DUSTSTONES ON MARS: SOURCE, TRANSPORT, DEPOSITION, AND EROSION

NATHAN T. BRIDGES

Applied Physics Laboratory, MP3-E171, 11101 Johns Hopkins Road, Laurel, Maryland 20723 USA

e-mail: nathan.bridges@jhuapl.edu

AND

DANIEL R. MUHS

US Geological Survey, MS 980, Box 25046, Federal Center, Denver, Colorado 80225 USA

ABSTRACT: Dust is an abundant material on Mars, and there is strong evidence that it is a contributor to the rock record as "duststone," analogous in many ways to loess on Earth. Although a common suite of dust formation mechanisms has operated on the two planets, fundamental differences in environments and geologic histories have resulted in vastly different weighting functions, causing distinct depositional styles and erosional mechanisms. On Earth, dust is derived predominantly from glacial grinding and, in nonglacial environments, by other processes, such as volcanism, eolian abrasion, and fluvial comminution. Hydrological and biological processes convert dust accumulations to loess deposits. Active hydrology also acts to clean dust from the atmosphere and convert loess into soil or erode it entirely. On Mars, glacial production of dust has been minor, with most fine particles probably produced from ancient volcanic, impact, and fluvial processes. Dust is deposited under arid conditions in which aggregate growth and cementation are the stabilizing agents. Thick accumulations result in duststone.

KEY WORDS: dust, loess, Mars, eolian processes, analogs

INTRODUCTION

The geologic record of a planetary body is reflective of its dominant rock-forming processes, preservation mechanisms, and degrees of erosion and exposure. In the sense that Earth and Mars are both "terrestrial" planets and the laws of physics are universal, analogs are useful because they are amenable to detailed study that can be applied to other environments. However, in many cases, only "partial" analogs are appropriate given the vastly different physical, chemical, and geological conditions between planets. Just as the suite of metamorphic facies on Earth resulting from plate tectonics is rare or absent within Mars' contiguous lithosphere, unique processes in the Martian environment have produced rocks with no direct terrestrial analog. As will be shown in this paper, the vast quantities of dust on Mars have likely formed "duststone." Although analogous to terrestrial loess in its mode of deposition and primary particle size and to yardangs in its erosion style, it stands on its own as a probable newly recognized rock type in the Solar System. Its geologic extent may be as significant as sandstone is on Earth. Major questions concern its genesis, evolution, and longevity. The intent of this paper is to show that duststones exist on Mars and that they are a component of that planet's global dust cycle. By integrating observational- and theoretical-based arguments, we discuss duststone formation and compare it to analogous processes on Earth. This paper is partly a review of the duststone hypothesis proposed by Bridges et al. (2010), but with the addition of a detailed comparison to the dust cycle and deposits on Earth.

To properly frame our discussion and arguments, definitions must be established. For the purposes of this paper, dust is defined as any relatively fine-grained particle capable of being suspended in a planet's atmosphere for a considerable length of time. Theoretically, the maximum size for suspension can be derived by balancing the threshold friction speed ($u*_t$) against the terminal fall velocity, with larger-size material predicted to saltate (Greeley and Iversen 1985). In practice, this predicts suspended material grain sizes that are too large. For example, the saltation/suspension boundary for basalt on Mars should be 175 μm (Bridges et al. 2010), despite the fact that eolian ripples, features formed by saltation and creep processes, contain particles less than 100 μm in diameter, which theory predicts should be suspended (Sullivan et al. 2008). This discrepancy between theory and observation may exist because of the lower atmospheric density on Mars compared to Earth, which causes particles to respond more slowly to turbulent eddies that drive suspension (Sullivan et al. 2005, 2011). The size range of suspended dust on Mars is estimated at 2 to 5 μm (Kahn et al. 1992, Lemmon et al. 2004), which is on the lesser side of the distribution of sizes found for loess, but similar to dust aerosols on Earth (Fig. 1). As will be discussed more comprehensively later, dust on the Martian surface commonly aggregates into larger particles and reaches sizes such that it can only move by saltation, or, upon further growth, it becomes immobile. Therefore, we define "dust" as particles that are ones to tens of micrometers in diameter that are suspended or deposited and "dust aggregates" as particles ~100 μm and larger that are on the surface and derived from dust. "Sand" is defined as grains ~100 μm and larger that undergo saltation and are the direct by-product of disintegrated rock.

TERRESTRIAL DUST AND LOESS: ORIGIN AND DEPOSITS

Definitions and Geomorphic Expressions

Although there has been considerable research comparing Martian and terrestrial dunes, the ways in which Martian dust and its depositional character contrast to that on Earth have received less attention. The general lack of geomorphic expression in terrestrial dust and loess deposits is at least part of the reason such analogs have not been extensively recognized or discussed. With increased sophistication in imaging of the Martian surface and atmosphere, it is now possible to make at least some comparisons between fine-grained eolian deposits on the two planets. In this section, we review dust and loess deposits on Earth, including their genesis, geologic record, and relation to climate forcing through time.

Fine-grained eolian deposits on Earth are classified as "loess" and

Sedimentary Geology of Mars
SEPM Special Publication No. 102, Copyright © 2012
SEPM (Society for Sedimentary Geology), Print ISBN 978-1-56576-312-8, CD/DVD ISBN 978-1-56576-313-5, p. 169–182.

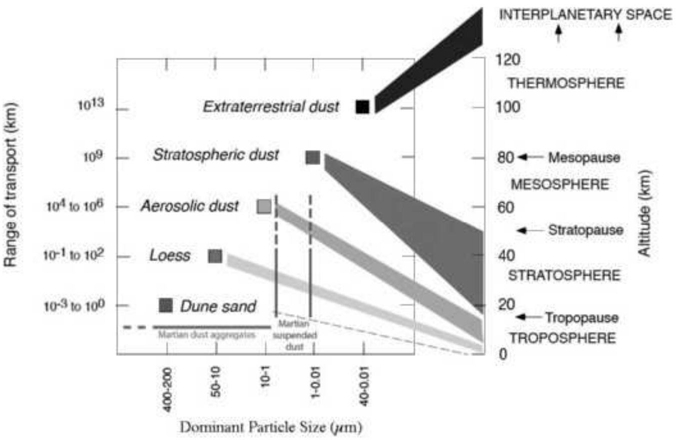

FIG. 1.—Diagram showing the continuum of dust on Earth (solid boxes), with range of transport in the initial phase (in contrast to multiple cycles of movement over long periods of time) as a function of dominant particle size and generalized zone of transport in the atmosphere. The vertical bars show the approximate particle size range of Martian suspended dust, and the horizontal line shows the size of dust aggregates on the planet. The Earth portion of this diagram was generated using concepts in Syers et al. (1969).

"aerosolic dust," unlike the simpler classification of all fine-grained material on Mars as "dust" (see previous). On Earth, these fine-grained materials can be viewed as intermediate members of a continuum of eolian deposits that range from sand to very-fine-grained (<1-μm-diameter) particles that are transported in the stratosphere (Fig. 1). Loess is dominated by silt-sized particles (2–50 μm particle diameters) and is coarser than the material referred to as aerosolic dust (particles <10 μm diameter), which is capable of long-range transport. Typically, loess contains 60–90% silt-sized particles and smaller amounts of sand and clay (< 2μm); it may cover as much as 5–10% of Earth's land surface (Pye 1987), including small areas of Africa and the Middle East, attesting to non-glacial origins for some deposits.

Loess deposits are expressed as distinctive sedimentary bodies. Thickness is highly variable and can range from a few centimeters to several hundred meters, with thickness and mean particle size decreasing downwind away from a source (Pye 1987; Muhs and Bettis 2000, 2003; Muhs et al. 2004, 2008). The deposits commonly have little or no geomorphic expression (Pye 1987), being draped over preexisting landforms as a mantle, with thickest accumulations in protected, low-lying areas or on flat, stable upland divides and thinnest accumulations occurring on narrow, rounded hillcrests (Fig. 2). One exception to the general rule of little geomorphic expression of loess may be "pahas," which are streamlined loess landforms aligned with paleowinds that may be yardang-like features (Flemal et al. 1972). However, similar morphologies also have been tied to primary deposition (Lewis 1960), so, at best, geomorphic expressions of eolian erosion of loess are limited.

Unlike eolian sand or fluvial–marine sediments, primary structures in loess are subtle or absent altogether. Faint, horizontal laminations and cross-bedding are only rarely apparent (Fig. 3). Rather, most loess deposits are characterized by a massive (as opposed to structured) condition. Interparticle binding by phyllosilicate clay minerals or secondary calcite accumulations commonly result in a significant amount of material strength, allowing for the formation of vertical faces along riverbanks or stream banks and road cuts. Secondary structures in loess are more common than primary structures, and these consist of fractures, burrows, rhizoliths (root casts composed of Fe-oxides or carbonate), carbonate nodules or concretions, oxidation or reduction streaks or bands, and paleosols. Lithification or cementation of loess particles to form siltstone is unknown in Quaternary (past 2.6 Ma) loess deposits on Earth, but it has been hypothesized to explain the origin of some Paleozoic and Mesozoic siltstones (Johnson 1989; Soreghan et al. 1997, 2002; Chan 1999).

Loess Composition

Loess has a mineralogy that usually includes quartz, plagioclase, K-feldspar, mica, calcite (and sometimes dolomite), and phyllosilicate clay minerals (smectite, chlorite, mica, and kaolinite) (Pye 1987). Heavy minerals are usually present, but in small amounts. Bulk geochemical studies show that the dominant constituent in loess is SiO_2, ranging from ~45% to 75%, but more typically in the range 55% to 65% (Pye 1987, Muhs and Bettis 2003, Muhs et al. 2008). The high SiO_2 content of loess deposits reflects a dominance of quartz, but smaller amounts of feldspars and clay minerals also contribute to this value. Clay mineralogy in loess shows considerable variability. In North America, for example, clay minerals of loess are a direct reflection of source sediment. In Alaska, loess deposits are ultimately derived from metamorphic rocks in major mountain ranges and are rich in mica and chlorite (Muhs and Budahn 2006). In the midcontinent, loess in the Mississippi River drainage basin is derived to a great extent from Paleozoic carbonate rocks (explaining the high calcite and dolomite) and shales, which explain the high mica content in the clay fraction (Grimley 2000). Loess in the Great Plains region is rich in smectite, derived mainly from Tertiary volcaniclastic siltstone that is also rich in smectite (Aleinikoff et al. 1999, 2008; Muhs et al. 1999, 2008).

Origin of Loess

The "glacial" model of loess formation is a traditional view that silt-sized particles are produced dominantly by glacial grinding of rocks, deposition in till, reworking by fluvial processes as outwash, and finally entrainment, transportation, and deposition by wind (Fig. 4a; see also discussion in Muhs and Bettis 2003). This model has led to the view that loess deposits are primarily markers of global glacial periods. Support for the classical glacial model of silt production and later eolian entrainment as loess comes from modern observations in regions such as Alaska, Canada, and Iceland. In these regions and elsewhere, glaciers exist now, and it is possible to observe loess entrainment from rivers that drain glaciated terrain (e.g., Muhs et al. 2004).

There is also strong evidence for nonglacial loess (Wright et al. 1998; Wright 2001a, 2001b; Smith et al. 2002; Crouvi et al. 2010). "Desert" loess is a term used loosely to describe eolian silt generated in and derived from arid or semiarid regions that were not glaciated. The debate on desert vs. glacial loess centers on whether silt-sized particles can be produced by mechanisms other than glacial grinding, particularly in deserts. This has particular applicability for Mars, where glacial activity has been at most a minor contributor to dust-size particle production over the history of the planet (see following). A variety of mechanisms can, in principle, produce silt-sized particles in arid regions, and these are summarized in a highly simplified model (Fig. 4b). These processes include frost shattering at high altitudes, comminution (particle size reduction by crushing or grinding) by fluvial and mass-movement transport, chemical weathering, salt weathering, eolian abrasion, and ballistic impacts (Wright et al. 1998; Wright 2001a, 2001b; Smith et al. 2002; Crouvi et al. 2010). As with the glacial environment, it is possible to observe silt entrainment in modern desert environments. A good example is the yearly transport of fine silts and clays out of Africa by wind, derived from the Sahara

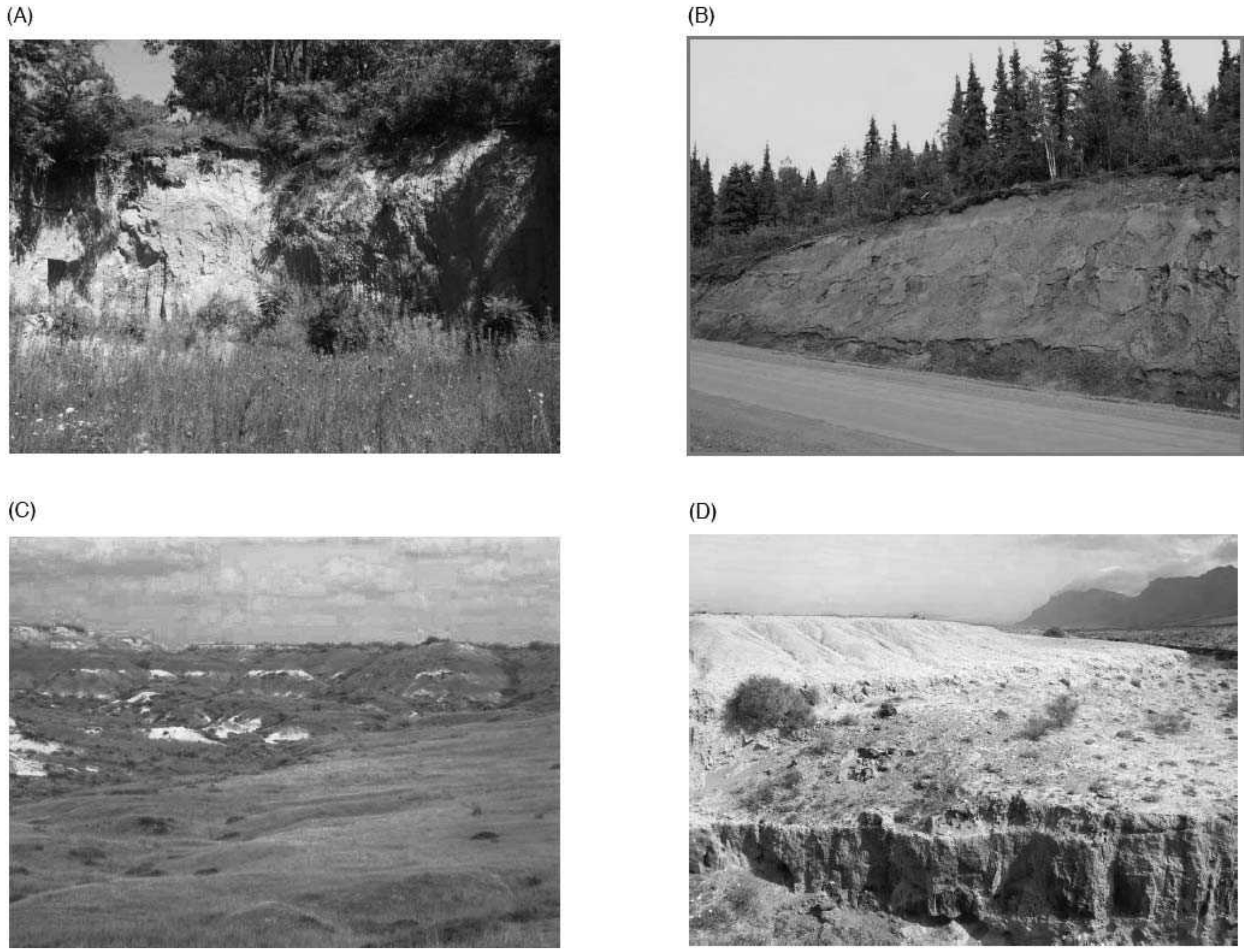

FIG. 2.—Gallery of loess deposits on Earth. (A) Upper part of last-glacial-age (Peoria) loess exposed in a quarry near Morrison, Illinois; thickness of section exposed is ~8 m. (B) Holocene (upper ~1 m) and last-glacial-age (lower ~8 m) loess exposed in a road cut south of the Yukon River, central Alaska. (C) Multiple loess units (gray bands) with intercalated calcic paleosols (white bands) spanning ~170 ka, exposed in the Ruhama Badlands, northern Negev Desert, Israel (see Wieder et al. 2008); thickness of section is ~15 m. (D) Loess-like, African-dust–derived eolian silt deposits (whitish, upper part of exposure) overlying basaltic tephra of Quaternary age (gray band, in foreground and lower part of exposure), near Famara, Lanzarote, Canary Islands, Spain; thickness of exposure is ~4 m. All photographs are by D.R. Muhs.

and Sahel regions, where glaciation has played no role in silt particle formation. Indeed, although fine particle transport by wind is highly seasonal in Africa, this continent is, overall, the largest source of dust on Earth today.

In the glacial loess-versus-desert loess debate, there is generally little consideration of silt particle inheritance from sedimentary rocks or volcanic tephra. Notable exceptions include sites in Argentina, where an Andean volcaniclastic source has been recognized (Zárate and Blasi 1993), and Australia, where siltstones may be the ultimate source of much of the silt-sized dust in arid basins (McTainsh 1989). Silt is abundant in the sedimentary rock record, and estimates show that fully half of the detrital quartz in the world's sedimentary rocks is composed of silt-sized particles (Blatt 1987). In the Great Plains of North America, sedimentary rock (volcaniclastic siltstone) is the most important source of silt-sized particles in Peoria Loess of last-glacial age (Aleinikoff et al. 1999, 2008; Muhs et al. 1999, 2008). Tephra is commonly composed of silt-sized particles, and where there are active volcanoes, silt-sized tephra is one of the components of loess, such as is the case in Holocene loess in southern Alaska (Muhs et al., 2004). Thus, in both the "glacial" and "desert" models of silt particle formation, inheritance from tephra or siltstone can be added as processes (Fig. 4).

It is therefore likely that loess in many regions has origins from both glacial and nonglacial processes. There is no question that glaciers are efficient producers of silt. Drainage basins in Alaska that have glaciers have significantly higher sediment yield than nearby basins that lack glaciers (Hallet et al. 1996). In Iceland, several meters of loess, mostly from glaciogenic sources, have been produced entirely within Holocene time. On the other hand, laboratory experiments demonstrate that eolian abrasion and fluvial comminution are very efficient silt producers (Wright 2001b), and it has been shown that at least some component of loess in deserts is derived from abrasion of sand-sized

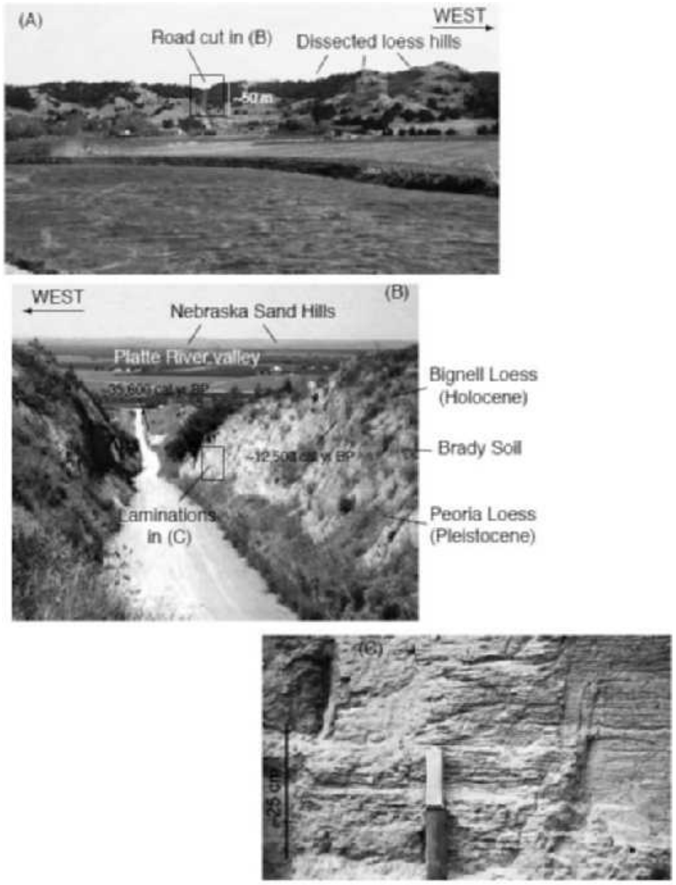

FIG. 3.—Bignell Hill, in the Loess Hills area of western Nebraska, where a road cut (A) exposes the thickest known exposure (~48 m) of last-glacial (ca. 35 ka to ca. 12 ka) loess in the world, overlain by ~2 m of Holocene loess (B). Middle part of last-glacial loess displays rare occurrence of primary structures (laminations) in loess (C). See Muhs et al. (2008) for details of Bignell Hill locality; all photographs are by D.R. Muhs.

particles by wind to silt-sized particles (Crouvi et al. 2010). This issue is relevant for studies of Martian dust and dust deposits because some processes of silt production on Earth do not occur now on Mars and perhaps only operated billions of years ago.

Geologic Records of Loess and Dust

The geologic record of loess on Earth consists of sequences of relatively unaltered loess with intercalated buried soils, or paleosols. Soil formation takes place when vegetation stabilizes the surface, usually because of a decrease in sedimentation rate. Such conditions may arise because of a decrease in wind strength, elimination of the loess–sediment source, or both. In Europe and much of North America, where loess has a glaciogenic origin, it is no surprise that loess in the stratigraphic record represents glacial periods, and paleosols represent interglacial periods (Muhs and Bettis 2003, Muhs 2007). In China, although the immediate sources of loess are desert basins upwind from the Loess Plateau, the loess is actually of glaciogenic origin, derived from mountain glaciers that surround the desert basins (Sun 2002). Thus, in China as well as in regions of Europe and North America near glaciated terrain, loess represents glacial periods, and paleosols represent interglacial periods that can be correlated with the deep-sea oxygen isotope record of glacial–interglacial cycles. Elsewhere in North America (e.g., the Great Plains), where loess is nonglaciogenic, ages of loess also correspond to glacial periods (Muhs et al. 1999, 2008; Roberts et al. 2003). Therefore, although loess in this region is not dependent on glacial sources, climatic conditions during glacial periods are apparently favorable for loess entrainment, transportation, and deposition.

Geologic records of aerosolic dust flux occur in a number of depositional settings for sediments, including deep-ocean basins, lakes, large ice sheets, and soils. Deep-sea sediments are one of the great archives of glacial–interglacial cycles of the Quaternary. Foraminifera in deep-sea cores faithfully record the ocean temperature changes and fluxes in the oxygen isotope composition of seawater that accompany the shifts between glacial and interglacial climates. The same deep-sea cores may also retain long-range–transported aerosolic dust, if fluvial transport and other means of sediment origin can be ruled out or accounted for. Several decades of study have shown that both the Pacific and Atlantic Oceans have long records of eolian sediment transport from Asia and Africa, respectively (Rea 2007). The coincidence of dust flux maxima with glacial periods indicates that glacial-climate conditions favored greater dust production and transport.

Thus, loess records from both glacial and nonglacial sources, the eolian record in deep-sea sediments, and the eolian record in polar ice caps all indicate that glacial periods on Earth correspond to times of greater dust flux. It is likely that stronger winds, lower vegetation density, drier climates in dust source areas, and a decreased intensity of the hydrological cycle all combined to enhance dust flux repeatedly in glacial periods of the Quaternary (Mahowald et al. 1999). During such times, which approach conditions on Mars, Earth was a far dustier planet than it is today. Quaternary glacial–interglacial cycles on Earth are controlled, to a great extent, by astronomical cycles of precession and tilt of Earth's axis and, to a lesser extent, orbital eccentricity (Berger 1992). These cycles are particularly critical for the distribution of solar insolation in summer and therefore affect ice-sheet growth and decay in the Northern Hemisphere of Earth. As discussed later herein, astronomical cycles also exert controls on Martian dust cycles.

THE ORIGIN OF MARTIAN DUST

The first geologic process discovered on Mars is also its most pervasive: Global dust storms were first observed telescopically in the nineteenth century and tend to occur annually (Martin et al. 1992, Newman 2001). Spacecraft observations show that smaller-scale dust storms and vortices are also frequent. It is estimated that 2.9×10^{12} kg yr^{-1} of dust is exchanged between the surface and atmosphere (Pollack et al. 1977). If evenly coated over the surface, this would form a layer just a few micrometers thick. However, as on Earth, the distribution and thickness of dust are nonuniform. Based on radar measurements, accumulations of dust in high-albedo, low-thermal inertia regions are estimated at up to 1 to 2 m from radar measurements (Christensen 1986), and in Arabia, thicknesses of 20 m and greater are based on the lack of small impact craters (Mangold et al. 2009) (a map showing Arabia and other Mars locations presented in the text is shown in Fig. 5). The polar layered deposits at both poles, with a thickness of a few kilometers, have a significant dust component (Kahn et al. 1992, Thomas et al. 1992, Smith et al. 1999, Phillips et al. 2008). Fundamental questions include: How did so much dust form, and how is it preserved on the surface today?

Based on remote sensing and in situ measurements, the composition of Martian dust is dominated by framework silicates, mostly feldspar, with lesser amounts of sulfate, pyroxene, olivine, amorphous phases, hematite, and magnetite—a mineralogy consistent with the mechanical

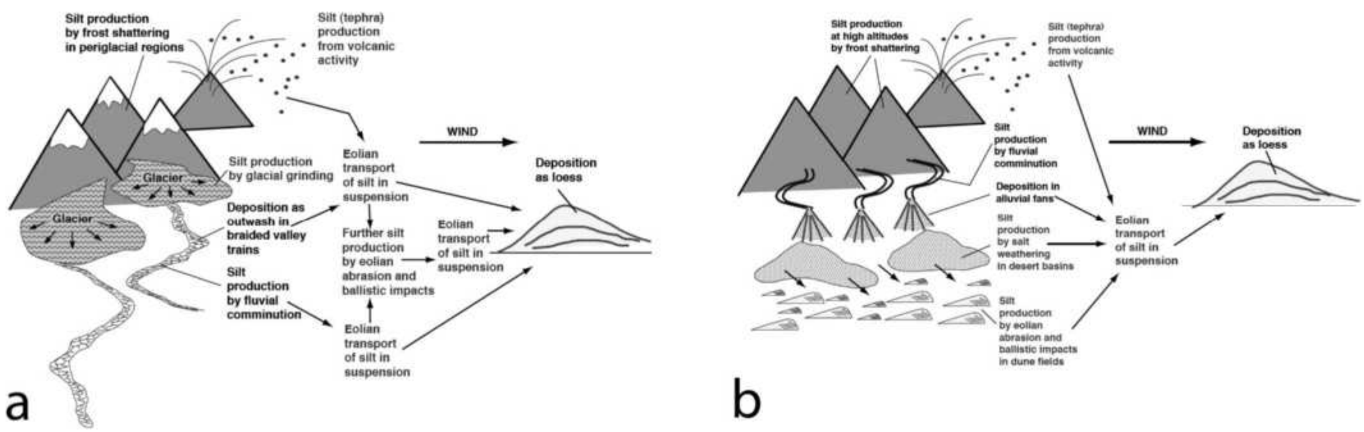

FIG. 4.—(a) Classical model of "glacial" loess formation on Earth wherein silt-sized particles are produced primarily by glacial grinding, delivered to outwash streams, and finally entrained by wind. (b) Model of "desert" loess formation on Earth wherein silt-sized particles are produced by a variety of nonglaciogenic processes before eventual entrainment by wind. Processes hypothesized to be active on both Earth and Mars are shown in gray text. Redrawn from Muhs and Bettis (2003).

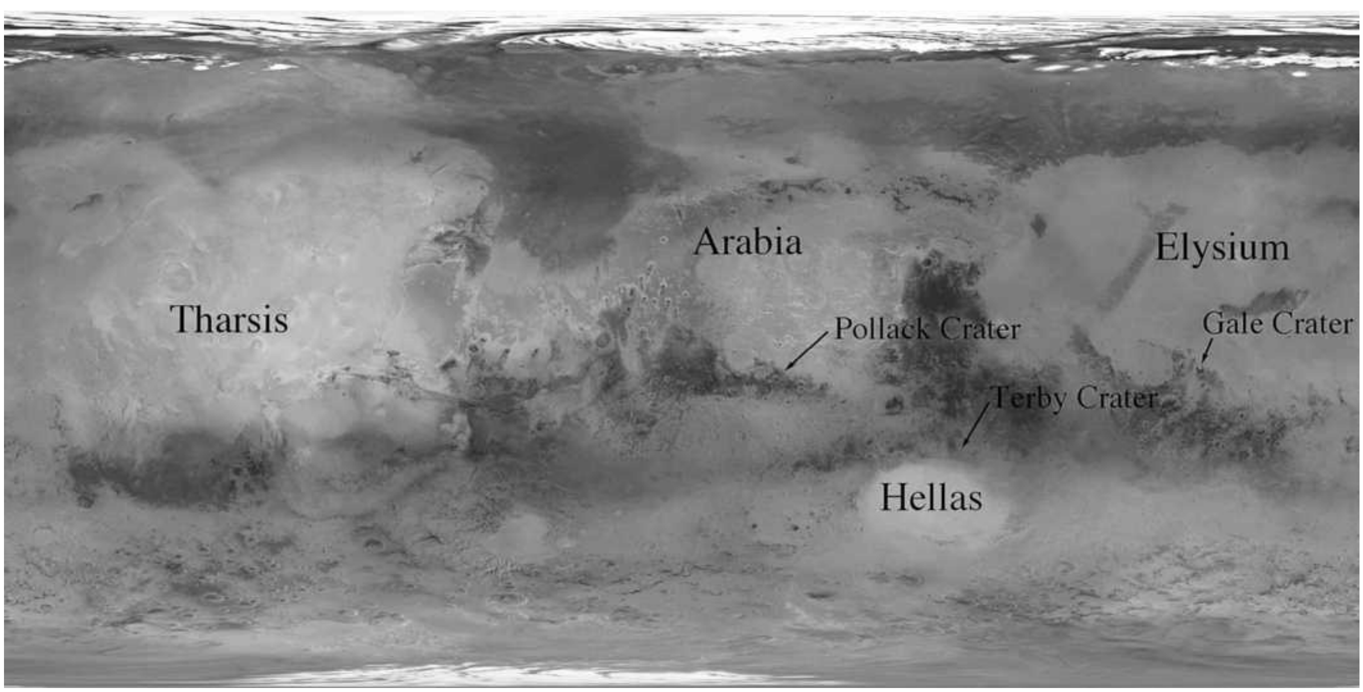

FIG. 5.—Mars map showing geographic features mentioned in the text.

weathering of basalt (Hamilton et al. 2005). This is distinct from the dominantly granitic (or rocks derived from granites) source for the majority of terrestrial loess. The presence of olivine in Martian dust in particular indicates that water did not play a dominant role in dust formation, with genesis occurring during dry periods of Martian history (Goetz et al. 2005).

There are several processes that can potentially produce Martian dust (Fig. 6). Mars has abundant volcanoes and lava flows, some of which have a morphology consistent with pyroclastic eruptions, such that ancient tephra was probably a significant contributor. Impacts have pulverized the surface, releasing large quantities of ejecta, including dust-size material. Meteoritic infall provides a steady source of particulate materials. Based on the impact cratering record, ~8.3 cm of material has been added to Mars, if distributed evenly over the surface, since the early–mid Noachian (>4 Ga) (Yen et al. 2006). The flux of nonmelted meteoritic material that reaches the Martian surface is estimated at 8.6×10^6 kg yr^{-1} (Flynn 1996), yielding ~8 cm of material (assuming basalt density), consistent with the cratering-derived estimate. Erosion associated with outflow channels, valley networks, other fluvial features, and possibly glacial grinding has disaggregated rocks and coarse soil components.

All of these processes were largely confined to the Noachian and

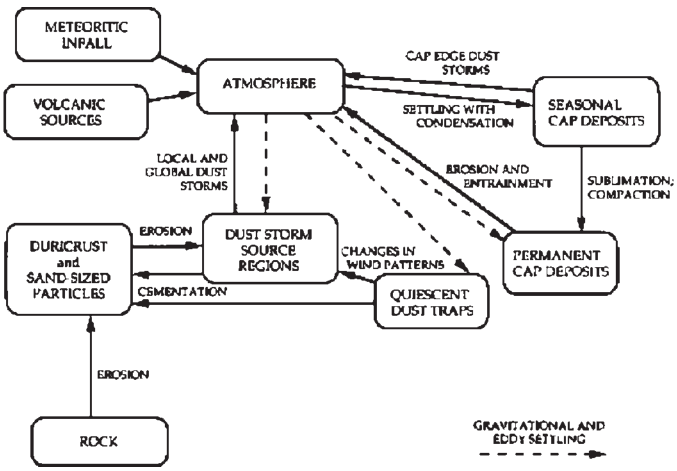

FIG. 6.—Schematic diagram of the Martian dust cycle. Reprinted with permission from Kahn et al. (1992).

Hesperian Periods before ca. 3.5 to 1.8 Ga (Tanaka et al. 1992). These high-energy processes (except for minor meteoritic infall) do not occur today, such that current rates of dust production are considerably lower. Probably the most significant dust-forming mechanism in the modern era is the abrasion and deflation of the relatively soft rock that constitutes yardangs and eolian mantles over much of Mars. Much of this material is likely duststone, the erosion of which recycles dust back into the Martian eolian system. More minor processes include eolian abrasion of rocks and soil by sand, rounding and disintegration of this sand in saltation and abrasion, thermal cycling of rocks, and possibly salt weathering (Wells and Zimbelman 1997). On Earth, precipitation cleanses dust from the atmosphere, and oceans and other water bodies serve as permanent sinks. On land, vegetation acts to stabilize dust and form soils. With Mars lacking these characteristics, dust is not easily removed from the eolian system.

MARTIAN DUST TRANSPORT AND DEPOSITION

Numerous wind tunnel and theoretical studies have demonstrated that the threshold for lofting dust that is ones to tens of micrometers in size by wind, regardless of the planetary surface, exceeds that needed for fine sand (Greeley and Iversen 1985). This is because dust is more easily sheltered by roughness elements, it resides partially within a laminar sublayer, and its low weight makes it more susceptible to electrostatic and other cohesive forces (Iversen et al. 1976, Shao and Lu 2000). Friction speeds needed to raise dust on Mars are on the order of 10 m s^{-1}, which is about an order of magnitude greater than on Earth (Greeley and Iversen 1985). These speeds correspond to free-stream winds over 100 m s^{-1} up to several kilometers per second—conditions rarely, if ever, achieved. So, under standard conditions, linear winds on their own cannot raise dust, and other mechanisms must be considered to account for the significant load of suspended dust in the atmosphere.

Some of the highest wind speeds on Mars are attained by katabatic flow off the retreating seasonal polar cap, combined with diurnal winds flowing from high elevations (Lee et al. 1982, Magalhaes and Gierasch 1982, Kahn et al. 1992). The magnitude of these winds is enhanced in the Southern Hemisphere, where the proximity of perihelion in late spring to the summer solstice produces strong gradients (Wells and Zimbelman 1997). These winds can exceed the threshold friction speeds of ~2 m s^{-1} necessary to saltate sand. The flux of dust lofted by saltating sand is proportional to $(u^*/u^*_t)^3$ (White 1979, Newman et al. 2002) or $(u^*/u^*_t)^4$ (Westphal et al. 1987, Haberle et al. 2003), such that even winds slightly above sand threshold can cause large dust fluxes. Vortices induced by surface–atmosphere thermal instabilities can also saltate sand and provide enhanced lift to raise dust, producing dust devils (Greeley et al. 2004). More minor contributors to dust suspension include removal of dust on pebbles and rocks, which, being higher in the wind speed profile and not sheltered by nearby roughness elements, are subjected to greater shear stress (White et al. 1997), and small-scale surface disruptions such as dust-rich landslides (Sullivan et al. 2001). On the other hand, dust aggregates with sizes equivalent to sand are easily moved (Fig. 7). This likely accounts for the presence of ripples in dust-rich regions (Bridges et al. 2010) and is a component of the duststone model that will be presented herein.

In local events, such as dust devils, dust is transported only a few kilometers. However, global storms can raise dust as high as 30 to 40 km or more, where it is subjected to transport by Hadley cells from the southern latitudes in the summer to the middle and temperate northern latitudes (Toon et al. 1980, Pollack et al. 1981, Magalhaes 1987, Haberle et al. 1993). In the latter stages of global storms, the high-altitude dust acts to dampen the lapse rate (temperature change with altitude), decreasing convective turbulence and causing dust to fall out (Haberle 1986, Wells and Zimbelman 1997). Grains can also serve as condensation nuclei for CO_2 and H_2O, enhancing dust deposition in polar regions (Kahn et al. 1992).

On Mars today, the coincidence of perihelion with southern spring results in global storms originating in the Southern Hemisphere and net dust deposition in the equatorial and northern latitudes via the cross-equatorial Hadley cell. Dust is spread nonuniformly, with enhanced deposition in preexisting high-albedo, low-thermal-inertia, low-rock-abundance regions, specifically, Tharsis, Elysium, and Arabia (Fig. 5). Why these regions serve as dust sinks is not completely understood. One hypothesis is that they may have initially formed in areas of low wind velocity, allowing dust to bury sand and rocks, thereby removing sand saltation triggers and roughness elements that could enhance turbulence to otherwise drive dust into suspension, thereby increasing the net accumulation rate (Christensen 1988).

The argument of perihelion has a period of ~51,000 years, such that the hemisphere where global storms originate changes every 25,000 years or so, with alternate epochs perhaps having a net north–south transport (Christensen 1986, Kahn et al. 1992, Kieffer and Zent 1992). Under this model, net dust sources and sinks alternate in hemispheric polarity in response to orbital precession. If true, then remnant, thick dust/duststone deposits may exist over much of Mars. These and other hypotheses of Martian dust removal and deposition are simplifications, as terrestrial experience and detailed models show that erosion and deposition are fundamentally driven by the mass balance per unit volume of surface (Exner 1925, Kubatko and Westerink 2007). In detail, it may be that some areas of "dust sinks" on Mars undergo periods of erosion and abrasion, and some "sources" may locally accumulate sediment. In our study of Mars, we are presently limited to observations that reflect broad-scale processes integrated over long time periods.

EVIDENCE FOR DUSTSTONE

Three classes of observations provide compelling evidence for the existence of duststone on Mars. Those are outlined next, along with interpretations based on these observations.

(1) Thermophysical and Geochemical Properties in Some Regions of Mars Are Consistent with Duststone

Observations: Cohesion and geochemical properties are correlated in Martian soil, with cohesive cloddy materials and surface crusts enriched in sulfur and chlorine (Clark et al. 1982; Rieder et al. 1997, 2004; Moore et al. 1999). The formation of this "duricrust" is believed

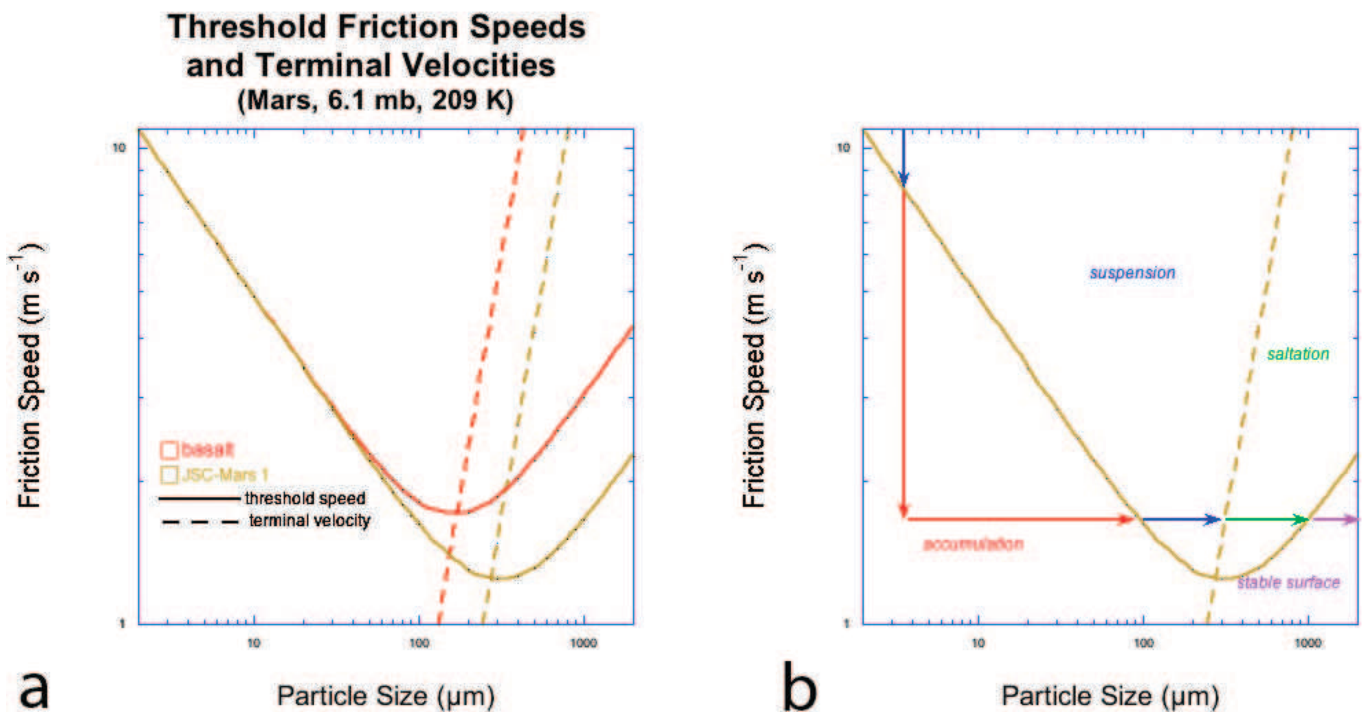

FIG. 7.— (a) Friction speed versus particle size for basalt and JSC-Mars-1, a Martian soil stimulant (Allen et al. 1998) with properties similar to that of Martian dust (Bridges et al. 2010). The solid curves are calculated threshold friction speeds using the parameterization of Shao and Lu (2000). The dotted lines represent the terminal fall velocity that theoretically corresponds to the suspension–saltation boundary. Details of these calculations are in Bridges et al. (2010). (b) Curves for JSC-Mars 1, with labels designating the dominant dynamic behavior of the particles within zones bounded by the threshold curve and terminal fall line. The bulk density of the particles is assumed to be constant. When winds fall below threshold, dust comes out of suspension and accumulates. Similarly, any saltating sand will stop moving, resulting in a stable surface. Arrows represent the evolution of dust particles: (1) Dust stays suspended in the atmosphere, falling to the surface in quiescent periods when friction speeds drop below threshold (blue and red vertical arrows, with blue showing zone at which the dust can be resuspended and red representing lower wind speeds for which dust accumulates). (2) Dust on the surface clumps into aggregates (red horizontal arrow). (3) The dust grows in size until it is above the threshold curve and can be resuspended (blue horizontal arrow). (4) Eventually the aggregates reach sufficient size such that they are too heavy to get suspended (so below the terminal velocity line, yet above the threshold curve). These particles, shown by the green arrow, will undergo saltation. (5) Further aggregation will again put the particles below the threshold curve, such that they remain stable (purple arrow). Cementation and burial harden the aggregates into "duststone."

to occur via the mobilization of S and Cl by water-vapor diffusion through the soil (Jakosky and Christensen 1986a) and may account for variations in thermal inertia outside the lowest inertia areas (Jakosky and Christensen 1986a, 1986b; Christensen and Moore 1992). Mapped at coarse scale by the Thermal Emission Spectrometer (TES), duricrust is not exposed everywhere, but rather it makes up one of four distinct thermophysical units, with the corresponding "Unit C" having moderate to high thermal inertia and intermediate albedo (Mellon et al. 2000). This unit consistently borders "Unit A," in which low inertia and high albedo indicate a dust cover. Tharsis is relatively rich in Cl and depleted in Si and Fe in its upper ~50 cm, as indicated by Gamma Ray Spectrometer (GRS) data (Karunatillake et al. 2009).

Interpretations: The thermophysical properties of Unit C, combined with its association with Unit A, are consistent with thick cemented dust (Mellon et al. 2000). In the dusty Unit A regions, a dust cover thicker than a couple of centimeters (the diurnal skin depth) can mask any underlying indurated material. Therefore, cemented dust could exist in these regions as well, or be exposed as outcrops at spatial scales finer than the TES data (~3 km per pixel). In support of this hypothesis, Tharsis, within Mellon et al.'s Unit A, exhibits large seasonal variability and diurnal differences in the apparent thermal inertia that can be attributed to a layered surface, such as loose dust covering an indurated layer, although horizontal mixing and subpixel slope effects are other possible causes (Putzig and Mellon 2007a, 2007b). Finally, in Tharsis, the chlorine-rich composition detected by GRS is consistent with an indurated soil composition.

(2) Dust Aggregates, the Precursors to Duststone, Are Common on Mars

Observations: Images from the Microscopic Imager on the Mars Exploration Rovers (MERs) show dust aggregates that reach sizes of several hundred micrometers (Herkenhoff et al. 2004, 2006, 2008; Sullivan et al. 2008), equivalent to that of sand. The much thinner and drier atmosphere on Mars makes dust much more prone to electrostatic forces than in the Earth environment (Greeley 1979, Krinsley and Leach 1981), such that aggregates and "cloddy material" (as classified at the *Viking* landing sites) can form (Shorthill et al. 1976, Greeley 1979). In addition to those at the landing sites, aggregates may make up low-thermal-inertia dunes and materials of the north polar erg (Herkenhoff and Vasavada 1999), a region also proposed to contain

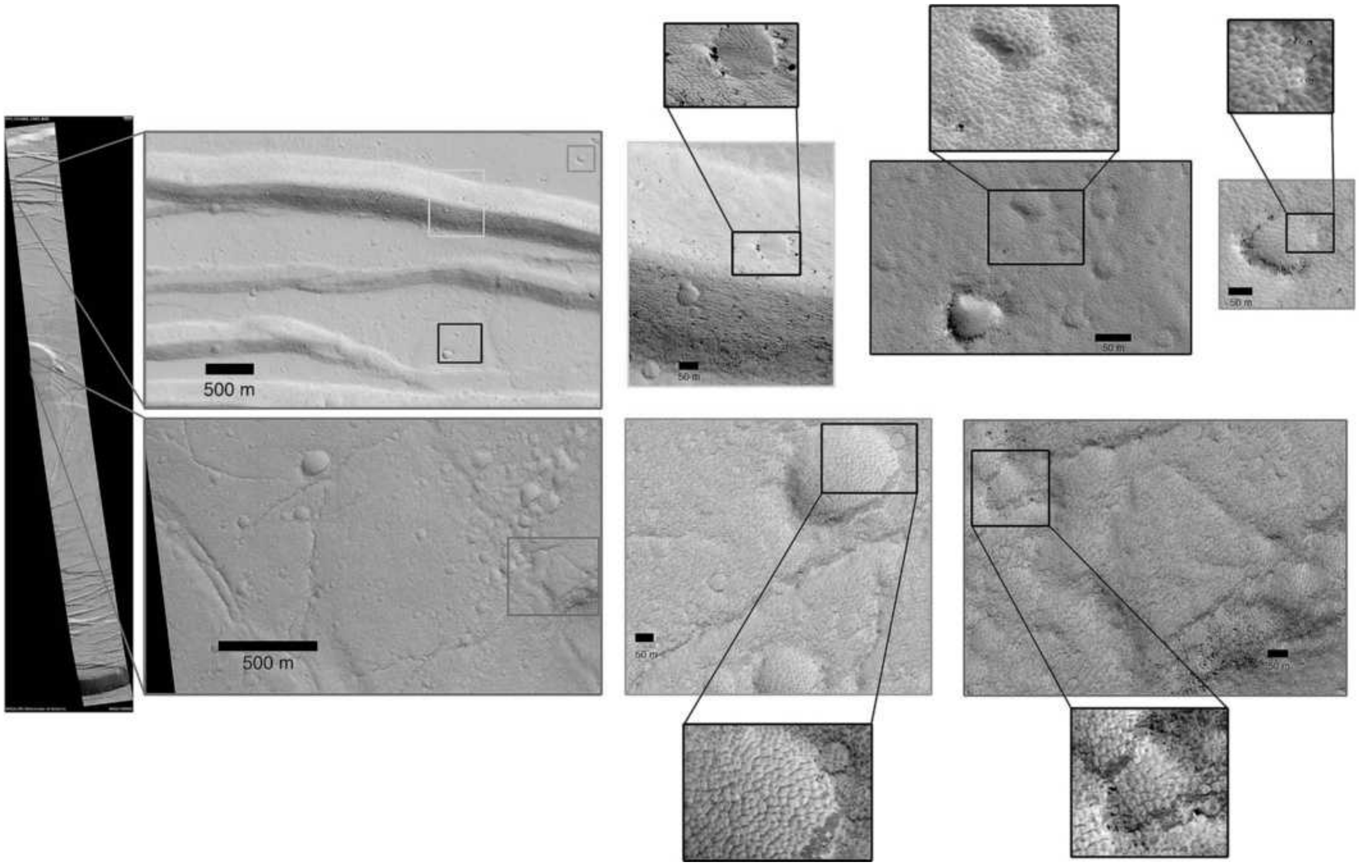

FIG. 8.—Reticulate bed forms within the Olympus Mons caldera. Note that the bedforms superpose underlying structural, volcanic, and crater topography. HiRISE footprint is at left. Boxes at left show locations of enlargements to the right. Black boxes within those show outline of stretched enlargements above and below (HiRISE image PSP_004966_1985).

"loess" deposits (Tanaka et al. 2008). Layered deposits in the Hellas Basin and nearby Terby Crater have been proposed to form from fine-grained material such as dust or loess (Wilson et al. 2007). It has also been conjectured that "parna sheets," analogous to clay aggregate lunette duneforms on Earth, may exist on Mars, although none have been definitively identified (Greeley and Williams 1994).

Interpretation: Dust aggregates are common on Mars, especially in areas for which remote-sensing observations indicate dust-rich compositions. Aggregates will continue to grow until they are disrupted by wind forces. In other words, dust that is ones to tens of micrometers in size that is found in the atmosphere is not likely to remain in this state on the surface, but rather to clump together with other dust particles and aggregates.

(3) There Is Geomorphological and Remote-Sensing Evidence for Thick Dust Deposits

Observations: High-resolution orbital images from the High Resolution Imaging Science Experiment (HiRISE) camera show that regions on Mars classified as dusty–Unit A (e.g., Tharsis, Arabia, and Elysium) contain suites of fine-scale reticulate bedforms intermixed with wind-sculpted, bright bedrock (Bridges et al. 2010) (Figs. 8, 9). The bedrock is commonly composed of blunt-to-tapered outcrops for which plan-view shape and texture are analogous to terrestrial yardangs. Earth-based radar measurements restrict the depth of a cover of fines to 1 to 2 m on average (Christensen 1986), consistent with a dusty surface above a more indurated substrate. In Tharsis, images of stratigraphic cross sections showing contacts with underlying basalt demonstrate that the light-toned, presumably dust-rich, bedform surfaces sit atop a mantle at least several meters thick (Keszthelyi et al. 2008) (Fig. 10). In Arabia, thickness of dusty material is estimated at 20 m or more (Mangold et al. 2009).

Interpretations: Because the thermophysical properties of these areas indicate a dust-rich composition, and particles with diameter of ones to tens of micrometers are suspended by winds above threshold, Bridges et al. (2010) proposed that the reticulate bedforms were formed from the saltation of dust aggregates hundreds of micrometers in size (Fig. 7b). Furthermore, because aggregates approaching a millimeter in size are below threshold, it was also suggested that continued growth would also act to stabilize the particles. This, combined with the formation of duricrust and subsequent burial by younger dust deposits, would form a semidurable "duststone" that, when exposed, would be subjected to wind abrasion processes.

The putative duststone rock may exist in other areas of Mars (Fig. 11). Buried and exhumed terrain is common on the planet (Malin and Edgett 2001, Frey et al. 2002, Edgett 2005), and much of the middle latitudes is covered by a mantle composed of ice-permeated dust that is many meters thick (Mustard et al. 2001). In some craters that were probably wholly or partially exhumed, light-toned rock masses are apparent. This includes Pollack Crater (containing "White Rock"; Fig.

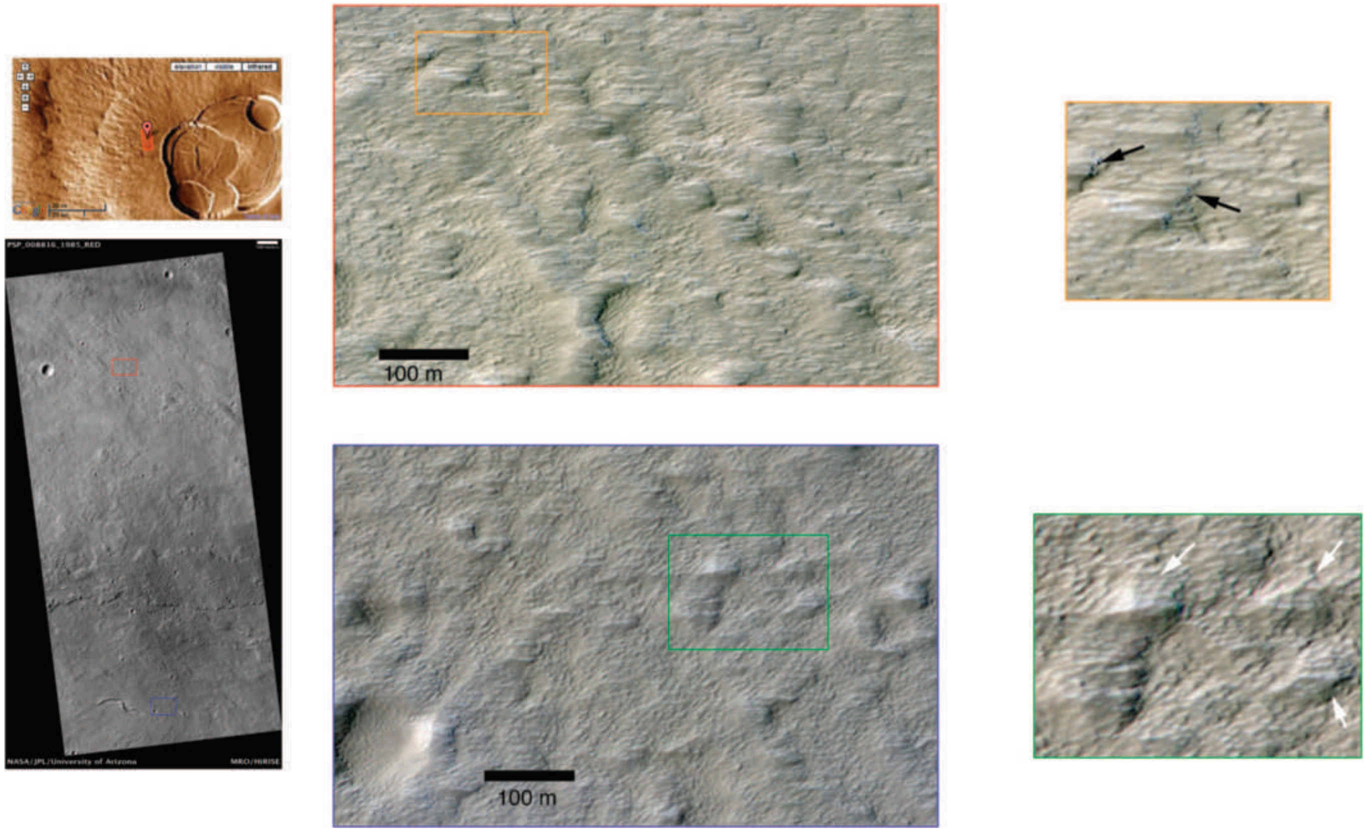

FIG. 9.—Details of yardangs on the western flank of Olympus Mons in HiRISE image PSP_008816_1985. Note the blunt to tapered edge of most yardangs and the wind tails in the lee of rocks, going from right to left, consistent with downslope winds. Black arrows in the upper right subframe show rocks and linear ridges to the west interpreted as wind tails. The white arrows in the lower right subframe show yardangs. Upper-left frame shows the location of the HiRISE footprint, at lower left.

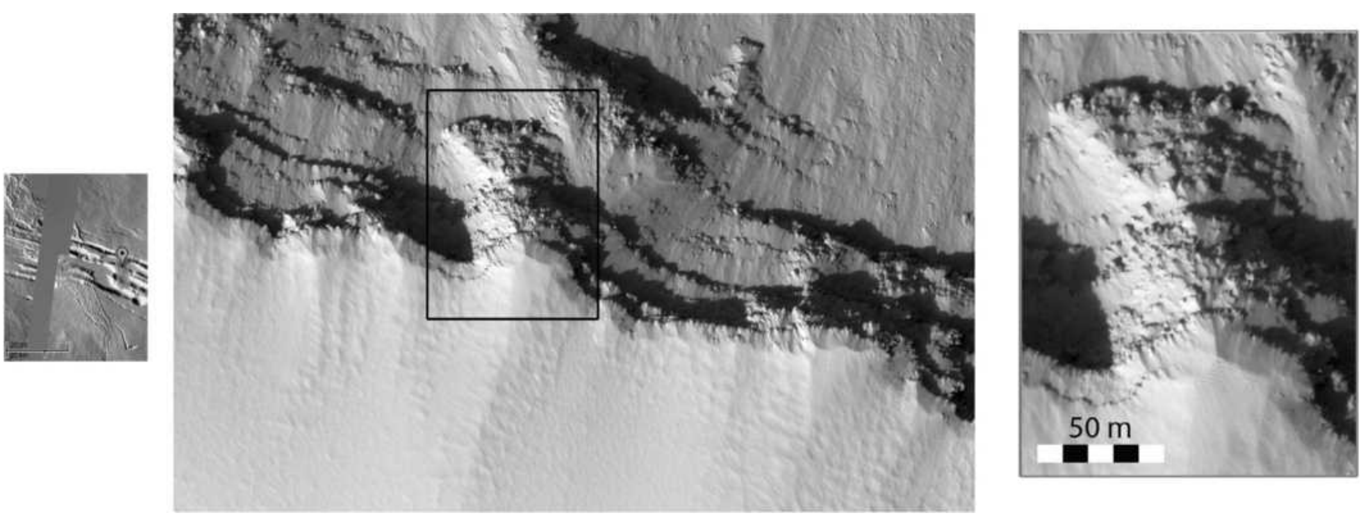

FIG. 10.—HiRISE image PSP_005651_1840 at the edge of the northern flank of Pavonis Mons. Left frame shows the location of the HiRISE footprint. Middle image shows details of bedrock outcrops, with enlargement of the red boxed area at right showing a thick dust cover.

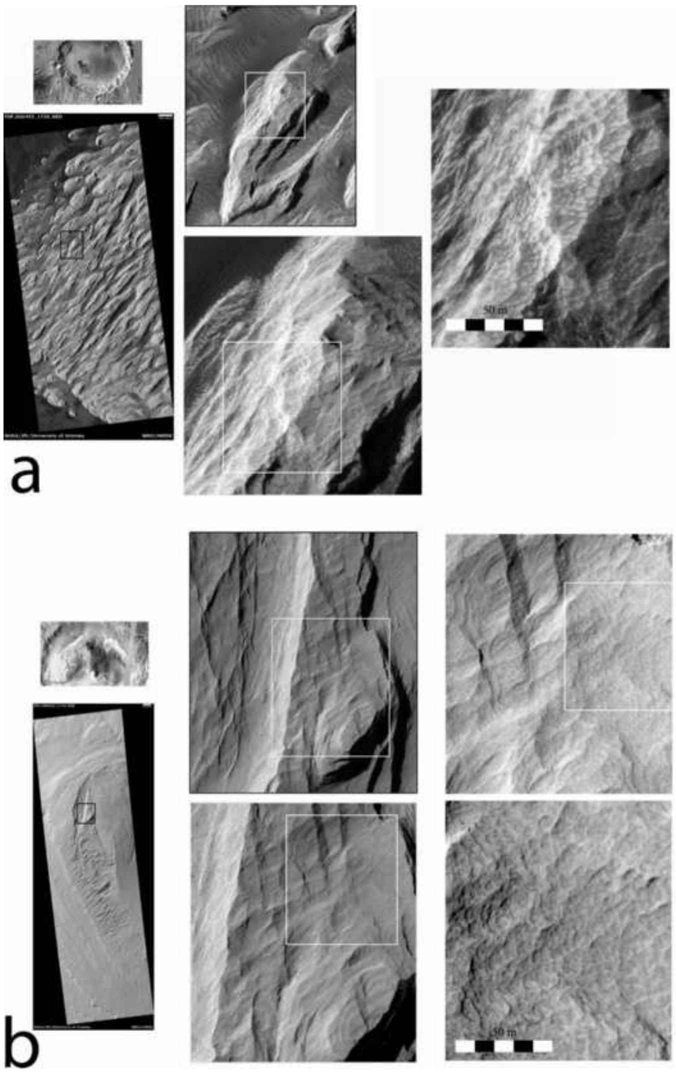

FIG. 11.—Two examples of possible duststone outside of low-thermal-inertia regions of Tharis, Arabia, and Elysium: (a) "White Rock" in Pollack Crater. Upper-left frame shows the location of the HiRISE footprint at lower left. A succession of boxes shows increasing detail of the rock texture (HiRISE image PSP_002455_1720). (b) The upper mound of Gale Crater. Upper-left frame shows the location of the HiRISE footprint at lower left. A succession of boxes shows increasing detail of the rock texture (HiRISE image PSP_008002_1750). Note that both White Rock and the Gale upper mound have a scalloped surface, with patterns similar to the reticulate bed forms and scoured surfaces seen in Figures 8 and 9, respectively.

11a) and the upper rock masses of the Gale Crater mound (Fig. 11b), both of which have remote-sensing properties consistent with indurated dust (Ruff et al. 2001, Milliken et al. 2010). A light-toned outcrop in Becquerel Crater and the upper Gale mound also contain nonconformable, rhythmically bedded rock masses with stratigraphic bundles consistent with astronomically-forced mechanisms (Lewis et al. 2008, Milliken et al. 2010) such as is expected for cyclical air-fall dust.

THE MARTIAN DUSTSTONE MODEL

The large volume of dust on Mars and the unique environment of its deposition compared to Earth account for the formation of duststone on that planet as opposed to our own. We present a hypothesis, modified after Bridges et al. (2010), for the formation of duststone that we believe is consistent with current data and suggest several measurements that can be performed by future missions to test it. The major stages of formation are:

(1) Dust is deposited on the Martian surface from atmospheric fallout. In "sinks," accumulation exceeds removal, and thick deposits form. Dust clumps into aggregates via electrostatic processes. It eventually reaches a large enough size that it can no longer be suspended when subjected to high-speed winds and instead saltates and forms small-scale ripples, such as those currently seen in the dusty Tharsis, Elysium, and Arabia regions. Many aggregates will be destroyed and comminuted to smaller particles. Depending on the pore space of the growing aggregates, others may remain above the threshold curve and be removed by the winds. However, a fraction, those with great enough density and those not destroyed, will grow to sufficient size such that they are below the threshold curve, resulting in a stable surface. This stable population increases over time (Fig. 7b).

(2) Once the surface is dominated by particles having a size below threshold, it becomes further stabilized through the formation of duricrust through multiple cycles of water-vapor diffusion (Jakosky and Christensen 1986a).

(3) Some stabilized surfaces will be buried by new dust deposits, which are, in turn, stabilized as well. Compaction of the underlying dust converts the cemented material into harder duststone.

(4) If subsequently exposed, the duststone will be subjected to wind abrasion, depending on the thickness and extent of local particulate materials. In many settings, this will be variable, as seen in Tharsis today, with some exposed duststone undergoing erosion and other units being sheltered. A similar scenario is envisioned for putative duststones in craters such as Pollack and Gale, although in these cases the nature of the former burying material is less certain as these areas are not currently dust sinks, although they may have been in past epochs. Sand serves as the main abrasive agent where it is present, given the much greater effectiveness of its ability to erode rock (Laity and Bridges 2009). However, in areas like Tharsis, where sand appears lacking, duststone should only be abraded if it is weakly cemented, because the main abrading material is probably dust aggregates.

(5) Over time, lithified duststone is eroded, forming scalloped yardangs and remnant rock masses. Many of the eroded particulates probably rejoin the eolian system as suspended grains or as saltating sand-size particles that are incorporated into the soil or break apart and go into suspension.

This model does not consider the role of ice. It has been proposed that the desiccated mantle covering much of the Martian midlatitudes is air-fall dust cemented by ice derived from diffused atmospheric water vapor under favorable orbital configurations (Mustard et al. 2001). This is analogous to the polar layered deposits being composed of dust–ice mixtures (Thomas et al. 1992). Such an ice-rich mantle is more like permafrost than bedrock, although desiccation is predicted to produce loose loess-like materials (Mustard et al. 2001), so that remnant masses could be considered duststone.

Finally, not all light-toned rocks on Mars are duststone. For example, the morphology of the wind-eroded Medusae Fossae Formation and its location near the Tharsis and Elysium volcanic constructs are consistent with a pyroclastic origin (Ward 1979, Scott and Tanaka 1982, Bradley et al. 2002, Mandt et al. 2008, de Silva et al. 2010). Fluvial processes have probably contributed to the formation of the lower parts of the light-toned layer mound in Gale Crater (Milliken et al. 2010, Thomson et al. 2011). It is likely that other light-toned rocks on Mars have similarly diverse origins. Our hypothesis is that many, but not all, light-toned rocks, especially those associated with thick dust deposits, are duststone.

CONTRIBUTION TO THE MARTIAN ROCK RECORD

The rates of erosion of duststone, and the ways in which these balance against rates of formation, are not known. Given that the dust cycle on Mars seems to have been occurring since the Noachian, it is likely that cycles of deposition, stability, and erosion have been ongoing ever since. Duststones will be protected from erosion for the duration over which they are buried, such that there could be very old formations beneath the surface today or recently exposed formations. Given the low rate of primary dust production on Mars in the current epoch, the rates of duststone formation versus erosion are directly tied into the question of whether dust abundance overall is increasing or decreasing. Answers to this and related questions cannot be found until more data on rock formation and exposure ages are provided by samples or in situ measurements.

With greater confidence we can say that the range of rock types currently forming on Mars is far more limited than it was in the Noachian and Hesperian. Volcanic and fluvial processes that produced suites of igneous and water-based sedimentary rocks are infrequent today. Thus, any preserved Noachian and Hesperian duststones are probably intermixed with strata formed by other geologic processes. Given their air-fall nature, duststones could serve as temporal marker beds analogous to ash and ash-derived clays on Earth. In the modern era, Martian rock production is limited to the movement of sand and the dust cycle, combined with processes to stabilize and lithify these materials. We therefore consider duststone, perhaps with limited sandstone production, as the dominant rock-forming processes in the current era. This idea can be tested by future in situ missions that visit areas believed to contain duststone.

COMPARISONS BETWEEN THE DUST CYCLES ON EARTH AND MARS, AND CONSIDERATIONS OF OTHER PLANETARY BODIES

When we define the dust cycle as describing the production of particles with diameters of ones to tens of micrometers, followed by their deposition and removal or incorporation into the geologic record, there are important commonalities and differences between Mars and Earth. Building upon the previous sections, we compare and contrast the dust cycles on the two planets and conclude with some general statements about the role of dust on other planetary bodies. Some of the ideas here are broad generalizations and are presented as working models that, we hope, will spur further research into the role of dust in the planetary context.

Some geologic processes that Earth and Mars have in common are those that act to comminute rock into finer material that can subsequently be suspended and deposited into accumulations of many meters or more. Although hydrological and particularly glacial processes have been more prevalent on Earth, with the latter considered one of the main producers of loess, the mechanism by which dust is formed has little impact on its subsequent evolution. In other words, it is sufficient to state that dust is formed in abundance on Earth and Mars, and that the modes of deposition, stabilization, accumulation, and removal are the steps in the process where environmental effects exert the strongest biases.

The deposition of dust on Mars is much more uniform than on Earth, where thicknesses are correlated to distance from source. Because Martian dust can be transported vast distances without being flushed out by precipitation or being affected by ocean-induced global circulation patterns, deposition can occur on global scales. Even the net south–north hemispheric bias in the current epoch is counterbalanced, over the long term, by the Martian precessional cycle.

Processes that remove dust compete against those acting to stabilize it, and these mechanisms are significantly different on Mars and Earth. Wind can obviously remove dust on both planets, but the much higher threshold speeds needed in Mars' low-density atmosphere make areas of low wind intensity or surface roughness much more prone to serve as dust sinks than is the case on Earth, where dust in virtually all locations can be resuspended. On Earth, fluvial erosion is an additional and significant removal agent and one that is completely absent on Mars. Therefore, unconsolidated dust is much more likely to remain stable on Mars.

On the other hand, dust stabilization on Mars seems to require some combination of aggregation, cementation, and burial. No liquid water is involved. In contrast, the active terrestrial hydrosphere and biosphere acts to alter primary dust and lesser amounts of sand and clay into a weakly cemented sediment, commonly bound by carbonate cement and clays. Such a "loessification" process is not expected on Mars today. Pre-Quaternary loessite, though probably requiring water for its diagenesis, may be a closer analog to Martian dust deposits in that it is better cemented and buried.

The erosion of dust deposits on the two planets is likewise quite different. Although loess can be derived from arid settings on Earth, it rarely accumulates in significant amounts in deserts, tending to get transported to areas where it can be stabilized by vegetation. Subsequently, pedogenesis begins, and soil formation dominates until further loess deposition takes place. Even pre-Quaternary loessite has identifiable paleosols (Soreghan et al. 1997, 2002). As such, the erosional expression of loess on Earth is poorly defined geomorphically, although there is no question that there are stratigraphic records of loess erosion seen as unconformities (see Muhs et al. [2003] for examples in Alaska). This is in marked contrast to Mars, where all dust accumulates in dry settings, and subsequent erosion is confined to eolian abrasion and deflation, forming yardangs. In Earth's deserts, yardangs can be composed of sandstone (quartz, carbonate, or gypsum-dominated) or fine-grained sediments, such as silts and clays.

The role of dust on Mars is unique in the Solar System. Unlike on Earth, it is almost entirely unaffected by hydrological and biological alteration processes that can accumulate into thick sequences. Once deposited, it becomes part of the sedimentary strata, as dust or duststone, or reenters the eolian system through deflation or abrasion. On Earth, about 23% of the surface is sandstone (Ronov 1983), or about twice the amount of loess coverage. Given the dustiness on Mars, the volume of duststone could be as proportionally significant as sandstone is on Earth. This can be verified or refuted by future missions and data analysis.

The differences between Mars and the other worlds in our Solar System with solid surfaces and significant atmospheres (e.g., Venus and Titan) are even more distinct. Venus lacks a hydrosphere, and dust production from volcanic sources is strongly inhibited by the great atmospheric pressure (93 bars), which reduces volatile exsolution (Head and Wilson 1986). Although *Magellan* radar shows evidence for wind features associated with dust (Greeley et al. 1992), production of fine particles is probably limited to impact processes. Vast deposits of dust like those on Mars or in loess on Earth are not seen. Titan has dunes that are probably composed of frozen organics (Barnes et al. 2008). Dust-size organics or other materials may be present on the surface, but the methane "hydrological" cycle (Tokano et al. 2006) may act to flush it out. Expanding our horizons to extra-solar planets, Mars-like bodies with thin atmospheres having limited or no hydrologic cycles could also have dust that is continually produced and accumulates over time. These may yet be discovered in future investigations.

CONCLUSIONS

Dust formation, transport, deposition, stabilization, and erosion are active geologic processes on Earth and Mars. Dust accumulates where it can be stabilized and will not be subjected to resuspension. On Earth, stabilization is driven by hydrological and biological processes. On Mars, accumulation occurs in dust sinks that may switch hemispheric polarity in response to the precessional cycle. The comparison to loess is therefore informative in understanding the ways in which dust on two terrestrial planets can accumulate and then have divergent evolutionary paths. On Mars, which lacks liquid water and biology, dust is stabilized by growth of micrometer-sized dust into aggregates and cementation. Continual accumulation will compact these materials, forming duststone. The role of dust on the two planets is similar in that thick accumulations can form, but the different environments result in divergent evolutionary paths.

ACKNOWLEDGMENTS

Comments from D. Rubin, J. Pigati, and an anonymous reviewer significantly improved this paper.

REFERENCES

Aleinikoff JN, Muhs DR, Bettis EA III, Johnson WC, Fanning CM, Benton R. 2008. Isotopic evidence for the diversity of late Quaternary loess in Nebraska: glaciogenic and non-glaciogenic Sources. *Geological Society of America Bulletin* 120:1362–1377.

Aleinikoff JN, Muhs DR, Sauer R, Fanning, CM. 1999. Late Quaternary loess in northeastern Colorado, Part II—Pb isotopic evidence for the variability of loess sources. *Geological Society of America Bulletin* 111: 1876–1883.

Allen CC, Morris RV, Jager KM, Golden DC, Lindstrom DJ, Lindstrom MM, Lockwood JP. 1998. Martian regolith simulant JSC Mars-1. *Lunar and Planetary Science* XXIX:1690.

Barnes JW, Brown RH, Soderblom L, Sotin C, Le Mouèic S, Rodriguez S, Jaumann R, Beyer RA, Buratti BJ, Pitman K, Baines KH, Clark R, Nicholson P. 2008. Spectroscopy, morphometry, and photoclinometry of Titan's dunefields from Cassini/VIMS. *Icarus* 195:400–414.

Berger AL. 1992. Astronomical theory of paleoclimates and the last glacial–interglacial cycle. *Quaternary Science Reviews* 11:571–581.

Blatt H. 1987. Oxygen isotopes and the origin of quartz. *Journal of Sedimentary Petrology* 57:373–377.

Bradley BA, Sakimoto SEH, Frey H, Zimbelman JR. 2002. Medusae Fossae Formation: new perspectives from Mars Global Surveyor. *Journal of Geophysical Research–Planets* 107. DOI:10.1029/2001JE001537

Bridges NT, Banks ME, Beyer RA, Chuang FC, Noe Dobrea EZ, Herkenhoff KE, Keszthelyi LP, Fishbaugh KE, McEwen AS, Michaels TI, Thomson BJ, Wray JJ. 2010. Eolian bedforms, yardangs, and indurated surfaces in the Tharsis Montes as seen by the HiRISE camera: evidence for dust aggregates. *Icarus* 205:165–182.

Chan MA. 1999. Triassic loessite of north-central Utah: stratigraphy, petrophysical character and paleoclimate implications. *Journal of Sedimentary Research* 69:477–485.

Christensen PR. 1986. Regional dust deposits on Mars: physical properties, age, and history. *Journal of Geophysical Research* 91:3533–3545.

Christensen PR. 1988. Global albedo variations on Mars: implications for active eolian transport, deposition, and erosion. *Journal of Geophysical Research* 93:7611–7624.

Christensen PR, Moore HJ. 1992. The Martian surface layer. *In* Kieffer HH, Jakosky BM, Snyder CW, Matthews MS (Editors). *Mars*: University of Arizona Press, Tucson, Arizona. p. 686–729.

Clark BC, Baird AK, Weldon RJ, Tsusaki DM, Schnabel L, Candelaria MP. 1982. Chemical composition of Martian fines. *Journal of Geophysical Research* 87:10059–10067.

Crouvi O, Amit R, Enzel Y, Gillespie AR. 2010. Active sand seas and the formation of desert loess. *Quaternary Science Reviews* 29:2087–2098.

de Silva SL, Bailey JE, Mandt KE, Viramonte JM. 2010. Yardangs in terrestrial ignimbrites: synergistic remote and field observations on Earth with applications to Mars. *Planetary and Space Science* 58:459–471.

Edgett KS. 2005. The sedimentary rocks of Sinus Meridiani: five key observations from data acquired by the Mars Global Surveyor and Mars Odyssey orbiters. *Mars: The International Journal of Mars Science and Exploration* 1:5–58.

Exner FM. 1925. Uber die Wechselwirkung zwischen Wasser und Geschiebe in Flussen. *Sitzunsberichte Adademie der Wessenschaften in Wien, Mathematische–Naturwissenschaftliche Klasse Abt. 2A*, 134:183–228.

Flemal RC, Odom IE, Vail RG. 1972. Stratigraphy and origin of paha topography of northwestern Illinois. *Quaternary Research* 2:232–243.

Flynn GJ. 1996. The delivery of organic matter from asteroids and comets to the early surface of Mars. *Earth, Moon, and Planets* 72:469–474.

Frey HV, Roark JH, Shockey KM, Frey EL, Sakimoto SEH. 2002. Ancient lowlands of Mars. *Geophysical Research Letters* 29. DOI:10.1029/2001GL013832

Goetz W, Bertelsen P, Binau CS, Gunnlaugsson HP, Hviid SF, Kinch KM, Madsen DE, Madsen MB, Olsen M, Gellert R, Klingelhöfer G, Ming DW, Morris RV, Rieder R, Rodionov DS, de Souza PA Jr, Schröder C, Squyres SW, Wdowiak T, Yen A. 2005. Indication of drier periods on Mars from the chemistry and mineralogy of atmospheric dust. *Nature* 436:62–65.

Greeley R. 1979. Silt–clay aggregates on Mars. *Journal of Geophysical Research* 84:6248–6284.

Greeley R, Arvidson RE, Elachi C, Geringer MA, Plaut JJ, Saunders RS, Schubert G, Stofan ER, Thouvenot EJP, Wall SD, Weitz CM. 1992. Aeolian features on Venus: preliminary Magellan results. *Journal of Geophysical Research* 97:13319–13345.

Greeley R, Iversen JD. 1985. *Wind as a Geological Process*. Cambridge University Press, Cambridge, UK. 333 p.

Greeley R, Whelley PL, Neakrase LDV. 2004. Martian dust devils: directions of movement inferred from their tracks. *Geophysical Research Letters* 31. DOI:10.1029/2004GL021599

Greeley R, Williams SH. 1994. Dust deposits on Mars: the "Parna" Analog. *Icarus* 110:165–177.

Grimley DA. 2000. Glacial and nonglacial sediment contributions to Wisconsin Episode loess in the central United States. *Geological Society of America Bulletin* 112:1475–1495.

Haberle RM. 1986. Interannual variability of global dust storms on Mars. *Science* 234:459–461.

Haberle RM, Murphy JR, Schaeffer J. 2003. Orbital change experiments with a Mars general circulation model. *Icarus* 161:66–89.

Haberle RM, Pollack JB, Barnes JR, Zurek RW, Leovy CB, Murphy JR, Lee H, Schaeffer J. 1993. Mars atmospheric dynamics as simulated by the NASA Ames general circulation model, 1. The zonal–mean circulation. *Journal of Geophysical Research* 98:3093–3123.

Hallet B, Hunter L, Bogen J. 1996. Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications. *Global and Planetary Change* 12:213–235.

Hamilton VE, McSween HY, Hapke B. 2005. Mineralogy of Martian atmospheric dust inferred from thermal infrared spectra of aerosols. *Journal of Geophysical Research* 110. DOI:10.1029/2005JE002501

Head JW, Wilson L. 1986. Volcanic processes and landforms on Venus—theory, predictions, and observations. *Journal of Geophysical Research* 91:9407–9446.

Herkenhoff KE, Golombek MP, Guinness EA, Johnson JB, Kussack A, Richter L, Sullivan R, Gorevan S. 2008. In-situ observations of the physical properties of the Martian surface. *In* Bell JF III (Editor), *The Martian Surface: Composition, Mineralogy, and Physical Properties*: Cambridge University Press, Cambridge, UK. p. 451–467.

Herkenhoff KE, Squyres SW, Arvidson R, Bass DS, Bell JF III, Bertelsen P, Cabrol NA, Gaddis L, Hayes AG, Hviid SF, Johnson JR, Kinch KM, Madsen MB, Maki JN, McLennan SM, McSween HY, Rice JW Jr, Sims M, Smith PH, Soderblom LA, Spanovich N, Sullivan R, Wang A. 2004. Textures of the soils and rocks at Gusev Crater from Spirit's Microscopic Imager. *Science* 305:824–826.

Herkenhoff KE, Vasavada AR. 1999. Dark material in the polar layered deposits and dunes on Mars. *Journal of Geophysical Research* 104:16487–16500.

Herkenhoff KE, Squyres SW, Anderson R, Archinal BA, Arvidson RE, Barrett

JM, Becker KJ, Bell JF III, Budney C, Cabrol NA, Chapman MG, Cook D, Ehlmann BL, Farmer J, Franklin B, Gaddis LR, Galuska DM, Garcia PA, Hare TM, Howington-Kraus E, Johnson JR, Johnson S, Kinch K, Kirk RL, Lee EM, Leff C, Lemmon M, Madsen MB, Maki JN, Mullins KF, Redding BL, Richter L, Rosiek MR, Sims MH, Soderblom LA, Spanovich N, Springer R, Sucherski RM, Sucharski T, Sullivan R, Torson JM, Yen A. 2006. Overview of Microscopic Imager investigation during Spirit's first 450 sols in Gusev Crater. *Journal of Geophysical Research* 111. DOI:10.1029/JE002574.

Iversen JD, Pollack JB, Greeley R, White BR. 1976. Saltation threshold on Mars: the effect of interparticle force, surface roughness, and low atmospheric density. *Icarus* 29:381–393.

Jakosky BM, Christensen PR. 1986a. Are the Viking Lander sites representative of the surface of Mars? *Icarus* 66:125–133.

Jakosky BM, Christensen PR. 1986b. Global duricrust of Mars - Analysis of remote-sensing data. *Journal of Geophysical Research* 91:3547–3559.

Johnson SY. 1989. Significance of loessite in the Maroon Formation (Middle Pennsylvanian to Lower Permian), Eagle Basin, northwest Colorado. *Journal of Sedimentary Petrology* 59:782–791.

Kahn RA, Martin TZ, Zurek RW, Lee SW. 1992. The Martian dust cycle. *In* Kieffer HH, Jakosky BM, Snyder CW, Matthews MS (Editors). *Mars*: University of Arizona Press, Tucson, Arizona. p. 1017–1053.

Karunatillake S, Wray JJ, Squyres SW, Taylor GJ, Gasnault O, McLennon SM, Boynton W, El Maary MR, Dohm JM. 2009. Chemically striking regions on Mars and stealth revisited. *Journal of Geophysical Research* 114. DOI:10.1029/2008JE003303

Keszthelyi L, Jaeger W, McEwen A, Tornabene L, Beyer RA, Dundas C, Milazzo M. 2008. High Resolution Imaging Science Experiment (HiRISE) images of volcanic terrains from the first 6 months of the Mars Reconnaissance Orbiter Primary Science Phase. *Journal of Geophysical Research* 113. DOI:10.1029/2007JE002968

Kieffer HH, Zent AP. 1992. Quasi-periodic climate change on Mars. *In* Kieffer HH, Jakosky BM, Snyder CW, Matthews MS (Editors). *Mars*: University of Arizona Press, Tucson, Arizona. p. 1180–1218.

Krinsley D, Leach R. 1981. Properties of electrostatic aggregates and their possible presence on Mars. *Precambrian Research* 14:167–178.

Kubatko EJ, Westerink JJ. 2007. Exact discontinuous solutions to Exner's bed evolution model: simple theory for sediment bores. *Journal of Hydraulic Engineering* 133. DOI:10.1061/(ASCE)0733-9429(2007)133:3(305)

Laity JE, Bridges NT. 2009. Ventifacts on Earth and Mars: analytical, field, and laboratory studies supporting sand abrasion and windward feature development. *Geomorphology* 105:202–217.

Lee SW, Thomas PC, Veverka J. 1982. Wind streaks in Tharsis and Elysium: implications for sediment transport by slope winds. *Journal of Geophysical Research* 87:10025–10041.

Lemmon MT, Wolff MJ, Smith MD, Clancy RT, Banfield D, Landis GA, Ghosh A, Smith PH, Spanovich N, Whitney B, Whelley P, Greeley R, Thompson S, Bell JF III, Squyres SW. 2004. Atmospheric imaging results from the Mars Exploration Rovers: Spirit and Opportunity. *Science* 204:1753–1756.

Lewis KW, Aharonson O, Grotzinger JP, Kirk RL, McEwen AS, Suer TA. 2008. Quasi-periodic bedding in the sedimentary rock record of Mars. *Science* 322:1532–1535.

Lewis PF. 1960. Linear topography in the southwestern Palouse, Washington–Oregon. *Annals of the Association of American Geographers* 50:98–111.

Magalhaes JA. 1987. The Martian Hadley circulation: comparison of 'viscous' model predictions to observations. *Icarus* 70:442–468.

Magalhaes J, Gierasch P. 1982. A model of Martian slope winds: implications for eolian transport. *Journal of Geophysical Research* 87:9975–9984.

Mahowald N, Kohfeld K, Hansson M, Balkanski Y, Harrison SP, Prentice IC, Schulz M, Rodhe H. 1999. Dust sources and deposition during the last glacial maximum and current climate: A comparison of model results with paleodata from ice cores and marine sediments. *Journal of Geophysical Research* 104:15895–15916.

Malin MC, Edgett KS. 2001. Mars Global Surveyor Mars Orbiter Camera: interplanetary cruise through primary mission. *Journal of Geophysical Research* 106:23429–23570.

Mandt K, de Silva SL, Zimbelman JR, Crown DA. 2008. The origin of the Medusae Fossae Formation, Mars: insights from a synoptic approach. *Journal of Geophysical Research* 113. DOI:10.1029/2008JE003076

Mangold N, Ansan V, Masson P, Vincendon C. 2009. Estimate of the eolian dust thickness in Arabia Terra, Mars: implications of a thick mantle (≥20 m) for hydrogen detection. *Géomorphologie: Relief, Processus, Environnement*:23–32.

Martin LJ, James PB, Dollfus A, Iwasaki K, Beish JD. 1992. Telescopic observations: visual, photographic, polarimetric. *In* Kieffer HH, Jakosky BM, Snyder CW, Matthews MS (Editors). *Mars*: University of Arizona Press, Tucson, Arizona, p. 34–70.

McTainsh GH. 1989. Quaternary aeolian dust processes and sediments in the Australian region. *Quaternary Science Reviews* 8:235–253.

Mellon MT, Jakosky BM, Kieffer HH, Christensen PR. 2000. High-resolution thermal inertia mapping from the Mars Global Surveyor Thermal Emission Spectrometer. *Icarus* 148:437–455.

Milliken RE, Grotzinger JP, Thomson BJ. 2010. Paleoclimate of Mars as captured by the stratigraphic record of Gale Crater. *Geophysical Research Letters* 37. DOI:10.1029/2009GL041870

Moore HJ, Bickler DB, Crisp JA, Eisen HJ, Gensler JA, Haldemann AFC, Matijevic JR, Reid LK, Pavlics F. 1999. Soil-like deposits observed by Sojourner, the Pathfinder rover. *Journal of Geophysical Research* 104:8729–8746.

Muhs DR. 2007. Loess deposits, origins, and properties. *In* Elias S (Editor). *The Encyclopedia of Quaternary Sciences*: Elsevier, Amsterdam. p. 1405–1418.

Muhs DR, Ager TA, Bettis EA III, McGeehin J, Been JM, Begét JE, Pavich MJ, Stafford TW Jr, Stevens DSP. 2003. Stratigraphy and paleoclimatic significance of late Quaternary loess–paleosol sequences of the last interglacial–glacial cycle in central Alaska. *Quaternary Science Reviews* 22:1947–1986.

Muhs DR, Aleinikoff JN, Stafford TW Jr, Kihl R, Been J, Mahan SA, Cowherd S. 1999. Late Quaternary loess in northeastern Colorado, part I—age and paleoclimatic significance. *Geological Society of America Bulletin* 111:1861–1875.

Muhs DR, Bettis EA III. 2000. Geochemical variations in Peoria Loess of western Iowa indicate paleowinds of midcontinental North America during last glaciations. *Quaternary Research* 53:49–61.

Muhs DR, Bettis EA III. 2003. Quaternary loess–paleosol sequences as examples of climate-driven sedimentary extremes. *In* Chan MA, Archer AW (Editors). *Extreme Depositional Environments: Mega End Members in Geologic Time:* Geological Society of America, Boulder, Colorado, Special Paper 370, p. 53–74.

Muhs DR, Bettis EA III, Aleinikoff J, McGeehin JP, Beann J, Skipp G, Marshall BD, Roberts HM, Johnson WC, Benton R. 2008. Origin and paleoclimatic significance of late Quaternary loess in Nebraska: evidence from stratigraphy, chronology, sedimentology, and geochemistry. *Geological Society of America Bulletin* 120:1378–1407.

Muhs DR, Budahn R. 2006. Geochemical evidence for the origin of late Quaternary loess in central Alaska. *Canadian Journal of Earth Sciences* 43:323–337.

Muhs DR, McGeehin JP, Beann J, Fisher E. 2004. Holocene loess deposition and soil formation as competing processes, Matanuska Valley, southern Alaska. *Quaternary Research* 61:265–276.

Mustard JF, Cooper CD, Rifkin MK. 2001. Evidence for recent climate change on Mars from the identification of youthful near-surface ground ice. *Nature* 412:411–414.

Newman CE. 2001. Modeling the dust cycle in the Martian atmosphere [unpublished Ph.D. thesis]: St. Anne's College, University of Oxford, UK, 494 p.

Newman CE, Lewis SR, Read PL, Forget F. 2002. Modeling the Martian dust cycle 1. Representations of dust transport processes. *Journal of Geophysical Research* 107. DOI:10.1029/2002JE001910

Phillips RJ, Zuber MT, Smrekar SE, Mellon MT, Head JW, Tanaka KL, Putzig NE, Milkovich SM, Campbell BA, Plaut JJ, Safaeinili A, Seu R, Biccari D, Carter LM, Picardi G, Orosei R, Surdas Mohit P, Heggy E, Zurek RW, Egan AF, Giacomoni E, Russo F, Cutigni M, Pettinelli E, Holt JW, Leuschen CJ, Mariangeli L. 2008. Mars north polar deposits: stratigraphy, age, and geodynamical response. *Science* 320:1182–1185.

Pollack JB, Colburn D, Kahn R, Hunter J, Van Camp W, Carlston CE, Wolf MR.

1977. Properties of aerosols in the Martian atmosphere, as inferred from Viking Lander imaging data. *Journal of Geophysical Research* 82:4479–4496.

Pollack JB, Leovy CB, Greiman PW, Mintz Y. 1981. A Martian general circulation experiment with large scale topography. *Journal of Atmospheric Science* 38:3–29.

Putzig NE, Mellon MT. 2007a. Apparent thermal inertia and the surface heterogeneity of Mars. *Icarus* 191:68–94. DOI:10.1016/j.icarus.2007.05.013

Putzig NE, Mellon MT. 2007b. Thermal behavior of horizontally mixed surfaces on Mars. *Icarus* 191:52–67. DOI:10.1016/j.icarus.2007.03.022

Pye K. 1987. *Aeolian Dust and Dust Deposits*: Academic Press, San Diego, California. 334 p.

Rea DK. 2007. Eolian records, deep-sea sediments. *In* Elias S (Editor). *The Encyclopedia of Quaternary Sciences*: Elsevier, Amsterdam. p. 643–649.

Rieder R, Economou T, Wanke H, Turkevich A, Crisp J, Brückner J, Dreibus G, McSween HY. 1997. The chemical composition of Martian soil and rocks returned by the Mobile Alpha Proton X-ray Spectrometer: preliminary results from the X-ray mode. *Science* 278:1771–1774.

Rieder R, Gellert R, Anderson RC, Brückner J, Clark BC, Dreibus G, Economou T, Klingelhöfer G, Lugmair GW, Ming DW, Squyres SW, d'Uston C, Wänke H, Yen A, Zipfel, J. 2004. Chemistry of rocks and soils at Meridiani Planum from the Alpha Particle X-ray Spectrometer. *Science* 306:1746–1749.

Roberts HM, Muhs DR, Wintle AG, Duller GAT, Bettis EA. 2003. Unprecedented last glacial mass accumulation rates determined by luminescence dating of loess from western Nebraska. *Quaternary Research* 59:411–419.

Ronov AB. 1983. *The Earth's Sedimentary Shell: Quantitative Patterns of its Structure, Composition and Evolution*: American Geological Institute, Alexandria, Virginia, Reprint Series 5, 80 p.

Ruff SW, Christensen PR, Clark RN, Kieffer HH, Malin MC, Bandfield JL, Jakosky BM, Lane MD, Mellon MT, Presley MA. 2001. Mars "White Rock" feature lacks evidence of an aqueous origin: results from Mars Global Surveyor. *Journal of Geophysical Research* 106:23921–23927.

Scott DH, Tanaka KL. 1982. Ignimbrites of Amazonis Planitia region of Mars. *Journal of Geophysical Research* 87:1170–1190.

Shao Y, Lu H. 2000. A simple expression for wind erosion threshold friction velocity. *Journal of Geophysical Research* 105:22437–22443.

Shorthill RW, Hutton RE, Moore HJ, Scott RF, Spitzer CR. 1976. Physical properties of Martian surface from Viking 1 Lander: preliminary results. *Science* 193:805–809.

Smith BJ, Wright JS, Whalley WB. 2002. Sources of non-glacial, loess-size quartz silt and the origins of "desert loess." *Earth-Science Reviews* 59:1–26.

Smith DE, Zuber MT, Solomon SC, Phillips RJ, Head JW, Garvin JB, Banerdt WB, Muhleman DO, Pettengill GH, Neumann GA, Lemoine FG, Abshire JB, Aharonson, O, Brown CD, Hauck SA, Ivanov AB, McGovern PJ, Zwally HJ, Duxbury TC. 1999. The global topography of Mars and implications for surface evolution. *Science* 284:1495–1503.

Soreghan GS, Elmore RD, Katz B, Cogoini M, Banerjee S. 1997. Pedogenically enhanced magnetic susceptibility variations preserved in Paleozoic loessite. *Geology* 25:1003–1006.

Soreghan GS, Elmore RD, Lewchuk MT. 2002. Sedimentologic–magnetic record of western Pangean climate in upper Paleozoic loessite (lower Cutler beds, Utah). *Geological Society of America Bulletin* 114:1019–1035.

Sullivan R, Arvidson R, Bell JF III, Gellert R, Golombek M, Greeley R, Herkenhoff K, Johnson J, Thompson S, Whelley P, Wray J. 2008. Wind-driven particle mobility on Mars: insight from Mars Exploration Rover observations at "El Dorado" and surroundings at Gusev Crater. *Journal of Geophysical Research* 113. DOI:10.1029/2008JE003101

Sullivan R, Banfield D, Bell JF, Calvin W, Fike D, Golombek M, Greeley R, Grotzinger J, Herkenhoff K, Jerolmack D, Malin M, Ming D, Soderblom LA, Squyres SW, Thompson S, Watters WA, Weitz CM, Yen A. 2005. Aeolian processes at the Mars Exploration Rover Merididiani Planum landing site. *Nature* 436. DOI:10.1038/nature03641.

Sullivan R, Bandfield D, Collins LR, Heineck JT, Korda DT. 2011. Determining the minimum saltation grain size on Mars. *Lunar and Planetary Science* XLXII:2651.

Sullivan R, Thomas P, Veverka J, Malin M, Edgett KS. 2001. Mass movement slope streaks imaged by the Mars Orbiter Camera. *Journal of Geophysical Research* 106:23607–23633.

Sun J. 2002. Provenance of loess material and formation of loess deposits on the Chinese Loess Plateau. *Earth and Planetary Science Letters* 203:845–859.

Syers JK, Jackson ML, Berkheiser VE, Clayton RN, Rex RW. 1969. Eolian sediment influence on pedogenesis during the Quaternary. *Soil Science* 107:421–427.

Tanaka KL, Rodriguez JAP, Skinner JA, Bourke MC, Fortezzo CM, Herkenhoff KE, Kolb EJ, Okubo CH. 2008. North polar region of Mars: advances in stratigraphy, structure, and erosional modification. *Icarus* 196:318–358.

Tanaka KL, Scott DH, Greeley R. 1992. Global stratigraphy. *In* Kieffer HH, Jakosky BN, Snyder CW, Matthews MS (Editors). *Mars*: University of Arizona Press, Tucson, Arizona. p. 345–382.

Thomas P, Squyres S, Herkenhoff K, Howard A, Murray B. 1992. Polar deposits of Mars. *In* Kieffer HH, Jakosky BM, Snyder CW, Matthews MS (Editors), *Mars*, University of Arizona Press, Tucson, 767–795.

Thomson BJ, Bridges NT, Milliken R, Baldridge A, Hook SJ, Crowley JR, Marion GM, de Souza Filho CR, Brown AJ, Weitz CM. 2011. Constraints on the origin and evolution of the layered mound in Gale Crater, Mars using Mars Reconnaissance Orbiter data. *Icarus* 214:413–432. DOI:10.1016/j.icarus.2011.05.002

Tokano T, McKay CP, Neubauer FM, Atreya SK, Ferri F, Fulchignoni M, Niemann HB. 2006. Methane drizzle on Titan. *Nature* 442:432–435.

Toon OB, Pollack JB, Ward W, Burns JA, Bilski K. 1980. The astronomical theory of climatic change on Mars. *Icarus* 44:552–607.

Ward AW. 1979. Yardangs on Mars: evidence of wind erosion. *Journal of Geophysical Research* 84:8147–8166.

Wells GL, Zimbelman JR. 1997. Extraterrestrial arid surface processes. *In* Thomas DSG (Editor). *Arid Zone Geomorphology: Process, Form, and Change in Drylands*, 2nd ed.: John Wiley & Sons Ltd, Chichester, West Sussex, UK. p. 559–690.

Westphal DL, Toon OB, Carlson TN. 1987. A two-dimensional numerical investigation of the dynamics and microphysics of Saharan dust storms. *Journal of Geophysical Research* 92:3027–3049.

White BR. 1979. Soil transport by winds on Mars. *Journal of Geophysical Research* 84:4643–4651.

White BR, Lacchia BM, Greeley R, Leach RN. 1997. Aeolian behavior of dust in a simulated Martian environment. *Journal of Geophysical Research* 102:25629–25640.

Wieder M, Gvirtzman G, Porat N, Dassa M. 2008. Paleosols of the southern coastal plain of Israel. *Journal of Plant Nutrition and Soil Science* 171:533–541.

Wilson SA, Howard AD, Moore JM, Grant JA. 2007. Geomorphic and stratigraphic analysis of Crater Terby and layered deposits north of Hellas basin, Mars. *Journal of Geophysical Research* 112. DOI:10.1029/2006JE002830

Wright J. 2001a. "Desert" loess versus "glacial" loess: quartz silt formation, source areas and sediment pathways in the formation of loess deposits. *Geomorphology* 36:231–256.

Wright J. 2001b. Making loess-sized quartz silt: data from laboratory simulations and implications for sediment transport pathways and the formation of 'desert' loess deposits associated with the Sahara. *Quaternary International* 76/77:7–19.

Wright J, Smith BJ, Whalley WB. 1998. Mechanisms of loess-sized quartz silt production and their relative effectiveness: laboratory simulations. *Geomorphology* 23:15–34.

Yen AS, Mittlefehldt DW, McLennan SM, Gellert R, Bell JF, McSween HY, Ming DW, McCoy TJ, Morris RV, Golombek M, Enonomou T, Madsen MB, Wdowiak T, Clark BC, Jolliff BL, Schröder C, Brückner J, Zipfel J, Squyres SW. 2006. Nickel on Mars: constraints on meteoritic material at the surface. *Journal of Geophysical Research* 111. DOI:10.1029/2006JE002797

Zárate M, Blasi A. 1993. Late Pleistocene–Holocene eolian deposits of the southern Buenos Aires Province, Argentina: a preliminary model. *Quaternary International* 17:15–20.