13

Planetary Aeolian Processes and Landforms

We describe the near-surface wind profile, its relation to environmental conditions, and how it can be quantified. The freestream wind speed can be converted to a friction wind speed, which relates to the flow at the atmosphere–surface interface and thus to the entrainment of sediment. The minimum wind speed for entrainment of aeolian sediment depends on gravity and grain size, so that threshold wind speed differs for varying planetary conditions. The difference in transport mechanism for grains leads to different depositional morphologies, which provide clues to the wind speed, wind direction, and sediment availability. Erosional landforms likewise provide information on near-surface atmospheric processes and surface sediments, as well as bedrock lithologies. The study of aeolian landforms thus informs our understanding of the atmosphere, surface geology, and sedimentology on other planets.

13.1 Bringing the Atmosphere Down to the Surface (and Why We Care)

In Chapter 12 we learned about atmospheric flow and the planetary boundary layer. But what happens when atmospheric flow comes into contact with the surface of a planet? Contextually, this question is both strange and obvious. It's strange because the large majority of planetary objects in the Solar System do *not* have both an atmosphere and a surface. For example, the giant planets, which constitute the greatest mass in our planetary system after the Sun, have atmospheres but no discrete surfaces. Conversely, the millions of asteroids, which are the most numerous planetary objects, have surfaces but no atmospheres. So for these massive and common populations, respectively, the question makes no sense. At the same time, the answer is obvious to anyone who resides on Earth. We are all familiar with wind-driven processes and the resultant landforms, the most familiar of which perhaps are sand dunes. Dunes and other wind-derived landforms are termed "**aeolian**," after Aeolus, the Greek god of the winds, and they have been found on virtually every planetary body that has a solid surface and a gaseous envelope (Figure 13.1). This gaseous envelope can take the form of a dense and stable atmosphere, as on Venus, or of a very tenuous, even transitory gaseous covering (sometimes called an "**exosphere**"), as on comets and other smaller bodies.

If aeolian processes require this rather uncommon confluence of both a solid surface and a gaseous envelop, why do we care about them? In part, we care because we live on Earth, where aeolian processes occur everywhere, influencing our everyday lives and even human history. In fact, scientific research dedicated to aeolian processes was initiated by Ralph Bagnold, a British engineer who conducted military countermeasures in North African deserts in World War II. Following this insightful field work, Bagnold spent decades conducting innovative wind tunnel research into sand movement and dune formation, thus laying the foundation for understanding aeolian processes. Despite their requirement for a somewhat limited confluence of conditions, aeolian processes are prevalent on our nearest planetary neighbors. Aeolian landforms have been documented on Mars, where aeolian sand transport is pervasive, and to a lesser extent on Venus, for which current data are less extensive and of lower resolution than for Mars. In addition, aeolian processes occur in the outer Solar System, such as on Titan and Triton, the largest satellites of Saturn and Neptune, respectively. They even occur on comets and on Pluto, which means they likely occur on the millions of other **Kuiper belt** objects in the outer Solar System.

Beyond their prevalence on terrestrial bodies, what information can aeolian landforms give us? Wind-driven landforms tell us both about the atmosphere (or exosphere) of the body and about the surface, as well as

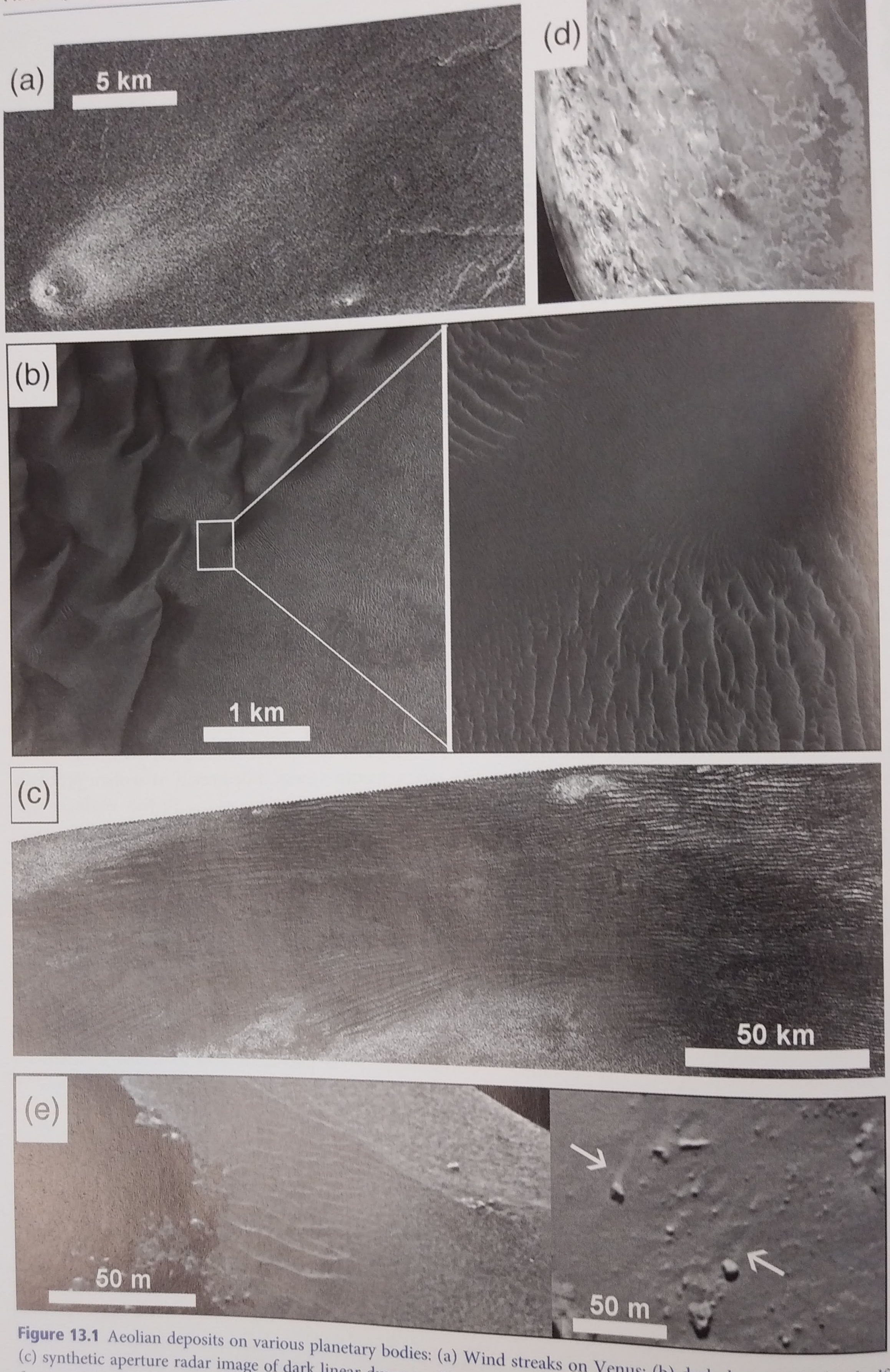

Figure 13.1 Aeolian deposits on various planetary bodies: (a) Wind streaks on Venus; (b) dark dunes and other bedforms on Mars; (c) synthetic aperture radar image of dark linear dunes on Titan; (d, upper right) dark wind streaks on Triton; (e) dune-like forms (left) and possible wind streaks (right) on Comet Churyumov-Gersimenko. NASA images.

BOX 13.1 USING WIND TUNNELS FOR AEOLIAN EXPERIMENTS

Following Bagnold's pioneering use of wind tunnels for investigating sand motion, wind tunnels have been used extensively to better understand aeolian processes and sedimentation. Most wind tunnels are terrestrial in nature, facilities that control ambient air to precise conditions to investigate processes that occur on Earth. Some terrestrial wind tunnels are portable, which enables geologists to run semi-controlled experiments under natural conditions in the field. Other wind tunnels are found in laboratories, where conditions, although more artificial, may be more precisely controlled. Examples of terrestrial laboratory wind tunnels are found at the Arizona State University in the USA and in the Trent University in Peterborough, Canada.

Planetary wind tunnels have also been constructed for investigating extraterrestrial aeolian processes. In the Planetary Aeolian Laboratory at the NASA Ames Research Center, facilities exist for simulating low-pressure atmospheric conditions, such as are found at the surface on Mars (the Mars Surface Wind Tunnel, or MARSWIT). Another high-pressure wind tunnel exists, originally constructed to simulate conditions on Venus, now refurbished for simulating surface atmospheric conditions on Titan (the Titan Wind Tunnel, or TWT – Figure 13.2).

Wind tunnels permit simulation of specific conditions relevant to different planetary processes. For example, one of the most basic aeolian processes is **entrainment** – getting stationary grains into motion. Entrainment on planetary bodies with thin atmospheres occurs largely through impact of upwind grains, whereas under thick atmospheres, entrainment occurs through fluid flow. To numerically model this difference, however, requires experimental data with which to test and tune the model. Conducting entrainment experiments at different pressures in wind tunnels provides such data.

Wind tunnels can also be used to investigate erosional aeolian processes. For thinner (low-density, lower-pressure) atmospheres, the wind must be moving more quickly to generate the force necessary to move grains, compared to thicker (higher-density, higher-pressure) atmospheres. Thus, aeolian **abrasion** should be more effective under thin atmospheres. Wind tunnel experiments provide data to quantify this difference in abrasion effectiveness with atmospheric density, and thus support interpretations of **ventifacts** and **yardangs** (discussed in Sections 13.6.2 and 13.6.1, respectively).

Simply put, aerodynamic forces are dominant in aeolian processes under high-density atmospheres, whereas ballistic (gravity-driven) interactions are dominant under low-density atmospheres. Having planetary wind tunnels that enable simulation of both of these extraterrestrial conditions has left us an impressive legacy in understanding the effects of different conditions on aeolian transport. As we expand our awareness of the pervasiveness of aeolian processes, laboratory simulation of planetary aeolian conditions will continue to provide fundamental information for interpreting these extraterrestrial planetary landforms (Figure 13.1).

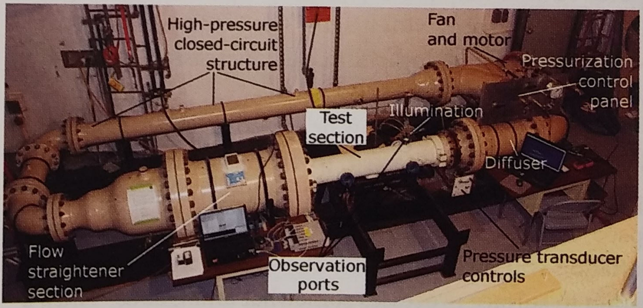

Figure 13.2 The Titan Wind Tunnel, with components of the structure labeled. For experiments, the test section is rolled toward the viewer, the test plate is inserted and covered with sediment, and the test section is rolled back into place. After sealing, the wind tunnel is pressurized. Lastly, wind is generated with the fan at the back right of the image, and observations of sand motion are made through the observation ports.

about the existence, characteristics, and cycling of both volatiles and sediments. In this way, the study of aeolian geology provides critical clues to the development and evolution of planetary surfaces. To access this information, however, we first have to understand how the wind interacts with and shapes these surfaces.

13.2 The Near-Surface Wind Profile

Imagine wind flowing across a land surface. Personal experience tells us that the wind speed decreases with proximity to the surface, as a result of the land surface exerting friction on the near-surface wind. The region of decreasing wind speed is referred to as the **boundary layer**.

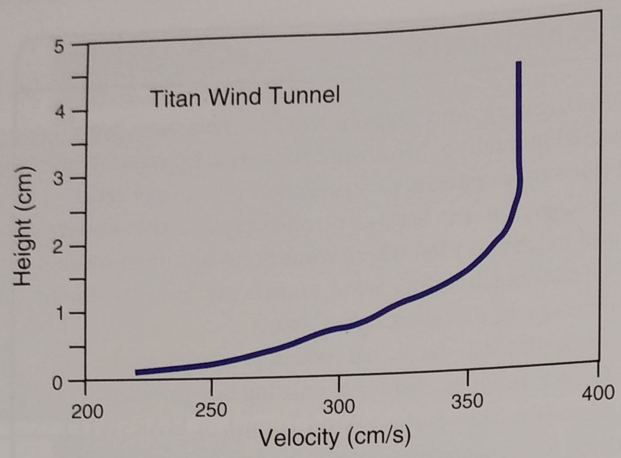

Figure 13.3 Wind speed profile from the Titan Wind Tunnel, with a typical logarithmic shape showing an increase in wind speed with distance from the surface.

This friction so retards the flow of the gas molecules closest to the surface that, at some theoretical height above the surface – termed "roughness height" and symbolized as z_0 – the wind flow ceases. For a relatively smooth surface, the lowest portion is denoted the laminar sublayer, where gas molecules move in layers (laminae) without turbulence. In contrast, the flow of gas molecules higher above the surface experiences less friction. This reduced friction effect means less slowing of the wind speed, so that above the boundary layer the wind speed no longer decreases with height. The result is a characteristic wind profile of logarithmically increasing speed with height (Figure 13.3).

The wind speed that can be experienced or measured by instruments is referred to as the **freestream wind speed**. Although measurement of the freestream wind speed is quite useful (think of weather forecasts), its variation with distance above the surface makes its application to surface processes somewhat ambiguous. Thus, a more independent measure of wind speed at the surface is desirable. For this purpose, the aeolian community uses the **friction wind speed** – think of this as a shear speed, a quantity that represents the strength of the gas flow at the interface with the surface. It is derived from (and directly proportional to) the freestream wind speed, but is independent of height. The quantitative relationship between freestream wind speed (u) and friction wind speed (u^*) is given by the "law of the wall," namely:

$$u_z = \frac{u^*}{\kappa} \ln \frac{z}{z_0} \tag{13.1}$$

where z is the height above the surface at which the freestream wind speed is measured and κ is the von Karmán constant (~0.4). In theory, the equation can be inverted to solve for friction wind speed (u^*) from the freestream wind speed measured at a single height, although in practice deriving z_0 requires measurements of the freestream wind velocity at two heights.

13.3 The Physics of Particle Entrainment

The wind speed required for entrainment of particles depends on the interplay of a number of forces, as well as the size of the particles.

13.3.1 Force (Torque) Balance: The Conditions for Entrainment

This near-surface wind exerts a force on the surface. For a granular surface, the wind forces acting on each grain include both a drag force (F_D) that results from the flow of gas over the grain surface, and a lift force (F_L) that results from the greater wind velocity over the top than the bottom of the grain. At the same time, the grain experiences gravity (F_g) as well as an interparticle force (F_{ip}) that results from water absorption, electrostatic attraction, and van der Waals forces. Gravity – or actually weight – increases for more massive (e.g., larger) grains. The interparticle forces increase with decreasing grain size (i.e., smaller particles stick together better), because smaller grains have a larger surface area-to-mass ratio, and the interparticle forces are a function of the surface area. Even dry, clay sticks together better than sand!

A balance of these forces provides the conditions under which the movement – or entrainment – of the grain by the wind would occur. In the natural world, grains are almost always perched on other grains, as in a sand dune, so that a grain at rest lies on a non-smooth surface. Thus, the forces acting on the grain must be expressed at torques (force multiplied by a lever arm distance) around a pivot point (Figure 13.4).

The individual torques can be expressed as:

$$\begin{aligned} r_D F_D &\sim K_D \rho_a D^3 u^{*2} \\ r_g F_L &\sim K_L \rho_a D^3 u^{*2} \mathrm{Re}^* \\ r_g F_g &\sim K_g (\rho - \rho_a) g D^4 \\ r_{ip} F_{ip} &\sim K_{ip} D^2 \ (\text{assumes } F_{ip} \propto D) \end{aligned} \tag{13.2}$$

in which $K_{\text{subscript}}$ denotes some empirically derived constant, ρ is the grain density, ρ_a is the atmospheric or gas density, and D is the grain diameter. Re* denotes the **Reynolds number** calculated with the friction wind speed (u^*), or u^*D/ν, where ν is the kinematic viscosity. The variable "r" with different subscripts denotes the distance of the lever arm for each of the corresponding forces.

Entrainment occurs when the wind-derived torques just barely exceed the torques holding the grain in place, that is, when

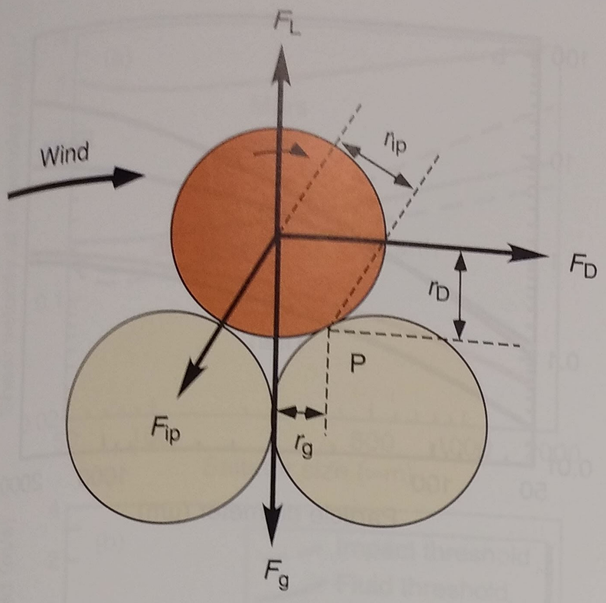

Figure 13.4 Force (torque) balance diagram, showing the various forces and their lever arms acting on a stationary grain before entrainment.

$$r_D F_D + r_g F_L \approx r_g F_g + r_{ip} F_{ip} \tag{13.3}$$

Solving this torque balance for u_* gives the minimum, or threshold, friction wind speed at which entrainment occurs. The equation for this threshold wind speed is:

$$u_{ft}^* = A\sqrt{\frac{\rho - \rho_a}{\rho} gD} \tag{13.4}$$

where u_{ft}^* is the threshold (minimum) friction wind speed for entrainment occurring only under the flow of the gas or "fluid." (As described in the next section, entrainment can also occur through the impact of other grains.) The variable A has a lot of hidden complexity because the interparticle forces are more complicated than a simple dependence on D. The equations for A contain (D^n) where n is ~2. Because of this exponent, the threshold wind speed curve as a function of D has a U-shape (Figure 13.5a). The minimum value (the bottom of the U) is termed the optimum diameter, or the grain size that is most easily moved by the wind.

13.3.2 Entrainment by Fluid and by Impact

The above derivation for the threshold wind speed considers only the flow of the gas. As gas is a fluid, this threshold speed is referred to as the fluid threshold speed. However, in the natural world, grains are also moved by the impact of other grains. The impacts of these other grains impart force to the bed, and thus affect the simple force balance shown in Figure 13.4.

The effects of impact vary with the density of the planetary atmosphere. For thinner atmospheric densities, as on Earth or Mars, the wind speeds have to be greater in order to impart the minimum necessary force to the grain to cause entrainment (Figure 13.6a). Thus, the impact of these fast-moving grains adds significantly to the force imparted by the wind to the grains on the surface, and this "splash" effect contributes to the movement of the grains. As a result, the force from the wind does not have to be as great to entrain the grains on the surface. Under these conditions, the impact threshold wind speed is less than the fluid threshold wind speed. On Earth, the impact threshold is roughly 90 percent of the fluid threshold. On Mars, with its even less-dense atmosphere, the differential is even greater: the impact threshold is roughly 10 percent of the fluid threshold. This significant differential means that grains may be entrained by a passing gust of wind, but because they then impact the grains around them, they may produce a cascading effect that – even at low wind speeds – results in greater movement than would be expected based on the wind speed alone. This effect is likely a significant factor in the global dust storms observed on Mars.

In contrast, on planetary bodies with dense atmospheres, such as Venus or Titan, the wind speed necessary to entrain the grains can be considerably slower. Because the grains are moving more slowly, they do not impart much additional force when they impact the grains on the surface. Thus, under denser atmospheric conditions, the wind speed for impact threshold is no less, and may be even numerically greater, than for the fluid threshold. Under these conditions, the wind speed for fluid threshold has to be maintained for continued transport (Figure 13.6b).

13.4 Aeolian Transport of Sediment

The movement of sediment particles occurs by several mechanisms. Before we explore these processes, we must consider how particle size affects grain transport.

13.4.1 Terminal Velocities for Sand versus Dust

Once grains of sediment have been entrained into motion through the atmosphere, they tend to fall. This simple and rather obvious statement actually provides an important means for distinguishing types of sediment, because different sediment sizes fall at different rates. Grains that fall more quickly do not travel as far (although they have more of an abrasive effect on the land surface over which they travel). Grains that fall more slowly can travel farther, and in some cases, much farther.

The terminal rate (or velocity) at which grains fall can be calculated using a force balance approach, in this case,

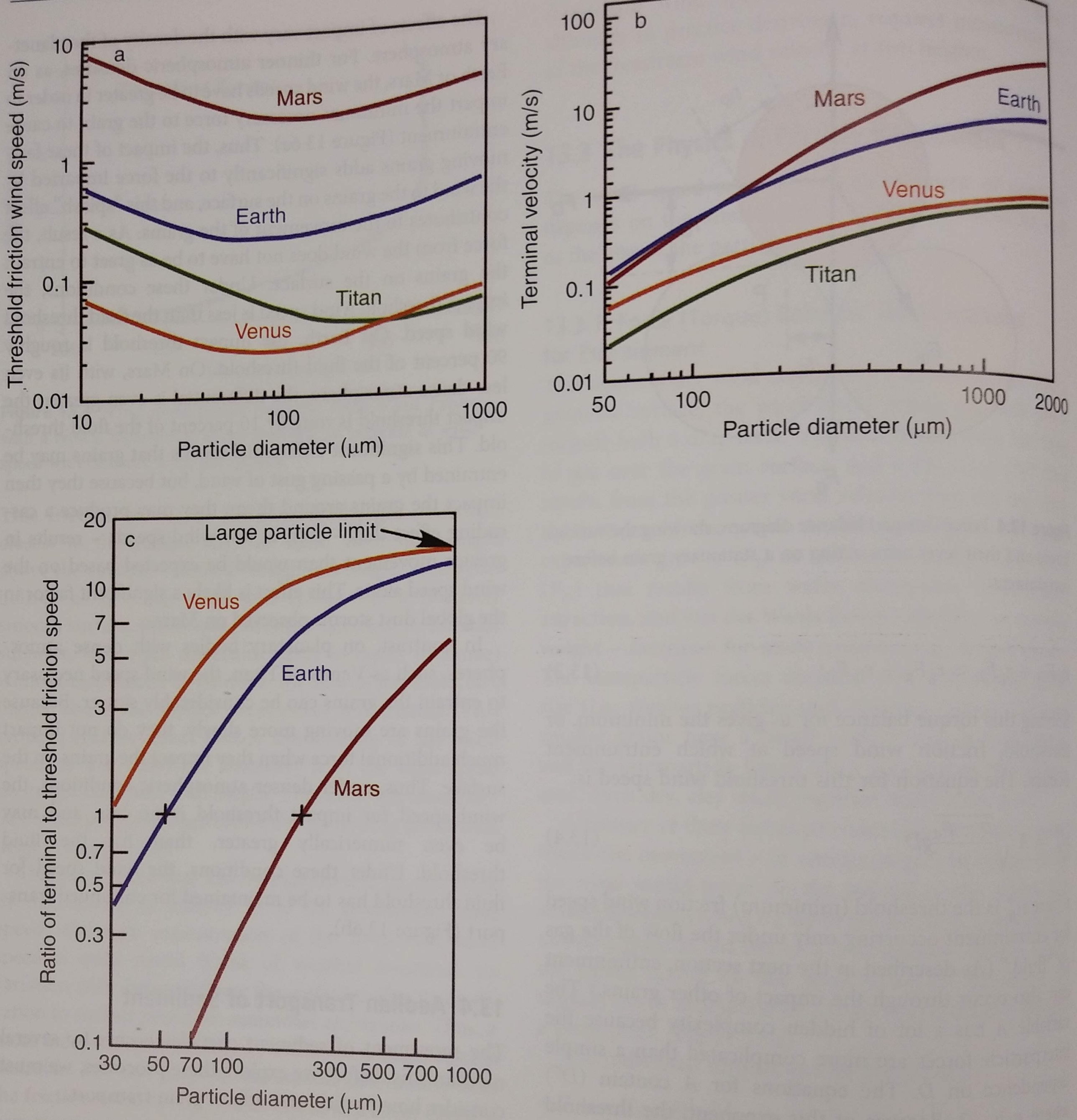

Figure 13.5 (a) Threshold friction wind speed curves for fluid entrainment on different planetary bodies as a function of particle size, derived from Equation 13.4. (b) Terminal velocities on different planetary bodies as a function of particle size, derived from Equation 13.6. (c) Ratio of terminal to threshold friction wind speeds. Grain sizes with values above 1 indicate transport through the atmosphere by suspension, whereas grain sizes with values above 1 indicate transport over the surface.

between the weight of a falling body and the drag exerted on it by the deflection of the atmosphere. Solving this force balance for velocity gives the equation

$$u_t = \sqrt{\frac{4\left(\rho - \rho_a\right)}{3\,\rho_a}\frac{gD}{C_D}} \qquad (13.5)$$

where u_t is the terminal velocity and C_D is a drag coefficient, largely a function of shape, which equates to ~0.4 for a sphere. This equation shows, as expected, that bodies fall faster if they are larger, denser, or under greater gravitational acceleration, and that they fall more slowly if the atmosphere is denser and/or the drag for the body shape is

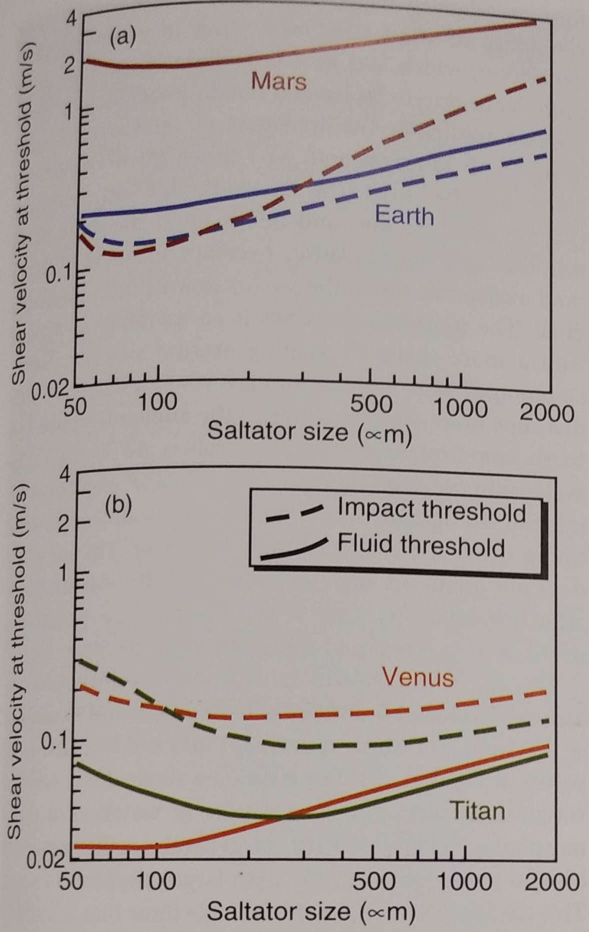

Figure 13.6 Threshold friction wind speed curves for fluid and impact entrainment on a variety of planetary bodies. (a) Planets with less dense atmospheres (Earth and Mars) have impact thresholds that are less than their fluid threshold, such that the wind speed may decrease (significantly for Mars) and the sand grains will remain in motion. (b) Planets with more dense atmospheres (Venus and Titan) have impact thresholds that are greater than their fluid thresholds. Modified from Kok et al. (2012).

greater. This approximation holds best for grains in the turbulent region, where the atmosphere creates drag due to deflection around the falling body. This condition implies that the falling body is relatively large, and under this condition, the atmospheric viscosity is not a factor (as shown by its absence from Equation 13.5).

For very small particles, the atmospheric viscosity (instead of deflection of the atmosphere around the body) provides the drag. In this case, the terminal velocity is approximated using **Stokes' Law**, which is

$$u_t = \frac{(\rho - \rho_a) g D^2}{18\mu} \qquad (13.6)$$

For small particles, this equation shows that the terminal velocity is inversely proportional to the viscosity of the gas and directly proportional to the square of the diameter (or surface area) of the particle. The square of a small number – such as the size of a ~10^{-6} m cloud droplet – is very small, showing that the terminal velocity of small grains is very slow. Together, these equations can be combined to yield a continuous curve for terminal velocity (Figure 13.5b).

Armed with this understanding of how grain size affects grain transport, we can now distinguish between major classes of grains. Larger sediment readily falls back to Earth. Here, we define "readily" as the condition in which the threshold friction velocity of the wind exceeds the terminal velocity of the grain, or where

$$u_t / u^*_{\mathrm{ft,it}} < 1 \qquad (13.7)$$

in which u_t is terminal velocity of the grain under the specific gravity and atmospheric viscosity conditions, and $u^*_{\mathrm{ft,it}}$ is threshold friction velocity for fluid or impact entrainment, governed largely by the atmospheric density. These larger grains move in more frequent contact with the surface, produce specific bedforms when they are deposited, and because they interact with the surface during transport, they cause more erosion. Conversely, if the terminal velocity is small compared to the wind speed that entrained the grains, or

$$u_t / u^*_{\mathrm{ft,it}} > 1 \qquad (13.8)$$

then the grains tend to stay in suspension (Figure 13.4c). Suspended grains travel farther and, because they fall more uniformly out of suspension, then tend to blanket landscapes without forming discrete bedforms.

13.4.2 Transport Mechanisms

Based on this understanding of how grain size affects transport, we can now be more specific about the different processes by which sediment is transported by the wind (Figure 13.7). Grains that move along the surface are transported by multiple, interrelated processes. The least energetic process is **creep**, in which grains are pushed, slid, or rolled over the surface by the force of the wind. With increasing wind speed, grains begin to bounce or hop over the surface in a process called **saltation** (from the Latin for jump). Saltating particles, when they impact, can cause other particles to be entrained in low hops for short distances, a process called reptation (related to the word reptile, which is from the Latin for crawl).

Smaller grains move in suspension, with infrequent or no interaction with the surface. Although small grains require relatively high wind speeds to be entrained

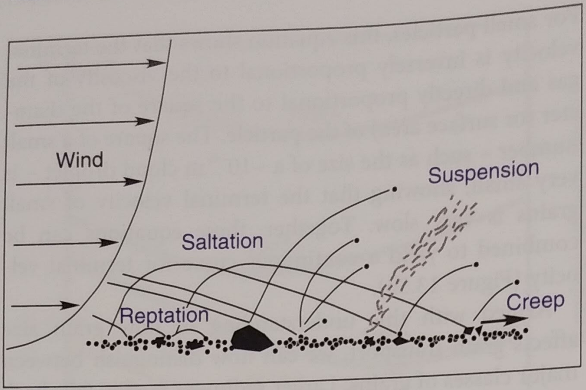

Figure 13.7 Diagram of sediment transport modes.

(remember the discussion of Figure 13.5), they require less energy than larger grains for transport because of their lesser mass (weight). Thus, dust is transported in suspension over very long distances. Dust from the Sahara Desert in Africa (Figure 13.8a) has been found in South America, and dust storms on Mars can become near-global in extent (Figure 13.8b). Dust can also be moved in dust devils, localized vortices that can exert sufficient lift to entrain dust (Figure 13.8c). Dust devil tracks on Mars are common in orbital images, showing both the frequency of dust devils and the prevalence of surface dust (Figure 13.8d). Dust devils in action have even been imaged by Mars rovers!

13.5 Aeolian Deposition and Planetary Landforms

When the wind flow slows to a speed below that necessary to keep grains in transport, the grains are deposited onto the surface. However, the mode of transport that the grains *were* experiencing before deposition strongly governs the resultant depositional landforms. The reason for the reduction in wind speed also influences the morphology and location of the deposition.

13.5.1 Depositional Landforms for Sand

As distinctive landforms that are often visible from orbit, sand dunes can provide valuable clues to the near-surface atmospheric processes (e.g., wind speed, direction), sediment locations and the transport pathways by which they arrive at those locations, and surrounding surface characteristics. Dunes, which build up through the progressive accumulation of sand, form different morphologies based on the conditions under which the sand was deposited.

This accumulation of sand commonly occurs in the lee of flow obstacles, such as rock outcrops. Dunes can also form on relatively flat terrain, where surface roughness or variability in wind flow may result in an initial sand deposition, which acts as a flow obstacle, causing more sand deposition in its lee and building the dune through positive feedback. The transport of sand over a dune begins with saltation and, to a lesser extent, reptation/creep up the stoss (or upwind) side of the dune (Figure 13.9). As the sand builds up at the dune crest, this localized accumulation becomes unstable and the sand avalanches down the lee (or downwind) side of the dune. The morphologic result is an asymmetric deposit with a more shallowly sloping upwind side (on Earth, common upwind slope values are 10–15°) and a lee side that dips more steeply, closer to the **angle of repose** (on Earth, common downwind slope values are 30–35°). The avalanching of sand down the lee side also engenders an internal stratigraphy to the dunes, specifically, downwind tilting strata also near the angle of repose. The lee slope does not attain an angle of repose due to deposition by other mechanisms, such as airfall, in which grains are lofted over the crest and settle out onto the lee face.

This net movement of sand from the upwind to the downwind side of a dune results in downwind migration of the dune. This dune movement may not be equal at all points in the dune field or even on a single dune, and the variation in sand transport results in variation in dune morphology (Figure 13.10). In general, smaller amounts of sand move more quickly than larger amounts of sand. This tendency means that for a single dune that is initially transverse to the wind, the ends (or horns) of the dune advance more quickly than the main body of sand. The bending of these dune tips downwind results in curved morphology in plan view, recognizable as a barchan dune. Under wind regimes that are more variable in direction, other morphologies, such as dome or star dune morphologies, result. Linear (or longitudinal) dunes occur where wind directions are bimodal, transporting sand first from one side of the dune, then from the other.

Thus, dune morphology and orientation (Figure 13.11) are fundamentally controlled by variability in wind direction and by sand availability (Figure 13.10). Sand availability is a combination of both the presence of sand-sized sediment and the readiness of that sediment for transport. For example, moist sand is less readily transportable than dry sand, so that sand availability may vary by season or even time of day. Where sand availability is high, dunes are larger and/or more closely spaced, even forming continuous ridges of dunes. If sand availability is low, the resultant dunes tend to be smaller, fewer in number, and/or more dispersed. While variability in the direction of winds above threshold influences dune morphology, the direction (or directions) of those winds

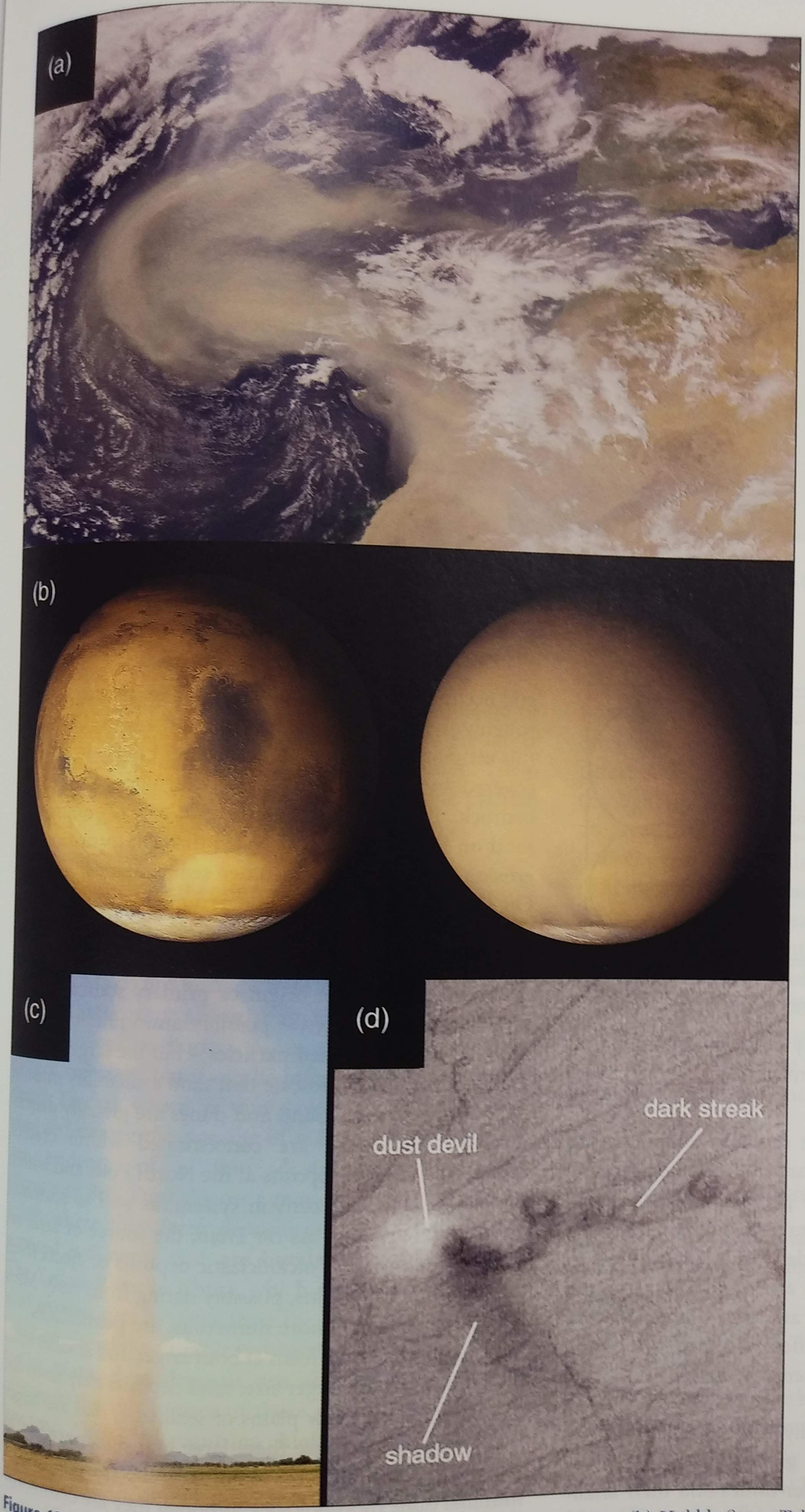

Figure 13.8 (a) Satellite image of a dust storm on the West Coast of Africa. (b) Hubble Space Telescope images of Mars showing a relatively dust-free atmosphere (left) and a global dust storm (right). (c) Ground image of a dust devil in Arizona. (d) Satellite image of dust devil and its track on Mars. NASA images.

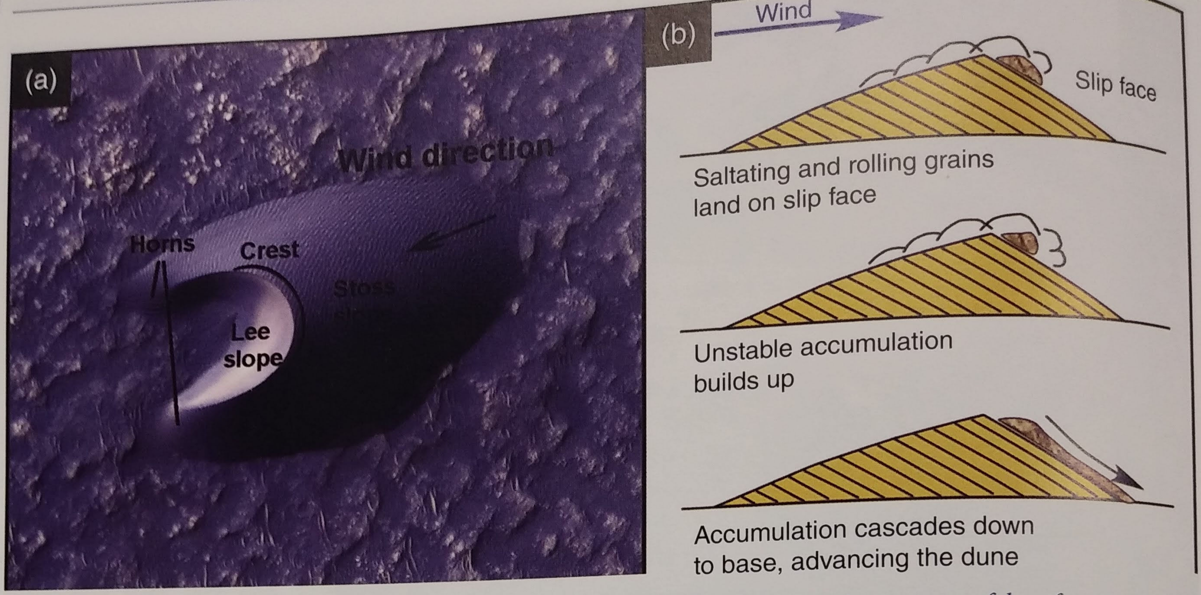

Figure 13.9 (a) Nomenclature for the parts of a dune on Mars (HiRISE image). (b) The granular processes of dune formation and the resulting internal stratigraphy.

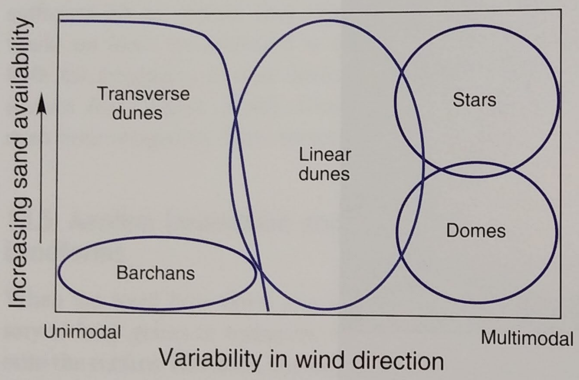

Figure 13.10 Plot of sand availability versus wind speed, showing stability fields for dune morphologies.

govern(s) the orientation of the resultant dunes. As the names imply, transverse dunes are transverse or perpendicular to the direction of the prevailing winds above threshold, whereas linear dunes are oriented between the two directions of the bimodal winds. In addition to wind direction and variability and sand supply, dune morphology and orientation are also influenced by anchoring or impedance. This effect of impedance means that certain dune types are particularly useful wind direction indicators.

Sand dunes, both individually and as fields, can be quite extensive in size. Dunes grow through the overtaking of larger dunes by smaller dunes and their amalgamation. Individual sand dunes reach ~200 m in height and occupy ~15–20 percent of the Sahara Desert, which is roughly the same size as the continental USA. Of course, this estimate means that ~75 percent of the Sahara is *not* covered by sand, but by areas of stone pavement or bedrock. Bedrock outcrops provide a ready source for sand on Earth, which is recycled through formation and erosion of sandstone and other sedimentary lithologies. On Titan, linear (or longitudinal) dunes cover most of the low latitudes, with individual dunes attaining lengths of tens of kilometers and heights similar to those in the Sahara on Earth. The origin or primary source of sand on Titan is not known – possibly atmospheric aerosols accreted into sand-sized particles – but the large size and extent of the dunes indicate that sand availability must be (or have been) high. Sand and dunes are globally distributed on Mars, and are concentrated within craters, around the layered deposits at the North Pole, and within the Valles Marineris canyon system, as well as scattered on intercrater plains. As on Titan, the source of sand on Mars is not known. Volcaniclastic deposits or ejecta from impacts into lava plains, possibly dating from early Mars when impacts were more numerous, are possibilities.

Sand deposits may form at both larger and smaller sizes than dunes. At the larger size, sand deposits may form flat or topographically low plains of sediments that are larger than sand size. Because these large grain sizes are not as readily transported by the wind, sand sheets exhibit no or minimal dune morphologies, although they are commonly found on the margins of dune fields. At the smaller size,

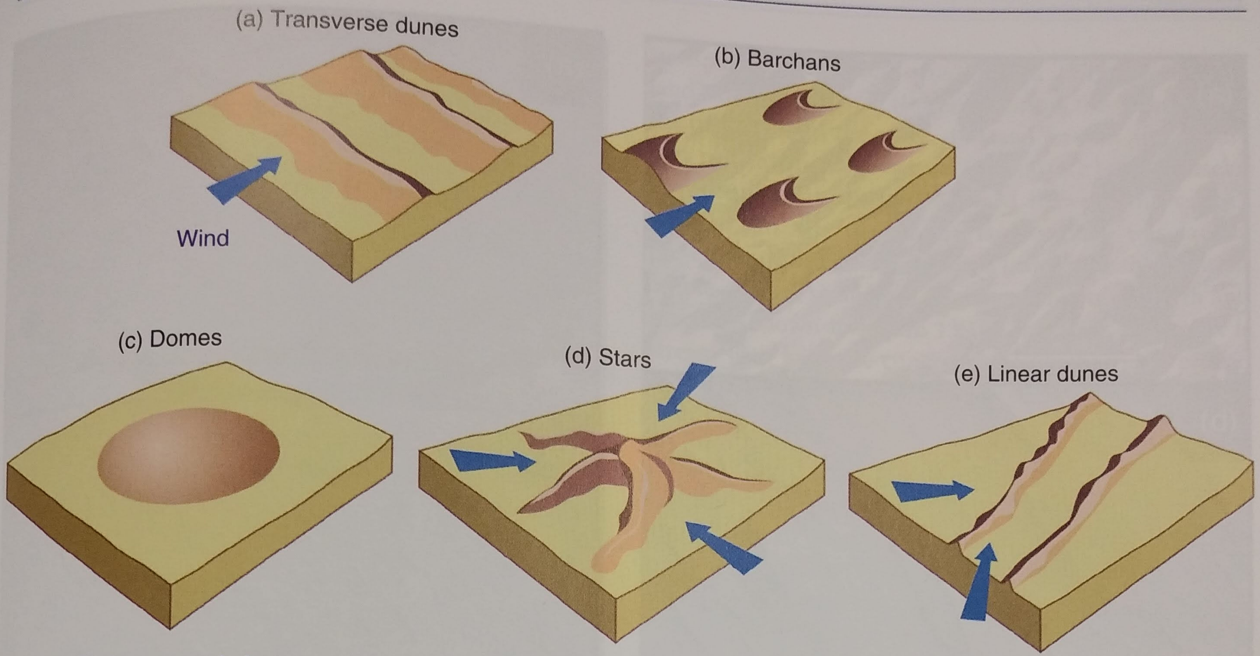

Figure 13.11 Dune morphologies are strongly controlled by wind direction and sand supply. Adapted from McKee (1979).

sand deposits form ripples as the products of reptation or creep. Mars exhibits some aeolian deposits that are larger than ripples on Earth but smaller than dunes. These "transverse aeolian ridges" may be specific to the martian conditions of a thin atmosphere and low gravity.

13.5.2 Depositional Landforms for Dust

When fine-grained dust settles out of suspension, it tends to blanket the landscape. Thus, in contrast to sand dunes, dust deposits do not form discrete landforms but rather a mantle that covers pre-existing landforms. Because it rarely develops landforms, this mantle can be difficult to discern remotely, but it can be inferred from a variety of clues. Tracks, where a thin layer of dust has been removed during the passage of a dust devil or a rover, are evidence of a thin dust mantle on Mars (Figure 13.8d). Dust deposits may be quite thick, as on Earth where they form deposits of loess (fine glacially scoured sediment) that are hundreds of meters thick. The Yellow River in China, which drains formerly glaciated regions, gets its name from the enormous volumes of loess or glacial dust that it transports.

On Mars, dust deposits are also areally extensive, although the thicknesses of these deposits are not well known. Although requiring higher wind speeds to entrain, dust is nevertheless extensively mobilized on Mars. Regional or global dust storms occur every few years (the frequency varies with orbital configuration). Because dust responds differently to heating than does sand or rock, thermal inertia maps can be used to detect surficial dust; they show that dust is distributed throughout the equatorial and low-latitude zones on Mars.

Dusty mantles can also include ice. The mid-latitudes of Mars show land surfaces that are currently undergoing disaggregation. This disaggregating terrain is interpreted as an ice-rich dust mantle that was emplaced during the last glacial (high-obliquity) period on Mars, but that is currently undergoing disaggregation or breakdown due to the current interglacial (low-obliquity) period. As this breakdown releases fine grains, dust loading in the martian atmosphere may increase.

13.6 Planetary Erosional Landforms

In addition to creating landforms by depositing material, aeolian sediments can also create landforms by removing material.

13.6.1 Yardangs

Yardangs (from the Turkic word meaning "steep bank") are erosional landforms composed of sedimentary rock that has been abraded by wind-borne sand into a streamlined form (Figure 13.12a). Yardangs are extensive on Mars (Figure 13.12b), perhaps because the high threshold

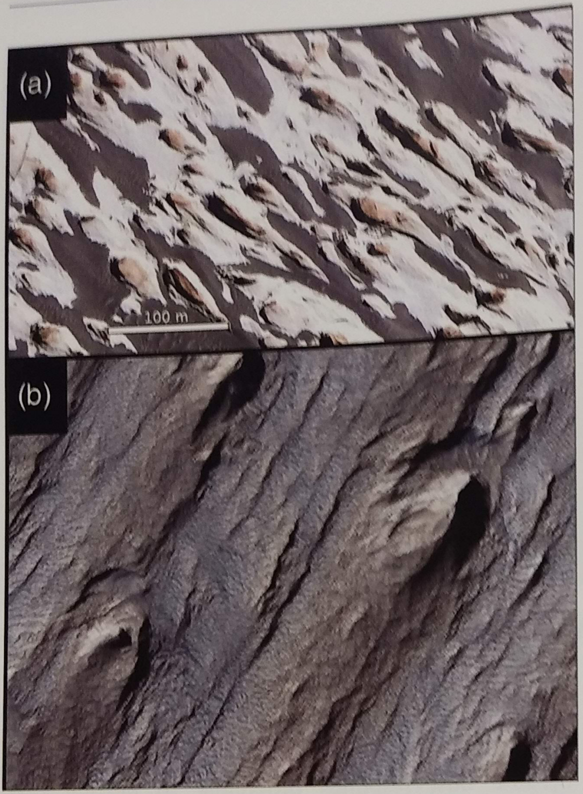

Figure 13.12 Erosional landforms. Yardangs (a) in the central Andes mountains, Chile (Google Earth), and (b) in Tithonium Chasma, Mars (NASA image). The scale bar in (a) applies also to (b).

wind speed, which gives wind-borne sand a very high kinetic energy, is more erosive than under denser atmospheres (e.g., on Earth). In addition, extensive sedimentary layers provide the necessary bedrock for yardang formation. Yardangs have also been hypothesized to exist on Titan and Pluto.

13.6.2 Ventifacts

Whereas yardangs are landscape in scale, ventifacts occur at sub-outcrop scale. Ventifacts result from abrasion of crystalline rock, although less defined examples can also form in sedimentary lithologies. Thus, like yardangs, they serve as current or paleo-wind indicators. Abrasion of crystalline boulders results in faces (or facets) with polished surfaces, pits, subparallel grooves, and/or flutes where the facets join. Such ventifacts have been observed on Earth (Figure 13.13a) and by rovers on Mars (Figure 13.13b), where the facets' dip slopes and groove direction give evidence of the direction of wind with speeds above threshold and an upwind source of available sand.

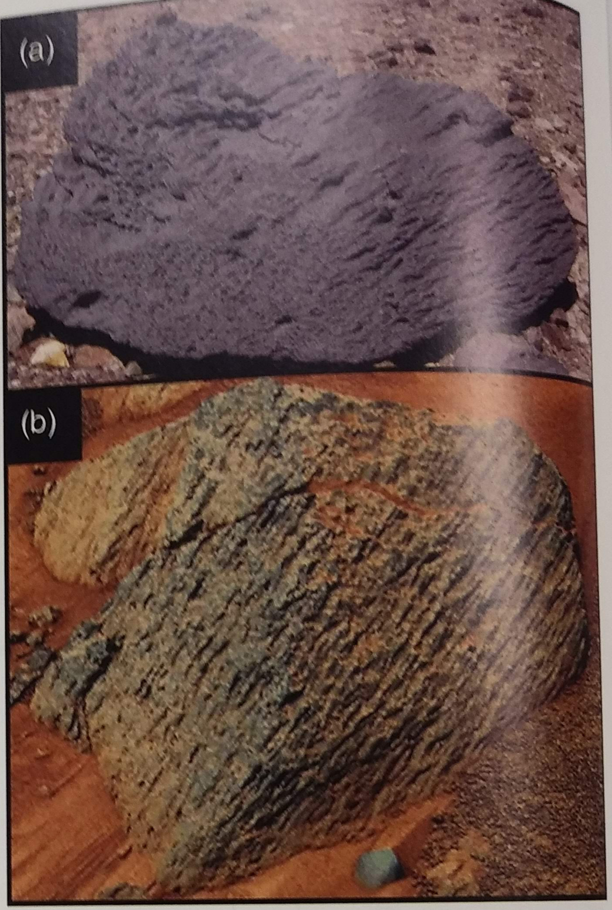

Figure 13.13 Ventifacts (a) in Death Valley, California (released under terms of the GNU FDL), and (b) near the Spirit rover landing site in Gusev crater, Mars (NASA image).

13.7 Combined or Ambiguous Planetary Landforms

Beyond known depositional or erosional landforms, the wind also creates landforms whose mechanism of formation is ambiguous.

13.7.1 Stone Pavements

Stone pavements are composed of extensive surficial sheets of coarse sediment, generally grains that are pebble-size or larger. Two models of formation are proposed for pavements. In one, the pavement is a lag deposit left behind by aeolian winnowing of more easily transportable (e.g., sand-sized) grains. In this model, the land surface is deflated (lowered) during the formation of the pavement. In the other, dust and fine silt accumulate on a pebble- or cobble-strewn land surface but, because of their smaller sizes, sift downward between the coarser particles. In this model, the land surface was largely pre-existing but has been inflated (raised) by the addition of fine-grained sediments. Stone pavements on Earth and Mars are shown in Figure 13.14. The Mars pavement is strewn with hematite concretions ("blueberries") eroded out of the rock during deflation.

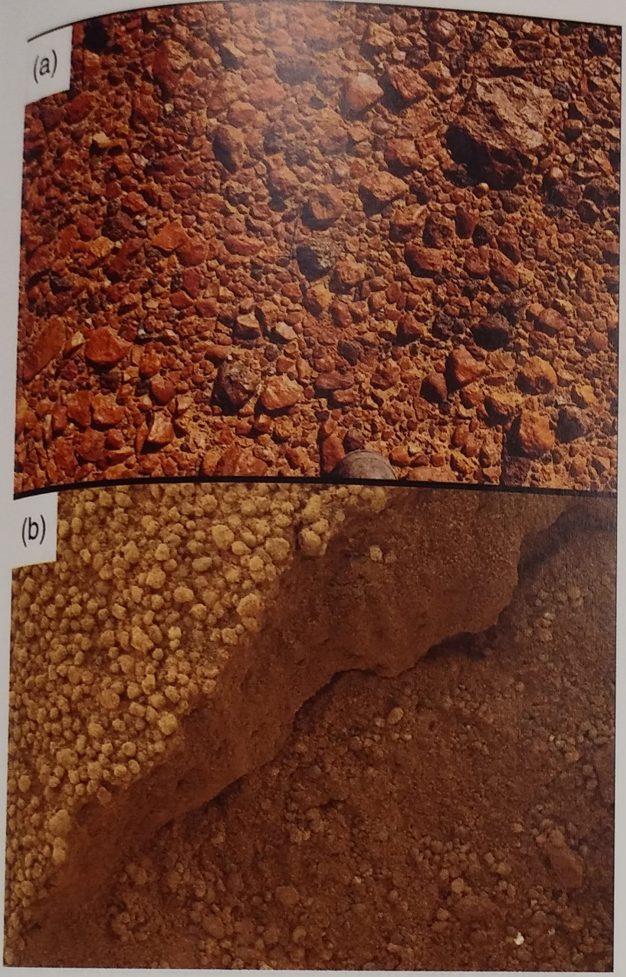

Figure 13.14 Stone pavements (a) in Australia (Mark Marathon, https://commons.wikimedia.org), and (b) in Gale Crater, Mars (NASA image).

13.7.2 Wind Streaks

Elongated areas of different albedo than the surrounding landscape can be wind streaks. These aeolian features occur downwind of obstacles and may result either from deposition of sediment due to a reduction of wind speed in the obstacle lee or from the removal of sediment due to increased turbulence. Thus, wind streaks on Mars are commonly bright due to deposition of dust, but may also be dark where a dust covering has likely been removed by movement of sand by turbulent eddies. Wind streaks that appear bright in radar images of Venus are evidence of larger grain sizes in the lee of the obstacle, whereas wind streaks that appear darker are smoother, perhaps being covered with smaller (e.g., silt) grain sizes (Figure 13.1a). In other situations, wind streaks may form downwind of a sediment source, for example on Triton (Figure 13.1d), where they are interpreted as deposits from geyser-like eruptions of gas, dust, and ice from the polar cap.

Summary

While atmospheric phenomena themselves, such as clouds, provide evidence of atmospheric conditions (Chapter 12), aeolian landforms give valuable clues to atmospheric processes occurring at the surface, including wind direction, strength, and variability. These processes can fundamentally affect the geology of planetary surfaces, redistributing sediments and changing surface albedo. These changes, in turn, cause variations in surface heating, which feed back into atmospheric processes, and provide records of sediment transport and redistribution. Dune morphologies provide evidence both of winds above threshold for extended periods of time and of sand availability in the upwind direction, whereas the orientation of the dunes provides evidence for the direction of these winds and the location of the sand source. Wind streaks provide similar information about the wind, whereas dust is more strongly linked to climate. Yardangs and ventifacts provide evidence not only of wind and sediment supply, but also of the bedrock lithology. Thus, the study of aeolian landforms is an important source of information for understanding planetary geologic processes.

Having considered how flowing gases influence planetary geology, we will next address the same question for flowing liquids.

Review Questions

1. What conditions are necessary for aeolian landscapes? Does your answer to this question differ between erosional and depositional landscapes? For different planets?
2. What information do aeolian landscapes provide? Does your answer to this question differ between erosional and depositional landscapes? For different planets?
3. How can aeolian landscapes be used to understand sediment transport paths?
4. How can aeolian landscapes be used to constrain atmospheric models?
5. Compare and contrast aeolian processes on Earth, Mars, Venus, and Titan.

SUGGESTIONS FOR FURTHER READING

Bagnold, R. A. (1941) *The Physics of Blown Sand and Desert Dunes.* London: Methuen and Co. Originally published by Methuen and Co, reprinted by Dover Publications. A classic in terrestrial aeolian science, no worse for age and inspiring in its technological creativity and the resultant insights.

Greeley, R., and Iversen, J. (1985) *Wind as a Geologic Process.* Cambridge: Cambridge University Press, An older but still useful "first view" of planetary aeolian processes.

Lorenz, R. A., and Zimbelman, J. R. (2014) *Dune Worlds: How Windblown Sand Shapes Planetary Landscapes.* New York: Springer-Praxis. A recent peripatetic review of sand movement by wind, how it can be studied, and what it can tell us.

McKee, E. D., ed. (1979). *A Study of Global Sand Seas.* Washington, DC: USGS. https://pubs.er.usgs.gov/publication/pp1052. A nice survey of global-scale deposits on Earth.

REFERENCES

Bridges, N. T., Greeley, R., Haldemann, A. F. C., et al. (1999) Ventifacts at the Pathfinder landing site, *Journal of Geophysical Research*, **104**, 8595–8615.

Burr, D. M., Bridges, N. T., Marshall, J. R., et al. (2015) Higher-than-predicted saltation thresholds on Titan. *Nature*, **517**, 60–66.

Ewing, R. C., Hayes, A. G., and Lucas, A. (2015) Sand dune patterns on Titan controlled by long-term climate cycles. *Nature Geoscience*, **18**, 15–19.

Greeley, R., Bender, K. C., Saunders, S., et al. (1997) Aeolian processes and features on Venus. In *Venus 2: Geology, Geophysics, Atmosphere, and Solar Wind Environment*, eds. Bougher, S. W., Hunten, D. M., and Phillips, R. J. Tucson, AZ: University of Arizona Press, pp. 13–62.

Greeley, R., Whelley, P. L., Neakrase, L. D. V., et al. (2008), Columbia Hills, Mars: aeolian features seen from the ground and orbit. *Journal of Geophysical Research*, **113**, E06S06.

Kok, J. F., Parteli, E. J. R., Michaels, T. I., et al. (2012), The physics of wind-blown sand and dust, *Reports on Progress in Physics*, **75**(10), 72. DOI: 10.1088/0034-4885/75/10/106901.

McKee, E. D., ed. (1979). *A Study of Global Sand Seas.* Washington, DC: USGS.