et al.*, 2006]. Previous CRISM studies of Juventae Chasma (Fig. 5.2b) have identified several sedimentary and igneous lithologies (e.g., ILDs, wall, floor rock) of variable compositions (e.g., LCP, HCP, olivine, monohydrated and polyhydrated sulfates) [Bishop *et al.*, 2009; Parente *et al.*, 2011]. One prominent ILD, informally named mound C [Catling *et al.*, 2006], has a small ($\sim 110 \text{ km}^2$) topographically-isolated dune field and a mantling unit atop its summit $\sim 3 \text{ km}$ above the chasm

floor (Figs. 5.2a, 5.11). The dark-toned (IRA albedo ~ 0.13), spatially heterogeneous mantling unit is a regional deposit that commonly occurs atop VM wall units [Malin and Edgett, 2001], but it also is found superposed atop ILDs [Catling *et al.*, 2006; Chojnacki and Hynek, 2008]. Catling *et al.* [2006] suggested this mantling unit (Fig. 5.11a) is a local sediment source for dunes. We tested this hypothesis using multiple data sets. THEMIS data provide moderate thermal inertia values for both surfaces (~ 320 tiu for dunes and ~ 440 tiu for mantling unit), with the dunes having values consistent with very coarse sand (Fig. 5.12b). Dark-toned detrital materials emerging from the flanks of mantling units show thermal inertia values intermediate to those previously listed. Ratioed spectra of summit dunes and the mantling unit display broad absorption features from 1.1 to 1.5 μm that are due to olivine (Fig. 5.11b). Farther south, unratioed spectra of both landforms also match except for a slight deviation from one another at wavelengths >2.1 μm . Dunes here and others on or around mound C also possess weak hydration features longward of ~ 1.9 and ~ 2.1 μm in CRISM data and are attributed to hydrated sulfates (Fig. 5.2b, 5.11b; see below), a common component of adjacent variable-toned ILD layers. Spectra of the dark-toned talus off the east flank of mound C (Fig. 5.11) show absorption features that are similar to those in the mantling unit spectrum. East of these occurrences, dark streaks and dune slip faces are oriented to the east and southeast and away from mound C (Fig. 5.11a; Chapter IV, Fig. 4.11c). CRISM analyses of other dune surfaces among chaotic terrain ~ 50 km east of mound C also shows similar prominent olivine spectral features (e.g., Chapter IV, Fig. 4.19a). Thus, both spectral and geomorphic evidence supports the hypothesis that dunes at the summit of mound C and lower-lying areas to the east are partially derived from the mantling unit.

Westward, a small valley is bounded by mound C to the east and wall termini to the west (Figs. 5.2a, 5.12). As the valley opens to the south, barchanoid dunes grade into sand sheets of the southern erg. Dunes possessing slip faces tend to face south. Mass-wasting fans composed of mid-toned material are observed on the western ILD flanks (Fig. 5.12a; IRA albedo ~ 0.24) and contrasts with the darker-toned (IRA albedo ~ 0.14) detrital material forming on the eastern flanks (e.g., Fig. 5.11b). Closer inspection of the lowest portions of the mound reveals sub-meter materials forming these fans that meet the dune field perimeter (Figs. 5.11, 5.12). Thermal inertias and IRA albedos decrease westward and downslope from the western edge of mound C and into the erg (Fig. 5.12b, Profile A–A'). In these locations, layered deposits and talus fans possess relatively high thermal inertia values (>500 tiu), whereas barchanoid dunes adjacent to talus have values consistent with coarse sand (~ 350 tiu). The degree of induration, rather than particle size differences, may be an important factor for the decreasing thermal inertia trend. Sand sheets further south possess even lower values that correspond with fine sand (~ 270 tiu; Chapter IV, section 4.2.3). Barchanoid dunes and erg sand sheets to the south of mound C do not display olivine absorption features in CRISM multispectral observations. Instead, spectra of dunes, southern sand sheets, ILD, and the ILD-derived talus fans show features consistent with HCP (Figs. 5.2b, 5.12a). In addition, several lower wall locations show morphologic evidence for mass-wasting near dunes and may serve as supplementary sources for erg sand (Fig. 5.2a; Table 5.3).

Capri Chasma possesses a modest population of small dune fields ($\sim 4\text{--}70\text{ km}^2$, 220 km^2 total), primarily surrounding ILD and chaotic materials far from walls (Fig. 5.2a). These barchanoid, transverse, and barchan dunes occur tightly adjacent to the layered deposit Capri Mensa, usually within ~ 3 km or a mean dune field width. Several small barchan dunes are

observed with slip faces oriented toward the main dune field (away from then mensa; Fig. 5.13a, inset). Capri Mensa is known to have some of the more prominent spectral signatures associated with water-altered minerals in VM. For example, the area covered in Fig. 5.13a has been identified as having monohydrated sulfates, polyhydrated sulfates, red crystalline hematite, and opaline silicates [Bibring and Langevin, 2008; Flahaut *et al.*, 2010; Roach *et al.*, 2010a]. These previous detections have always been associated with bedrock components or associated talus rather than with aeolian materials. We examined local dunes for evidence of alteration products using CRISM and HiRISE data. Local dune spectra do not show broad spectral features associated with mafic minerals. Instead, the spectra have sharp absorption features at 1.94, 2.13, and 2.4 μm (Fig. 5.13c), similar to spectra of the monohydrated sulfate kieserite ($\text{MgSO}_4 \cdot \text{H}_2\text{O}$). This spectral signature is distinct from both polyhydrated sulfates, due to a lack of a 1.45- μm feature, and from szomolnokite ($\text{FeSO}_4 \cdot \text{H}_2\text{O}$), because of the absence of bound Fe^{2+} in CRISM VNIR spectra (not shown) [Cloutis *et al.*, 2006; Bishop *et al.*, 2009]. To test whether the sulfate signature is truly associated with dune sand or underlying deposits, values of CRISM IRA albedo and band depth at 2.1 μm (BD2100) were extracted along a transect (Fig. 5.13d). The deepest hydration features are found in spectra from the low-albedo dune crests. Based on these diagnostic features, we identified dunes to be partially composed of kieserite. Examination of CRISM observations a few kilometers to the east also show that ILD cliffs and fans all reveal absorptions at 2.1 μm , but with features at ~ 1.92 and ~ 2.39 μm (Fig. 5.13c). These spectra are more complex and may have spectral contributions from both monohydrated and polyhydrated magnesium sulfates (e.g., hexahydrite). Note that the discrepancies between the dune and ILD spectral slopes are due to using different denominators (floor vs. mantling unit, respectively) and do not affect band positions. Other dune fields show similar hydration features as adjacent Capri

Mensa fans (Fig. 5.2b) in CRISM hyperspectral and multispectral data (Figs. 5.2, 5.4d; Table 5.3), confirming that this relationship is not an isolated occurrence. In contrast to this sand composition, the spectra of dunes to the north and east (farthest from the Capri Mensa) have broader mafic features and no evidence of hydration (Fig. 5.2b).

HiRISE color ratio images of other local dunes show a notable dichotomy with dune crests appearing dark-blue to purple and lighter-toned surfaces occurring along smooth dune flanks and lower-order bedforms (Fig. 5.13a, inset). We hypothesize this difference is due to physical fractionation between different sand compositions and tested this by measuring the calibrated HiRISE I/F values. HiRISE color band ratios of the lighter-toned “yellow” sands are consistent with ferric minerals (i.e., high IR/RED and IR/BG ratios), whereas the darker “purple” sands possess values consistent with ferrous minerals (Fig. 5.13a). Kieserite does not contain bound iron, so its presence or absence does not affect HiRISE color values.

Ganges Chasma has the highest concentration of dunes ($\sim 12,000 \text{ km}^2$) in VM, including the largest ($\sim 8000 \text{ km}^2$) non-polar erg on Mars (e.g., Figs. 5.1b, 5.2a) [Chojnacki and Moersch, 2009; Chojnacki *et al.*, 2012b]. The majority of the Ganges dune fields are located surrounding the sulfate-bearing Ganges Mensa and other layered deposits (Chapter IV, Figs. 4.15d–4.15f). Eastward, light-toned ILD yardangs formed from easterly winds partially composed of hydrated sulfates and mafic minerals possess fans of detrital materials emulating on their flanks adjacent to aeolian bedforms (Chapter IV, Fig. 4.19b). In most cases, an unambiguous spectral relationship between dunes and ILDs was not evident; analysis of CRISM spectra showed that dune sand is dominantly mafic with evidence of pyroxene, olivine, unaltered glasses, and various mixtures (Chapter IV, Fig. 4.19b). A few exceptions showed weak but discernable hydration

features in spectra of Ganges Mensa-dune field contacts, where mass-wasting of mid-toned ILD material was apparent (Fig. 5.2; Table 5.3; Chapter IV, Fig. 4.15f).

4.1.3 Landslide PSSs

Several VM dune fields are superposed on Late Amazonian (1 Ga–200 Myr) landslides located in Ius, Melas, Coprates, and Hebes chasmata (Figs. 5.1b, 5.8a, 5.9d; Chapter IV, Fig. 4.12). For instance, a small dune field atop the massive Ius Labes landslide (Figs. 5.2a, 5.14) is dated at ~200 Myr years old [Quantin *et al.*, 2004]. Duneform slip faces in this field are broadly oriented toward the east-southeast and parallel to the steep (~30–40°) eroding landslide scarp (Fig. 5.14a–5.14c). Small duneforms are present near the brink of a steep scarp, several 100s of meters above the main field (Fig. 5.14d, inset). Multispectral data of dark-toned landslide scarp materials and dunes show absorption features at 1.1 and 2.2 μm , indicating HCP-bearing basalt (Figs. 5.4e, 5.14a, inset). To the east, CRISM VNIR images of the scarp and aeolian bedform surfaces display variability in iron mineralogy. The dunes, adjacent eroding outcrop, and dark streak all produce near identical continuum-removed spectra (Fig. 5.14d, inset).

Most instances of landslide dune field provenancial relationships are relatively ambiguous. Landslide surfaces are often partially mantled in dust [Chojnacki *et al.*, 2006; Ferguson *et al.*, 2006a], making spectral interpretations difficult. Also, the fact that landslides are primarily composed of wall materials makes distinctions between landslide-derived and wall-derived sediments unclear. For example, the Coprates Chasma landslide dune field shown in Fig. 5.15 appears below a landslide scarp, with falling dunes on the terrace above (also see Fig. 5.7e). This landslide, dated at ~1 Gy years old [Quantin *et al.*, 2004], extends north where it terminates at the base of the northern wall. CRISM analysis (Fig. 5.15d, blue and green spectra) of landslide surfaces shows broad absorption features associated with olivine and pyroxene, but

also sharper features near 1.96 and 2.3 μm indicating Al and Fe-Mg phyllosilicates, all common wall constituents [Murchie *et al.*, 2009b; Flahaut *et al.*, 2012]. Dune sand spectra appear with broad features near 1 μm are comparable to other observations of olivine-bearing Coprates dunes (e.g., Figs. 5.3a, 5.6b) and adjacent landslide surfaces, but notable differences exist. In the VNIR spectra of these dunes broad features occur at $\sim 0.8 \mu\text{m}$, but are without a drop-off in reflectance at shorter wavelengths, which indicates the presence of ferric iron and alteration [Cloutis and Bell, 2004]. At longer wavelengths, an additional asymmetric absorption occurs in the 2.2–2.3- μm region (Fig. 5.15d, inset). This band is most readily attributed to an Mg-O-H bend [Clark *et al.*, 1990] and likely indicates alteration, likely of olivine (section 5.3) [Orofino *et al.*, 2006].

Other landslides with superposed dune fields in Ius, Hebes, Ganges, and Coprates chasmata show photogeologic evidence as local sand sources for dunes (Fig. 5.1b; Table 5.3). However, due to the lack of CRISM data coverage in these locations and the reasons noted above, the associations are ambiguous. In addition, the spatial relationships and morphologic evidence (e.g., slip face orientation) imply certain dune fields have migrated onto landslides and may have no provenancial relationship with the slides on which they are now superposed (Chapter IV).

4.1.4 Regional Intra-Rift Dune Fields

Large collections of contiguous dunes may provide the sediment for spatially separated, smaller, more mobile dunes [Bagnold, 1941; Pye and Tsoar, 1990]. Climbing dunes in Melas and Hebes chasmata are $\sim 10\text{--}20 \text{ km}$ upslope from large floor dunes (Fig. 1b). Melas Chasma climbing dunes are superposed on spurs and gullies 10 km to the south and $\sim 1000 \text{ m}$ higher in elevation than the PSS floor dune field (Chapter IV, Fig. 4.11a) [Chojnacki *et al.*, 2010]. An analogous situation was found in southern Hebes Chasma, where a train of climbing barchans

appears within a gully 20 km to the south, ~2000 m higher in elevation than dunes and sand sheets at the canyon bottom (Chapter IV, Fig. 4.13a). In both cases, floor and climbing dune slip faces are oriented southward. Spectra of climbing dunes are a close match to spectra of floor dunes and northern walls with characteristics attributable to weathered Fe-glasses (Figs. 5.2, 5.4g–5.4h; Table 5.3).

4.2 Possible Extra-Rift Extended Sand Pathways

4.2.1 Context of Low-Albedo Materials and Possible Pathways

Our survey of potential sand transport pathways into VM was conducted within the area displayed in Fig. 5.1a, which encompasses all terrain within ~1000 km of the VM dune fields. To highlight possible low-albedo pathways, a TES albedo view of the region is provided with threshold values of 0.12 and below masked in white (Fig. 5.16). Albedos at or below this value are primarily associated with craters to the north of eastern VM (e.g., Mutch crater), eastern chaotic regions (e.g., Aurorae Chaos), plains surrounding Holden crater, the rugged terrain of Thaumasia Planum, plains of southern Solis Lacus, and pits within Noctis Labyrinthus. No albedo values below the threshold are detected in the Tharsis (northwest) and Phoenicis Lacus (west) regions (Figs. 5.1a, 5.16). Our survey of these regions did not produce evidence for extensive dune-forming deposits, dark streaks, or sand sheets in these areas, nor have previous studies reported them here [Hayward *et al.*, 2007a; Tirsch, 2009]. Plateau regions north of Ganges Chasma and Aurorae Chaos show albedo patterns consistent with dark streaks extending southward of craters toward the rift rim. The plains to the south of Coprates and Capri chasmata show more extensive patterns of low-albedo surfaces. Albedos of the plateau terrain directly adjacent of the rift rim are generally high (~0.2) except for Ganges, Capri, and eastern Coprates chasmata, particularly to the south.

Interconnected dunes, dark streaks, and sand sheets within the region may form sand transport pathways (Fig. 5.1a). The largest collection of particulate extra-rift material includes twenty-two extra-rift dune fields on the plateau surrounding VM (Figs. 5.17a–5.17b). Most of these occurrences are small (2–40 km²), with only three dune fields larger than 100 km² in area (~1600 km² total; Table 5.2). Dune fields are dominantly (21 of 22) found in small craters (20–80 km in diameter) and possess primarily barchanoid, barchan, and transverse morphologies (Figs. 5.18a–5.18c). No evidence for dunes emerging out of these craters (e.g., climbing dunes) was found, although many craters to the northeast of VM do show prominent southward-oriented dark streaks (Figs. 5.16, 5.17a). Numerous other small, low-albedo deposits (e.g., sand sheets, sand patches) are found within shallow troughs, craters and other depressions, in addition to lower-order bedforms (e.g., ripples, transverse aeolian ridges (TARs)), but contiguous collections of dunes are generally absent. Likewise, regional dark-toned crater central peaks and pits [Tirsch, 2009; Quantin *et al.*, 2012] are often without prominent aeolian deposits.

Potential major sand corridors are found with the canyons that connect VM with the extreme northern (Maja Valles), eastern (Aurorae Chaos), and western (Noctis Labyrinthus) edges of the rift (Figs. 5.1, 5.17a, 5.17c). Maja Valles outflow channel does contain several small (<1 km²) isolated duneforms and dark streaks scattered amongst chaotic blocky materials [Chapman *et al.*, 2003], but no major dune fields. Noctis Labyrinthus and Aurorae Chaos host nine wide-spread dune fields (Figs. 5.17c–5.17d, 5.18d–5.18f). Except for two larger fields (~130 km²) in western Aurorae Chaos, most of these are smaller dune fields (3–26 km², ~420 km² total; Table 5.2). Both dune populations are located at lower elevations, removed from canyon walls. Additional smaller corridors connected with Ganges Chasma occur with Allegheny and Elaver valles, that flowed eastward, and a similar unnamed valley that flowed

northward to Orson Welles crater, hereafter referred to as the “Orson Welles Valles” (Figs. 5.1b, 5.17a). Allegheny and Elaver valles are both heavily degraded, without many low-albedo deposits (dunes or otherwise), and often have dust devil tracks, indicating dusty surfaces. Orson Welles Valles does have discrete patches of dark-toned surfaces (Fig. 5.16), but these are typically associated with trough wall, not significant aeolian materials.

4.2.2 Paleowind Directional Analysis

Dunes to the northeast of the rift in Lunae Palus and north Coprates possess slip faces facing to the south-southwest and toward VM (Figs. 5.17a, 5.18a–5.18b). Dune fields located in small craters ($< \sim 30$ km) are positioned in the southwest to southeast quadrants of their host craters, and the DCA position is consistent with variable winds from the north. In addition, all major dark streaks are oriented toward the south-southwest direction indicating they formed from north-northeasterly winds. Larger linear low albedo patterns south of Mutch, Orson Welles, and other large craters are also oriented in north to south (Fig. 5.16).

Of the seven dune fields southeast of VM, only three were used for paleowind analyses, as the remaining occurrences show dunes severely degraded and without prominent slip faces (e.g., Figs. 5.18c), possibly due to lack of activity. The slip faces that are present in these areas, like those of the dunes to the north of VM, dominantly face southeast to southwest (away from VM; Figs. 5.17b). For instance, the majority of Holden crater dune slip faces appear south facing [Hayward *et al.*, 2007b]. Dune fields are either located in the crater centers or in the southern halves, the latter consistent with a northerly wind regime based on DCA orientations. No prominent dark streaks are located in these regions.

The Noctis Labyrinthus and Aurorae Chaos dune slip face orientations are more variable than was the case with plateau dunes, likely due to local topographic effects. The two

easternmost Noctis Labyrinthus dune fields with a sufficient number of unidirectional dune types (Figs. 5.17c, 5.18d) exhibit mean slip face directions nearly orthogonal to one another. Dunes in the eastern portions of Aurorae Chaos (Figs. 5.17d, 5.18e) possess slip faces that are dominantly south-southeast-facing (away from VM), broadly similar to plateau dunes to the north and south of VM (Figs. 5.17a–5.17b). In contrast, the two largest Aurorae Chaos dune fields, located in the western portions of the chaos (Figs. 5.17d, 5.18f), show mean slip face directions oriented to the west-southwest, the same orientation as adjacent dark streaks. These dunes are ~300 km east of eastern Ganges Chasma dune fields, with an unobstructed potential pathway between them.

4.2.3 Composition

In CRISM spectra of dune fields to the north of the rift, HCP was commonly identified, either alone or with olivine (Fig. 5.17a; Table 5.1). In some case, CRISM data of dune surfaces possessed no suitable spectra, possibly due to surface dust contamination. Dune spectra to the south of VM are consistent with spectra of olivine and pyroxene (Fig. 5.17b). These sand mineralogies are similar to compositions reported for other tropical latitude dune fields, with the notable absence of LCP (Chapter IV). No clear compositional spatial pattern is evident, except for the common occurrences of dunes with HCP to the north of eastern VM.

Most spectra of the Noctis Labyrinthus dune fields are ambiguous, likely due to dust mantlings. One exception is a small dune field in a collapse pit (Figs. 5.17c, 5.18d), for which spectra include absorptions attributable to HCP and olivine. Spectra of the Aurorae Chaos dune fields (Fig. 5.17d) are consistent with basaltic sand. Additional compositional studies of regional extra-rift landforms (e.g., crater peaks, layered deposits) often note low-albedo bedrock and particulate surfaces of basaltic compositions [*Mustard et al.*, 2005; *Tirsch et al.*, 2011; *Quantin et al.*, 2012], similar to our analysis.

5.0 Discussion: Synthesis of Results

5.1 Local or Regional source(s) with Short Sand Pathways

Multiple intra-rift local lithologies and other regional landforms are all candidates as sand sources for VM dune fields. In support to the evidence presented in section 4.1, we provide additional arguments for intra-rift sourcing.

Wall Sources

Dunes in Valles Marineris are often proximal to eroding wall materials with similar low albedos. Previous studies of VM wall stratigraphy was shown them to have ubiquitous horizontal layers, which are interpreted as evidence of Noachian-aged voluminous flood volcanism [McEwen *et al.*, 1999; Beyer and McEwen, 2005], volcanoclastic ash deposits [Hynek *et al.*, 2003; Bandfield *et al.*, 2013], and/or intrusive magmatism [Williams *et al.*, 2003], all related to the Tharsis formation. Compositional mapping has confirmed a mafic petrogenesis for VM wall rock, with olivine and pyroxene as major bedrock components [Murchie *et al.*, 2009a; Clenet *et al.*, 2011; Flahaut *et al.*, 2011]. Spurs-and-gullies possess some of the greatest thermophysical heterogeneity of all martian surfaces (at the 100-m scale); resistant spurs often have thermal inertia values consistent with bedrock (>1200 tiu), whereas gullies typically have depressed values (<300 tiu) suggestive of dust, talus, or aeolian material [Edwards *et al.*, 2009; Chojnacki *et al.*, 2010]. Spurs often form at the upper portions of wall sequences directly below the plateau and extend downslope, terminating either near gullies or abruptly into scarps or shelves several kilometers above the floors. Gullies rarely extend up to the plateau edge, instead they terminate below the top few kilometers of steep, resistant spurs and layers [Beyer and McEwen, 2005]. The largest gully in a given sequence may either terminate into scarps or

extend downward to exit onto the canyon floor. In any case, examples of steep ($>30^\circ$), near-continuous potential sand pathways from the middle wall down to the canyon floors are present.

The best examples of local wall outcrops providing sediment for VM dunes fields are found in Coprates Chama (Fig. 5.2a; Table 5.3; section 4.1.1). The falling dune fields atop the freestanding east Coprates massif are topographically isolated from potential extra-rift sand sources (Figs. 5.2a, 5.7). Elevation and slope constraints place falling dune source layers in the upper ~ 2 km of the wall, above which no major bedforms occur. In most cases, HRSC-derived slopes above and below wall dunes are at or above the angle of repose for sand-sizes ($\sim 32^\circ$), indicating dune sediment transport will be unidirectionally downslope. No climbing dunes are observed in this portion of the rift and slopes are too steep ($>35^\circ$) for transport above the bottom third of the wall [Chojnacki *et al.*, 2010]. The transport directions of dunes, as inferred by slip faces orientations and mesoscale atmospheric modeling results [Rafkin and Michaels, 2003; Toigo and Richardson, 2003], are often in-line with local gradients and leading away from the proposed wall outcrop sources (Figs. 5.3a, 5.5, 5.6a, 5.7). In locations such as these, where winds align with the downhill gradient, saltation path lengths will elongate, resulting in enhanced sediment flux [Pye and Tsoar, 1990]. Falling dunes have uniformly higher thermal inertias than floor dunes, implying coarser sand sizes and less mature sediment. The decreasing thermophysical properties (thermal inertia and CRISM albedo) in the (inferred) downwind direction from lower walls (Figs. 5.3a, 5.6a, 5.9d) [Chojnacki *et al.*, 2012b] are interpreted as evidence for the transport of finer, darker-toned sand particles downwind of the source and/or removal of light-toned secondary minerals. Because IRA albedo decreases with thermal inertia along these proposed sediment pathways suggests the trend is not solely due to variable dust coverage. Impact comminution would require much greater distances (>500 km) to detect

[Greeley and Kraft, 2001; Rogers and Christensen, 2003]. Spectra of local falling dunes and floor dunes are similar to spectra of adjacent upslope upper and lower wall outcrops, respectively (Figs. 5.2b, 5.3a, 5.4a–5.4c, 5.6b, 5.7). In addition, certain falling and floor dunes are spectral matches (Figs. 5.2a, 5.4c, 5.7d). These spectral matches are significant because VM wall rock is spectrally heterogeneous; in this and previous work, analyses of many CRISM wall observations identified nearly a dozen spectral endmembers which are distinct from dune spectra [Bishop *et al.*, 2009; Poulet *et al.*, 2009; Roach *et al.*, 2010b; Flahaut *et al.*, 2012]. These spectral pairings, along with photogeologic and thermophysical evidence, imply relatively discrete, local (<10 km) sand pathways from wall gullies down to chasm floors.

Based on mapping in Coprates Chasma and results described above, the following sediment transport system is proposed (Fig. 5.19), in which: (1) upper wall outcrops dominantly produce basaltic-bearing sediment for falling dunes located in gullies; (2) falling dunes found on lower walls provide sand for adjacent floor dunes; (3) lower wall outcrops provide additional sediment for floor dunes. In many cases, upper wall units were identified as the principal (possibly only?) source for adjacent falling dunes (*e.g.*, Figs. 5.3, 5.7b–5.7c, 5.8–5.10), whereas lower wall units as the dominant source for adjacent floor dunes (*e.g.*, Figs. 5.3, 5.5–5.6, 5.7d–5.7e, 5.8–5.10). Floor dunes may have benefited from ancient extra-rift sources, but no clear evidence for this scenario is found in the current epoch (section 5.2).

Similar to Coprates, lower wall sources in Melas Chasma likely provide sand for floor dune fields, but are without intermediate falling dunes (Figs. 5.2a, 5.8–5.9; Table 5.3). These northeastern walls outcrops may not just be a local source for nearby dunes, but also be the ultimate source for the majority of Melas Chasma dune fields (Fig. 5.2). The floor dune fields in south of the chasma are ~100–200 km south of the walls units. Although, topographic variations

likely effect migrations patterns, Melas dune fields have means slip face directions broadly facing south (Figs. 5.8–5.10; Chapter IV, Fig. 4.11a). Like dunes and lower walls in the north, these dunes were determined to be partially composed of weathered glasses (Fig. 5.2b, 5.4h, 5.9a). As weathered glasses have not been reported elsewhere in the canyon (other than dunes and outcrops reported herein), these northern walls might have served many Melas Chasma aeolian bedforms as a distant intra-rift source (Fig. 5.2b). All chasmata show photogeologic evidence of low-albedo sediment derivation from wall units, but will require more detailed study to be properly characterized.

Layered Deposit Sources

The VM region is of particular interest to planetary scientists due to the high density of Late Noachian- to Early Hesperian-aged layered deposits of an unclear petrogenesis. Proposed origins of ILDs include aeolian, lacustrine, mass-wasting, glacial, and volcanic systems, or various combinations of these processes [Lucchitta *et al.*, 1992; Malin and Edgett, 2000; Chapman and Tanaka, 2001; Hynek *et al.*, 2003; Chojnacki and Hynek, 2008; Rossi *et al.*, 2008]. Images of ILDs tend to show friable, thinly-layered deposits with various tones and moderate thermal inertia surfaces (300–600 tiu), which is consistent with a sedimentary origin [Ferguson *et al.*, 2006a; Chojnacki and Hynek, 2008]. In some chasmata, ILDs cover a major proportion of the floor area and can reach elevations of ~7 km, rivaling adjacent canyon walls [Lucchitta, 1999]. Correlated with these light-toned ILDs are spectrally-detected water-altered minerals, such as crystalline gray hematite and various hydrated sulfates [Christensen *et al.*, 2001a; Bishop *et al.*, 2009; Murchie *et al.*, 2009b]. In addition, some ILD layers are also partially comprised of mafic mineral groups [Mangold *et al.*, 2008; Murchie *et al.*, 2009a]. Evidence for mass-wasting of ILD flanks is prevalent, with fans of dark- to mid-toned materials of variable compositions

(e.g., pyroxenes, Fe-oxides, sulfates) emerging from higher-lying strata [*Chojnacki and Hynek*, 2008; *Mangold et al.*, 2008]. This ILD-derived detrital material may source adjacent aeolian bedforms (section 4.1.2).

Olivine-bearing dunes at the summit of Juventae Chasma's mound C have similar compositions and thermophysical properties as a regional mantling unit (Figs. 5.11–5.12), earlier hypothesized to be a local bedform sediment source [*Catling et al.*, 2006]. Additional dark-toned material is observed being shed downslope to the east, where other olivine-bearing ergs are located. In contrast to this trend, western dunes, sand sheets to the south, and ILD-sourced fans have a nearly identical HCP-bearing composition, consistent with local sourcing (Fig. 5.12). CRISM observations of sand sheets and barchans on the extreme eastern edge of the erg produce spectra of LCP-bearing sand (Figs. 5.2b, 5.4f; Chapter IV). The existence of two dune populations in close proximity to one another (within ~10 km) and a third on the opposite side of the chasma, each with different spectral signatures (i.e., olivine, HCP, sulfate, and LCP), argues for their having different, relatively unmixed sediment sources, and a low degree of sediment homogenization (Fig. 5.2b).

Mapping in Juventae Chasma provides evidence of a very different sediment transport system (Fig. 5.20; Table 5.3) than was proposed for Coprates, namely: (1) low-albedo ILD mantling units provide olivine sand for summit and erg dunes to the southeast; light-toned ILD materials produce (2) HCP- and (3) sulfate-bearing sediment for dunes and sand sheets to the west and south; (4) wall outcrops provide LCP-bearing sand for eastern erg dunes. Collectively, these occurrences represent a regional source for Juventae erg sand. Other wall and chaos material sources are likely linked to dune sediment production, but are less constrained (Fig. 5.2a; Table 5.3).

Multiple observations of Capri Chasma dune fields imply dominantly local sediment derivation, in part, from adjacent sulfate-bearing Capri Mensa to the south (Figs. 5.2, 5.4d, 5.13; Table 5.3). Spectral evidence for kieserite, a previously unreported and likely uncommon composition for martian or terrestrial aeolian sediment, is most apparent in the dunes inferred to be directly downwind (west) from the ILDs. Sulfates are relatively soft minerals; for instance, kieserite has a hardness of 3.5 (cf. olivine at 6.5–7) [Nesse, 2012]. Therefore, sulfate sand is unlikely to survive repeated saltation and reptation over great distances and sand dunes bearing these minerals are limited in range from their sediment source. Gypsum (hardness of 2 [Nesse, 2012]) dunes of White Sands, New Mexico are locally sourced from the Lake Otero playa and their particle size distribution is observed to decrease rapidly downwind of the source [Muhs, 2004; Ghrefat *et al.*, 2007; Kocurek *et al.*, 2007]. The nearest known location of kieserite-bearing ILDs outside of VM is ~1800 km east in Aram Chaos [Lichtenberg *et al.*, 2010] and is too distant to be a plausible. This robust evidence of rare kieserite-bearing dunes that are spatially coincident with monohydrated sulfate-bearing ILDs supports a genetic relationship. In addition, the discovery of monohydrated sulfate-bearing dunes broadens the formational environments in which dunes may be formed. In contrast, spectra of dunes east of Capri Mensa (Fig. 5.2a) have features consistent with basaltic minerals. Additional evidence from HiRISE color images supports the fractionation of dune sediment compositions, and may indicate other unknown sources are providing sediment for dunes (Fig. 5.13a; section 5.4). Finally, we note that chasmata with the highest densities of layered deposits (e.g., Candor, Ophir) and of similar compositions do not have dune fields, suggesting that not all ILDs are inherently linked to sediment production for high-order bedforms.

Landslide Sources

The deformation and subsidence of the chasmata interiors have resulted from gravity-driven mass-wasting processes, ranging from Late Amazonian chasma-filling landslides to smaller localized talus fans; these occurrences represent the most recent major rift-forming events [Scott and Tanaka, 1986]. Emplaced landslide heads typically start several kilometers below the plateau, and slides extend downslope onto the canyon floor and butt up against floor material, other landslides, or the adjacent walls [Lucchitta *et al.*, 1992]. Thermophysical measurements and visible inspection of landslides' hummocky surfaces indicate variable aeolian wind-stripping and dust deposition occurs in concentrated areas, likely dictated by topography [Chojnacki *et al.*, 2006; Mellon *et al.*, 2008].

Valles Marineris dune fields that are superposed on Late Amazonian (<1 Ga) landslides may also be sourced locally (section 4.1.3). One Ius Chasma (Fig. 5.2a) landslide, dated at 150 My [Quantin *et al.*, 2004], shows photogeologic evidence of aeolian sediment production with dark material emerging from landslide scarps adjacent to downslope duneforms (Fig. 5.14). Dune slip face orientations are consistent with dune transport broadly away from the scarp. CRISM data of both the scarp and dunes show absorption features indicative of HCP-bearing basalt (Figs. 5.4e, 5.14a). Collectively, these observations are consistent with local derivation of dune sediment from landslide materials. Given that this dune field is constrained to have formed in the last 150 My (Chapter IV), sediment production is likely accelerated by the high relief, slopes, and exposure of local strata. Dune sediment derived from impact and aeolian erosion alone appears unlikely, based on some simple calculations. For the average Amazonian erosion rate derived from Mars Exploration Rover data (1 nm/y) [Golombek *et al.*, 2006], only 15 cm of sediment would be produced over the 150-Myr-age of the landslide, whereas the dunes adjacent

to the slides (Fig. 5.14a) are 30-40 m tall, based on slip face measurements and assumption of slip face angle (32°) [Bourke *et al.*, 2006].

Another landslide dune field in Coprates Chasma (Fig. 5.15) has a more ambiguous relationship with surrounding geology. These altered olivine-bearing dunes (section 5.3) are located atop a landslide partially composed of mafic and phyllosilicate minerals, compositions similar to adjacent wall spur and landslide surfaces. In these locations, pinpointing a specific source outcrop is nearly impossible, because of the nearly continuous distribution of dune sand from the southern wall terrace and northward across a landslide of composed similar minerals. Regardless, we propose these dunes were derived locally, either from walls or the underlying slide, based on compositional, morphological, and spatial evidence. Other examples in Ius, Hebes, and Coprates chasmata (Fig. 5.1b) have dunes in close association to landslide material of analogous basaltic composition and may imply a genetic relationship, but will require further study.

Other Intra-Rift Dune Field Sources

Climbing dune fields in Melas and Hebes chasmata may have been sourced from larger, floor dunes in the inferred upwind directions (section 4.1.4; Table 5.3; Chapter IV, Figs. 4.11a, 4.13a) [Chojnacki *et al.*, 2010]. On Earth, locations at the bottom of canyon walls are zones where orthogonal anabatic wind flow accelerates, promoting upslope sand transport [Pye and Tsoar, 1990; Raupach and Finnigan, 1997; Wiggs *et al.*, 2002]. For northerly canyon bottom winds, which encounter the southern walls of Hebes and Melas chasmata, modeled atmospheric flow tends to increase [Rafkin and Michaels, 2003; Toigo and Richardson, 2003]. Also, floor and climbing dune pairs are spectral matches (weathered glasses; Figs. 5.2b, 5.4g–5.4h). Photogeologic and compositional evidence at these two sites implies floor dune sand is at least a

partial source for smaller, climbing duneforms. Terrestrial dune mobility is inversely correlated with size (i.e., smaller dunes are more mobile), as is likely the case on Mars [*Chojnacki et al.*, 2011a; *Bridges et al.*, 2012a, 2012b]. These examples in VM agree with the notion for higher mobility of smaller climbing dunes. In addition, these occurrences represent intra-rift regional sources, distinct from the smaller, more discrete local sources described above.

5.2 Extended Sand Pathways From External Sources

In addition to evaluating intra-rift local and regional PSSs, we also searched for extended sand pathways that may connect external sand source(s) with VM (section 4.2).

Possible Plateau Sand Pathways

The 22 dune fields and smaller isolated aeolian deposits (e.g., sand sheets, sand patches) on the plateau surrounding VM (Figs. 5.1a, 5.17a–5.17b; Table 5.2) are small in area (1–70 km² each, ~1600 km² total) and do not currently constitute a significant reservoir that could supply VM dunes with sediment [*Chojnacki et al.*, 2012a]. The total area of these deposits is <8% of the dune field area mapped within VM (Chapter IV). The examples north of eastern VM are dominantly crater-related dune fields, typically separated by 100–200 kilometers, and with paleowind indicators (i.e., slip faces, dark streaks) suggestive of transport southward toward VM (Figs. 5.17a, 5.18a–5.18b). Southward oriented dark streaks extend several crater diameters and imply they were formed by a persistent wind regime [*Edgett*, 2002]. Dune slip face orientations in south Coprates and Margaritifer Sinus are also broadly facing south, implying transport of material away from the rift (Figs. 5.17b, 5.18c). Likewise, the majority of dune fields located in small craters (<~30 km) are positioned in the southern halves of their craters and, based on DCA position, are suggestive of winds from the north. The evidence of a regional north to northeasterly wind regime is consistent with the findings of previous studies comparing results

from global circulation models (GCMs) to morphologies of dune fields [*Hayward et al.*, 2007a; *Gardin et al.*, 2012] and dark streaks [*Edgett*, 1994]. Potential pathways northward into VM are particularly problematic atop the low albedo plateau regions nearest the southern Coprates and Capri rim, where crater chains, fossae, and catenae would impede dune migration (Figs. 5.1b, 5.16, 5.17b). CRISM analysis (Figs. 5.2b, 5.17) did not reveal a strong correlation between dune sand composition in and outside of VM. Although some plateau dunes are composed of basaltic minerals, similar to those found in some VM dunes (e.g., HCP, olivine), these are also the most commonly detected sand compositions on Mars (Chapter IV) [*Tirsch et al.*, 2011]. Unfortunately, no evidence for alteration products (e.g., sulfate, weathered glass) that might provide a more unique mineralogic marker was detected in spectra of extra-rift dunes.

Given a persistent northerly wind regime, sediment supply, and adequate transport pathways or corridors (e.g., Orson Welles Valles), Lunae Palus and north Coprates aeolian sand might have been transported southward into Ganges and Coprates chasmata (Figs. 5.1a, 5.17a). However, sand deposits in these locations show little evidence for escape out of their hosting craters, as some dune systems do [*Silvestro et al.*, 2010a; *Tirsch et al.*, 2011]. Although dark-toned sands sourced from dunes show evidence for escape out of steep ($\sim 50^\circ$) crater alcove walls, that sand typically forms dark streaks rather than dunes (e.g., Victoria crater [*Geissler et al.*, 2008]), as a result of limited sand supply. In a similar fashion, sand may escape out chasmata reentrants. For example, the plateau south of Hebes Chasma possesses dark streaks [*Chojnacki et al.*, 2006] directly south of climbing dunes located several kilometers below in chasma gullies (Chapter IV, Fig. 4.13a). Thus, the evidence for sand escape from chasmata indicates that such escape is volumetrically limited. In any rate, these apparently crater-confined dunes and sand sheets may be derived from local layers within crater walls, as demonstrated

elsewhere by *Tirsch et al.* [2011]. Dune fields to the south of VM are unlikely to be part of extended sand pathways into VM, based on their small areal size, paleowind indicators, and inferred level of activity. Although it is highly probable that dune-forming sediment has blown into the rift since its Noachian genesis, currently no dune fields or major sand sheets are located within ~200 km of the VM rim (Fig. 5.1). Thus, extended sand pathways into VM from the plateaus do not exist in the current epoch, nor are these aeolian deposits a likely sediment source. Similarly, analysis of dune fields across Noachis Terra by *Fenton* [2005] did not produce evidence for extended transport pathways and distant sources. As in that case, long sand pathways might have been prevalent in the past, but were erased or buried, or their supply dwindled.

Possible Adjoining Region Sand Pathways

Spatially isolated dune fields in the adjoining regions of Noctis Labyrinthus and Aurorae Chaos are relatively small in area; the total area of these nine dune fields is <2% of VM dune field area (Figs. 5.1a, 5.17c–5.17d, 5.18d–5.18e; Table 5.2). TES thermal inertia values of dunes in these regions are lower than for dunes in VM (i.e., smaller grain sizes) (Chapter IV), which make them inconsistent as a source, as sand sizes decrease with transport. Noctis Labyrinthus dune fields are without viable pathway to the western-most Ius dune fields because large landslides that cover major portions of canyon floors prohibit west-east transit (Figs 5.2a, 5.18d). The smaller dune fields in eastern Aurorae Chaos are separated from eastern VM by rugged chaotic materials and there is no clear pathway westward (Fig. 5.18e). In both cases, paleowind indicators are inconsistent with dune field transport toward central VM.

Perhaps the most viable sand pathway or corridor into VM is the eastern edge of Ganges Chasma, where the unobstructed canyon opens northeastward into Hydraotes and Aurorae chaos

(Figs. 5.1, 5.17d, 5.18f). The western-most Aurorae Chaos dunes are ~300 km east of those in eastern Ganges and possess paleowind indicators (i.e., slip faces and dark streaks) suggestive of westward transport (Figs. 5.17d, 5.18f). Light-toned ILD yardangs in this part of the rift also likely formed from persistent easterly winds (e.g., Chapter IV, Fig. 4.19b). These paleowind indicators are consistent with predicted equatorial easterly “trade” winds at this latitude (~8° S) [e.g., *Haberle*, 1986]. Spectra of these Aurorae Chaos dunes are consistent with HCP-bearing basalt and are analogous, although not unique, to many dune spectra in eastern VM (Chapter IV, Figs. 4.19c, 4.20g). Ganges Chasma ergs contain some of the largest accumulations of dune sand on Mars (Chapter IV) and likely indicate persistent sand transport and deposition into these areas, possibly from eastern chaos regions. Terrestrial ergs primarily form downwind of large regional sources, in addition to smaller local sources [*Bagnold*, 1941; *Wilson*, 1973]. Existing dunes, drifts, and dark streaks in western Aurorae Chaos may represent the remnants of a longer-term sand pathway that fed Ganges Chasma ergs and dunes in the ancient past. Additionally, local sources (described above) and possible ancient influxes of sand from the northern plateau might have contributed to these massive ergs. Other ancient sediment influxes related to Noachian- or Hesperian-aged regional fluvial or aqueous activity are plausible [*Lucchitta et al.*, 1992; *Chapman and Tanaka*, 2001; *Murchie et al.*, 2009b], but remain unknown. Although we find no compelling evidence for contemporary developed sand pathways leading into VM, we suggest ancient pathways into eastern VM are most likely to have occurred westward from the eastern chaos regions and southward from the northern plateau region.

5.3 Spectral Similarities and Differences Between Sand and Sources

Spectral signatures between sand and source material can yield useful information on possible chemical or physical changes that occurred during the weathering that initially produced

the sand from the source, during transport, or after becoming incorporated into a dune field. On Earth, the composition of clastic sediments rarely correlates with the source rock composition on a one-to-one basis because of rapid chemical weathering and physical fractionation [Johnsson, 1993; Modi, 2011; McGlynn, 2012]. Even minor amounts of chemical weathering under arid conditions and hydrodynamic sorting over short distances (<10 km) can alter the geochemistry of sediments enough to where provenancial signatures are lost [Modi, 2011]. Weathering on Mars, which is assumed to be water-limited and acidic, is expected to attack olivine first and more rapidly than other phases [McSween *et al.*, 2004; Hurowitz *et al.*, 2006; Stopar *et al.*, 2006]. Basaltic minerals that have been exposed to acidic and oxidizing conditions for even short periods of time (~30 days) can undergo alteration that results in large changes in their spectral slope, concavity, and absorption band position [Michalski *et al.*, 2006; Orofino *et al.*, 2006; Horgan *et al.*, 2011]. However, in situ geochemical characterization of many primary lithologies (e.g., picritic basalt) exposed on the surface for billion of years are relatively pristine, implying that weathering in these locations is not extensive [McSween *et al.*, 2006; Brückner *et al.*, 2008].

The clearest examples of local sand sources have remarkably similar spectra to adjacent dunes, indicating minor compositional differences between the two (Figs. 5.3a, 5.4a–5.4b, 5.4e, 5.4g–5.4h, 5.6b, 5.7c–5.7e, 5.9a, 5.11b, 5.12a, 5.14a). Because typically the same denominator spectrum was chosen for both targets, and both were from the same CRISM columns, ratioed spectra can be qualitatively compared without discrepancies caused by using different denominators or from instrument artifacts. The lack of evidence for alteration is a somewhat surprising result; examples of relatively pristine mafic mineral spectra are evident. The presence of VM dune sand composed of unaltered olivines, pyroxenes, and Fe-bearing glasses suggests that these sediments have not undergone a large degree of chemical weathering, nor have rinds or

coatings built up sufficiently to alter their spectral attributes [McSween *et al.*, 2006; Minitti *et al.*, 2007; Horgan *et al.*, 2011]. This fact provides some constraints on the timing of their surface exposure relative to other events in VM. The proposed accumulation of persistent VM lakes and subsequent chaotic floods in the Early Hesperian [Carr, 2006; Lucchitta, 2010] would have had to occur prior to dune emplacement and exposure of their mafic sources.

The largest spectral distinction between targets is the overall VNIR reflectance values or albedo. With few exceptions, near-infrared albedos decrease from bedrock to talus surfaces by ~5% and from talus to dune surfaces by ~5–10% (Figs. 5.3a, 5.4, 5.6, 5.7c–5.7d, 5.9a, 5.11b, 5.12a, 5.14a). We suggest this trend is the result of a complicated mix of one or more effects at the CRISM 18- to 200-m scales: partial dust coverage on talus slopes, lighter-toned boulders and cobbles mixed in with talus, aeolian transport of lower albedo mafic sand progressively away from sources, and removal of physically-weaker alteration products that are generally higher in albedo. Although illumination effects may also have a role, this trend is noted to occur at multiple orientations and shows no correlation with solar azimuth. VNIR laboratory measurements of basalts and pyroxenes show an increase in albedo for particles greater than coarse sand sizes (E. Cloutis, unpublished data, 2006), consistent with our findings. This trend contrasts previous laboratory studies [Cloutis *et al.*, 1986; King and Ridley, 1987a; Cloutis and Gaffey, 1991; Clark *et al.*, 1993; Orofino *et al.*, 2006] that have noted an overall increase in reflectance with decreasing particle size for the size range from ~0.45 μm to ~400 μm , and prompts further study. In any case, this decreasing albedo trend is not attributed to geochemical alteration, but rather physical processes and particle size differences.

Although unaltered basaltic minerals and glasses were the most common dune compositions detected in VM (Chapter IV), several examples of altered or secondary sand

compositions were also detected (~23% of detected VM dune compositions; Chapter IV). Occasionally, ferric iron was detected in VNIR spectra of dunes which otherwise displayed basaltic spectral shapes at longer wavelengths. One isolated example of altered olivine-bearing dunes in Coprates Chasma (Fig. 5.15d) was indicated by the spectral signature of ferric iron and a 2.3- μm spectral feature - alteration features not clearly present in spectra of the adjacent surfaces. Similar examples of altered olivine have been reported in the Nili Fossae region as part of bedrock surfaces [Mustard *et al.*, 2007; Ehlmann *et al.*, 2009] and in laboratory studies [Orofino *et al.*, 2006], but altered olivine sediment forming aeolian bedforms have not been described. If the adjacent, unaltered, olivine-bearing wall or landslide surfaces were the sand source, this would indicate chemical alteration to dune sediment over relatively short distances (~10s of kilometers). However, this is peculiar, considering the spectrally pristine examples of olivine in neighboring wall and dune surfaces (Figs. 5.3a, 5.6b).

Many spectra of Melas and Coprates chasmata dunes exhibit characteristics attributable to weathered or altered glasses, whereas others dunes in VM have features of unaltered, glass-rich basalt (Fig. 5.2b; Chapter IV). These occurrences of altered and unaltered glasses are spatially removed from one another. Dunes and wall sources partially composed of weathered glasses have nearly identical spectral composition (Figs. 5.4g–5.4h, 5.7c, 5.7d, 5.9a). It would appear these materials underwent chemical weathering *in situ* (as part of wall stratigraphy) prior to dune formation. Regardless of their alteration history, the presence of weathered glasses indicates yet another alteration mineral class discovered in VM and broadens the geologic context of the region.

Although unambiguous examples of Capri Chasma kieserite-bearing dunes were detected, spectra of their most likely sand sources (Capri Mensa talus fans) were noted to have

slightly different features at ~ 1.92 and ~ 2.39 μm (Fig. 5.13c). Capri Mensa is known to harbor a complicated stratigraphic sequence of both monohydrated and polyhydrated magnesium sulfates [Roach *et al.*, 2010a]. Spectra of the magnesium sulfates series ($\text{MgSO}_4 \cdot n\text{H}_2\text{O}$), with multiple hydrated states ($n = 0-7$), show absorption features that overlap as they broaden with increasing hydration [Dalton *et al.*, 2005] and may account for the observed trends. However, it is unclear why the polyhydrated phase is not detected in dune spectra despite appearing readily available for transport at the base of mensa. Dehydration of sulfate phases within the dune field may be a possibility given enough time and persistent hyper-arid conditions. Although previous analysis of paired VNIR observations of VM layered deposits did not detect a change in the assemblage's hydration state over short time periods (i.e., 1–2 Mars Year) [Roach *et al.*, 2009], those observations were of bedrock, whereas the environment of saltating sand might induce phase change more readily. Laboratory research indicates dewatering of polyhydrated sulfates forms amorphous sulfates rather than kieserite [Vaniman *et al.*, 2004; Freeman *et al.*, 2007], which may be difficult to detect in spectra. On the other hand, spectral analysis of the Olympia Undae gypsum sand sea shows hydration band depths do not diminish over short distances (<100 km) from their sources units [Horgan *et al.*, 2009]; although seasonal processes at those high latitudes and much higher daytime temperatures in VM make clear comparison difficult. Additionally, sulfate mineral hardness tends to decrease slightly with increased hydration (e.g., kieserite = 3.5, hexahydrite = 2.0 [Nesse, 2012]), which may be a factor in terms of sand particle durability. Thus, dehydration of more-hydrated sulfate phases and/or preferential fractionation of kieserite into dune sand (see below) are viable possibilities for the spectral discrepancy between dunes and the talus source.

5.4 Evidence for Intra-Dune Sand Fractionation

Density and particle size fractionation can arise from variations in settling velocities, grain entrainment thresholds, and grain transport velocities [Hughes *et al.*, 2000], and can profoundly affect compositional variability of sediment. Even given a single starting source lithology of basalt, compositional modification can occur by hydrodynamic sorting (e.g., olivine phenocrysts that segregate into larger bin sizes), and potentially mimic alteration pathways [McGlynn, 2012; McGlynn *et al.*, 2012]. A terrestrial example of aeolian sediment fractionating into different order bedforms at relatively small scales (~ 10 m) is observed within the perimeter of the Grand Falls dune field, Arizona. There, locally sourced (~ 10 km) ash and lapilli form coarse-grained ripples, whereas river wash delivered, quartz-rich sediment forms the bulk of surrounding barchan dunes [Hayward *et al.*, 2010a]. A larger scale example of density-driven aeolian fractionation can be found in southern Peru, where one small (~ 15 m), dark-toned barchan is composed of $\sim 50\%$ magnetite-bearing sand, but is surrounded by larger quartz-rich dunes [Gay, 1999]. Provided sufficient spatial resolution, these types of observations might be found occurring within martian aeolian systems.

The spatial resolutions of CRISM targeted and HiRISE color observations allow for greater discrimination between adjacent surface compositions than has been possible before. For example, certain dune spectra in eastern Coprates Chasma have narrow $\sim 1\text{-}\mu\text{m}$ absorptions, indicating olivine-bearing basalts, whereas other dune surface spectra ~ 1 km away possess broader $\sim 1\text{-}$ and $\sim 2.3\text{-}\mu\text{m}$ features attributed to HCP (Fig. 5.6b). Although other adjacent lower wall surface minerals (e.g., LCP, Fe/Mg-smectite [Flahaut *et al.*, 2012]) encapsulated in that figure appear to be available for downslope transport and incorporation into the dunes, their spectral signatures are not apparent in the dunes. This site is also noted to have monotonically

decreasing thermophysical properties downwind (inferred) of the source scarp (Fig. 5.6a). We suggest these heterogeneities and trends are the result of distinct source layers within the scarp producing variable talus-sand compositions and/or the downwind aeolian fractionation of these mafic minerals.

Melas Chasma color HiRISE images of lower wall and aeolian bedform surfaces show notable differences in color and inferred composition. Spur-derived, mid-toned talus fans primarily overlie darker-toned scree that have unknown sources (Fig. 5.10b). Abundant bedforms are located at the base of these talus fans with TARs that are both over- and underlying dark dunes, depending on the location. As represented in the false-color image of Fig. 5.10, these distinctive tan (ferric) and purple (ferrous) talus surfaces may be the sources for lower-lying TARs and dunes (Figs. 5.10d–5.10e), respectively, as they are nearly spectrally identical within HiRISE’s limited portion of the spectrum (Fig. 5.10f). Although not diagnostic, our observations of Melas Chasma bedform sediment fractionation imply compositional distinctions between TARs and dunes. Due to the small scale of TARs, these distinctions would not be observable with other available orbital instruments and provide insight into the possible origin of local TARs. Analyses of other color HiRISE images in VM consistently produce this trend (i.e., ferric values for TARs and ferrous for dark-toned dunes) and confirm this is not an isolated case.

Analysis of Capri Chasma dunes also shows evidence for a large degree of compositional heterogeneity (Figs. 5.2b, 5.13). CRISM observations in some locations show subtle variations in hydration features and, in some locations, are absent; hydration band depths are deeper in spectra of darker-toned dune crests relative to spectra of troughs (Fig. 5.13d). This trend is attributed to coarser grains forming dune crests, as found in terrestrial examples [Ghrefat *et al.*, 2007] and inferred from orbit in other martian sulfate-bearing dunes [Roach *et al.*, 2007; Horgan

et al., 2009]. As discussed above, observations of Capri Mensa show evidence for mass-wasting of both monohydrated and polyhydrated magnesium sulfates, but the monohydrated sulfate appears to be the preferred mobile phase. HiRISE color band ratios have values consistent with minerals containing both ferric- and ferrous-iron in false-color images of the two distinctive yellow and purple sand surfaces (Fig. 5.13a). Thermal inertias of the “yellow” sand are lower than the “purple” sand, consistent with finer particle sizes (i.e., ~400 μm vs. ~1200 μm , respectively). Evidence for hydrated mineral, iron oxidation state, and grain size differences or partitioning of Capri Chasma sand populations is consistent with aeolian compositional fractionation, likely due to variable mineral densities and the sorting capacity of the wind. At a much larger scale, Juventae Chasma shows evidence for several different sand compositions, represented as the spectral endmembers (i.e., olivine, LCP, HCP, and sulfate) nearest to their suspected source(s) (Fig. 5.2; Table 5.3; section 5.1). This evidence is interpreted to represent fractionation at the basinal scale. Additional examples presented in Chapter IV support the fractionation of dune sand at multiple scales (dune, dune field, and basin). These results are contrary to interpretations of the aeolian-reworking of Gusev crater dark soils and sand that, despite potentially billions of years of exposure to aeolian transport and fractionation, are insufficient to produce well sorted sediments [McGlynn *et al.*, 2011; McGlynn, 2012].

Previous terrestrial and martian studies have found olivine grains in volcanoclastic sand to be coarser than other components [Edgett and Lancaster, 1993; Sullivan *et al.*, 2008; Mangold *et al.*, 2011; Tirsch *et al.*, 2012]. For instance, certain aeolian deposits of the Ka’u Desert, Hawaii, are noted to have medium to coarse grained olivine sand forming dunes along with smaller basaltic glasses and minerals [Tirsch *et al.*, 2012]. Investigations of the El Dorado ripple field by the Spirit rover showed a slight enrichment in 200–300 μm olivine grains relative to smaller

pyroxene grains and lithic fragments [Sullivan *et al.*, 2008]. Those authors suggested olivine to be enriched due to slightly greater particle durability to physically weathering.

To test for a possible correlation between sand composition and inferred particle size, THEMIS thermal inertias were collected from sand surfaces with CRISM-derived compositions. Thermal inertias were collected from both the dune stoss sides and relatively flat areas to provide a representative sample of the dune. Histograms of THEMIS thermal inertia values of sand with spectral characteristics of olivine, basalt (i.e., olivine and pyroxene), pyroxene (both HCP and LCP), sulfate, and weathered glass are shown in Fig. 5.21. Consistent with earlier studies (Chapter IV) [Aben, 2003; Fenton *et al.*, 2003; Fergason *et al.*, 2006a; Chojnacki *et al.*, 2010], most dune surfaces have thermal inertia values (~220–400 tiu) corresponding to medium to coarse sand sizes. Basalt, pyroxene, and sulfate sand compositions typically have intermediate thermal inertias (240–280 tiu), whereas olivine dunes have higher values (300 tiu mean). Although there is a range of thermal inertia values associated with all sand compositions and the end-members overlap, olivine thermal inertia values are higher than those derived for the other sand compositions. This trend is interpreted as evidence of a coarser, mechanically stronger, olivine sand population relative to other compositionally distinct sand surfaces in VM. Field studies of Icelandic locally-derived sand sheets found that olivine is the coarser sand population relative to sands composed of pyroxene, plagioclase, and glass [Baratoux *et al.*, 2011; Mangold *et al.*, 2011]. In addition, Mangold *et al.* [2011] proposed this enrichment in olivine sand in their field site was due to: 1) sub-surface crystallization creating a pure mineral phase of individual euhedral crystal structures, 2) greater hardness values for olivine (6.5–7) relative to glasses and pyroxenes (5–6), 3) low chemical weathering rates, and 4) the slightly (<10%) greater density of

olivine relative to other present phases. Some of these factors may apply to martian examples described herein as well.

In addition, the evidence of olivine-bearing sand in VM and elsewhere (Chapter IV) [Sullivan *et al.*, 2008; Tirsch, 2009; McGlynn *et al.*, 2012], is counter to the interpretations of earlier studies of olivine-bearing surfaces [Edwards *et al.*, 2008; Bandfield *et al.*, 2011]. Bandfield *et al.* [2011] proposed that olivine is highly correlated (spatially) with bedrock, as indicated by elevated thermal inertias (>1200 tiu), and the fact that sandy/silty soils are largely olivine-poor, invoking ancient or ongoing chemical weathering as the cause. Documentation herein suggests substantial chemical weathering has not occurred on these olivine sand dunes since their exposure, nor is all martian olivine found exclusively associated with bedrock.

5.5 Source Material(s) Formation and Sediment Production Mechanisms

The diversity of primary and secondary sand minerals and the geologic contexts in which the Valles Marineris dunes are found suggest a variety of formation mechanisms for their sources. Primary sources include the upper and lower wall layers of VM, which are principally composed of basaltic lithologies, some which have been aqueously altered [Murchie *et al.*, 2009b; Flahaut *et al.*, 2012]. VM wall segments have been earlier interpreted as consisting of Noachian-aged massively-layered flood basalts [Witbeck *et al.*, 1991; McEwen *et al.*, 1999], volcanoclastic ash deposits [Hynek *et al.*, 2003; Bandfield *et al.*, 2013], with additional magmatic intrusions (e.g., dikes) [Williams *et al.*, 2003; Flahaut *et al.*, 2011]. Some of these wall units are spatially coincident with falling and floor dune source layers and our compositional mapping is consistent with a magmatic or volcanoclastic petrogenesis, depending on the location.

Other primary sources include the Juventae Chasma mantling unit that is a wide-spread, but heterogeneous unit superposed atop of wall and ILD surfaces [Lucchitta *et al.*, 1992; Malin

and Edgett, 2001; Chojnacki and Hynek, 2008]. Based on its spatial distribution, mafic composition, low-albedo, thermal inertia, and “pasted-on” appearance we interpret this unit as consisting of partially-consolidated sub-aerially deposited volcanoclastic materials, possibly deposited in relation to Tharsis explosive volcanism. Following weathering, terrestrial ash- and glass-rich deposits of phreatic eruptions are readily mobilized by the wind and can form adjacent bedforms, given sufficient conditions (wind strength/duration) [Edgett and Lancaster, 1993; Hayward *et al.*, 2010a; Tirsch *et al.*, 2012]. Weakly-indurated mantling deposits at the top of the mound C stratigraphy are the most exposed to atmospheric flow and resulting aeolian abrasion (sand blasting), and have greater gravitational potential energy for mass-wasting. In addition to mantling deposits, the thermophysical properties and morphology of poorly consolidated layers of some VM walls segments have been interpreted as evidence for extensive Noachian-aged volcanoclastic ash materials [Bandfield *et al.*, 2013]. Martian volcanoclastic ash deposits exposed in crater wall layers have also been proposed as sources for other dune populations [Fenton, 2005; Tirsch *et al.*, 2011] and these occurrences in VM may be similar.

Secondary sources include the Late Noachian- to Early Hesperian-aged, sulfate-bearing layered deposits in Capri, Juventae, and Ganges chasmata. The ultimate origin(s) of these ILDs is beyond the scope of this work (for a review see Chojnacki and Hynek [2008] and references therein), but the majority of observations and interpretations favor a sedimentary origin (e.g., aeolian, lacustrine). The spectral, photogeologic, and thermophysical link between ILD talus and sand strengthens the hypothesis that these ILDs are partially composed of ancient altered aeolian deposits and may be considered as sandstones, as previously proposed (see Chapter IV, section 5.5) [Nedell *et al.*, 1987; Malin and Edgett, 2000; Murchie *et al.*, 2009a].

We speculate that several mechanisms or processes are responsible for the breakdown and mobilization of rift aeolian source units, depending on the location and starting composition: thermal stress (i.e., thermal stress fatigue or thermal shock), impact gardening, aeolian erosion, and ancient weakening of bedrock due to aqueous alteration/diagenesis, all compounded by gravity-driven processes. Although these mechanisms are discussed in the context of findings described herein, we suggest these are broadly applicable to sediment production elsewhere on Mars. Studies of rock properties under cold and dry conditions found freeze-thaw and thermal stress fatigue to be effective, yet rarely recognized, mechanisms for rock flaking and breakdown [Hall, 1999; Hall and André, 2001]. It has been shown basalt is more susceptible to thermal stress fatigue than several other lithologies (e.g., quartzite, taconite), due to a higher thermal conductivity [Hall and André, 2001]. These thermal effects would be compounded for VM lithologies partially composed of secondary minerals (e.g., hydrated sulfates, phyllosilicates), due to their high thermal conductivity and greater potential for volume change.

Impact gardening or bolide comminution is a likely persistent mechanism by which the surface was fragmented and soils or sediments generated [Hartmann *et al.*, 2001; McGlynn *et al.*, 2011, 2012; McGlynn, 2012]. Although the peak impactor flux at Mars occurred early in its history [Hartmann and Neukum, 2001] and contemporary rates imply bolide impacts to be relatively minor components of surface erosion [Malin *et al.*, 2006], past impacts certainly had a role in the degradation of VM proposed source units, based on the prevalence of variable-sized and -aged craters [Lucchitta *et al.*, 1992; Quantin *et al.*, 2004]. It is estimated post-Noachian small impacts to have produced several 10s of meters of regolith, which would have aided bedrock degradation and sediment production [Hartmann *et al.*, 2001]. In addition to impact-generated sediment from pre-existing surfaces, some dark-toned sediments are likely impact

melts or tektitic glasses as redistributed as impact ejecta [Moroz *et al.*, 2009]. The impacts that created the small craters (km-scale) located on the surrounding plateau, wall, landslide, and canyon floor surfaces (Figs. 5.5, 5.7c, 5.8, 5.15) likely contributed to the production of dune and other bedform sediments.

Aeolian erosion (e.g., sand blasting) is likely to have occurred for the majority of Mars' history, as evident from large landforms in orbital images (e.g., yardangs, grooves) [Greeley *et al.*, 1992], and ancient sandstones (i.e., layered deposits) [Grotzinger *et al.*, 2005; Lewis *et al.*, 2008], small-scale rock features (e.g., ventifacts, flutes) examined at landing sites [Greeley, 2002; Greeley *et al.*, 2006]. Large yardangs formed from VM sulfate-bearing ILDs are examples of local disintegration of sedimentary layered units brought on by aeolian and mass-wasting processes (Figs. 5.11–5.13) [Catling *et al.*, 2006; Chojnacki and Hynek, 2008]. Given the prevalence for dunes, low-order bedforms, and other aeolian depositional or erosional features (e.g., crescentic scours) in close proximity to friable sedimentary sources, persistent wind erosion (and associated sediment production) is probable with respect to source unit degradation.

Ancient alteration of VM wall layers by ground water has apparently occurred forming a variety of secondary minerals [Murchie *et al.*, 2009a, 2009b; Ehlmann *et al.*, 2011]. Depending on the duration, aqueous alteration and diagenesis can profoundly affect primary lithologies' susceptibility to the other degradation mechanisms listed above [Johnsson, 1993]. One example of a possible wall rock alteration front is seen in Coprates Chasma (Fig. 5.3), which shows the indistinct transition from phyllosilicate- to olivine-bearing layers. The alteration mechanisms that formed this compositional contact here and elsewhere in VM may have had a role in the subsequent sourcing of sand via the weakening of bedrock due to aqueous diagenesis [Andrews-Hanna *et al.*, 2007; Murchie *et al.*, 2009a].

Topographic-setting is a first-order parameter influencing the nature and energy of transportational and depositional systems [Johnsson, 1993]. On Earth, high relief, steep sloped surfaces under dry climate conditions tend to experience moderate physical weathering, limited chemical weathering, low transportation rates, decreased transport times, and generally high sedimentation rates [Johnsson, 1993; Pelletier and Rasmussen, 2009]. These terrestrial generalizations are highly relevant to VM, where the high relief, slopes, and exposure of local strata may be driving factors in sourcing sediment for widespread aeolian bedforms.

6.0 Conclusions and Hypothesis Assessment

We have investigated the validity of the hypothesis that Valles Marineris dune fields are largely derived from a variety of local and regional (intra-rift) sediment sources, using multiple remote sensing data sets and techniques. We also investigated potential extended sand pathways leading into VM from unknown extra-rift sources. Qualitative and quantitative characterizations of local and regional PSSs and their paired dune fields are summarized in Table 5.3, which was constructed using the criteria outline in section 3.1. These results support our hypothesis: there is compelling evidence that several dune populations in VM are locally and regionally derived from intra-rift wall outcrops, ILD units, landslides, and other VM dune fields.

The investigation led to the following observations and interpretations:

- Evidence for mass-wasting of walls and ILDs (e.g., boulder tracks and associated boulders) are observed on steep-slopes adjacent to or within dune fields. Observations likely indicate these occurrences as persistent and broadly contemporaneous sources.
- Underlying slopes that connect proposed upper and lower wall sources to falling and floor dunes, respectively, often approach or exceed the angle of repose for sand-sized grains

(~32°). Most of these short (<10 km) pathways are relatively narrow (<1 km), which contribute to discrete dune sand pathways that are unidirectionally downslope.

- Several VM dune fields are superposed on steep-sloped ancient wall massifs or ILDs. The underlying units formed in the Noachian or Early Hesperian, which almost certainly indicates these dunes are topographically isolated from extra-rift sand sources and derived locally.
- In all cases, the thermal inertia (and the associated inferred particle size/degree of induration) of proposed sources is greater than that of dune sediment. Often, thermal inertias and albedos decrease downwind of the source, as inferred by morphology. These trends are attributed to be the result of preferential transport of finer, more mobile material that is dark-toned, and/or deflation of weaker particles of light-toned alteration products.
- The large degree of dune sand compositional heterogeneity, at multiple scales (dune, dune field, and basin), argues for distinct sediment sources and aeolian fractionation.
- Dozens of examples of dunes that are spectrally paired with intra-rift local and regional PSSs were provided. Little evidence for alteration of mafic minerals was detected, consistent with minimal chemical weathering. The greatest spectral distinction was the decrease in overall VNIR reflectance (albedo) from bedrock to talus to dune surfaces, which is attributed to physical processes.
- In Coprates and Melas chasmata, mapping is consistent with the principle sand source for dunes being altered and unaltered basaltic upper and lower wall. These primary source layers are inferred to be ancient extrusive and intrusive igneous units. The presence of various glassy basaltic sands may be evidence for wide-spread, sub-aerially deposited volcaniclastic units within the Noachian-aged walls.

- Capri, Juventae, and Ganges chasmata show evidence for sediment derivation from adjacent ILD units as stored (secondary) sources. Dunes containing “soft” secondary minerals, such as kieserite, are unlikely to have been produced from external sources due to the physical weathering during transport that would have occurred over vast distances and low durability of these minerals.
- Extra-rift dune fields, sand sheets, and sand patches are located on the plateaus surrounding VM and the adjoining areas of Noctis Labyrinthus and Aurorae Chaos. These isolated occurrences do not form interconnected networks of sand pathways into the rift in the current epoch.
- The best candidates for ancient sand pathways leading into VM are the east-to-west corridor into Ganges Chasma from Aurorae Chaos and the plateau north of eastern VM. These locations have dunes and dark streaks that shared a similar basaltic composition with certain VM dune fields and possessed paleowind indicators (e.g., slip faces, dark streaks) indicative of transport toward VM.

Thus, these points suggest great diversity in Valles Marineris dune sediment provenance and pathways. Furthermore, we suggest these examples of local and regional sourcing are more easily detected and volumetrically significant than many other locations on Mars, due to the extreme topography, compositional diversity of local lithologies, and extent of source units.

References

- Aben, L. K. (2003), Compositional and thermophysical analysis of Martian aeolian dunes, M.S. Thesis, Ariz. State Univ, Tempe, AZ.
- Ahmad, K., M. Khare, and K. K. Chaudhry (2005), Wind tunnel simulation studies on dispersion at urban street canyons and intersections—a review, *Journal of Wind Engineering and Industrial Aerodynamics*, 93(9), 697–717, doi:10.1016/j.jweia.2005.04.002.
- Anderson, J. A., S. C. Sides, T. L. Sucharski, and K. J. Becker (2004), Modernization of the Integrated Software for Imagers and Spectrometers, p. Abstract #2039.
- Andrews-Hanna, J., M. Zuber, R. Arvidson, and S. Wiseman (2010), Early Mars hydrology: Meridiani playa deposits and the sedimentary record of Arabia Terra, *J. Geophys. Res.*, 115, doi:10.1029/2009JE003485.
- Andrews-Hanna, J. C., R. J. Phillips, and M. T. Zuber (2007), Meridiani Planum and the global hydrology of Mars, *Nature*, 446(7132), 163–166, doi:10.1038/nature05594.
- Armstrong, J. C., and C. B. Leovy (2005), Long term wind erosion on Mars, *Icarus*, 176(1), 57–74, doi:10.1016/j.icarus.2005.01.005.
- Arvidson, R. E. (1974), Wind-blown streaks, splotches, and associated craters on Mars: Statistical analysis of Mariner 9 photographs, *Icarus*, 21(1), 12–27, doi:10.1016/0019-1035(74)90086-4.
- Arvidson, R. E. et al. (2006), Nature and origin of the hematite-bearing plains of Terra Meridiani based on analyses of orbital and Mars Exploration rover data sets, *J. Geophys. Res.*, 111(E12), doi:10.1029/2006JE002728.
- Bagnold, R. A. (1941), *The Physics of Blown Sand and Desert Dunes*, Methuen, London.
- Bandfield, J. L. (2002a), Global mineral distributions on Mars, *J. Geophys. Res.*, 107(E6), doi:10.1029/2001JE001510.
- Bandfield, J. L. (2002b), Spectroscopic study of the Moses Lake dune field, Washington: Determination of compositional distributions and source lithologies, *J. Geophys. Res.*, 107(E11), doi:10.1029/2000JE001469.
- Bandfield, J. L. (2009), Effects of surface roughness and graybody emissivity on martian thermal infrared spectra, *Icarus*, 202(2), 414–428, doi:10.1016/j.icarus.2009.03.031.
- Bandfield, J. L., V. E. Hamilton, and P. R. Christensen (2000), A Global View of Martian Surface Compositions from MGS-TES, *Science*, 287(5458), 1626–1630, doi:10.1126/science.287.5458.1626.
- Bandfield, J. L., A. Deanne Rogers, and C. S. Edwards (2011), The role of aqueous alteration in the formation of martian soils, *Icarus*, 211(1), 157–171, doi:10.1016/j.icarus.2010.08.028.
- Bandfield, J. L., C. S. Edwards, D. R. Montgomery, and B. D. Brand (2013), The dual nature of the martian crust: Young lavas and old clastic materials, *Icarus*, 222(1), 188–199, doi:10.1016/j.icarus.2012.10.023.
- Baratoux, D., N. Mangold, O. Arnalds, J.-M. Bardintzeff, B. Platevoët, M. Grégoire, and P. Pinet (2011), Volcanic sands of Iceland - Diverse origins of aeolian sand deposits revealed at Dyngjúsandur and Lambahraun, *Earth Surface Processes and Landforms*, 36(13), 1789–1808, doi:10.1002/esp.2201.
- Bateman, M. D., R. G. Bryant, I. D. L. Foster, I. Livingstone, and A. J. Parsons (2012), On the formation of sand ramps: A case study from the Mojave Desert, *Geomorphology*, 161–162, 93–109, doi:10.1016/j.geomorph.2012.04.004.

- Beyer, R. A., and A. S. McEwen (2005), Layering stratigraphy of eastern Coprates and northern Capri Chasmata, Mars, *Icarus*, 179(1), 1–23, doi:10.1016/j.icarus.2005.06.014.
- Bibring, J. P., and Y. Langevin (2008), Mineralogy of the Martian surface from Mars Express OMEGA observations, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. F. I. Bell, pp. 153–168, Cambridge University Press, Cambridge.
- Bibring, J. P., A. Soufflot, and J. Mustard (2004), OMEGA: Observatoire pour la Mineralogie, l'Eau, les Glaces, et l'Activite, ESA Special Publication, 1240, 37–49.
- Bibring, J. P. et al. (2005), Mars surface diversity as revealed by the OMEGA/Mars Express observations, *Science*, 307(5715), 1576–1581, doi:10.1126/Science.1108806.
- Bishop, J. L. et al. (2009), Mineralogy of Juventae Chasma: Sulfates in the light-toned mounds, mafic minerals in the bedrock, and hydrated silica and hydroxylated ferric sulfate on the plateau, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003352.
- Bishop, M. A. (2011), Aeolian scours as putative signatures of wind erosion and sediment transport direction on Mars, *Geomorphology*, 125(4), 569–574, doi:10.1016/j.geomorph.2010.10.029.
- Blasius, K. R., J. A. Cutts, J. E. Guest, and H. Masursky (1977), Geology of the Valles Marineris: First Analysis of Imaging From the Viking 1 Orbiter Primary Mission, *J. Geophys. Res.*, 82(28), 4067–4091, doi:10.1029/JS082i028p04067.
- Bourke, M., S. A. Wilson, and J. R. Zimbelman (2003), The variability of transverse aeolian ridges in troughs on Mars, vol. XXXIV, p. Abstract 2090.
- Bourke, M., M. Balme, R. Beyer, K. Williams, and J. Zimbelman (2006), A comparison of methods used to estimate the height of sand dunes on Mars, *Geomorphology*, 81(3-4), 440–452, doi:10.1016/j.geomorph.2006.04.023.
- Bourke, M., K. Edgett, and B. Cantor (2008), Recent aeolian dune change on Mars, *Geomorphology*, 94(1-2), 247–255, doi:10.1016/j.geomorph.2007.05.012.
- Bourke, M. C., J. E. Bullard, and O. S. Barnouin-Jha (2004), Aeolian sediment transport pathways and aerodynamics at troughs on Mars, *J. Geophys. Res.*, 109(E7), doi:10.1029/2003JE002155.
- Bourke, M. C., N. Lancaster, L. K. Fenton, E. J. R. Parteli, J. R. Zimbelman, and J. Radebaugh (2010), Extraterrestrial dunes: An introduction to the special issue on planetary dune systems, *Geomorphology*, 121, 1–14, doi:10.1016/j.geomorph.2010.04.007.
- Breed, C. S., M. J. Grolier, and J. F. McCauley (1979), Morphology and Distribution of Common “Sand” Dunes on Mars: Comparison With the Earth, *J. Geophys. Res.*, 84(B14), 8183–8204, doi:10.1029/JB084iB14p08183.
- Bridges, N. T., P. E. Geissler, A. S. McEwen, B. J. Thomson, F. C. Chuang, K. E. Herkenhoff, L. P. Keszthelyi, and S. Martinez-Alonso (2007), Windy Mars: A dynamic planet as seen by the HiRISE camera, *Geophys. Res. Lett.*, 34(23), doi:10.1029/2007GL031445.
- Bridges, N. T. et al. (2010), Aeolian bedforms, yardangs, and indurated surfaces in the Tharsis Montes as seen by the HiRISE Camera: Evidence for dust aggregates, *Icarus*, 205(1), 165–182, doi:10.1016/j.icarus.2009.05.017.
- Bridges, N. T., F. Ayoub, J.-P. Avouac, S. Leprince, A. Lucas, and S. Mattson (2012a), Earth-like sand fluxes on Mars, *Nature*, advance online publication, doi:10.1038/nature11022.
- Bridges, N. T. et al. (2012b), Planet-wide sand motion on Mars, *Geology*, 40(1), 31–34, doi:10.1130/G32373.1.

- Brückner, J., G. Dreibus, R. Gellert, S. W. Squyres, H. Wänke, A. Yen, and J. Zipfel (2008), Mars Exploration Rovers: chemical composition by the APXS, in *The Martian Surface*, edited by J. Bell and J. Bell, pp. 58–102, Cambridge University Press, Cambridge.
- Burns, R. G. (1993), Origin of electronic spectra of minerals in the visible to near-infrared region, in *Remote Geochemical Analysis: Elemental and Mineralogical Composition*, edited by C. M. Pieters and P. A. J. Englert, Cambridge University Press.
- Cantor, B. (2007), MOC observations of the 2001 Mars planet-encircling dust storm, *Icarus*, 186(1), 60–96, doi:10.1016/j.icarus.2006.08.019.
- Cantor, B. A., K. M. Kanak, and K. S. Edgett (2006), Mars Orbiter Camera observations of Martian dust devils and their tracks (September 1997 to January 2006) and evaluation of theoretical vortex models, *J. Geophys. Res.*, 111(E12), E12002, doi:10.1029/2006JE002700.
- Cardinale, M., G. Komatsu, S. Silvestro, and D. Tirsch (2012), The influence of local topography for wind direction on Mars: two examples of dune fields in crater basins, *Earth Surf. Process. Landforms*, n/a, doi:10.1002/esp.3289.
- Carr, M. (2006), *The surface of Mars*, [Rev. ed.], Cambridge University Press, Cambridge; New York.
- Catling, D., S. Wood, C. Leovy, D. Montgomery, H. Greenberg, C. Glein, and J. Moore (2006), Light-toned layered deposits in Juventae Chasma, Mars, *Icarus*, 181(1), 26–51, doi:10.1016/j.icarus.2005.10.020.
- Chapman, M. G., and K. L. Tanaka (2001), Interior trough deposits on Mars: Subice volcanoes?, *J. Geophys. Res.-Planets*, 106(E5), 10087–10100.
- Chapman, M. G., M. T. Gudmundsson, A. J. Russell, and T. M. Hare (2003), Possible Juventae Chasma subice volcanic eruptions and Maja Valles ice outburst floods on Mars: Implications of Mars Global Surveyor crater densities, geomorphology, and topography, *J. Geophys. Res.*, 108(E10), doi:10.1029/2002JE002009.
- Chojnacki, M., and B. M. Hynek (2008), Geological context of water-altered minerals in Valles Marineris, Mars, *J. Geophys. Res.*, 113(E12), doi:10.1029/2007JE003070.
- Chojnacki, M., and J. E. Moersch (2009), Valles Marineris Dune Fields: Thermophysical Properties, Morphology, and Provenance, p. Abstract #2486, Lunar and Planetary Institute, Houston.
- Chojnacki, M., B. M. Jakosky, and B. M. Hynek (2006), Surficial properties of landslides and surrounding units in Ophir Chasma, Mars, *J. Geophys. Res.*, 111(E4), doi:10.1029/2005JE002601.
- Chojnacki, M., J. Moersch, and D. Burr (2010), Climbing and falling dunes in Valles Marineris, Mars, *Geophys. Res. Lett.*, 37, doi:10.1029/2009GL042263.
- Chojnacki, M., D. M. Burr, J. E. Moersch, and T. I. Michaels (2011a), Orbital observations of contemporary dune activity in Endeavor crater, Meridiani Planum, Mars, *J. Geophys. Res.*, 116, doi:10.1029/2010JE003675.
- Chojnacki, M., J. E. Moersch, and D. M. Burr (2011b), Variable scale surface change of Valles Marineris dune fields and adjacent terrains, in *EPSC-DPS*, vol. 6, p. 1529.
- Chojnacki, M., J. E. Moersch, D. M. Burr, and J. J. Wray (2012a), Potential Sediment Sources and Pathways in Valles Marineris Dune Fields: Implications for Martian Aeolian Systems, in *Third International Planetary Dunes Workshop: Remote Sensing and Image Analysis of Planetary Dunes*, p. Abstract #7040, Lunar and Planetary Institute, Flagstaff, Arizona.

- Chojnacki, M., J. E. Moersch, D. M. Burr, and J. J. Wray (2012b), Valles Marineris Dune Fields: Sediment Pathways and Provenance, p. Abstract #2444, Lunar and Planetary Institute, Houston.
- Christensen, P. (1986), The spatial distribution of rocks on mars, *Icarus*, 68(2), 217–238, doi:10.1016/0019-1035(86)90020-5.
- Christensen, P. (1988a), Eolian intracrater deposits on Mars: Physical properties and global distribution, *Icarus*, 56(3), 496–518, doi:10.1016/0019-1035(83)90169-0.
- Christensen, P. R. (1988b), Global Albedo Variations on Mars: Implications for Active Aeolian Transport, Deposition, and Erosion, *J. Geophys. Res.*, 93(B7), 7611–7624.
- Christensen, P. R. et al. (1992), Thermal Emission Spectrometer Experiment: Mars Observer Mission, *J. Geophys. Res.*, 97(E5), 7719, doi:10.1029/92JE00453.
- Christensen, P. R., R. V. Morris, M. D. Lane, J. L. Bandfield, and M. C. Malin (2001a), Global mapping of Martian hematite mineral deposits: Remnants of water-driven processes on early Mars, *J. Geophys. Res.-Planets*, 106(E10), 23873–23885.
- Christensen, P. R. et al. (2001b), Mars Global Surveyor Thermal Emission Spectrometer experiment: Investigation description and surface science results, *J. Geophys. Res.*, 106(E10), 23823–23871, doi:10.1029/2000JE001370.
- Christensen, P. R. et al. (2003), Morphology and composition of the surface of Mars: Mars Odyssey THEMIS results, *Science*, 10.1126/science.1080885.
- Christensen, P. R. et al. (2004), The Thermal Emission Imaging System (THEMIS) for the Mars 2001 Odyssey Mission, *Space Science Reviews*, 110(1/2), 85–130, doi:10.1023/B:SPAC.0000021008.16305.94.
- Christensen, P. R., J. L. Bandfield, A. D. Rogers, T. D. Glotch, V. E. Hamilton, S. W. Ruff, and M. B. Wyatt (2008), Global Mineralogy Mapped from the Mars Global Surveyor Thermal Emission Spectrometer, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell, pp. 195–220, Cambridge University Press, Cambridge.
- Clancy, R. T., M. J. Wolff, B. A. Cantor, M. C. Malin, and T. I. Michaels (2009), Valles Marineris cloud trails, *J. Geophys. Res.*, 114, doi:10.1029/2008JE003323.
- Clark, R. N., T. V. V. King, M. Klejwa, G. A. Swayze, and N. Vergo (1990), High Spectral Resolution Reflectance Spectroscopy of Minerals, *J. Geophys. Res.*, 95(B8), 12653–12680, doi:10.1029/JB095iB08p12653.
- Clark, R. N., G. A. Swayze, A. J. Gallagher, T. V. V. King, and W. M. Calvin (1993), The U. S. Geological Survey, Digital Spectral Library: Version 1: 0.2 to 3.0 microns, U.S. Geological Survey Open File Report, 93-592, 1340 pages.
- Clenet, H., J. Flahaut, C. Quantin, P. Pinet, Y. Daydou, and P. Allemand (2011), Compositional Diversity of Mafic Rocks in the Vicinity of Valles Marineris, Mars, Using Modified Gaussian Model, vol. 38th Lunar and Planetary Science Conference, p. abstract 1674.
- Cloutis, E., and J. F. I. Bell (2004), Mafic silicate mapping on Mars: effects of palagonitic material, multiple mafic silicates, and spectral resolution, *Icarus*, 172(1), 233–254, doi:10.1016/j.icarus.2004.06.002.
- Cloutis, E. A., and M. J. Gaffey (1991), Spectral-compositional variations in the constituent minerals of mafic and ultramafic assemblages and remote sensing implications, *Earth, Moon, and Planets*, 53(1), 11–53, doi:10.1007/BF00116217.
- Cloutis, E. A., M. J. Gaffey, T. L. Jackowski, and K. L. Reed (1986), Calibrations of Phase Abundance, Composition, and Particle Size Distribution for Olivine-Orthopyroxene Mixtures

- From Reflectance Spectra, *Journal of Geophysical Research*, 91(B11), 11641–11653, doi:10.1029/JB091iB11p11641.
- Cloutis, E. A. et al. (2006), Detection and discrimination of sulfate minerals using reflectance spectroscopy, *Icarus*, 184(1), 121–157, doi:10.1016/j.icarus.2006.04.003.
- Dalton, J. B., O. Prieto-Ballesteros, J. S. Kargel, C. S. Jamieson, J. Jolivet, and R. Quinn (2005), Spectral comparison of heavily hydrated salts with disrupted terrains on Europa, *Icarus*, 177(2), 472–490, doi:10.1016/j.icarus.2005.02.023.
- Delamere, W. A., L. L. Tornabene, A. S. McEwen, K. Becker, J. W. Bergstrom, N. T. Bridges, E. M. Eliason, D. Gallagher, K. E. Herkenhoff, and L. Keszthelyi (2010), Color imaging of Mars by the High Resolution Imaging Science Experiment (HiRISE), *Icarus*, 205(1), 38–52, doi:10.1016/j.icarus.2009.03.012.
- Dohm, J. M., J.-P. Williams, R. C. Anderson, J. Ruiz, P. C. McGuire, G. Komatsu, A. F. Davila, J. C. Ferris, D. Schulze-Makuch, and V. R. Baker (2009), New evidence for a magmatic influence on the origin of Valles Marineris, Mars, *Journal of Volcanology and Geothermal Research*, 185(1-2), 12–27, doi:10.1016/j.jvolgeores.2008.11.029.
- Edgett, K. S. (1994), The Sand Component of the Modern Martian Aeolian Sedimentary System, Ph.D. Dissertation, Ariz. State Univ., Tempe, AZ.
- Edgett, K. S. (2002), Low-albedo surfaces and eolian sediment: Mars Orbiter Camera views of western Arabia Terra craters and wind streaks, *J. Geophys. Res.*, 107(E6), doi:10.1029/2001JE001587.
- Edgett, K. S., and N. Lancaster (1993), Volcaniclastic aeolian dunes: terrestrial examples and application to martian sands, *Journal of Arid Environments*, 25(3), 271–297, doi:10.1006/jare.1993.1061.
- Edgett, K. S., and M. C. Malin (2000a), Examples of Martian Sandstone: Indurated, Lithified, and Cratered Eolian Dunes in MGS MOC Images, p. Abstract #1071, Lunar and Planetary Institute, Houston.
- Edgett, K. S., and M. C. Malin (2000b), New views of Mars eolian activity, materials, and surface properties: Three vignettes from the Mars Global Surveyor Mars Orbiter Camera, *J. Geophys. Res.*, 105(E1), 1623–1650, doi:10.1029/1999JE001152.
- Edwards, C. S., P. R. Christensen, and V. E. Hamilton (2008), Evidence for extensive olivine-rich basalt bedrock outcrops in Ganges and Eos chasmas, Mars, *J. Geophys. Res.*, 113(E11), doi:10.1029/2008JE003091.
- Edwards, C. S., J. L. Bandfield, P. R. Christensen, and R. L. Fergason (2009), Global distribution of bedrock exposures on Mars using THEMIS high-resolution thermal inertia, *J. Geophys. Res.*, 114(E11), doi:10.1029/2009JE003363.
- Ehlmann, B. L. et al. (2009), Identification of hydrated silicate minerals on Mars using MRO-CRISM: Geologic context near Nili Fossae and implications for aqueous alteration, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003339.
- Ehlmann, B. L., J. F. Mustard, S. L. Murchie, J.-P. Bibring, A. Meunier, A. A. Fraeman, and Y. Langevin (2011), Subsurface water and clay mineral formation during the early history of Mars, *Nature*, 479(7371), 53–60, doi:10.1038/nature10582.
- Al-Enezi, A., K. Pye, R. Misak, and S. Al-Hajraf (2008), Morphologic characteristics and development of falling dunes, northeast Kuwait, *Journal of Arid Environments*, 72(4), 423–439, doi:10.1016/j.jaridenv.2007.06.017.
- Fenton, L. K. (2005), Potential sand sources for the dune fields in Noachis Terra, Mars, *J. Geophys. Res.*, 110(E11), doi:10.1029/2005JE002436.

- Fenton, L. K. (2006), Dune migration and slip face advancement in the Rabe Crater dune field, Mars, *Geophysical Research Letters*, 33, doi:10.1029/2006GL027133.
- Fenton, L. K., and R. A. Beyer (2012), Reconstructing the Formative Winds of a Dune Field in Ganges Chasma, Mars, by Bootstrapping with the Rule of Maximum Gross Bedform-Normal Transport, p. Abstract #7028, Lunar and Planetary Institute, Houston.
- Fenton, L. K., and R. K. Hayward (2010), Southern high latitude dune fields on Mars: Morphology, aeolian inactivity, and climate change, *Geomorphology*, 121(1-2), 98–121, doi:10.1016/j.geomorph.2009.11.006.
- Fenton, L. K., J. L. Bandfield, and A. W. Ward (2003), Aeolian processes in Proctor Crater on Mars: Sedimentary history as analyzed from multiple data sets, *J. Geophys. Res.*, 108(E12), doi:10.1029/2002JE002015.
- Fenton, L. K., A. D. Toigo, and M. I. Richardson (2005), Aeolian processes in Proctor Crater on Mars: Mesoscale modeling of dune-forming winds, *J. Geophys. Res.*, 110(E6), doi:10.1029/2004JE002309.
- Fenton, L. K., P. E. Geissler, and R. M. Haberle (2007), Global warming and climate forcing by recent albedo changes on Mars, *Nature*, 446(7136), 646–649, doi:10.1038/nature05718.
- Fenton, L. K., T. I. Michaels, and R. A. Beyer (2012), Aeolian Sediment Sources and Transport in Ganges Chasma, Mars: Morphology and Atmospheric Modeling, vol. XLIII, p. Abstract 2441.
- Ferguson, R. L., P. R. Christensen, and H. H. Kieffer (2006a), High-resolution thermal inertia derived from the Thermal Emission Imaging System (THEMIS): Thermal model and applications, *J. Geophys. Res.*, 111(E12), doi:10.1029/2006JE002735.
- Ferguson, R. L., P. R. Christensen, J. F. Bell, M. P. Golombek, K. E. Herkenhoff, and H. H. Kieffer (2006b), Physical properties of the Mars Exploration Rover landing sites as inferred from Mini-TES–derived thermal inertia, *J. Geophys. Res.*, 111(E2), doi:10.1029/2005JE002583.
- Flahaut, J., C. Quantin, P. Allemand, P. Thomas, and L. Le Deit (2010), Identification, distribution and possible origins of sulfates in Capri Chasma (Mars), inferred from CRISM data, *J. Geophys. Res.*, 115(E11), doi:10.1029/2009JE003566.
- Flahaut, J., J. F. Mustard, C. Quantin, H. Clenet, P. Allemand, and P. Thomas (2011), Dikes of distinct composition intruded into Noachian-aged crust exposed in the walls of Valles Marineris, *Geophysical Research Letters*, 38(15), doi:10.1029/2011GL048109.
- Flahaut, J., C. Quantin, H. Clenet, P. Allemand, J. F. Mustard, and P. Thomas (2012), Pristine Noachian crust and key geologic transitions in the lower walls of Valles Marineris: Insights into early igneous processes on Mars, *Icarus*, doi:10.1016/j.icarus.2011.12.027.
- Freeman, J. J., A. Wang, and B. L. Jolliff (2007), Pathways to Form Kieserite from Epsomite at Mid to Low Temperatures, with Relevance to Mars, p. Abstract #1298, Lunar and Planetary Institute, Houston.
- Frey, H. (1979), Martian canyons and African rifts: Structural comparisons and implications, *Icarus*, 37(1), 142–155, doi:10.1016/0019-1035(79)90122-2.
- Fryberger, S. G. (1979), Dune forms and wind regime, vol. 1042, pp. 137–170, U.S. Geol. Surv. Prof. Pap.
- Gaddis, L. (1997), An overview of the Integrated Software for Imaging Spectrometers (ISIS), *Lunar Planet. Sci.*, XXVIII, abstract 1226.

- Gardin, E., P. Allemand, C. Quantin, S. Silvestro, and C. Delacourt (2012), Dune fields on Mars: Recorders of a climate change?, *Planetary and Space Science*, 60(1), 314–321, doi:10.1016/j.pss.2011.10.004.
- Gay, S. P. (1999), Observations regarding the movement of barchan sand dunes in the Nazca to Tanaca area of southern Peru, *Geomorphology*, 27, 279–293.
- Geissler, P. E. (2005), Three decades of Martian surface changes, *J. Geophys. Res.*, 110(E2), doi:10.1029/2004JE002345.
- Geissler, P. E. et al. (2008), First in situ investigation of a dark wind streak on Mars, *J. Geophys. Res.*, 113(E12), doi:10.1029/2008JE003102.
- Geissler, P. E., N. W. Stanzos, N. T. Bridges, and the HiRISE Science Team (2011), Shifting sands on Mars: 3 case studies, vol. XLII, p. Abstract 2537.
- Geissler, P. E., M. E. Banks, N. T. Bridges, S. Silvestro, and HiRISE Science Team (2012), HiRISE Observations of Sand Dune Motion on Mars: Emerging Global Trends, p. Abstract #7053, Lunar and Planetary Institute, Houston.
- Geissler, P. E., L. K. Fenton, and N. T. Bridges (2013), Dust “Drifts” on Mars, p. Abstract #2573, Lunar and Planetary Institute, Houston.
- Gendrin, A. et al. (2005), Sulfates in martian layered terrains: the OMEGA/Mars Express view, *Science*, 307(5715), 1587–1591, doi:10.1126/Science.1109087.
- Ghrefat, H. A., P. C. Goodell, B. E. Hubbard, R. P. Langford, and R. E. Aldouri (2007), Modeling grain size variations of aeolian gypsum deposits at White Sands, New Mexico, using AVIRIS imagery, *Geomorphology*, 88(1-2), 57–68, doi:10.1016/j.geomorph.2006.10.013.
- Golombek, M., K. Robinson, A. McEwen, N. Bridges, B. Ivanov, L. Tornabene, and R. Sullivan (2010), Constraints on ripple migration at Meridiani Planum from Opportunity and HiRISE observations of fresh craters, *J. Geophys. Res.*, 115, doi:10.1029/2010JE003628.
- Golombek, M. P. et al. (2006), Erosion rates at the Mars Exploration Rover landing sites and long-term climate change on Mars, *J. Geophys. Res.*, 111(E12), doi:10.1029/2006JE002754.
- Gorelick, N. S., M. Weiss-Malik, B. Steinberg, and S. Anwar (2003), JMARS: A Multimission Data Fusion Application, in 34th Lunar and Planetary Science Conference, p. Abstract #2057, Lunar and Planetary Institute, Houston, Texas.
- Greeley, R. (2002), Terrestrial analogs to wind-related features at the Viking and Pathfinder landing sites on Mars, *J. Geophys. Res.*, 107(E1), doi:10.1029/2000JE001481.
- Greeley, R., and J. E. Guest (1987), Geologic map of the eastern equatorial region of Mars (USGS I-1802-B, 1987, 1:15M scale).
- Greeley, R., and J. D. Iversen (1985), Wind as a geological process on Earth, Mars, Venus and Titan, edited by R. Greeley, Cambridge Univ. Press, New York.
- Greeley, R., and M. D. Kraft (2001), Survivability of Aggregate Sands on Mars, p. Abstract #1839, Lunar and Planetary Institute, Houston.
- Greeley, R., R. Leach, B. White, J. Iversen, and J. Pollack (1980), Threshold windspeeds for sand on Mars: Wind tunnel simulations, *Geophys. Res. Lett.*, 7(2), 121, doi:10.1029/GL007i002p00121.
- Greeley, R., N. Lancaster, S. W. Lee, and P. C. Thomas (1992), Martian Aeolian Processes, Sediments and Features, in Mars, edited by H. H. Kieffer, B. M. Jakosky, C. W. Snyder, and M. S. Matthews, pp. 730–766, University of Arizona Press, Tucson.
- Greeley, R., A. Skyeck, and J. B. Pollack (1993), Martian Aeolian Features and Deposits: Comparisons With General Circulation Model Results, *J. Geophys. Res.*, 98(E2), 3183–3196, doi:10.1029/92JE02580.

- Greeley, R. et al. (2006), Gusev crater: Wind-related features and processes observed by the Mars Exploration Rover Spirit, *J. Geophys. Res.*, 111(E2), doi:10.1029/2005JE002491.
- Grotzinger, J. P. et al. (2005), Stratigraphy and sedimentology of a dry to wet eolian depositional system, Burns formation, Meridiani Planum, Mars, *Earth and Planetary Science Letters*, 240(1), 11–72, doi:10.1016/j.epsl.2005.09.039.
- Haberle, R. M. (1986), The climate of Mars, *Scientific American*, 254(5).
- Haberle, R. M., J. B. Pollack, J. R. Barnes, R. W. Zurek, C. B. Leovy, J. R. Murphy, H. Lee, and J. Schaeffer (1993), Mars Atmospheric Dynamics as Simulated by the NASA Ames General Circulation Model 1. The Zonal-Mean Circulation, *J. Geophys. Res.*, 98(E2), 3093–3123, doi:10.1029/92JE02946.
- Hall, K. (1999), The role of thermal stress fatigue in the breakdown of rock in cold regions, *Geomorphology*, 31(1–4), 47–63, doi:10.1016/S0169-555X(99)00072-0.
- Hall, K., and M.-F. André (2001), New insights into rock weathering from high-frequency rock temperature data: an Antarctic study of weathering by thermal stress, *Geomorphology*, 41(1), 23–35, doi:10.1016/S0169-555X(01)00101-5.
- Hartmann, W. K., and G. Neukum (2001), Cratering chronology and the evolution of Mars, *Space Science Reviews*, 96(1–4), 165–194.
- Hartmann, W. K., J. Anguita, M. A. de la Casa, D. C. Berman, and E. V. Ryan (2001), Martian Cratering 7: The Role of Impact Gardening, *Icarus*, 149(1), 37–53, doi:10.1006/icar.2000.6532.
- Hayes, A. G., J. P. Grotzinger, L. A. Edgar, S. W. Squyres, W. A. Watters, and J. Sohl-Dickstein (2011), Reconstruction of eolian bed forms and paleocurrents from cross-bedded strata at Victoria Crater, Meridiani Planum, Mars, *J. Geophys. Res.*, 116, doi:10.1029/2010JE003688.
- Hayward, R. K., K. F. Mullins, L. K. Fenton, T. M. Hare, T. N. Titus, M. C. Bourke, A. Colaprete, and P. R. Christensen (2007a), Mars Global Digital Dune Database and initial science results, *J. Geophys. Res.*, 112(E11), doi:10.1029/2007JE002943.
- Hayward, R. K., K. F. Mullins, L. K. Fenton, T. M. Hare, T. N. Titus, M. C. Bourke, A. Colaprete, and P. R. Christensen (2007b), Mars Global Digital Dune Database: MC2 – MC29, USGS Open File Report.
- Hayward, R. K., T. N. Titus, T. I. Michaels, L. K. Fenton, A. Colaprete, and P. R. Christensen (2009), Aeolian dunes as ground truth for atmospheric modeling on Mars, *J. Geophys. Res.*, 114(E11), doi:10.1029/2009JE003428.
- Hayward, R. K., J. R. Zimbelman, L. K. Fenton, T. N. Titus, and G. E. Cushing (2010a), Mars Analog: Grand Falls Dune Field, Arizona, in *Second International Planetary Dunes Workshop: Planetary Analogs _ Integrating Models, Remote Sensing, and Field Data*, p. Abstract #2004, Lunar and Planetary Institute, Alamosa, Colorado.
- Hayward, R. K., L. K. Fenton, K. L. Tanaka, T. N. Titus, and P. R. Christensen (2010b), Mars global digital dune database: Dune volume estimates in the north polar region, vol. XLI, p. Abstract 1109.
- Hazen, R. M., D. Papineau, W. Bleeker, R. T. Downs, J. M. Ferry, T. J. McCoy, D. A. Sverjensky, and H. Yang (2008), Mineral evolution, *American Mineralogist*, 93(11–12), 1693–1720, doi:10.2138/am.2008.2955.
- Herkenhoff, K. E., M. P. Golombek, E. A. Guinness, J. B. Johnson, A. Kusack, L. Richter, R. J. Sullivan, and S. Gorevan (2008), In situ observations of the physical properties of the Martian surface, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell and J. Bell, pp. 451–467, Cambridge University Press, Cambridge.

- Hiroi, T., and C. M. Pieters (1994), Estimation of grain sizes and mixing ratios of fine powder mixtures of common geologic minerals, *J. Geophys. Res.*, 99(E5), 10867–10879, doi:10.1029/94JE00841.
- Hobbs, S. W., D. J. Paull, and M. C. Bourke (2010), Aeolian processes and dune morphology in Gale Crater, *Icarus*, 210(1), 102–115, doi:10.1016/j.icarus.2010.06.006.
- Horgan, B., and J. F. Bell (2012a), Widespread weathered glass on the surface of Mars, *Geology*, doi:10.1130/G32755.1.
- Horgan, B., P. Mann, J. Stromberg, and E. A. Cloutis (2011), Acid Alteration of Basalts, Andesites, and Anorthosites: Near-IR Spectra and Implications for Martian Soil Formation, in 42nd Lunar and Planetary Science Conference, p. Abstract #2415, Lunar and Planetary Institute, Houston.
- Horgan, B., L. K. Fenton, and P. R. Christensen (2012a), Comparing Active Modes of Mass Movement on Martian Dunes, p. Abstract #7052, Lunar and Planetary Institute, Houston.
- Horgan, B., D. Clarke, and J. F. I. Bell (2012b), Evidence for Glass-Rich Pyroclastics in Martian Dunes, in Third International Planetary Dunes Workshop: Remote Sensing and Image Analysis of Planetary Dunes, p. Abstract #7050, Lunar and Planetary Institute, Flagstaff, Arizona.
- Horgan, B., M. Chojnacki, J. Iai, D. Clarke, J. Joseph, and J. F. Bell III (2012c), Widespread Explosive Volcanism on Mars Inferred from the Global Distribution of Glass-Rich Sediments, in Geological Society of America, p. Paper No. 60–4, Geological Society of America, Charlotte, N.C.
- Horgan, B. H., J. F. Bell, E. Z. Noe Dobrea, E. A. Cloutis, D. T. Bailey, M. A. Craig, L. H. Roach, and J. F. Mustard (2009), Distribution of hydrated minerals in the north polar region of Mars, *J. Geophys. Res.*, 114(E1), doi:10.1029/2008JE003187.
- Horgan, B. H. N., and J. F. Bell (2012b), Seasonally active slipface avalanches in the north polar sand sea of Mars: Evidence for a wind-related origin, *Geophys. Res. Lett.*, 39(9), L09201, doi:10.1029/2012GL051329.
- Hughes, M. G., J. B. Keene, and R. G. Joseph (2000), Hydraulic Sorting of Heavy-Mineral Grains by Swash on a Medium-Sand Beach, *Journal of Sedimentary Research*, 70(5), 994–1004, doi:10.1306/112599700994.
- Hurowitz, J. A., S. M. McLennan, N. J. Tosca, R. E. Arvidson, J. R. Michalski, D. W. Ming, C. Schroder, and S. W. Squyres (2006), In situ and experimental evidence for acidic weathering of rocks and soils on Mars, 111(E2).
- Hynek, B. M., R. J. Phillips, and R. E. Arvidson (2003), Explosive volcanism in the Tharsis region: Global evidence in the Martian geologic record, *J. Geophys. Res.*, 108(E9), doi:10.1029/2003JE002062.
- Iversen, J. D., and K. R. Rasmussen (1994), The effect of surface slope on saltation threshold, *Sedimentology*, 41(4), 721–728, doi:10.1111/j.1365-3091.1994.tb01419.x.
- Iversen, J. D., and K. R. Rasmussen (1999), The effect of wind speed and bed slope on sand transport, *Sedimentology*, 46(4), 723–731, doi:10.1046/j.1365-3091.1999.00245.x.
- Iversen, J. D., and B. R. White (1982), Saltation threshold on Earth, Mars and Venus, *Sedimentology*, 29(1), 111–119, doi:10.1111/j.1365-3091.1982.tb01713.x.
- Jakosky, B. M. (1986), On the thermal properties of Martian fines, *Icarus*, 66, 117–124.
- Jakosky, B. M., B. M. Hynek, S. M. Pelkey, M. T. Mellon, S. Martinez-Alonso, N. E. Putzig, N. Murphy, and P. R. Christensen (2006), Thermophysical properties of the MER and Beagle II landing site regions on Mars, *J. Geophys. Res.*, 111(E8), E08008.

- Johnsson, M. J. (1993), The system controlling the composition of clastic sediments, Geological Society of America Special Paper, 284, 1–19.
- Kieffer, H. H., B. M. Jakosky, C. W. Snyder, and M. S. Matthews (Eds.) (1992), Mars, Space science series., University of Arizona Press, Tucson.
- King, T. V. V., and W. I. Ridley (1987a), Relation of the Spectroscopic Reflectance of Olivine to Mineral Chemistry and Some Remote Sensing Implications, *J. Geophys. Res.*, 92(B11), 11457–11469, doi:10.1029/JB092iB11p11457.
- King, T. V. V., and W. I. Ridley (1987b), Relation of the Spectroscopic Reflectance of Olivine to Mineral Chemistry and Some Remote-Sensing Implications, *Journal of Geophysical Research - Solid Earth and Planets*, 92(B11), 11457–11469.
- Kirk, R. L. et al. (2008), Ultrahigh resolution topographic mapping of Mars with MRO HiRISE stereo images: Meter-scale slopes of candidate Phoenix landing sites, *J. Geophys. Res.*, 113, doi:10.1029/2007JE003000.
- Kirk, R. L., E. Howington-Kraus, and M. R. Rosiek (2009), Build Your Own Topographic Model: A Photogrammetry Guest Facility for Planetary Researchers, in 40th Lunar and Planetary Science Conference, p. Abstract #1414, Lunar and Planetary Institute, Houston.
- Kocurek, G., and N. Lancaster (1999), Aeolian system sediment state: theory and Mojave Desert Kelso dune field example, *Sedimentology*, 46, 505–515, doi:10.1046/j.1365-3091.1999.00227.x.
- Kocurek, G., M. Carr, R. Ewing, K. G. Havholm, Y. C. Nagar, and A. K. Singhvi (2007), White Sands Dune Field, New Mexico: Age, dune dynamics and recent accumulations, *Sedimentary Geology*, 197(3-4), 313–331, doi:10.1016/j.sedgeo.2006.10.006.
- Kocurek, G., N. T. Bridges, K. S. Edgett, W. Goetz, K. W. Lewis, M. B. Madsen, D. M. Rubin, R. J. Sullivan, and MSL Science Team (2013), Rocknest Sand Shadow at the Curiosity Field Site: Morphology, Origin and Stabilization, p. Abstract #1375, Lunar and Planetary Institute, Houston.
- Kreslavsky, M. A., and J. W. Head (1999), Kilometer-scale slopes on Mars and their correlation with geologic units: Initial results from Mars Orbiter Laser Altimeter (MOLA) data, *J. Geophys. Res.*, 104(E9), 21911–21924, doi:10.1029/1999JE001051.
- Kreslavsky, M. A., and J. W. Head (2000), Kilometer-scale roughness of Mars: Results from MOLA data analysis, *J. Geophys. Res.*, 105(E11), 26695–26711, doi:10.1029/2000JE001259.
- Kreslavsky, M. A., and J. W. Head (2002), Mars: Nature and evolution of young latitude-dependent water-ice-rich mantle, *Geophysical Research Letters*, 29, doi:10.1029/2002GL015392.
- Lancaster, N. (2009), *Geomorphology of Desert Environments*, edited by A. J. Parsons and A. D. Abrahams, pp. 557–595, Springer Netherlands.
- Lancaster, N., and R. Greeley (1990), Sediment Volume in the North Polar Sand Seas of Mars, *J. Geophys. Res.*, 95(B7), 10921–10927, doi:10.1029/JB095iB07p10921.
- Lancaster, N., and V. P. Tchakerian (1996), Geomorphology and sediments of sand ramps in the Mojave desert, *Geomorphology*, 17(1-3), 151–165, doi:10.1016/0169-555X(95)00101-A.
- Langevin, Y. (2005), Summer Evolution of the North Polar Cap of Mars as Observed by OMEGA/Mars Express, *Science*, 307(5715), 1581–1584, doi:10.1126/science.1109438.
- Langevin, Y., F. Poulet, J.-P. Bibring, and B. Gondet (2005), Sulfates in the North Polar Region of Mars Detected by OMEGA/Mars Express, *Science*, 307(5715), 1584–1586, doi:10.1126/science.1109091.

- Lewis, K. W., O. Aharonson, J. P. Grotzinger, S. W. Squyres, J. F. Bell, L. S. Crumpler, and M. E. Schmidt (2008), Structure and stratigraphy of Home Plate from the Spirit Mars Exploration Rover, *Journal of Geophysical Research*, 113(E12), doi:10.1029/2007JE003025.
- Lichtenberg, K. A. et al. (2010), Stratigraphy of hydrated sulfates in the sedimentary deposits of Aram Chaos, Mars, *J. Geophys. Res.*, 115, doi:10.1029/2009JE003353.
- Lucchitta, B. (1999), Geologic map of Ophir and central Candor Chasmata (MTM -05072) of Mars, U.S. Geol. Surv. Misc. Invest. Ser. Map I-2568, scale 1:500,000.
- Lucchitta, B. K. (2010), 5 - Lakes in Valles Marineris, in *Lakes on Mars*, pp. 111–161, Elsevier, Amsterdam.
- Lucchitta, B. K., A. McEwen, G. D. Clow, P. Geissler, R. B. Singer, R. A. Schulz, and S. W. Squyres (1992), The Canyon System on Mars, in *Mars*, edited by H. H. Kieffer, B. M. Jakosky, C. W. Snyder, and M. S. Matthews, pp. 453–492, University of Arizona Press, Tucson.
- Malin, M. C., and K. S. Edgett (2000), Sedimentary rocks of early Mars, *Science*, 290(5498), 1927–1937.
- Malin, M. C., and K. S. Edgett (2001), Mars Global Surveyor Mars Orbiter Camera: Interplanetary cruise through primary mission, *J. Geophys. Res.*, 106(E10), 23429–23570, doi:10.1029/2000JE001455.
- Malin, M. C., K. S. Edgett, L. V. Posiolova, S. M. McColley, and E. Z. N. Dobrea (2006), Present-Day Impact Cratering Rate and Contemporary Gully Activity on Mars, *Science*, 314(5805), 1573–1577, doi:10.1126/science.1135156.
- Malin, M. C. et al. (2007), Context Camera Investigation on board the Mars Reconnaissance Orbiter, *J. Geophys. Res.*, 112(E5), doi:10.1029/2006JE002808.
- Mangold, N., A. Gendrin, B. Gondet, S. Lemouelic, C. Quantin, V. Ansan, J. Bibring, Y. Langevin, P. Masson, and G. Neukum (2008), Spectral and geological study of the sulfate-rich region of West Candor Chasma, Mars, *Icarus*, 194(2), 519–543, doi:10.1016/j.icarus.2007.10.021.
- Mangold, N., D. Baratoux, O. Arnalds, J.-M. Bardintzeff, B. Platevoet, M. Grégoire, and P. Pinet (2011), Segregation of olivine grains in volcanic sands in Iceland and implications for Mars, *Earth and Planetary Science Letters*, 310(3-4), 233–243, doi:10.1016/j.epsl.2011.07.025.
- Mauget, S. A., E. C. Cordero, and P. T. Brown (2011), Evaluating Modeled Intra- to Multidecadal Climate Variability Using Running Mann–Whitney Z Statistics, *J. Climate*, 25(5), 1570–1586, doi:10.1175/JCLI-D-11-00211.1.
- McEwen, A. et al. (2010), The High Resolution Imaging Science Experiment (HiRISE) during MRO's Primary Science Phase (PSP), *ICARUS*, 205(1), 2–37, doi:10.1016/j.icarus.2009.04.023.
- McEwen, A. S., M. C. Malin, M. H. Carr, and W. K. Hartmann (1999), Voluminous volcanism on early Mars revealed in Valles Marineris, *Nature*, 397(6720), 584–586, doi:10.1038/17539.
- McEwen, A. S. et al. (2007), Mars Reconnaissance Orbiter's High Resolution Imaging Science Experiment (HiRISE), *J. Geophys. Res.*, 112(E5), doi:10.1029/2005JE002605.
- McGlynn, I. O. (2012), Formation and Alteration of Basaltic Soils on Mars, PhD Dissertation, Univ. of Tenn., Knoxville, TN.
- McGlynn, I. O., C. M. Fedo, and H. Y. McSween (2011), Origin of basaltic soils at Gusev crater, Mars, by aeolian modification of impact-generated sediment, *J. Geophys. Res.*, 116, doi:10.1029/2010JE003712.

- McGlynn, I. O., C. M. Fedo, and H. Y. McSween (2012), Soil mineralogy at the Mars Exploration Rover landing sites: An assessment of the competing roles of physical sorting and chemical weathering, *J. Geophys. Res.*, 117(E1), E01006, doi:10.1029/2011JE003861.
- McGuire, P. C., J. L. Bishop, A. J. Brown, A. A. Fraeman, G. A. Marzo, M. Frank Morgan, S. L. Murchie, J. F. Mustard, M. Parente, and S. M. Pelkey (2009), An improvement to the volcano-scan algorithm for atmospheric correction of CRISM and OMEGA spectral data, *Planetary and Space Science*, 57(7), 809–815, doi:10.1016/j.pss.2009.03.007.
- McKee, E. D. (1979), Introduction to a study of global sand seas, in *A Study of Global Sand Seas*, U.S. Geol. Surv. Prof. Pap., 1052, 3–17.
- McKeown, N. K., J. L. Bishop, E. Z. Noe Dobrea, B. L. Ehlmann, M. Parente, J. F. Mustard, S. L. Murchie, G. A. Swayze, J.-P. Bibring, and E. A. Silver (2009), Characterization of phyllosilicates observed in the central Mawrth Vallis region, Mars, their potential formational processes, and implications for past climate, *J. Geophys. Res.*, 114, doi:10.1029/2008JE003301.
- McSween, H. Y., and K. Keil (2000), Mixing relationships in the Martian regolith and the composition of globally homogeneous dust, *Geochimica et Cosmochimica Acta*, 64(12), 2155–2166, doi:10.1016/S0016-7037(99)00401-9.
- McSween, H. Y. et al. (2004), Basaltic Rocks Analyzed by the Spirit Rover in Gusev Crater, *Science*, 305(5685), 842–845, doi:10.1126/science.3050842.
- McSween, H. Y. et al. (2006), Characterization and petrologic interpretation of olivine-rich basalts at Gusev Crater, Mars, *J. Geophys. Res.*, 111(E2), E02S10.
- McSween, H. Y., G. J. Taylor, and M. B. Wyatt (2009), Elemental Composition of the Martian Crust, *Science*, 324(5928), 736–739, doi:10.1126/science.1165871.
- McSween, H. Y., I. O. McGlynn, and A. D. Rogers (2010), Determining the modal mineralogy of Martian soils, *J. Geophys. Res.*, 115, doi:10.1029/2010JE003582.
- Mellon, M. T., B. M. Jakosky, H. H. Kieffer, and P. R. Christensen (2000), High-resolution thermal inertia mapping from the Mars Global Surveyor Thermal Emission Spectrometer, *Icarus*, 148(2), 437–455.
- Mellon, M. T., R. L. Fergason, and N. E. Putzig (2008), The thermal inertia of the surface of Mars, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell, pp. 399–427, Cambridge University Press, Cambridge.
- Michaels, T. I. (2011), *Aeolian Phenomena at Nili and Meroe Paterae*, vol. XLII, p. Abstract 2697.
- Michalski, J. R., M. D. Kraft, T. G. Sharp, and P. R. Christensen (2006), Effects of chemical weathering on infrared spectra of Columbia River Basalt and spectral interpretations of martian alteration, *Earth and Planetary Science Letters*, 248(3–4), 822–829, doi:10.1016/j.epsl.2006.06.034.
- Minitti, M. E., C. M. Weitz, M. D. Lane, and J. L. Bishop (2007), Morphology, chemistry, and spectral properties of Hawaiian rock coatings and implications for Mars, *J. Geophys. Res.*, 112(E5), E05015, doi:10.1029/2006JE002839.
- Modi, A. L. (2011), Processes controlling the composition of first-cycle sediments deposited in an arid-climate, with implications for provenance reconstruction studies, M.S. Thesis, Univ. of Tenn, Knoxville, TN.
- Moore, H. J., D. B. Bickler, J. A. Crisp, H. J. Eisen, J. A. Gensler, A. F. C. Haldemann, J. R. Matijevic, L. K. Reid, and F. Pavlics (1999), Soil-like deposits observed by Sojourner, the

- Pathfinder rover, *Journal of Geophysical Research*, 104(E4), 8729, doi:10.1029/1998JE900005.
- Moroz, L. V. et al. (2009), Spectral properties of simulated impact glasses produced from martian soil analogue JSC Mars-1, *Icarus*, 202(1), 336–353, doi:10.1016/j.icarus.2009.02.007.
- Muhs, D. R. (2004), Mineralogical maturity in dunefields of North America, Africa and Australia, *Geomorphology*, 59(1-4), 247–269, doi:10.1016/j.geomorph.2003.07.020.
- Muhs, D. R., R. L. Reynolds, J. Been, and G. Skipp (2003), Eolian sand transport pathways in the southwestern United States: importance of the Colorado River and local sources, *Quaternary International*, 104(1), 3–18, doi:10.1016/S1040-6182(02)00131-3.
- Murchie, S. et al. (2007), Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) on Mars Reconnaissance Orbiter (MRO), *J. Geophys. Res.*, 112(E5), doi:10.1029/2006JE002682.
- Murchie, S. et al. (2009a), Evidence for the origin of layered deposits in Candor Chasma, Mars, from mineral composition and hydrologic modeling, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003343.
- Murchie, S. L. et al. (2009b), A synthesis of Martian aqueous mineralogy after 1 Mars year of observations from the Mars Reconnaissance Orbiter, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003342.
- Murchie, S. L. et al. (2009c), Compact Reconnaissance Imaging Spectrometer for Mars investigation and data set from the Mars Reconnaissance Orbiter's primary science phase, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003344.
- Mustard, J. F., F. Poulet, A. Gendrin, J. P. Bibring, Y. Langevin, B. Gondet, N. Mangold, G. Bellucci, and F. Altieri (2005), Olivine and pyroxene, diversity in the crust of Mars, *Science*, 307(5715), 1594–1597.
- Mustard, J. F. et al. (2007), CRISM-OMEGA Observations of Phyllosilicate-Olivine Stratigraphy in Nili Fossae, Mars, in 38th Lunar and Planetary Science Conference, p. Abstract #2071, Lunar and Planetary Institute, League City, Texas.
- Nedell, S., S. W. Squyers, and D. W. Anderson (1987), Origin and evolution of the layered deposits in the Valles Marineris, Mars, *Icarus*, 70(3), 409–414, doi:10.1016/0019-1035(87)90086-8.
- Nesse, W. D. (2012), *Introduction to mineralogy*, Oxford University Press, New York.
- Neukum, G., and et. al. (2004), HRSC: The High Resolution Stereo Camera of Mars Express, *Eur. Space Agency Spec. Publ.*, 1240, 17–35.
- Orofino, V., R. Politi, A. Blanco, and S. Fonti (2006), Diffuse reflectance of altered olivine grains: Remote sensing detection and implications for Mars studies, *Planetary and Space Science*, in press.
- Parente, M. (2008), A New Approach to Denoising CRISM Images, p. Abstract #2528, Lunar and Planetary Institute, Houston.
- Parente, M., H. D. Makarewicz, and J. L. Bishop (2011), Decomposition of mineral absorption bands using nonlinear least squares curve fitting: Application to Martian meteorites and CRISM data, *Planetary and Space Science*, 59(5-6), 423–442, doi:10.1016/j.pss.2011.01.009.
- Parsons, A. J., and A. D. Abrahams (Eds.) (2009), *Geomorphology of Desert Environments*, Springer Netherlands, Dordrecht.
- Pelkey, S. M. et al. (2007), CRISM multispectral summary products: Parameterizing mineral diversity on Mars from reflectance, *J. Geophys. Res.*, 112(E8), doi:10.1029/2006JE002831.
- Pelletier, J. D., and C. Rasmussen (2009), Quantifying the climatic and tectonic controls on hillslope steepness and erosion rate, *Lithosphere*, 1(2), 73–80, doi:10.1130/L3.1.

- Piqueux, S., and P. R. Christensen (2009), A model of thermal conductivity for planetary soils: 1. Theory for unconsolidated soils, *J. Geophys. Res.*, 114(E9), doi:10.1029/2008JE003308.
- Poulet, F., and S. Erard (2004), Nonlinear spectral mixing: Quantitative analysis of laboratory mineral mixtures, *J. Geophys. Res.-Planets*, 109(E2).
- Poulet, F., N. Mangold, D. Loizeau, J. P. Bibring, Y. Langevin, J. Michalski, and B. Gondet (2008a), Abundance of minerals in the phyllosilicate-rich units on Mars, *Astronomy & Astrophysics*, 487(2), L41–U193, doi:10.1051/0004-6361:200810150.
- Poulet, F., R. E. Arvidson, C. Gomez, R. V. Morris, J.-P. Bibring, Y. Langevin, B. Gondet, and J. Griffes (2008b), Mineralogy of Terra Meridiani and western Arabia Terra from OMEGA/MEx and implications for their formation, *Icarus*, 195(1), 106–130, doi:10.1016/j.icarus.2007.11.031.
- Poulet, F., J.-P. Bibring, Y. Langevin, J. F. Mustard, N. Mangold, M. Vincendon, B. Gondet, P. Pinet, J.-M. Bardintzeff, and B. Platevoet (2009), Quantitative compositional analysis of martian mafic regions using the MEx/OMEGA reflectance data 1. Methodology, uncertainties and examples of application, *Icarus*, 201(1), 69–83, doi:10.1016/j.icarus.2008.12.025.
- Presley, M. A., and P. R. Christensen (1997), Thermal conductivity measurements of particulate materials 2. Results, *J. Geophys. Res.*, 102(E3), 6551–6566, doi:10.1029/96JE03303.
- Presley, M. A., and R. A. Craddock (2006), Thermal conductivity measurements of particulate materials: 3. Natural samples and mixtures of particle sizes, *J. Geophys. Res.*, 111(E9), doi:10.1029/2006JE002706.
- Presley, M. A., R. A. Craddock, and N. Zolotova (2009), The effect of salt crust on the thermal conductivity of one sample of fluvial particulate materials under Martian atmospheric pressures, *J. Geophys. Res.*, 114(E11), doi:10.1029/2009JE003355.
- Putzig, N., and M. Mellon (2007a), Apparent thermal inertia and the surface heterogeneity of Mars, *Icarus*, 191(1), 68–94, doi:10.1016/j.icarus.2007.05.013.
- Putzig, N., and M. Mellon (2007b), Thermal behavior of horizontally mixed surfaces on Mars, *Icarus*, 191, 52–67, doi:10.1016/j.icarus.2007.03.022.
- Pye, K., and H. Tsoar (1990), *Aeolian sand and sand dunes*, Unwin Hyman, London.
- Quantin, C., P. Allemand, N. Mangold, and C. Delacourt (2004), Ages of Valles Marineris (Mars) landslides and implications for canyon history, *Icarus*, 172(2), 555–572, doi:10.1016/j.icarus.2004.06.013.
- Quantin, C., J. Flahaut, H. Clenet, P. Allemand, and P. Thomas (2012), Composition and structures of the subsurface in the vicinity of Valles Marineris as revealed by central uplifts of impact craters, *Icarus*, 221(1), 436–452, doi:10.1016/j.icarus.2012.07.031.
- Rafkin, S., R. M. Haberle, and T. I. Michaels (2001), The Mars Regional Atmospheric Modeling System: Model Description and Selected Simulations, *Icarus*, 151(2), 228–256, doi:10.1006/icar.2001.6605.
- Rafkin, S. C. R., and T. I. Michaels (2003), Meteorological predictions for 2003 Mars Exploration Rover high-priority landing sites, *J. Geophys. Res.*, 108(E12), doi:10.1029/2002JE002027.
- Ramsey, M. S., P. R. Christensen, N. Lancaster, and D. A. Howard (1999), Identification of sand sources and transport pathways at the Kelso Dunes, California, using thermal infrared remote sensing, *Geological Society of America Bulletin*, 111(5), 646–662, doi:10.1130/0016-7606(1999)111<0646:IOSSAT>2.3.CO;2.
- Raupach, M. R., and J. J. Finnigan (1997), The influence of topography on meteorological variables and surface-atmosphere interactions, *Journal of Hydrology*, 190, 182–213.

- Roach, L. H. et al. (2007), CRISM Spectral Signatures of the North Polar Gypsum Dunes, p. Abstract #1970, Lunar and Planetary Institute, Houston.
- Roach, L. H., J. F. Mustard, S. L. Murchie, J. -P. Bibring, F. Forget, K. W. Lewis, O. Aharonson, M. Vincendon, and J. L. Bishop (2009), Testing evidence of recent hydration state change in sulfates on Mars, *J. Geophys. Res.*, 114, doi:10.1029/2008JE003245.
- Roach, L. H., J. F. Mustard, M. D. Lane, J. L. Bishop, and S. L. Murchie (2010a), Diagenetic haematite and sulfate assemblages in Valles Marineris, *Icarus*, 207(2), 659–674, doi:10.1016/j.icarus.2009.11.029.
- Roach, L. H., J. F. Mustard, G. Swayze, R. E. Milliken, J. L. Bishop, S. L. Murchie, and K. Lichtenberg (2010b), Hydrated mineral stratigraphy of Ius Chasma, Valles Marineris, *Icarus*, 206(1), 253–268, doi:10.1016/j.icarus.2009.09.003.
- Rogers, D., and P. R. Christensen (2003), Age relationship of basaltic and andesitic surface compositions on Mars: Analysis of high-resolution TES observations of the northern hemisphere, *J. Geophys. Res.*, 108(E4), doi:10.1029/2002JE001913.
- Rossi, A. P., G. Neukum, M. Pondrelli, S. van Gasselt, T. Zegers, E. Hauber, A. Chicarro, and B. Foing (2008), Large-scale spring deposits on Mars?, *J. Geophys. Res.*, 113(E8), doi:10.1029/2007JE003062.
- Ruff, S. W., and P. R. Christensen (2002), Bright and dark regions on Mars: Particle size and mineralogical characteristics based on Thermal Emission Spectrometer data, *J. Geophys. Res.*, 107, doi:10.1029/2001JE001580.
- Ruff, S. W., and P. R. Christensen (2007), Basaltic andesite, altered basalt, and a TES-based search for smectite clay minerals on Mars, *Geophysical Research Letters*, 34(10), doi:10.1029/2007GL029602.
- Russell, P. et al. (2008), Seasonally active frost-dust avalanches on a north polar scarp of Mars captured by HiRISE, *Geophys. Res. Lett.*, 35(23), L23204, doi:10.1029/2008GL035790.
- Sagan, C. et al. (1973), Variable Features on Mars, 2, Mariner 9 Global Results, *J. Geophys. Res.*, 78(20), 4163–4196, doi:10.1029/JB078i020p04163.
- Salvatore, M. R., J. F. Mustard, M. B. Wyatt, and S. L. Murchie (2010), Definitive evidence of Hesperian basalt in Acidalia and Chryse planitiae, *J. Geophys. Res.*, 115(E7), doi:10.1029/2009JE003519.
- Schultz, R. A., and J. Lin (2001), Three-dimensional normal faulting models of the Valles Marineris, Mars, and geodynamic implications, *J. Geophys. Res.*, 106(B8), 16549–16566, doi:10.1029/2001JB000378.
- Scott, D. H., and K. L. Tanaka (1986), Geology map of the western equatorial region of Mars, U.S. Geol. Surv. Misc. Invest. Map, I-1802-A, scale 1:15,000,000.,
- Seelos, F. P., S. L. Murchie, D. C. Humm, O. S. Barnouin, F. Morgan, H. W. Taylor, C. Hash, and CRISM Team (2011), CRISM Data Processing and Analysis Products Update: Calibration, Correction, and Visualization, p. Abstract #1438, Lunar and Planetary Institute, Houston.
- Silvestro, S., G. Di Achille, and G. G. Ori (2010a), Dune morphology, sand transport pathways and possible source areas in east Thaumasia Region (Mars), *Geomorphology*, 121(1-2), 84–97, doi:10.1016/j.geomorph.2009.07.019.
- Silvestro, S., L. K. Fenton, D. A. Vaz, N. T. Bridges, and G. G. Ori (2010b), Ripple migration and dune activity on Mars: Evidence for dynamic wind processes, *Geophys. Res. Lett.*, 37(20), L20203.

- Silvestro, S., D. A. Vaz, L. K. Fenton, and P. E. Geissler (2011), Active aeolian processes on Mars: A regional study in Arabia and Meridiani Terrae, *Geophysical Research Letters*, 38, doi:10.1029/2011GL048955.
- Silvestro, S., L. K. Fenton, T. I. Michaels, A. Valdez, and G. G. Ori (2012a), Interpretation of the complex dune morphology on Mars: dune activity, modelling and a terrestrial analogue, *Earth Surface Processes and Landforms*, 37(13), 1424–1436, doi:10.1002/esp.3286.
- Silvestro, S., L. K. Fenton, T. I. Michaels, A. Valdez, and G. G. Ori (2012b), Interpretation of the complex dune morphology on Mars: dune activity, modelling and a terrestrial analogue, *Earth Surf. Process. Landforms*, n/a, doi:10.1002/esp.3286.
- Skinner, J. A., T. M. Hare, and K. L. Tanaka (2006), Digital Renovation of the Atlas of Mars 1:15,000,000-Scale Global Geologic Series Maps, p. Abstract #2331.
- Smith, D. E. et al. (2001), Mars Orbiter Laser Altimeter: Experiment summary after the first year of global mapping of Mars, *J. Geophys. Res.*, 106(E10), 23689–23722, doi:10.1029/2000JE001364.
- Smith, M. D., B. J. Conrath, J. C. Pearl, and P. R. Christensen (2002), Thermal Emission Spectrometer Observations of Martian Planet-Encircling Dust Storm 2001A, *Icarus*, 157(1), 259–263, doi:10.1006/icar.2001.6797.
- Soderblom, L., and J. Bell (2008), Introduction and Historical Perspective: 1. Exploration of the Martian surface: 1992–2007, in *The Martian Surface: Composition, Mineralogy, and Physical Properties*, edited by J. Bell, pp. 3–20, Cambridge University Press, Cambridge.
- Solomon, S. C. et al. (2005), New Perspectives on Ancient Mars, *Science*, 307(5713), 1214–1220, doi:10.1126/science.1101812.
- Spirn, A. W. (1986), *Air Quality at the Street-Level: Strategies for Urban Design*, Cambridge: Harvard Graduate School of Design.
- Squyres, S. W. et al. (2004), The Opportunity Rover's Athena science investigation at Meridiani Planum, Mars, *Science*, 306(5702), 1698–1703.
- Stockstill-Cahill, K. R., F. S. Anderson, and V. E. Hamilton (2008), A study of low-albedo deposits within Amazonis Planitia craters: Evidence for locally derived ultramafic to mafic materials, *J. Geophys. Res.*, 113(E7), doi:10.1029/2007JE003036.
- Stopar, J. D., G. Jeffrey Taylor, V. E. Hamilton, and L. Browning (2006), Kinetic model of olivine dissolution and extent of aqueous alteration on mars, *Geochimica et Cosmochimica Acta*, 70(24), 6136–6152, doi:10.1016/j.gca.2006.07.039.
- Sullivan, R. et al. (2005), Aeolian processes at the Mars Exploration Rover Meridiani Planum landing site, *Nature*, 436(7047), 58–61, doi:10.1038/nature03641.
- Sullivan, R. et al. (2008), Wind-driven particle mobility on Mars: Insights from Mars Exploration Rover observations at “El Dorado” and surroundings at Gusev Crater, *J. Geophys. Res.*, 113(E6), doi:10.1029/2008JE003101.
- Thomas, P., and C. Weitz (1989), Sand dune materials and polar layered deposits on Mars, *Icarus*, 81(1), 185–215, doi:10.1016/0019-1035(89)90133-4.
- Tirsch, D. (2009), Dark aeolian sediments in Martian craters: Composition and sources, Ph.D. dissertation, Freie Univ., Berlin.
- Tirsch, D., R. Jaumann, A. Pacifici, and F. Poulet (2011), Dark aeolian sediments in Martian craters: Composition and sources, *J. Geophys. Res.*, 116(E3), doi:10.1029/2009JE003562.
- Tirsch, D., R. A. Craddock, T. Platz, A. Maturilli, J. Helbert, and R. Jaumann (2012), Spectral and petrologic analyses of basaltic sands in Ka’u Desert (Hawaii) - implications for the dark

- dunes on Mars, *Earth Surface Processes and Landforms*, 37(4), 434–448, doi:10.1002/esp.2266.
- Toigo, A. D., and M. I. Richardson (2003), Meteorology of proposed Mars Exploration Rover landing sites, *J. Geophys. Res.*, 108, doi:10.1029/2003JE002064.
- Tornabene, L. L., J. E. Moersch, H. Y. McSween, A. S. McEwen, J. L. Piatek, K. A. Milam, and P. R. Christensen (2006), Identification of large (2–10 km) rayed craters on Mars in THEMIS thermal infrared images: Implications for possible Martian meteorite source regions, *Journal of Geophysical Research*, 111(E10), doi:10.1029/2005JE002600.
- Tsoar, H., B. White, and E. Berman (1996), The effect of slopes on sand transport — numerical modelling, *Landscape and Urban Planning*, 34(3–4), 171–181, doi:10.1016/0169-2046(95)00235-9.
- Vaniman, D. T., D. L. Bish, S. J. Chipera, C. I. Fialips, J. William Carey, and W. C. Feldman (2004), Magnesium sulphate salts and the history of water on Mars, *Nature*, 431(7009), 663–665, doi:10.1038/nature02973.
- Vardoulakis, S., B. E. Fisher, K. Pericleous, and N. Gonzalez-Flesca (2003), Modelling air quality in street canyons: a review, *Atmospheric Environment*, 37(2), 155–182, doi:10.1016/S1352-2310(02)00857-9.
- Wiggs, G., J. Bullard, B. Garvey, and I. Castro (2002), Interactions Between Airflow and Valley Topography with Implications for Aeolian Sediment Transport, *Physical Geography*, 23(5), 366–380, doi:10.2747/0272-3646.23.5.366.
- Williams, J.-P., D. A. Paige, and C. E. Manning (2003), Layering in the wall rock of Valles Marineris: intrusive and extrusive magmatism, *Geophys. Res. Lett.*, 30(12), doi:10.1029/2003GL017662.
- Wilson, I. G. (1972), Universal discontinuities in bedforms produced by the wind, *Journal of Sedimentary Research*, 42(3), 667–669, doi:10.1306/74D725EF-2B21-11D7-8648000102C1865D.
- Wilson, I. G. (1973), *Ergs*, *Sedimentary Geology*, 10(2), 77–106, doi:10.1016/0037-0738(73)90001-8.
- Witbeck, N. E., K. L. Tanaka, and D. H. Scott (1991), *Geologic Map of Valles Marineris Region, Mars, MAP I-2010*, U.S. Geol. Surv., Reston, Va.,
- Wolff, M. J., M. D. Smith, R. T. Clancy, R. Arvidson, M. Kahre, F. Seelos, S. Murchie, and H. Savijärvi (2009), Wavelength dependence of dust aerosol single scattering albedo as observed by the Compact Reconnaissance Imaging Spectrometer, *J. Geophys. Res.*, 114, doi:10.1029/2009JE003350.
- Wyatt, M. B., and H. Y. McSween (2002), Spectral evidence for weathered basalt as an alternative to andesite in the northern lowlands of Mars, *Nature*, 417(6886), 263–266, doi:10.1038/417263a.
- Xu, Z. X., K. Takeuchi, and H. Ishidaira (2003), Monotonic trend and step changes in Japanese precipitation, *Journal of Hydrology*, 279(1–4), 144–150, doi:10.1016/S0022-1694(03)00178-1.
- Yen, A. S. et al. (2005), An integrated view of the chemistry and mineralogy of martian soils, *Nature*, 436(7047), 49–54, doi:10.1038/nature03637.
- Yen, A. S. et al. (2013), Evidence for a Global Martian Soil Composition Extends to Gale Crater, p. Abstract #2495, Lunar and Planetary Institute, Houston.
- Yue, S., and C. Wang (2002), The influence of serial correlation on the Mann–Whitney test for detecting a shift in median, *Advances in Water Resources*, 25(3), 325–333, doi:10.1016/S0309-1708(01)00049-5.

Zimbelman, J. R. (2010), Transverse Aeolian Ridges on Mars: First results from HiRISE images, *Geomorphology*, 121(1-2), 22–29, doi:10.1016/j.geomorph.2009.05.012.

Appendix V

Table 5.1 Extra-Rift CRISM Observations Used in This Study^a.

CRISM Targeted	CRISM Multispectral
FRT128F3	MSP3717_07
FRT148DB	MSP55CD_03
FRT16A9E	MSP55CD_03
FRT8236	MSP117BD_01
FRT9620	MSP8D60_01
FRT9905	
FRTCA5B	
FRTCF49	
HRL18642	

^aObservation type and resolution are as follows: full (FRT) 18 m/pixel, half resolution targeted (HRL) 36 m/pixel, half resolution short targeted (HRS) 36 m/pixel, multispectral window (MSW) 200 m/pixel, and multispectral survey (MSP) 100 m/pixel observation.

Table 5.2 Regional Extra-Rift Dune Fields Used in This Study.

Dune ID ^a	Region ^b	Area (km ²) ^c	Class
3012+512	Lunae Palus	24	Extra-Rift Dune Field
3035-746	N. Coprates	7	Extra-Rift Dune Field
3036-552	N. Coprates	7	Extra-Rift Dune Field
3040+022	Lunae Palus	5	Extra-Rift Dune Field
3047-895	N. Coprates	10	Extra-Rift Dune Field
3049+893	Lunae Palus	10	Extra-Rift Dune Field
3054+092	Lunae Palus	64	Extra-Rift Dune Field
3061+643	Lunae Palus	20	Extra-Rift Dune Field
3070-343	N. Coprates	7	Extra-Rift Dune Field
3074-094	N. Coprates	10	Extra-Rift Dune Field
3074-019	N. Coprates	20	Extra-Rift Dune Field
3076-304	N. Coprates	72	Extra-Rift Dune Field
3080+208	Lunae Palus	31	Extra-Rift Dune Field
3088-281	S. Coprates	38	Extra-Rift Dune Field
3134-018	N. Coprates	690	Extra-Rift Dune Field
3144-235	S. Coprates	5	Extra-Rift Dune Field
3204-217	Margaritifer Sinus	207	Extra-Rift Dune Field
3212-027	Margaritifer Sinus	64	Extra-Rift Dune Field
3229-271	Margaritifer Sinus	9	Extra-Rift Dune Field
3229-183	Margaritifer Sinus	2	Extra-Rift Dune Field
3240-189	Margaritifer Sinus	5	Extra-Rift Dune Field
3264-251	Margaritifer Sinus	322	Extra-Rift Dune Field
		1629	
2610-700	Noctis Labyrinthus	6	Intra-Rift Dune Field
2639-727	Noctis Labyrinthus	26	Intra-Rift Dune Field
2672-769	Noctis Labyrinthus	18	Intra-Rift Dune Field
2679-660	Noctis Labyrinthus	3	Intra-Rift Dune Field
3210-832	Aurorae Chaos	128	Intra-Rift Dune Field
3212-769	Aurorae Chaos	143	Intra-Rift Dune Field
3266-844	Aurorae Chaos	16	Intra-Rift Dune Field
3282-696	Aurorae Chaos	11	Intra-Rift Dune Field
3298-735	Aurorae Chaos	66	Intra-Rift Dune Field
Total		417	

^aDune field IDs where the first four digits are the dune field's centroid East longitude, last three digits are the dune field's latitude (both without a decimal), and the separating + or – sign indicating which hemisphere. This convention was also used for Valles Marineris Dune Fields.

^bRegions where the dune fields are located in.

^cArea of dune field polygons, as calculated by ENVI using MOLA-gridded data.

Table 5.3 Dune Field Potential Sand Source Locations and Associated Evidence.

Dune Field	Location	PSS ^a	Photogeologic ^b	Composition ^c	Topography ^d	Paleowind ^e	TP ^f	Longitude	Latitude	Elevation (m)
2821-089	SE Ius	LW/UW(?); 1	x		x	x		281.92	-9.34	-1000
2823-078_LS	NE Ius	LS; 2	x	x		x		282.38	-7.75	-2700
2845-155_CD	SE Hebes	DF (2846-126); 1	x	x	x			284.44	-1.31	-3900
2886-133_CD	SE Melas	DF (2886-131); 1		x		x		288.58	-13.23	-3400
2894-099	NE Melas	LW; 1	x	x	x			289.79	-9.64	-1000
2901-101_LS	NE Melas	LS/LW(?); 3	x	x	x	x	x	290.27	-10.13	-2567
2908-105	NE Melas	LW; 4	x	x	x	x		290.88	-10.34	-2000
2910-107_LS	NE Melas	LS; 2	x	x	x	x		291.43	-10.68	-3000
2971-046_ERG	Juventae	ILD, LW; 5	x	x	x	x	x	298.12	-3.67	-1720
2979-331	N Juventae	ILD; 2	x	x	x		x	298.32	-3.40	-1150
2984-347	N Juventae	ILD; 3	x	x	x	x		298.30	-3.67	-1800
3018-142	Coprates	DF (3019-144_FD); 3	x	x	x	x		302.08	-14.38	-4700
3019-144_FD	Coprates	UW; 4	x	x	x	x	x	301.86	-14.51	-750
3023-153	S of Coprates	LW; 4	x	x	x	x	x	302.25	-15.20	-1625
3031-149_LS	Coprates	LS/LW(?); 1	x	x				303.14	-14.90	-4800
3038-143_FD	Coprates massif	UW; 1	x	x	x	x		303.86	-14.31	-2200
3043-150	Coprates massif	LW; 1	x	x	x	x	x	304.53	-14.78	-3500
3049-148	Coprates massif	LW; 1	x	x	x		x	304.99	-14.87	-2800
3057-145_FD	Coprates massif	UW; 3	x	x	x	x		305.58	-14.58	-967
3062-148_FD	Coprates massif	UW; 2	x	x	x	x		306.12	-14.78	-1050
3064-143	NE Coprates	LW; 3	x	x	x	x	x	306.75	-14.20	-3700
3067-155	SE Coprates	LW; 2	x	x	x			306.65	-15.56	-3600
3107-068	N Ganges	ILD; 2	x	x	x			310.58	-7.20	-1455

Table 5.3 (continued)

Dune Field	Location	PSS ^a	Photogeologic ^b	Composition ^c	Topography ^d	Paleowind ^e	TP ^f	Longitude	Latitude	Elevation (m)
3113-068	N Ganges	ILD; 1	x	x	x	x		311.35	-7.17	-600
3116-131	Capri	ILD; 2	x	x	x			311.66	-13.10	-1900
3123-132	Capri	ILD; 1	x	x	x			312.23	-13.32	-1900
3135-127	Capri	ILD; 4	x	x	x	x		313.71	-12.79	-1656
3138-869_ERG	E Ganges	ILD; 3	x	x	x	x		313.62	-8.48	-3500

^aPotential Sediment Sources: upper walls (UW), lower walls (LW), landslide (LS), interior layered deposits (ILD), dune fields (DF, source dune field ID is provided), where the following number refers to the number of occurrence. A “?” indicates some ambiguity between PSSs. For PSS with multiple occurrences an average latitude, longitude, and elevation are given. Positive evidence in columns b–f are indicated by an “x”. Compare with Fig. 5.2.

^bPhotogeologic evidence of dark-toned sediment production found within HiRISE/CTX images.

^cA spectral match was found with CRISM data.

^dThe topographic gradient is such that gravity will assist sediment from the PSS downhill toward dunes.

^ePaleowind indicators (e.g., slip faces, dark streaks) indicated transport away from the PSS.

^fThermophysical properties (TP) indicate PSS have greater thermal inertia than dunes and dune thermal inertias decrease moving away from PSS.

Figure 5.1 MOC wide-angle mosaics colorized with MOLA topography of the distribution of dune fields surrounding and within Valles Marineris.

(a) A regional view of the terrain in and around VM. The 1:5 million-quadrangle regions are shown with the black rectangles and locations of interest are provided with italic text. (b) A closer view of the VM distribution of dune fields and their classification.

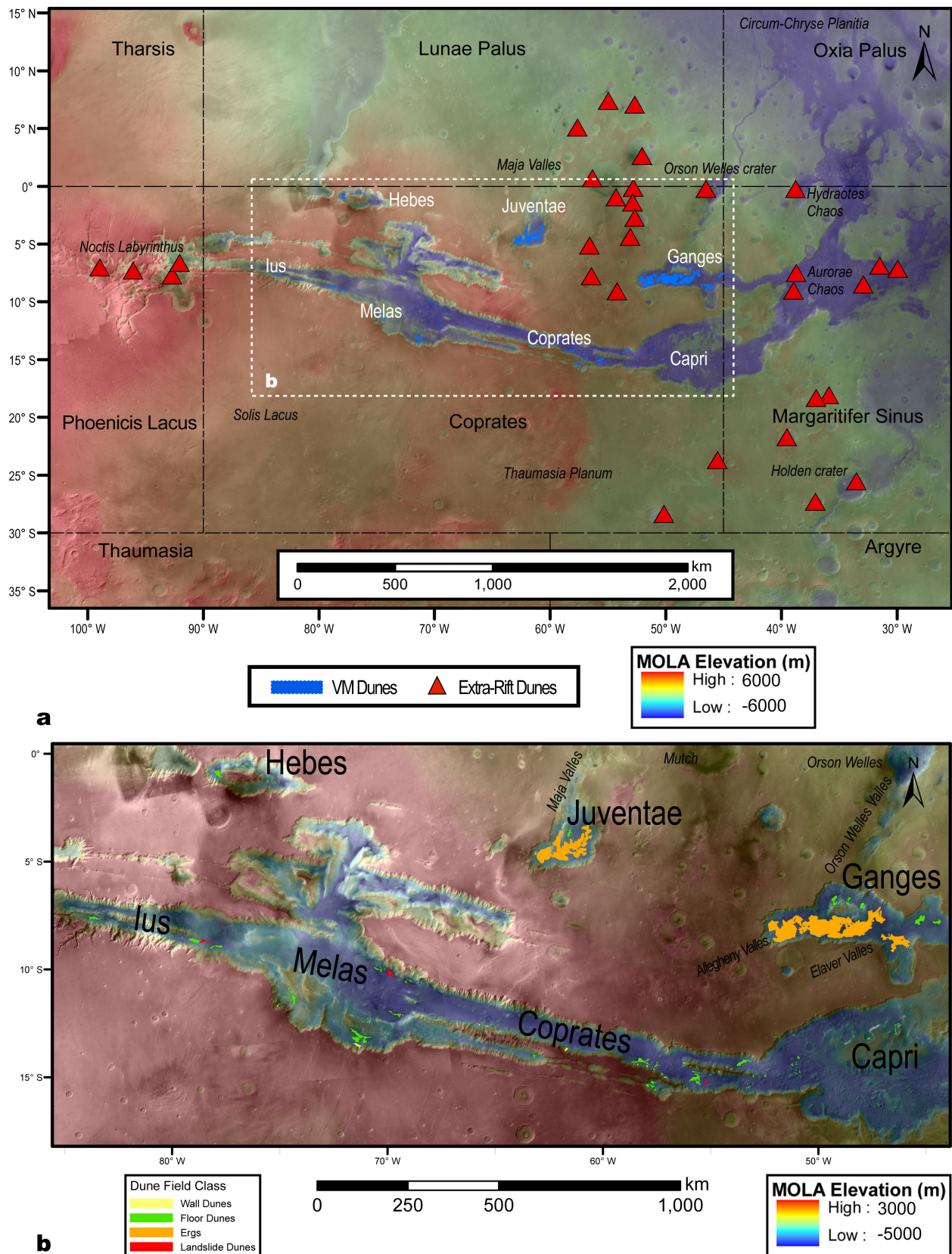


Figure 5.1

Figure 5.2 MOC wide-angle mosaics of Valles Marineris dune fields (blue), potential sediment sources, and dune compositions.

(a) A MOC wide-angle mosaic where context boxes are provided for remaining relevant figures. Dashed arrows represent perspective views. Colored circles represent PSSs classified in Table 5.3. (b) A MOC wide-angle mosaic of dune field compositional classes as determined with CRISM. CRISM observations are listed in Table 4.2 of Chapter IV. Compare with Chapter IV, Fig. 4.17.

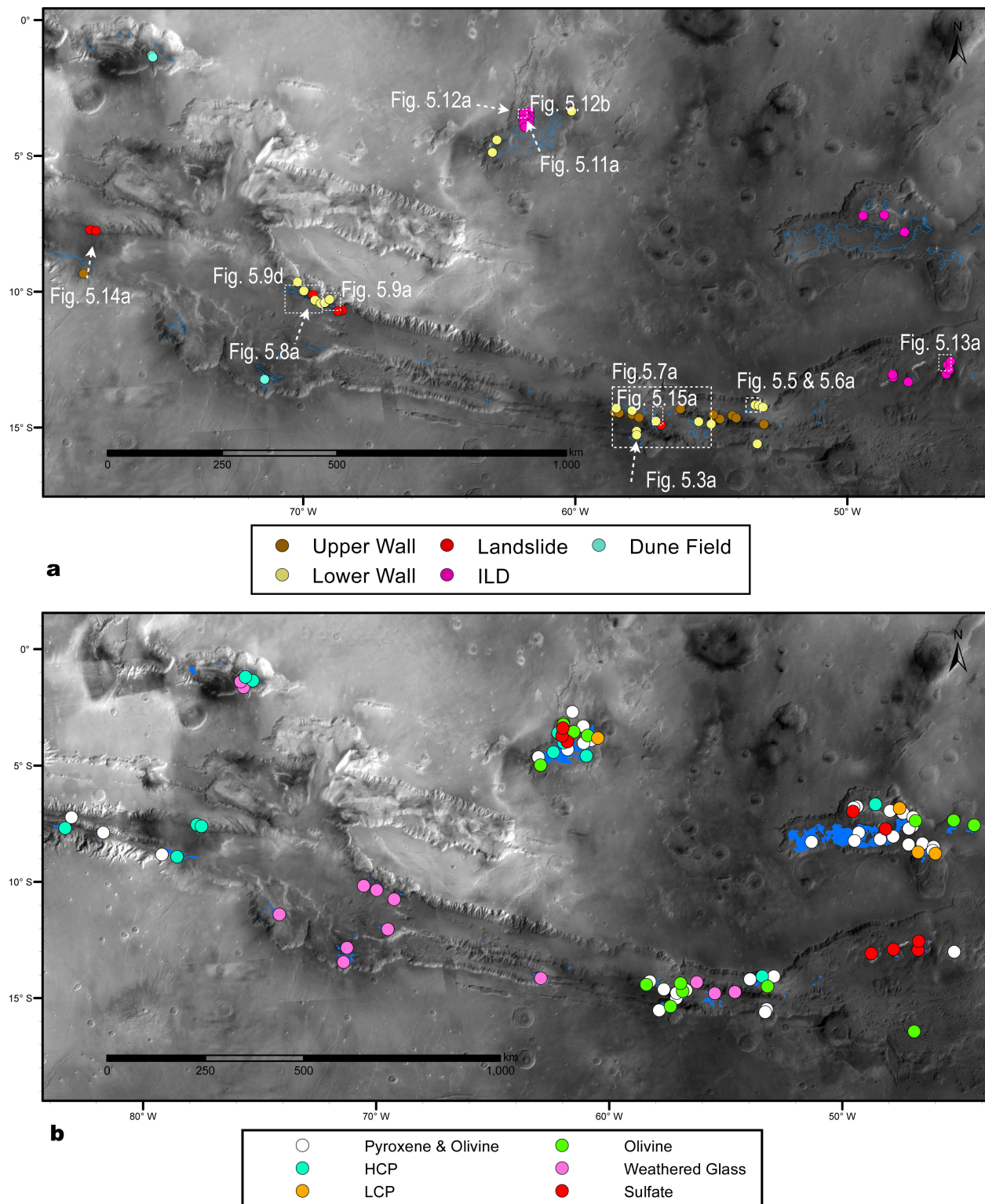


Figure 5.2

Figure 5.3 Dune and wall PSS relationships in collapse pits south of Coprates Chasma.

(a) A CTX perspective view using a HRSC DEM (orbit 1918, 75 m/pixel) colorized with CRISM FRT1FD85 of lower wall outcrops, talus, and dunes, where Fe-Mg phyllosilicate features appear as magenta (D2300), olivine as green (OLINDEX), and LCP as blue (LCPINDEX). (lower inset) Matching olivine spectra of lower spurs, talus, and dunes are shown. Also see Fig. 5.4a. (upper left inset) Spectra of wall spurs with absorptions that are diagnostic of various smectite clays. Vertical lines are placed at 1.4, 1.92, 2.2, and 2.3 μm . Slip face orientations and putative sediment sources/pathways are shown. (upper right inset) HRSC profiles show a consistent downslope trend along A–A' and steep slopes above dunes. The vertical lines in that plot indicate locations where the profile crosses wall rock-talus contacts (diamond). Annotations are the same for Figs. 5.6a, 5.9a, 5.9d, 5.11a, 5.12a, 5.13a, 5.14a, and 5.15a. (b–d) HiRISE ESP_023806_1645 grayscale (RED) and enhanced false-color (IRB) views of select locations. (d) Boulder tracks on steep $\sim 30\text{--}40^\circ$ talus slopes and occasionally boulders are found within the dune field (Chapter IV). North is toward the top in all images unless otherwise indicated.

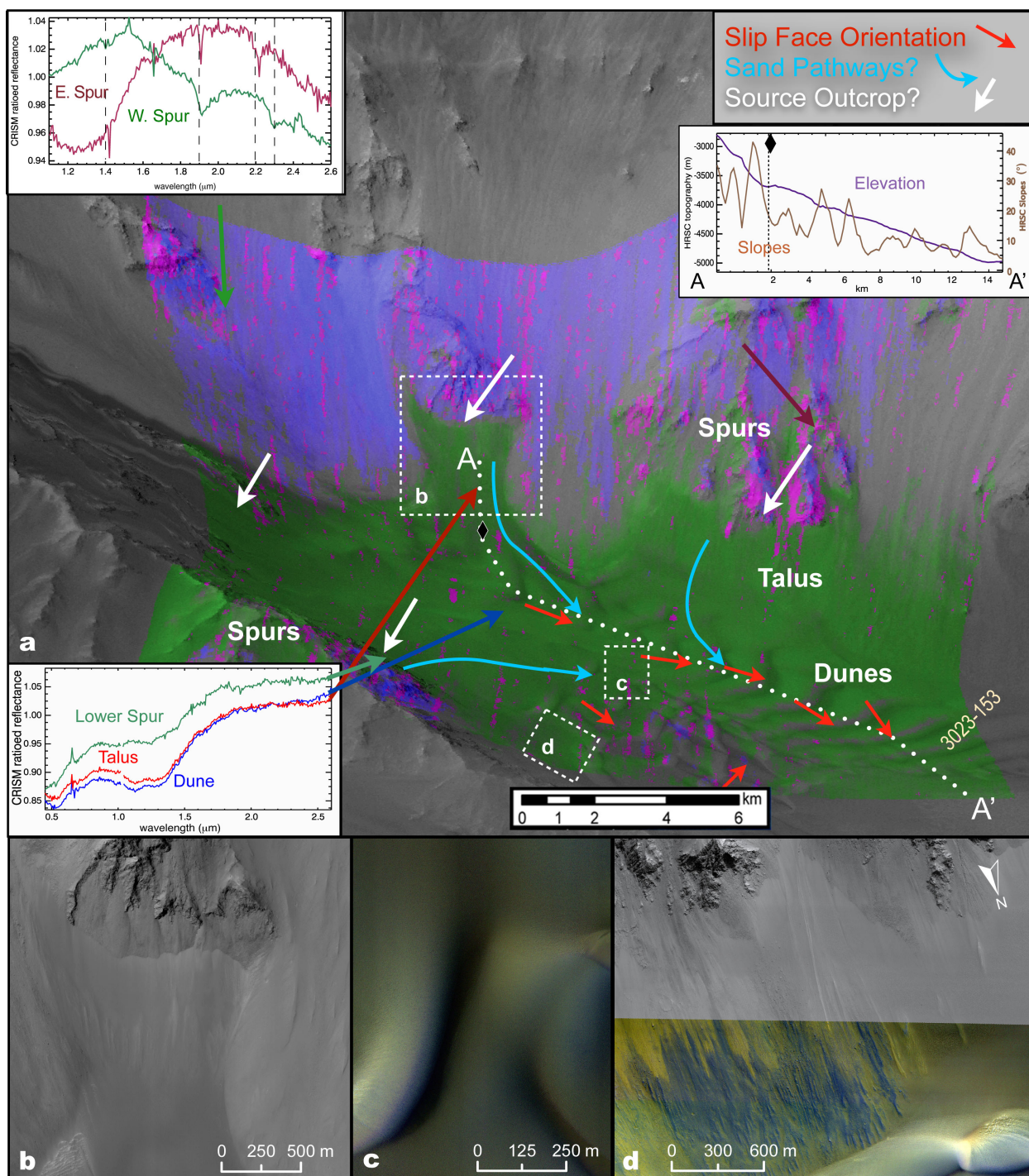


Figure 5.3

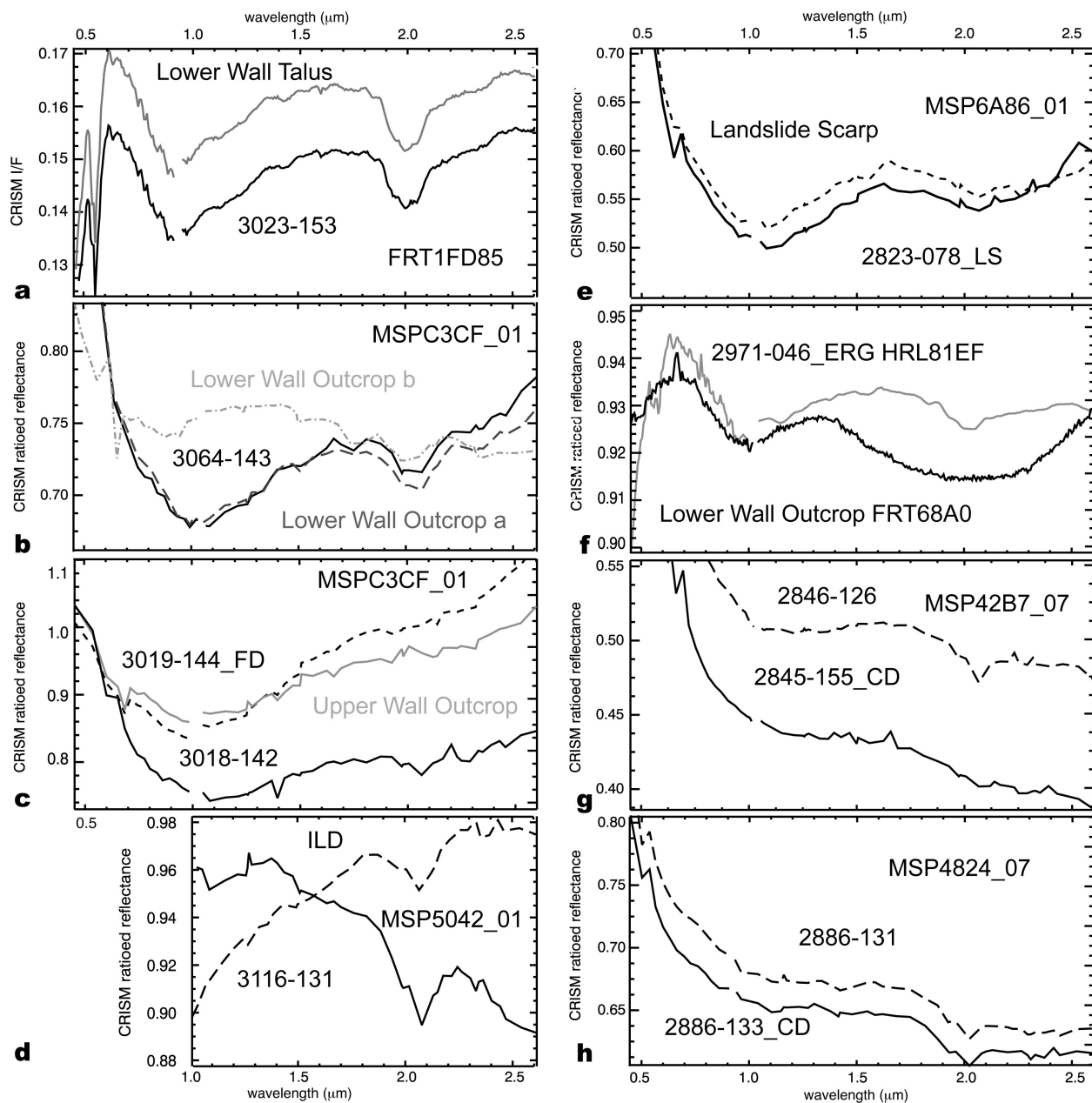


Figure 5.4 CRISM spectra of VM dunes and PSSs.

(a) Dune and talus unratioed spectra in Coprates Chasma (see Fig. 5.3a for location). (b) Dune and talus spectra in east Coprates Chasma. Note that not all wall surfaces match spectral shapes of dunes (see Fig. 5.6b for locations). (c) Coprates Chasma upper wall PSS, floors, and falling dunes (see Fig. 5.7a for location). (d) Dune and ILA spectra in west Capri Chasma (see Fig. 5.2a for location). (e) Dune and landslide scarp spectra in Ius Chasma (see Fig. 5.14a for location). (f) Dune and wall outcrop spectra in east Juventae Chasma. (g) Hebes and (h) Melas chasmata floor and climbing dune pairs.

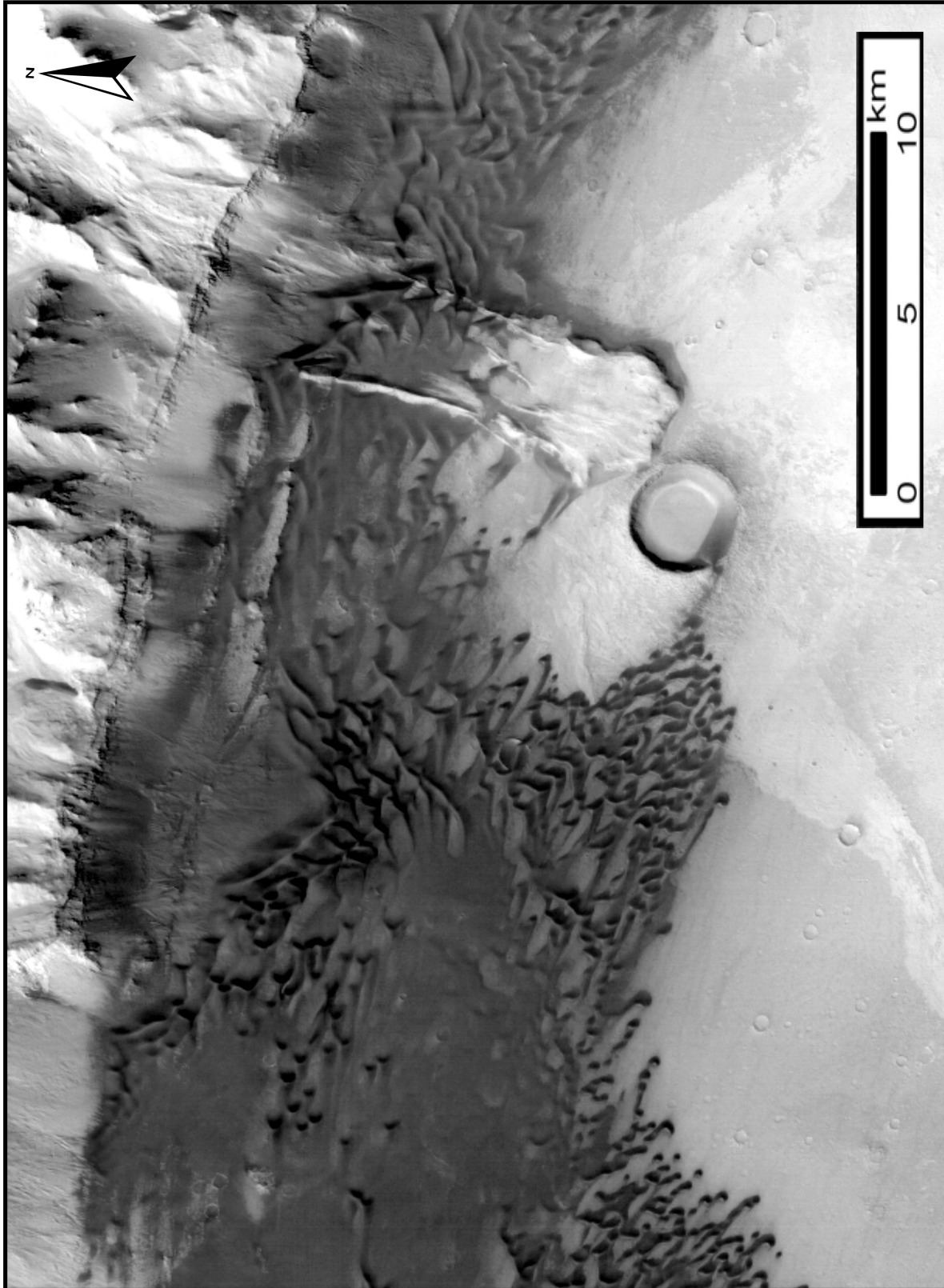


Figure 5.5 CTX image P22_009789_1656_XN_14S053W of dune and lower wall PSS relationships in east Coprates Chasma. Also see Fig. 5.6a and 5.6b for a broader- and finer-scale view, respectively.

Figure 5.6 Dune and lower wall PSS relationships in east Coprates Chasma.

(a) CTX mosaics colorized with THEMIS thermal inertia, where grain sizes from thermal inertia are given in larger white text. (insets) Segmented profile A–A' shows thermal inertia, IRA albedo, and HRSC-derived topography and slopes (orbit 1000, 50 m/pixel). Vertical lines are at the scarp-dune contact (diamonds). (b) A closer CTX perspective view using HRSC topography and colorized with CRISM FRT21CB0 mafic mineralogy (RGB: HCPINDEX, OLINDEX, and LCPINDEX). (right inset) Variable talus and scarp compositions (olivine and HCP) match that of certain sand dune surfaces, suggesting very localized sources and mixing of sediment. Also see Fig. 5.4b. (left insets) HiRISE ESP_025731_1655 RED and enhanced 3-color (RGB) views of select locations along steep $\sim 30\text{--}40^\circ$ talus-slopes above dunes.

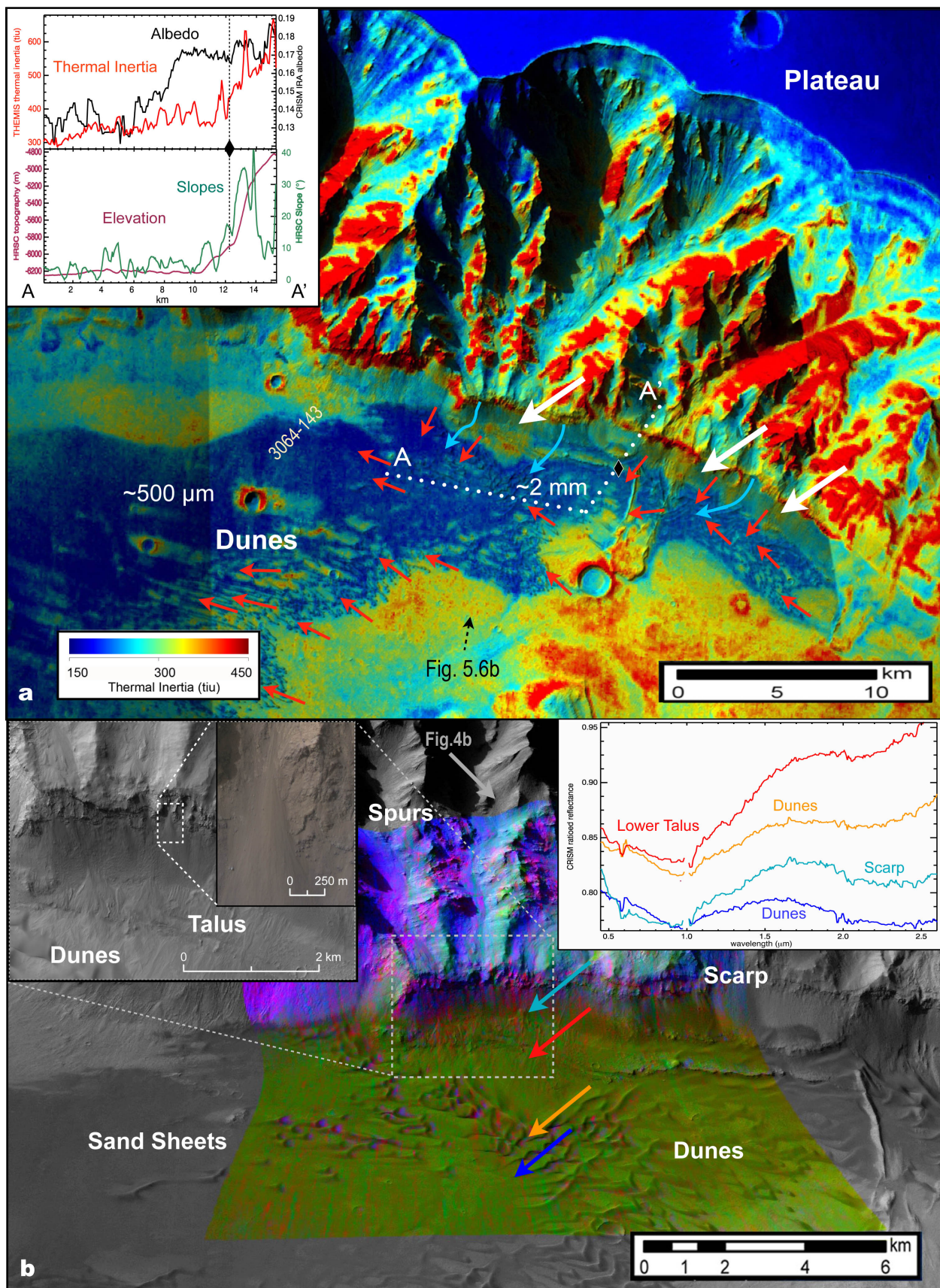


Figure 5.6

Figure 5.7 East Coprates Chasma dune field compositional relationships.

(a) THEMIS daytime infrared mosaic colorized with MOLA elevation and CRISM multispectral coverage (RGB: 2.5, 1.5, 1.08 μm) used in this study overlaid. Dune fields are shown in red and colored circles are CRISM-detected compositions and should be compared with Fig. VM 2b. (b) A CTX image of falling dunes, adjacent upslope outcrops, and their associated spectra (P07_003869_1655_XN_14S058W; CRISM MSP157CE_03). (c) Oblique view of a falling dune field and upper wall PSS on CTX (P16_007139_1653_XN_14S056W) over HRSC topography (orbit 438, 100 m/pixel) and their associated spectra in CRISM MSP3F10_03. (d) Floor dunes, adjacent lower wall outcrops, and their spectra (CTX P18_007930_1649_XI_15S055W; CRISM MSP4477_05). Also see Fig. 5.15a. (e) A CTX mosaic of falling dunes, lower-lying floor dunes, and their unratioed spectra in CRISM MSW2FFA_01.

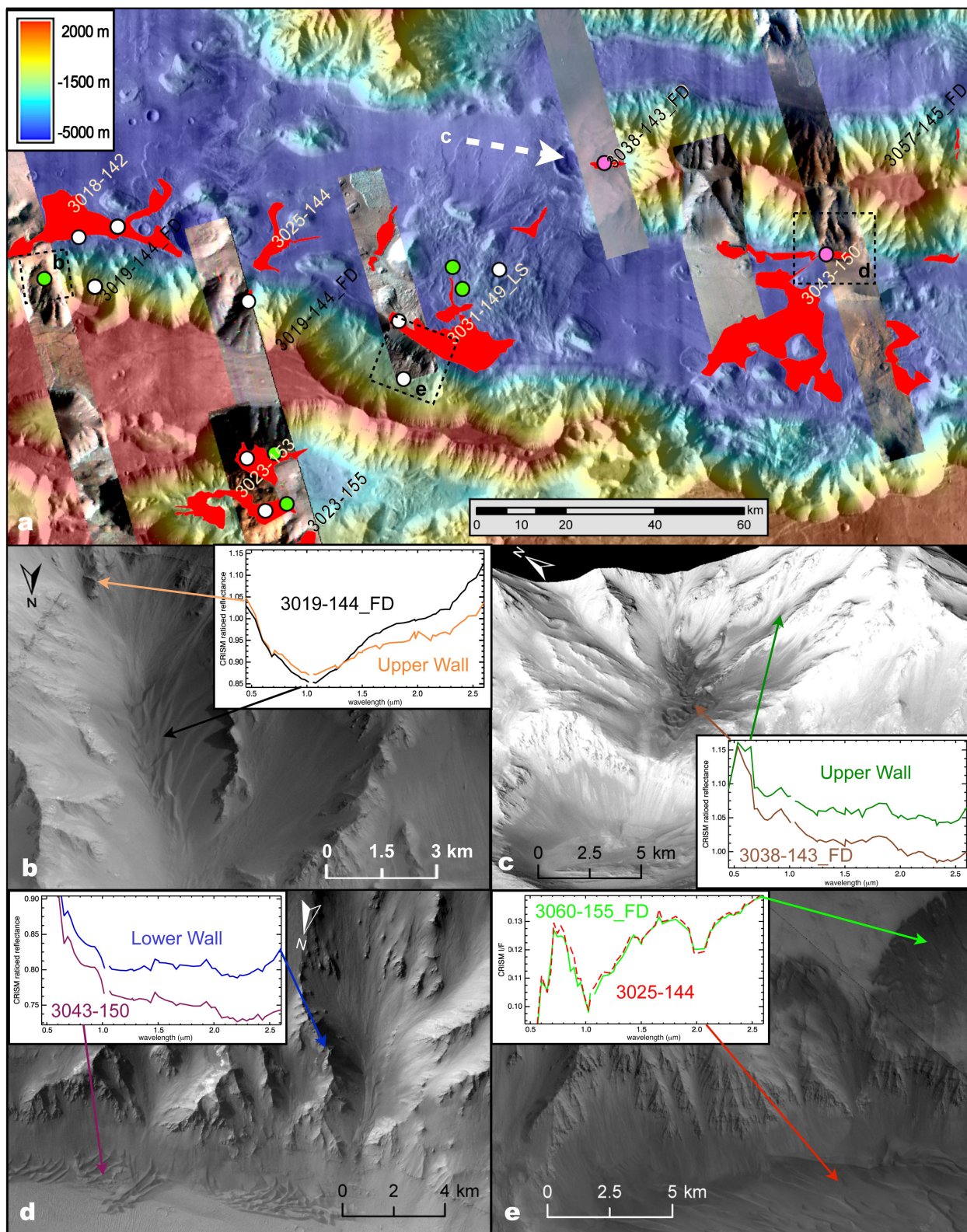


Figure 5.7

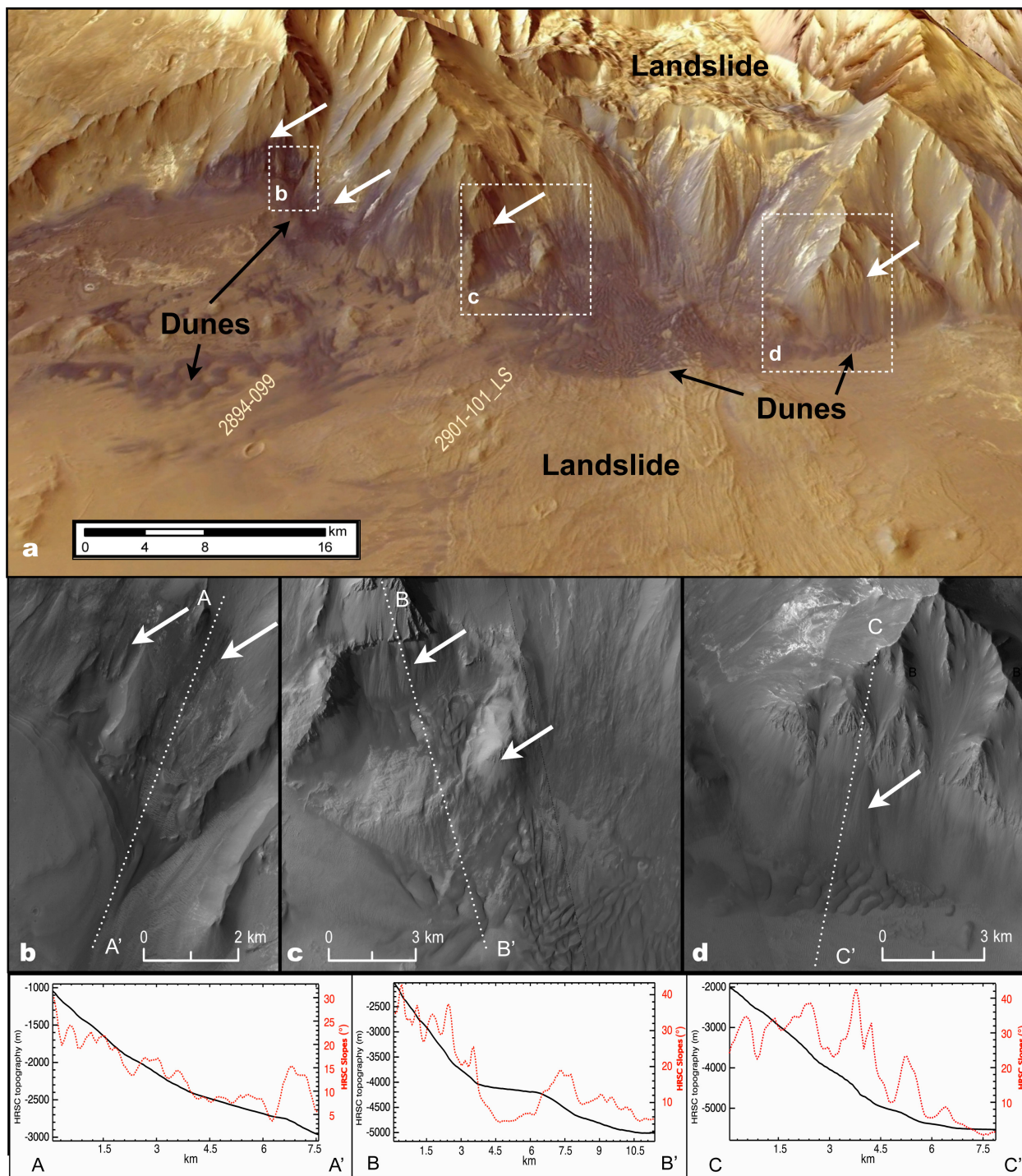


Figure 5.8 Morphologies of Melas Chasma dune fields, landslides, and lower wall outcrop PSSs (white arrows).

(a) Northward perspective view of HRSC color using HRSC topography (orbit 3195, 75 m/pixel). (b–d) Closer views of dunes and dark-toned eroding wall outcrop in CTX mosaic. Outcrop-derived fans of mid- to dark-toned material appear to provide sediment for local dunes, some of which have slip faces oriented away from the inferred source. HRSC topography (black) and slope (red) profiles display slopes at or above the angle of repose for coarse sand ($>32^\circ$). Compare with 5.9–5.10.

Figure 5.9 Composition and thermophysical properties of Melas Chasma dune fields, landslides, and lower wall outcrop PSSs.

(a) CTX mosaic with CRISM MSP1AAF1_05 (black) and MSP46F2_05 (green) spectra of dunes. Locations of 2901-101_LS and associated PSS spectra are shown with color-coded arrows in (d). (b) HiRISE PSP_010396_1695 showing PSSs and (c) HiRISE ESP_020378_1695 RGB of adjacent dunes. (d) CTX mosaics colorized with THEMIS (I35040006) thermal inertias of the largest VM landslide dune field. (inset) A closer view showing a lower wall PSS outcrop upslope of dunes. Note the similar thermal inertias of debris fans and dunes, values consistent with coarse sand. Compare with Fig. 5.8 and 5.10.

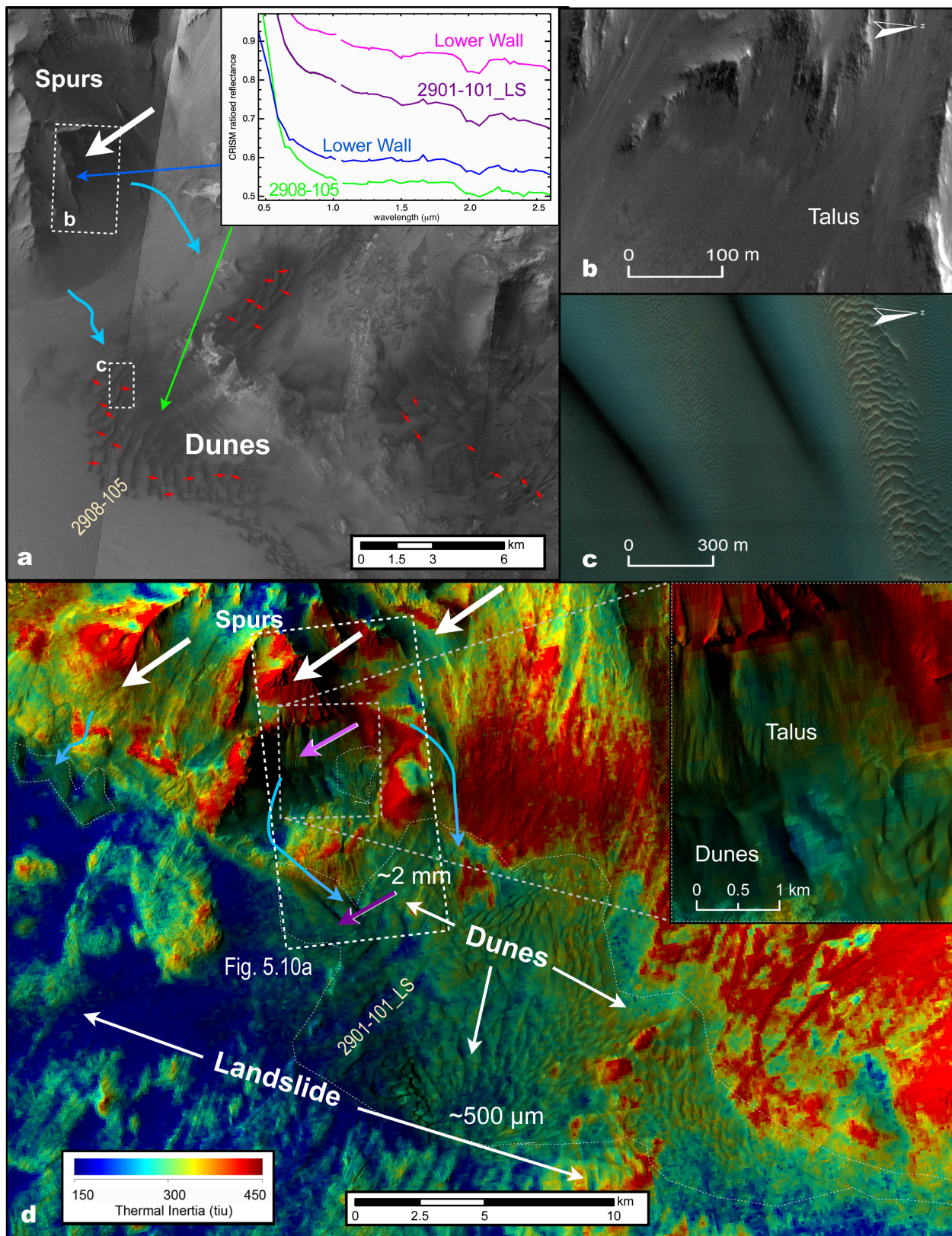


Figure 5.9

Figure 5.10 HiRISE views of Melas Chasma dune fields and lower wall outcrop. (a) HiRISE ESP_028857_1700 RED and IRB false-color over CTX images of dunes and outcrops, also viewed in Figs. 5.8c and 5.9d. HiRISE ESP_028923_1700 IRB false-color views of (b) outcrops and fans of mid- and dark-toned PSS talus, (c) boulder and boulder tracks (white dashed arrows) on intermediate slopes, (d) fan-dune contacts, and (e) lower dune and TAR surfaces. (f) Color ratios of select surfaces, where aeolian bedforms are represented with stars and outcrop-talus with circles. Locations are indicated with yellow and purple arrows in Figs, 5.10a–5.10e. The ferric-ferrous fields were derived from shown laboratory minerals spectra by *Delamere et al.* [2010].

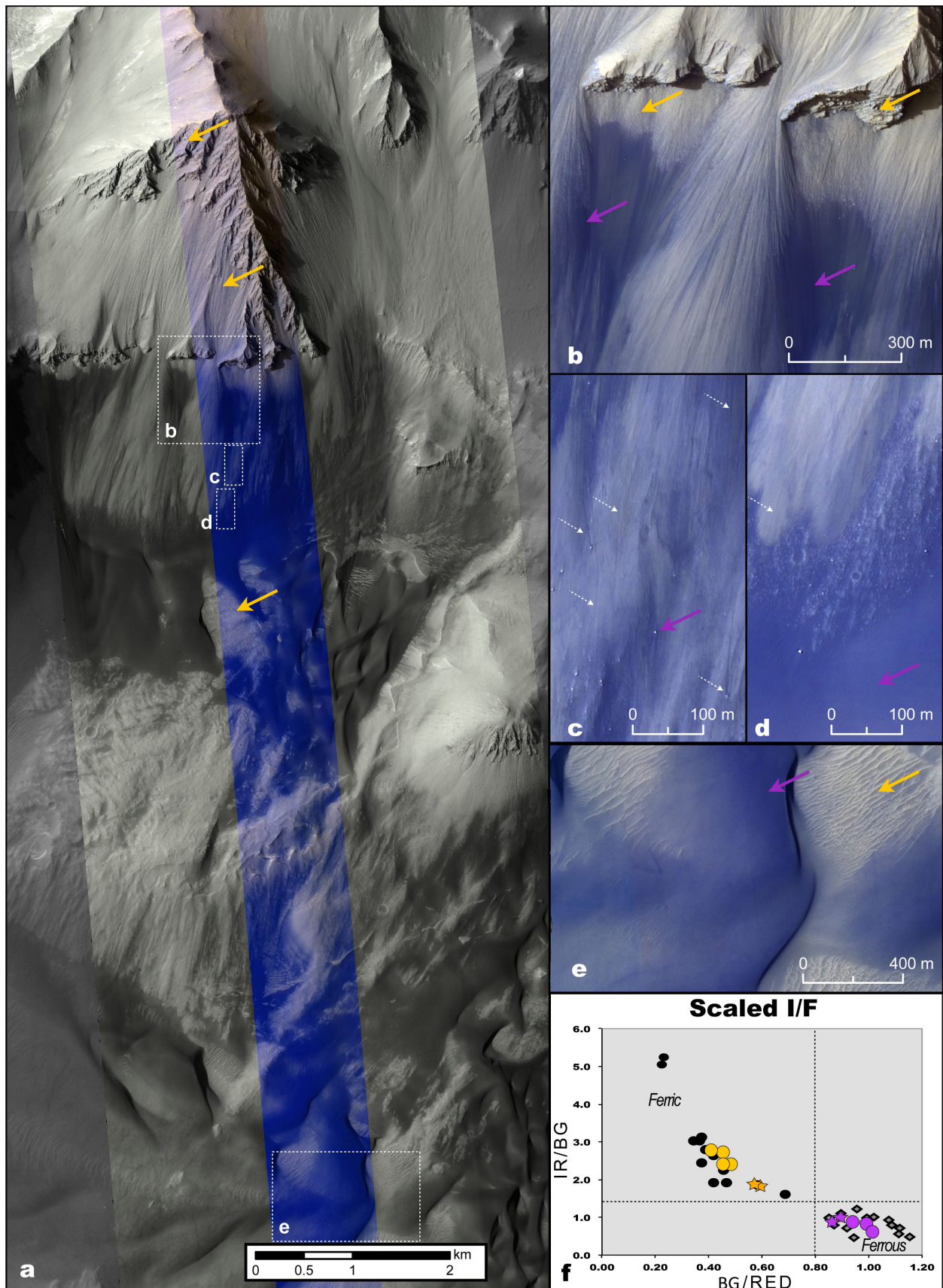


Figure 5.10

Figure 5.11 Juventae Chasma's mound C dune field and mantling unit PSS.

(a) A northward perspective view using CTX T01_000875_1765_XI_03S061W and HRSC topography (orbit 1070, 50 m/pixel). Dune fields are located on the mound summit and ~3 km below on either side on the chasm floor. (b) A nadir view of the summit dune field in the same CTX image colorized with CRISM HRL28A6, where hydration features appear as magenta (SINDEX), olivine as green (OLINDEX), and other mafic minerals (e.g., HCP) in blue (VAR). (insets) CRISM ratioed (HRL444C) and unratioed (HRL28A6) spectra of dunes and mantling units are nearly identical, with the former consistent with an olivine composition. Note the fans of dark material ~1300 m down section at the extreme right. (insets) HiRISE views of (upper, PSP_002946_1765) of mantling units and (lower, ESP_018624_1765) of dunes.

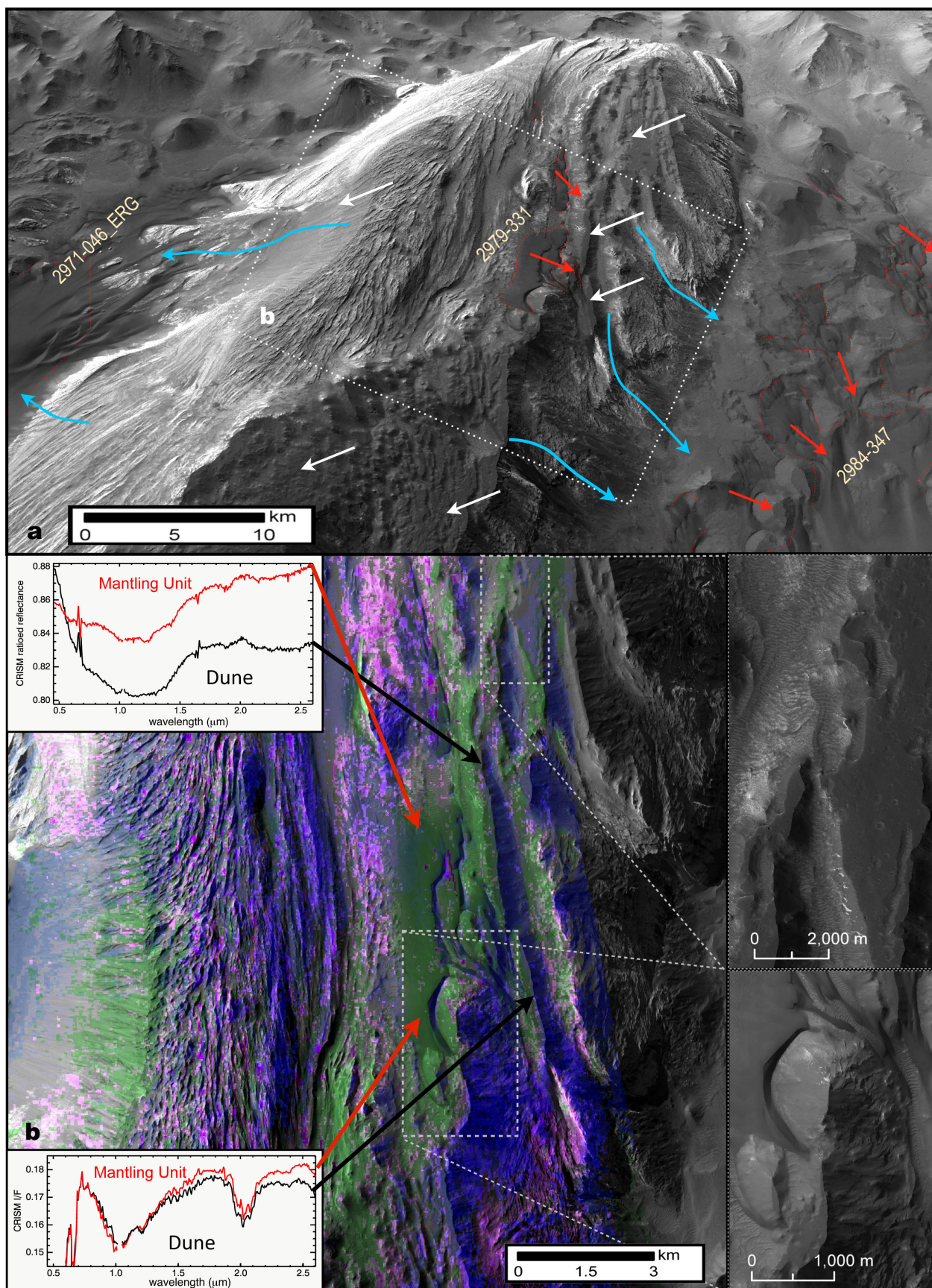


Figure 5.11

Figure 5.12 Juventae Chasma's erg and west mound C talus-fan relationships.

(a) A eastward perspective view using the same data sets as Fig. 5.11a. Dunes and sand sheet are located adjacent to ILD-derived fans of mid-toned material. (inset) CRISM multispectral observations MSP14D58_01 shows HCP associated with mound fans, dunes, and sand sheets ~10 km to the south (not shown). (inset) HiRISE ESP_016712_1760 view of eroding ILD flanks.

(b) CTX mosaics colorized with THEMIS thermal inertias of mound C and associated dune fields (I36899010). Profile A–A' shows THEMIS thermal inertia and CRISM IRA albedo. The vertical lines in that plot indicate profile locations of talus-ILD and ILD-mantling unit contacts (diamonds). (inset) CTX P15_007060_1770_XN_03S062W. Note the decreasing thermophysical properties downslope of the talus-ILD contact and the similar properties of the dunes and mantling unit.

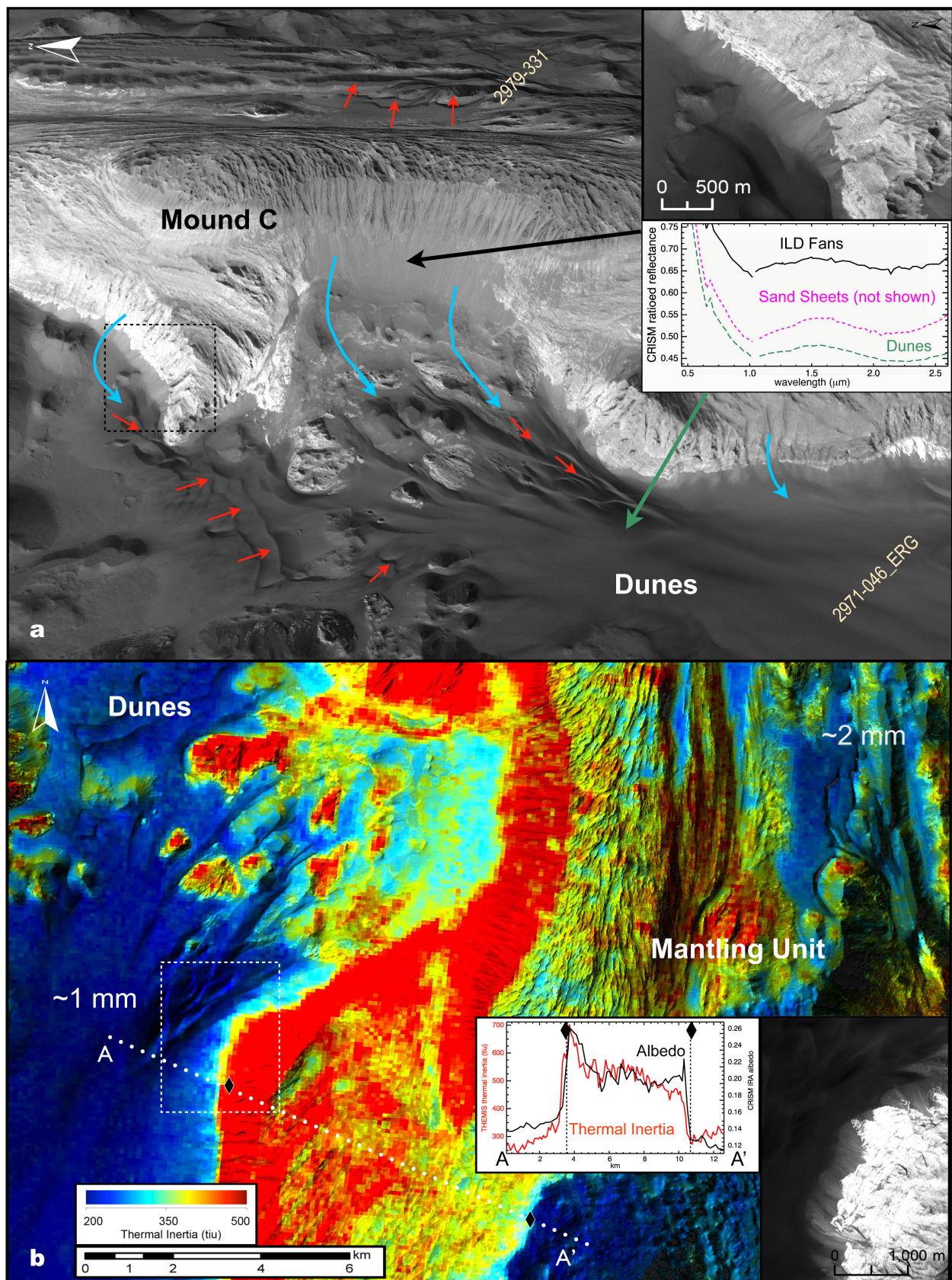


Figure 5.12

Figure 5.13 Morphologic and compositional relationships of east Capri Mensa dunes.

(a) A CTX mosaic of dune and ILD morphologies. (upper right inset) Contextual relationship of dunes, ILD, and ILD-derived fans in HiRISE ESP_020931_1675 RED and RGB. Note the small barchan dune with slip face oriented away from the PSS. (lower right inset) False-color views of dunes where purple and yellow surfaces (arrows) are consistent with colors of ferrous and ferric surfaces in ESP 022276_1670 IRB (see center inset and text). (b) CRISM HRL12E34 overlaid on CTX, where hydration features appear as red (BD2100), yellow (SINDEX), and blue (BD1900R), whereas mafic signatures appear in green (BDI2000). (c) Spectra indicating dunes, ILDs, and talus fans are partially composed of kieserite sand mixed with other mafic/sulfate materials. Lines are positioned at 1.97, 2.14, and 2.4 μm . Note the ILD and fan spectra were collected from FRT681A (locations shown in (a)) using a different denominator, resulting in positive slopes. (d) CRISM profiles along HiRISE ESP_013283_1670 showing IRA albedo and hydration trends across dune surfaces.

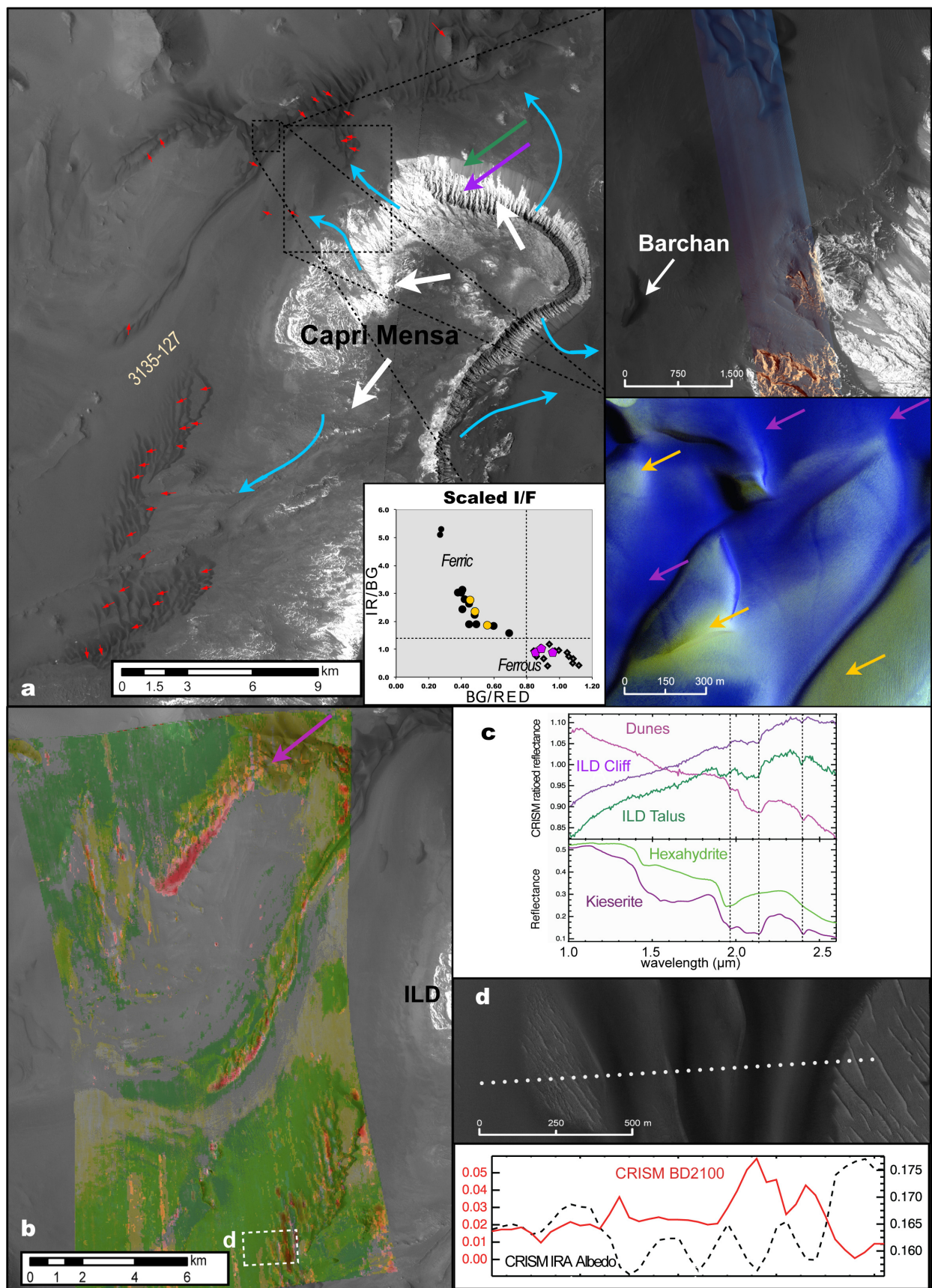


Figure 5.13

Figure 5.14 Ius Chasma landslide dune field morphologic and compositional relationships. (a) Oblique northward view using CTX B18_016739_1730_XN_07S077W with HRSC topography (orbit 2149, 50 m/pixel). Profile A–A' shows HRSC topography (black) and slope (red) profiles (lower left inset) where vertical lines in that plot indicate profile locations of scarp-dune and dune-landslide contacts (diamonds). (upper insets) Nadir views of landslide scarps and dark-toned dunes. Both morphologies display nearly identical spectra (HCP-bearing basalt) in CRISM multispectral observations (MSP1AA73_01). (b–c) HiRISE ESP_026444_1720 RED and IRB views of dunes and PSS. (d) CTX perspective view using HRSC topography and colorized with CRISM FRT1F86A VNIR, where 1- μ m Fe minerals appears as magenta (RPEAK1), variations of $\text{Fe}^{2+}/\text{Fe}^{3+}$ iron in green (BDI1000VIS), and dust as blue (RBR). (lower inset) Outcrop, dunes, and dark streak spectra all show similar absorptions. (upper inset) HiRISE ESP_023543_1720 IRB views of outcrops and bedforms.

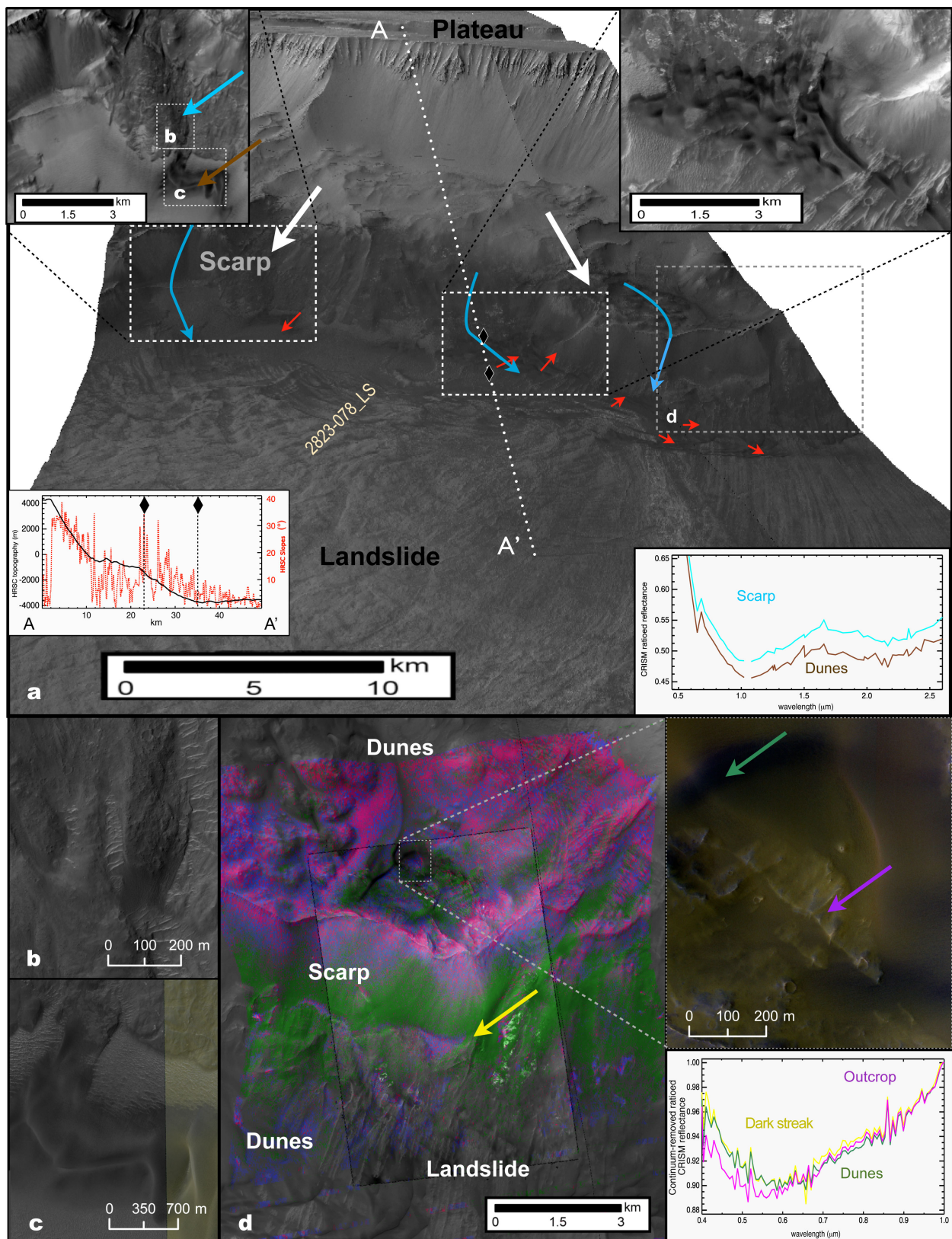


Figure 5.14

Figure 5.15 Coprates Chasma south wall, terraces, landslides, and dunes. (a) CTX mosaic with HiRISE views of (b) darker, presumably more active dunes (PSP_006783_1655) in comparison to (c) lighter-tone, possibly dust mantled dunes (PSP_007785_1645). (upper left inset) A topographic profile shows a hummocky landslide surface (HRSC orbit 438, 100 m/pixel). Lines show the extent of falling dunes, PSS, and continuous landslide dunes (diamonds). (d) Landslide and dune surfaces in a CTX image colorized with CRISM FRT93E3, where Fe-Mg phyllosilicate features appear as magenta (D2300), olivine as green (OLINDEX), and HCP as blue (HCPINDEX). HiRISE PSP_006783_1655 views of (upper left inset) small dunes and (lower left inset) other landslide surfaces that possess (right inset) spectral features consistent with altered olivine. Lines are positioned at 1.96 and 2.3 μm .

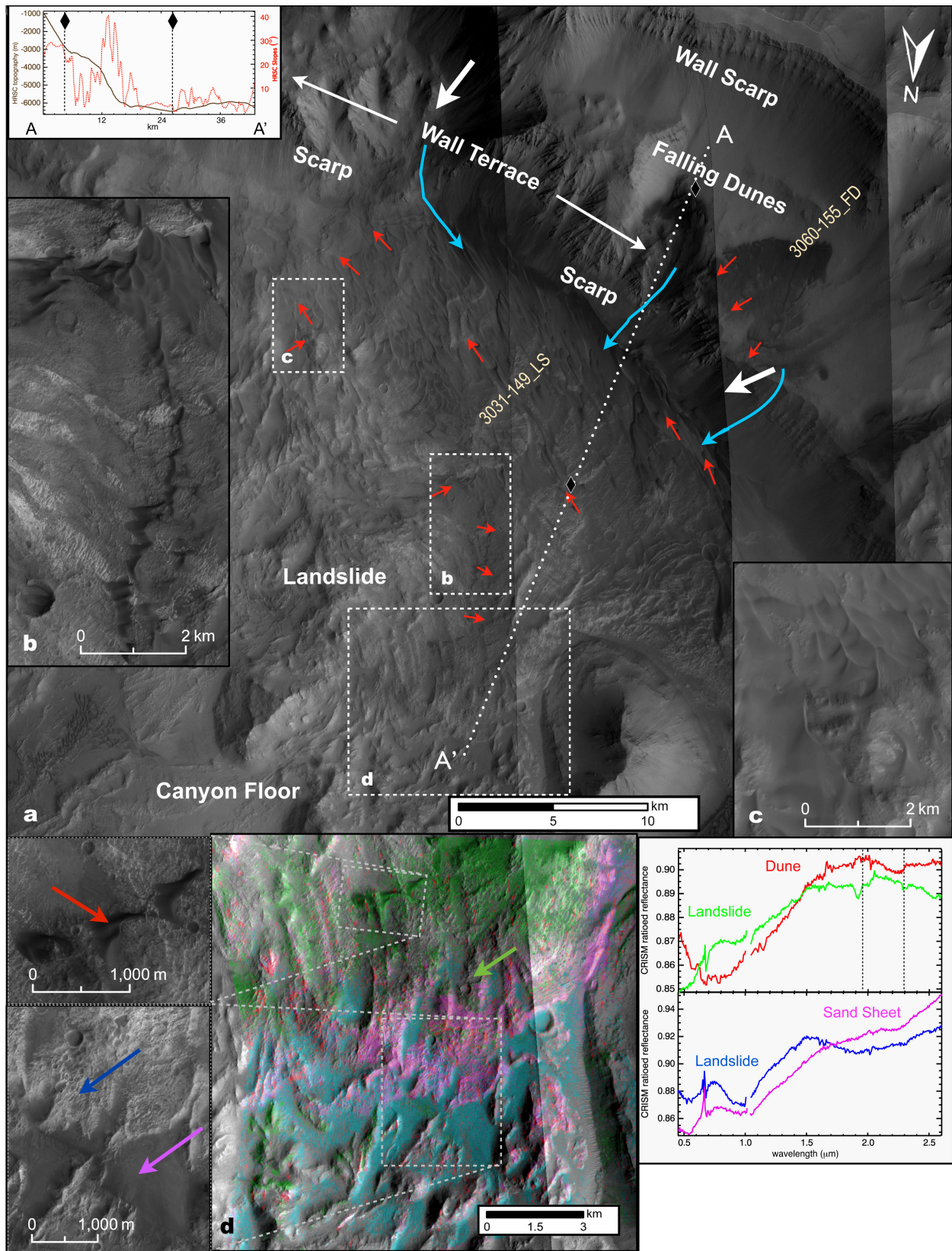


Figure 5.15

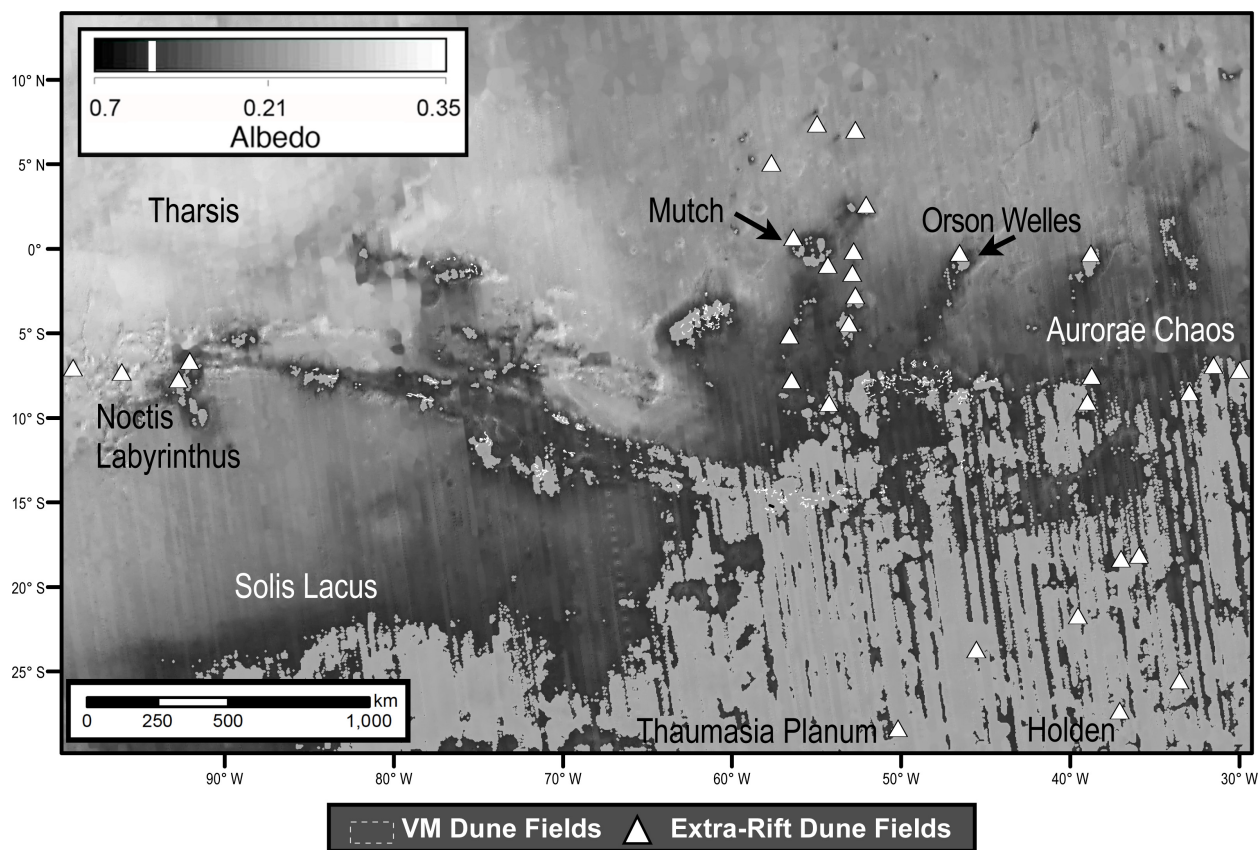


Figure 5.16 TES bolometric albedo map of the region and the distribution of extra-rift and VM dune fields. The threshold value of 0.12 and below is highlighted in white.

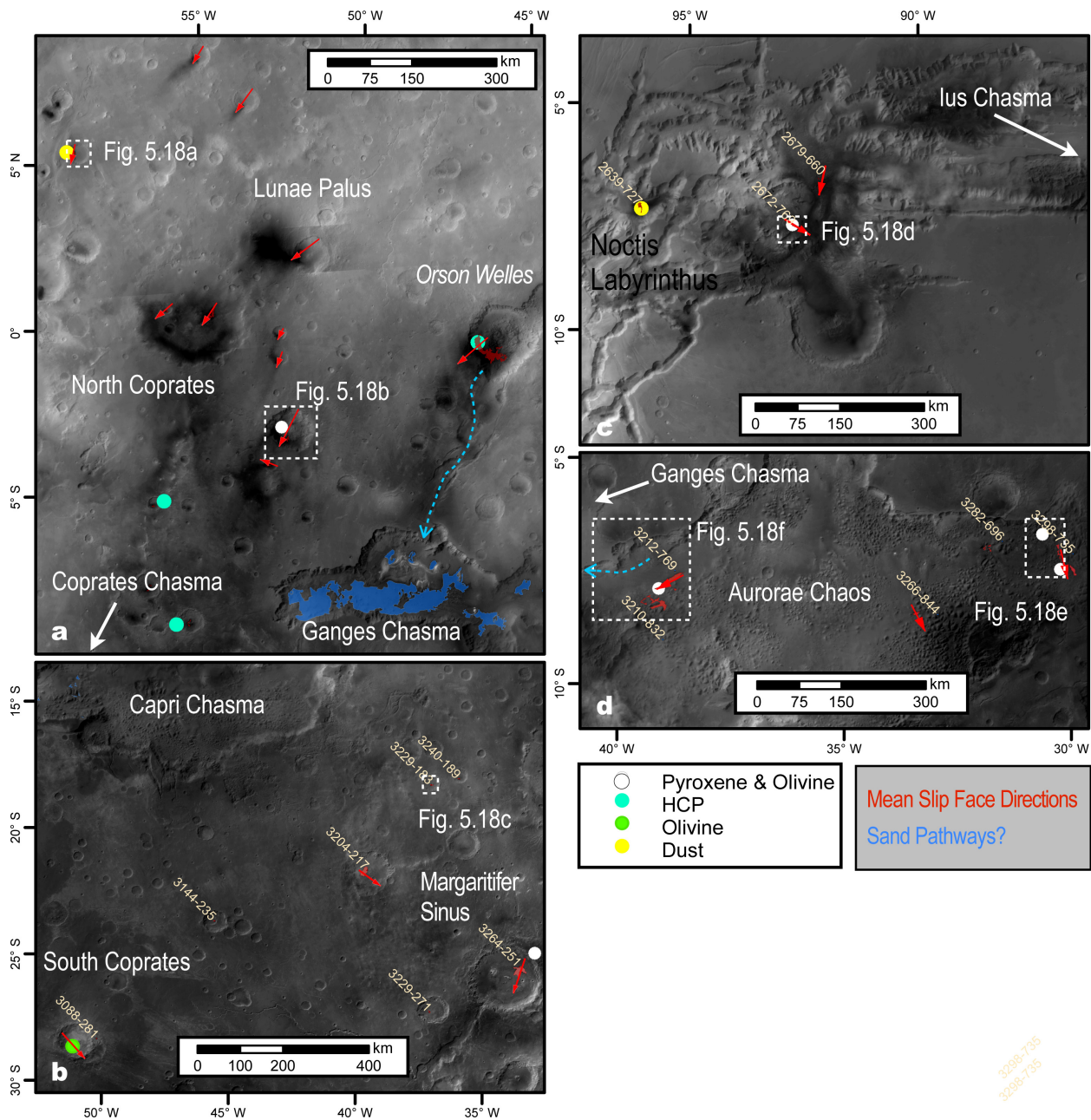


Figure 5.17 MOC wide-angle mosaics of extra-rift dunes and other low-albedo deposits. Extra-rift dune fields (a) north and (b) south of VM and in the adjoining lower-lying regions of (c) Noctis Labyrinthus and (d) Aurorae Chaos. For all dune fields (red polygons) possessing unidirectional dunes (e.g., barchans) slip faces were measured and averaged (red arrows) to approximate dune transport direction. Dune compositions are shown (circles) where CRISM observations are listed in Table 5.1. VM dune fields are shown in blue.

Figure 5.18 Context and high-resolution images of select extra-rift dune fields.

(a) A typical small dune field on a smooth crater floor in Lunae Palus. Transverse dunes grade into small barchans (inset). P12_005860_1849_XI_04N058W. (b) Larger dunes and sand sheets adjacent to crater wall topography just northwest of Ganges Chasma (inset, HiRISE ESP_023463_1765). B20_017305_1770_XN_03S052W. (c) A small crater-related dune field ~350 km SE of Capri Chasma. Dunes adjacent to alluvial fans appear severely degraded and without prominent slip faces, possibly due to inactivity (inset, HiRISE ESP_018966_1615). G02_018966_1628_XN_17S037W. (d) A small dune field in a steep-sided Noctis Labyrinthus collapse pit (inset, HiRISE ESP_012599_1720). P13_005980_1742_XI_05S093W. (e) Variable dune morphologies spread out over ~40 km of chaotic terrain in eastern Aurorae Chaos (inset, CTX G04_019784_1869_XN_06N053W). (f) Western Aurorae Chaos dunes and dark streaks ~300 km east of Ganges Chasma.

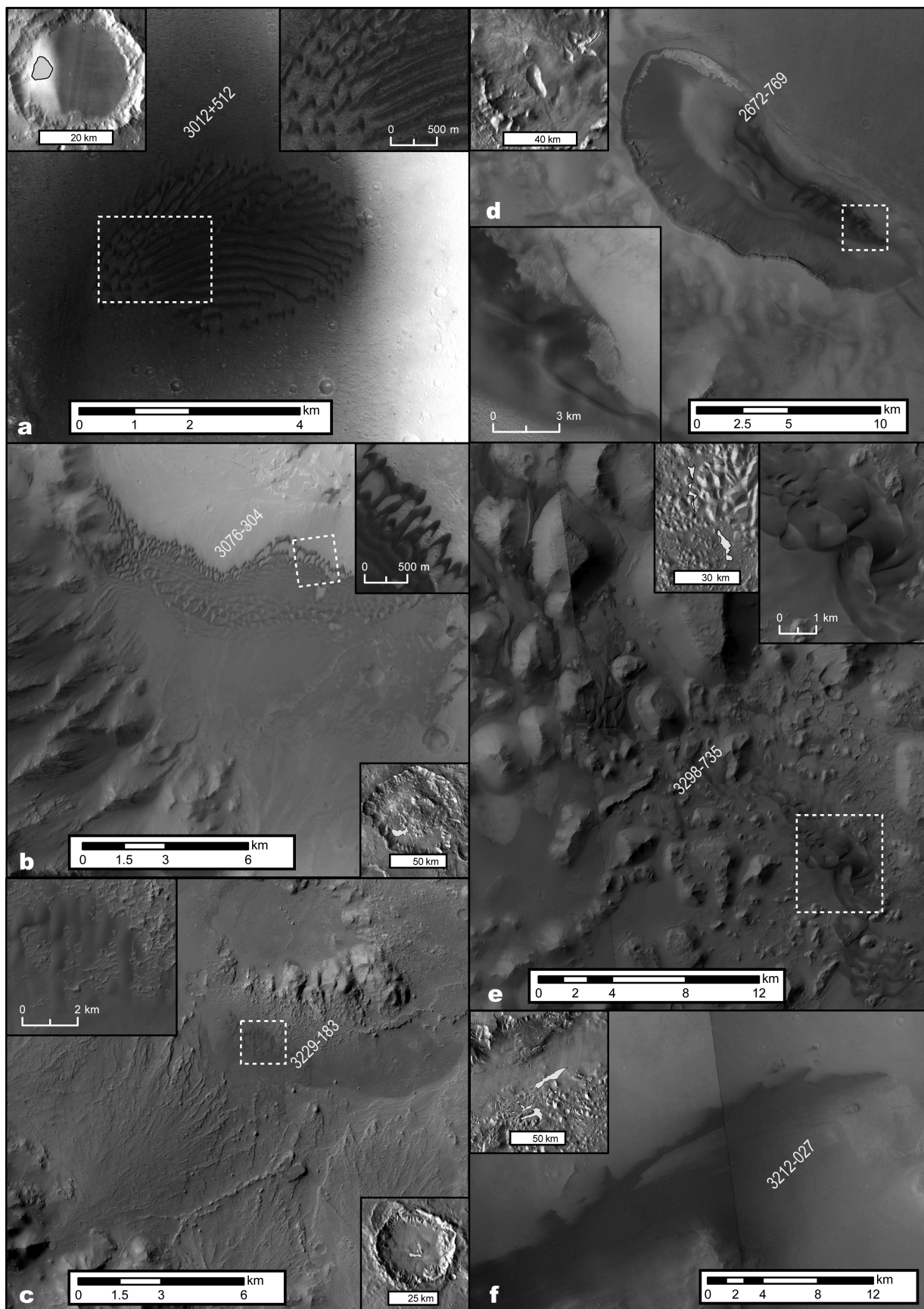


Figure 5.18

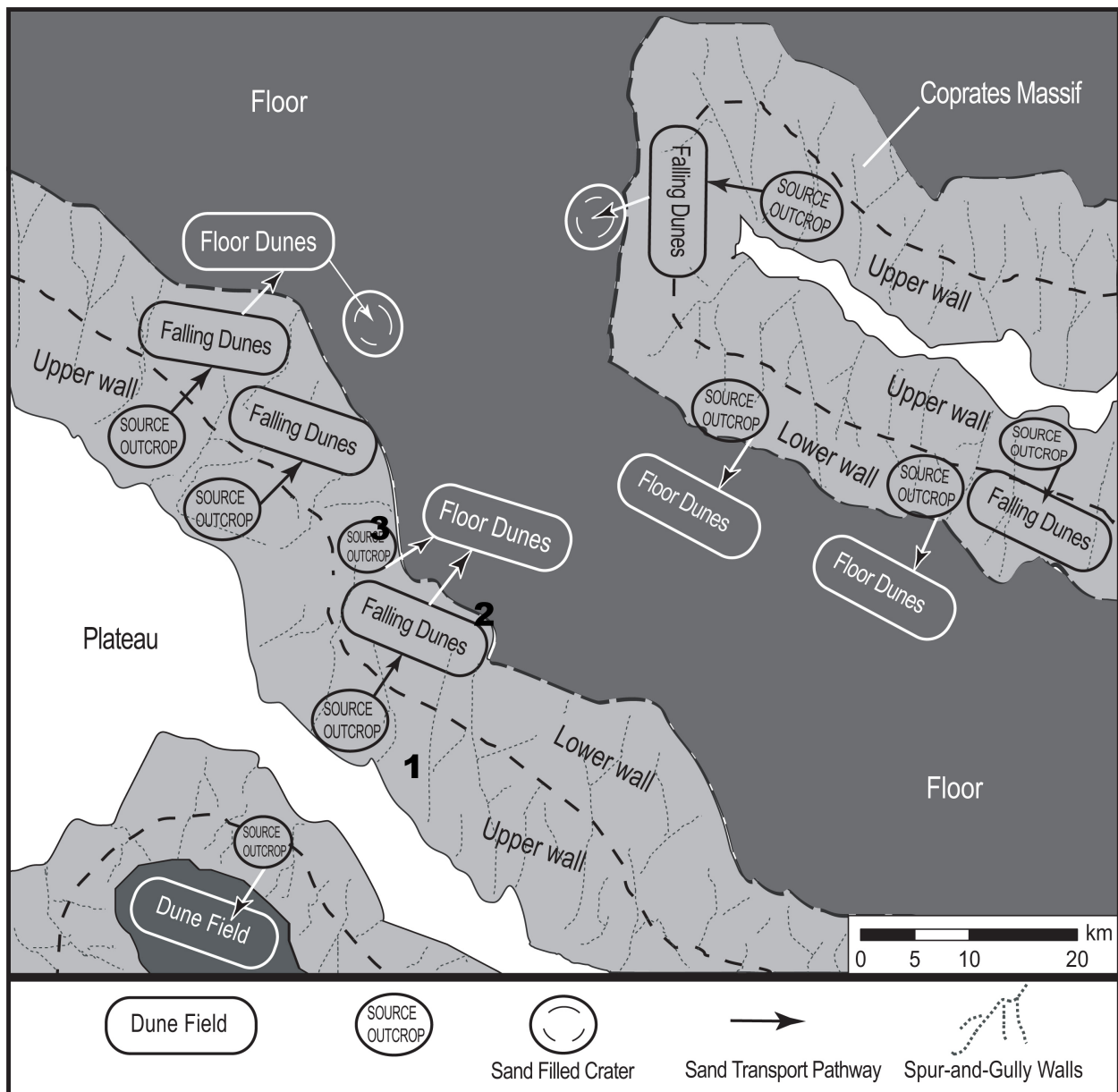


Figure 5.19 Schematic representation of the east Coprates Chasma sediment transport system, based on morphologic, compositional, and thermophysical mapping. See text and Figs. 5.3 and 5.5–5.7 for descriptions. (1) Upper wall outcrops provide sediment for falling dunes located in gullies and terraces; (2) lower wall falling dunes migrate downslope to provide sediment for adjacent floor dunes; (3) lower wall outcrops produce additional sediment for floor dunes.

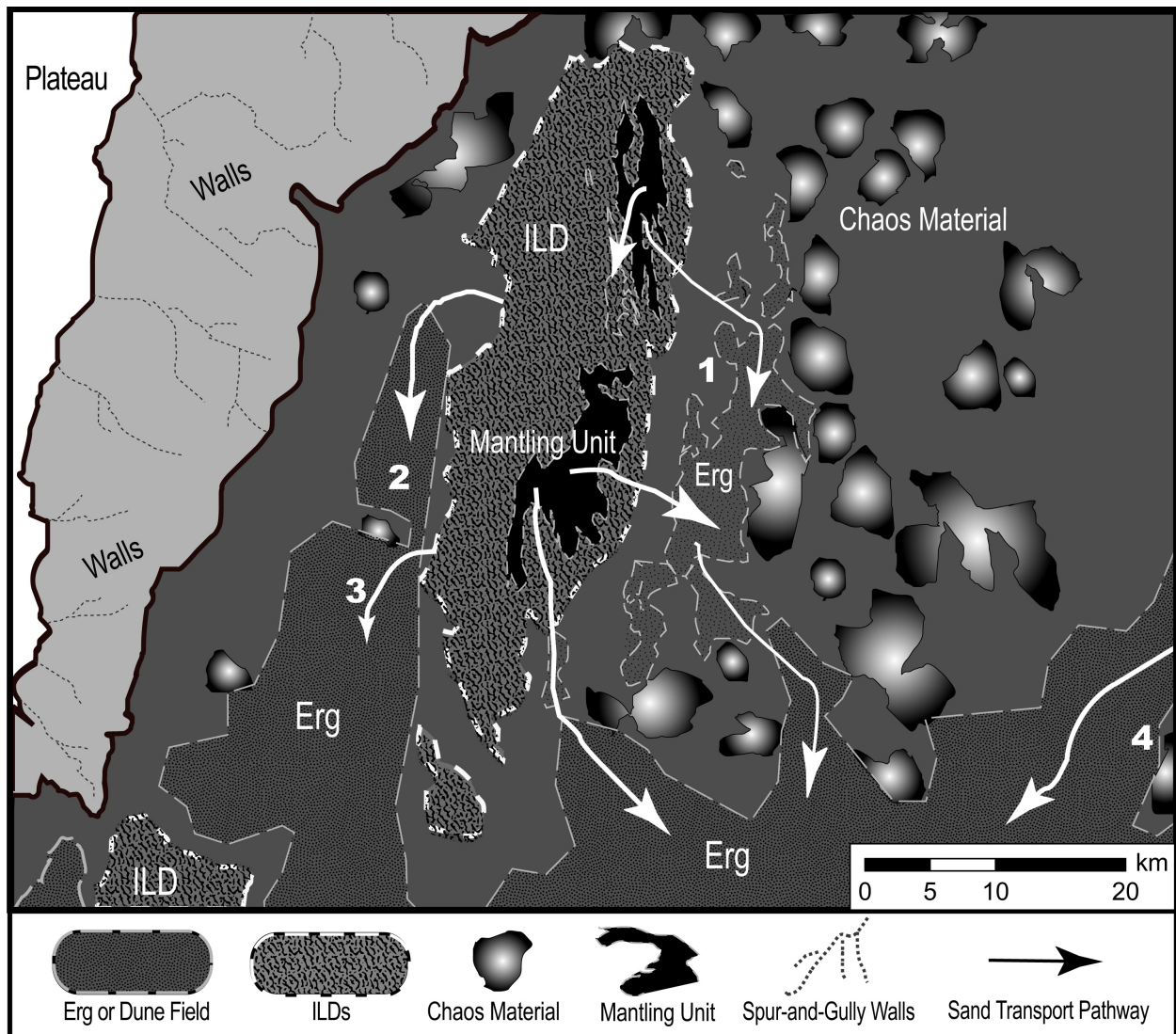


Figure 5.20 Schematic representation of the Juventae Chasma sediment transport system. See text and Figs. 5.11–5.12 for description.

(1) ILD mantling units provide olivine-bearing sediment for summit dunes and erg dunes to the southeast; ILD materials produce (2) HCP- and (3) sulfate-bearing sediment for dunes to the west and south; (4) wall outcrops provide LCP-bearing sand for eastern erg dunes.

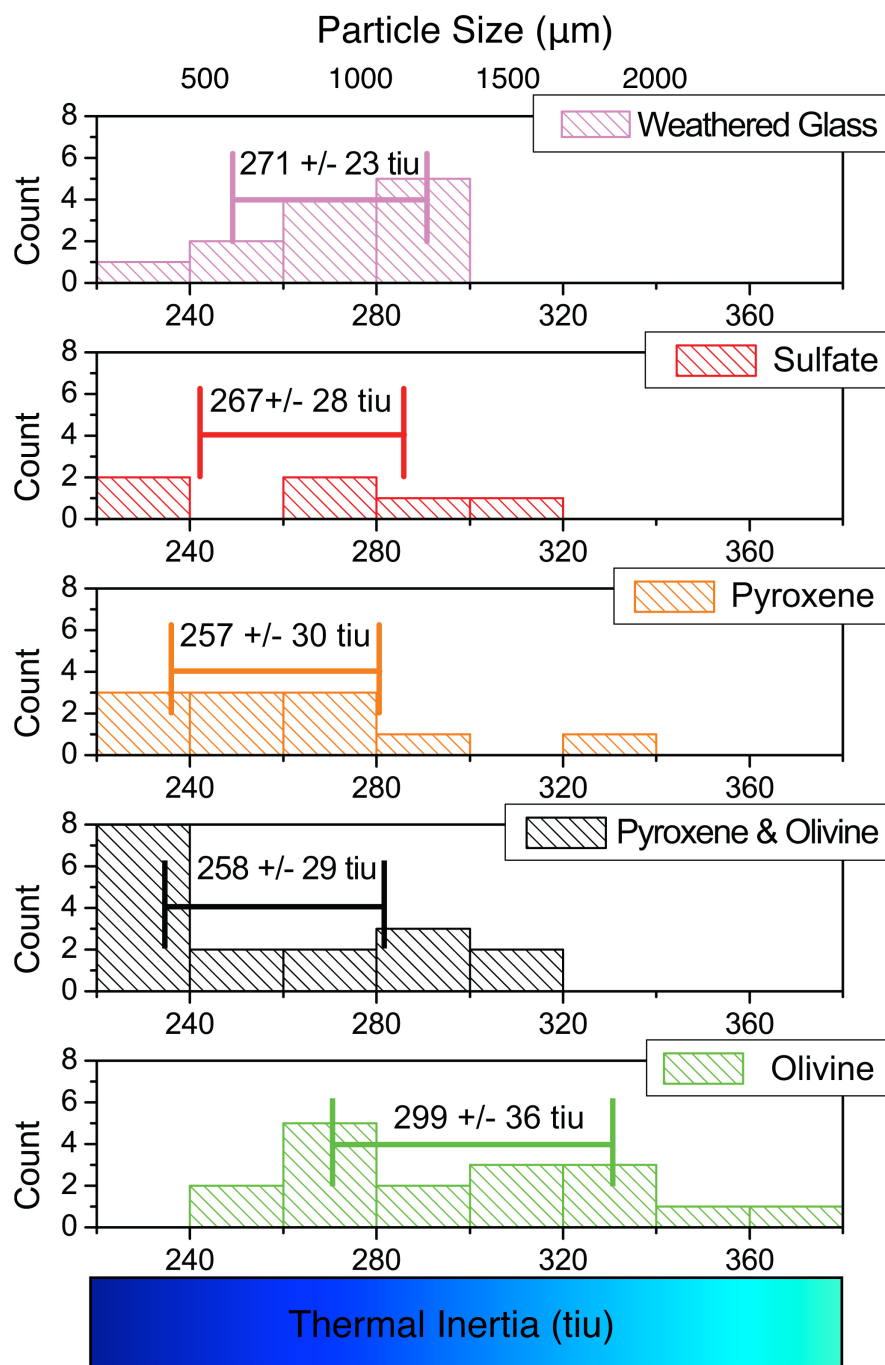


Figure 5.21 Thermal inertia histograms of CRISM-detected dune sand compositions. Mineral classes and number of measurements per class are: weathered glass (12); sulfate (6); pyroxene (11); pyroxene and olivine (basalt) (17); olivine (17). Mean THEMIS thermal inertia values and error bars showing one standard deviation are shown. Approximate particle sizes are shown above.

VITA

Matthew Chojnacki was born in Aurora, Colorado on December 28th, 1973. He went to primary, middle, and high school in the Front Range of the Rocky Mountains where he was introduced to geology and snow skiing. With the encouragement of his parents, those early experiences in the mountains and exposure to the Viking and Voyager planetary missions would have profound roles upon his career paths.

From 1993-2003 Matt accumulated undergraduate college credits at the University of Wyoming, University of Utah, and University of Colorado, Boulder. During most of that time (1993-2000) he was a member of the United States Ski Team, where he competed in the acrobatic discipline of freestyle aerial skiing. He traveled the world (13 countries on four continents) representing his country at three World Championships, one Olympics, and collected multiple National and International medals and titles. His professional snow-jumping career culminated on April 4th 2001 in Winter Park, Colorado where he set the Guinness World Record for most difficult aerial maneuver ever completed (quadruple-twisting quadruple backflip). Although formally retired from the U.S. Ski Team, Matt continues to participate in the sport's summer training apparatus of "water-ramping" in Park City, Utah. There, he performs in public exhibitions, fundraisers, and corporate shows, in addition to guest coaching development and elite ski team members.

Soon after retirement from the Ski Team, Matt would finally graduate in 2003 with a B.A. in Physics and a second B.A. in Planetary Science (*Magna Cum Laude*) from the University of Colorado, Boulder. He remained in Boulder with his two undergraduate advisers at the Laboratory for Atmospheric and Space Physics (LASP) as a research assistant (2004-2007). There he led two-peer review journal articles on Mars surface geology, participated with the

Mars Odyssey's Thermal Emission Imaging System (THEMIS) science team, and attended science conferences.

Matthew enrolled in the Earth and Planetary Sciences graduate program at the University of Tennessee, Knoxville, in the Fall of 2007. From that time until 2012 he took courses in geology, physics, and chemistry at UT in pursuit of a Doctor of Philosophy degree in Planetary Geology. He also taught several laboratory courses in introductory and intermediate geology. Additionally, he participated with the Mars Exploration Rover mission as a download lead for the rovers' primary science cameras (Pancam). Along with teaching and attending classes, Matt conducted research, which culminated in this dissertation. During his tenure at Tennessee, Matt published or submitted five 1st-author publications, eight as a co-author, and participated in nearly two-dozen science conferences, workshops, and team meetings. He also visited Washington D.C. as a student representative for the Division for Planetary Sciences (DPS) of the American Astronomical Society (AAS) Federal Relations Subcommittee (FRS) in 2012 and 2013 to communicate the priorities and concerns of the planetary community to government officials. He expects to be awarded his Doctor of Philosophy degree in May of 2013. Matt is currently working as a post-doctoral fellow for the Mars Reconnaissance Orbiter's High Resolution Imaging Science Experiment (HiRISE) camera at the University of Arizona, Tucson.