CHAPTER 11

MOVEMENT OF SEDIMENT BY THE WIND

INTRODUCTION

1 Everyone knows that winds on the Earth are commonly strong enough to erode, transport, and deposit sediment. What is perhaps less obvious is that the modes of sediment transport by the wind are greatly different from those of sediment transport by water flows. This great difference does not arise from any great difference in the structure of the wind at the lowermost levels in the atmosphere: you saw in Chapter 7 that low in the atmospheric boundary layer the dynamics of flow are the same in all essential respects as in turbulent shear flows above a solid boundary in water. The difference lies in the greatly different ratio of sediment density to fluid density, which is almost eight hundred times greater in air than in water; go back and look at Figure 8-5 in Chapter 8 to see where the point for ρ_s/ρ lies for quartz-density particles in air, relative to the point for quartz-density particles in water. This difference has profound effects on the nature of particle movement in the two fluid media. As discussed briefly in Chapter 8, the very large ratio of particle density to air density means that the trajectories of particles that are in transport by the wind are largely independent of the fluid turbulence, except for fine particles, in the silt and clay size range.

2 Another important difference between sediment transport by wind and sediment transport by water is that the wind is a more efficient size-sorting agent. For transport by water, it is broadly true that larger particles are more difficult to move than finer particles—silts are moved much more readily than gravels, for example—but the weakness of this effect is highlighted by the nearly equal mobility of a wide range of sand to gravel sizes in many flow settings, as discussed in Chapter 14. By contrast, the wind entrains dust and silt much more readily than sand, provided that the sediment is not bound to the substrate by cohesive forces, and gravel is much more difficult to move than sand. Except in the very strongest winds, all but the finest gravel sizes are invariably immobile, whereas water flows, even leaving rheological flows like debris flows out of account, can move even large boulders if the flow is sufficiently strong.

3 It is not an exaggeration for me to say that the modern era of study of sand movement by the wind started with R.A. Bagnold's work in the deserts of North Africa in the 1930s, which culminated in the publication of his little book (literally "little": 265 pages in a book measuring 22 cm by 14 cm) *The Physics of Blown Sand and Desert Dunes* in 1941. It is a classic, in the fullest sense of the term: it is an outstanding example of a magisterial work that sets the course of future work in a field of science for many decades. It is by far the most widely cited work on eolian sediment movement, and it remains essential reading for anyone who is seriously interested in the topic. Also, several extensive early wind-tunnel studies of eolian sand transport, with results that are still valuable today, are worthy of mention (Kawamura, 1951; Zingg, 1952, 1953; Horikawa and Shen, 1960; Belly, 1964). Chepil, in a long series of papers, (see especially Chepil, 1945, 1958, 1959), was the pioneer in modern studies of wind erosion of soils; some of his work bears directly upon the transport of loose sand by wind. After the appearance of a multitude of papers on saltation from the mid-1970s to the mid-

1990s, in large part from just a few groups of researchers (Greeley and co-workers; Willetts and co-workers; Anderson, Haff, and co-workers; see the list of references at the end of the chapter), the frequency of published works on saltation has decreased somewhat. You are likely to get that impression if you scan the list of references. For clear reviews of the eolian sediment movement, see Greeley and Iversen (1985), Anderson (1989), Anderson et al. (1991), and Willetts (1998).

4 Research in the field of eolian sediment transport, over the past several decades, has fallen fairly naturally into three overlapping areas: soil erosion; transport of sand by saltation; and the nature and dynamics of eolian bed forms (wind ripples and eolian dunes). (The adjective *eolian*, meaning *produced, eroded, carried, or deposited by the wind*, and spelled *aeolian* in British-style English, comes from the name of a minor Greek god, Aeolos, who was the keeper of the four winds; see the Encyclopedia Mythica or the Wikipedia on the Internet for more information.) This chapter deals with the second of those areas. Loess—deposits of windblown silt that is carried in suspension far from its source, for tens or even hundreds of kilometers—covers a far larger percentage of the Earth's surface than eolian sand, and it is important for agriculture in many parts of the world, but the topic of loess deposition is beyond the scope of these notes.

SALTATION

Introduction

5 The characteristic mode of motion of sand particles in air is ***saltation***: particles are launched from the bed, take arching trajectories of widely varying heights and lengths, and splash down onto the bed at low angles, commonly rebounding and/or putting other particles into motion. The term, introduced into geology by McGee (1908, p. 199), is derived from the Latin verb *saltare*, meaning to jump or leap. Movement by saltation has also been invoked for water transport of particles near the bed (see chapter 10), although the distinctiveness of saltation in water is not nearly as clear as in air.

6 Saltation in air became well known through the early experimental studies by Bagnold (1941), Chepil (1945), Zingg (1952), and others. In recent years there has been much attention to eolian saltation, in part because of the growing concern over desertification, and also in part because of the interest in how sediment is transported by wind on other planets—most especially, Mars. Early studies of saltation dealt in large part with the nature and dynamics of saltation trajectories. Later, especially during the late 1980s and early 1990s, emphasis tended to shift to a more unified consideration of the overall saltation system produced by a steady wind. In more recent years, this has extended to study of saltation in the unsteady winds characteristic of natural environments. Also, as computational power has grown it has become possible to develop increasingly sophisticated numerical models of saltation.

7 The study of saltation can be viewed as falling into several related areas:

- threshold for motion
- forces causing liftoff

- the geometry and dynamics of particle trajectories, including the distributions of jump height and jump distance
- the effects of wind velocity and of sediment size, sorting, and particle shape on mode of saltation and on saltation transport rates
- the effect of unsteadiness of the wind on saltation
- the effect of the saltation cloud on the structure of the near-surface wind

To some extent it is artificial to treat these topics separately, but nonetheless it seems helpful in developing clear understanding. Accordingly, each of these topics treated in sections below, after some comments about observing saltation.

8 Saltating particles are highly abrasive, because of their very large relative inertia—much greater than for water-borne particles. Both natural and artificial solid materials at heights within the saltation cloud, even hard rocks, are gradually abraded. Saltating sand also sculpts distinctive eolian landforms. Such topics are not within the scope of these notes.

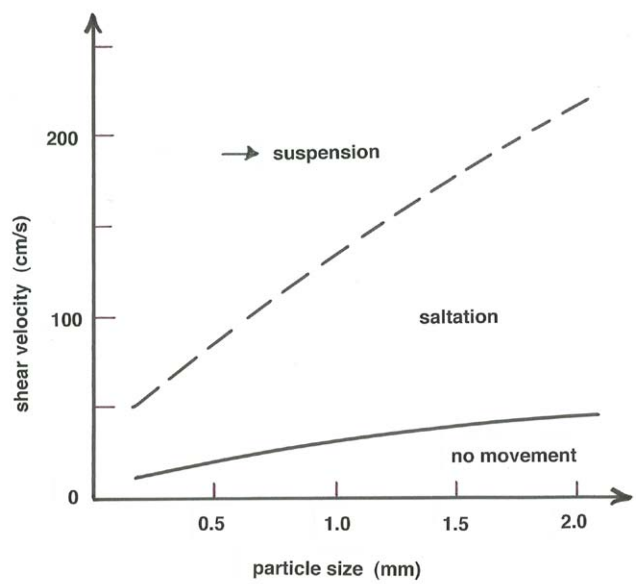

Figure 11-1. Cartoon graph showing the ranges of distinctive modes of eolian particle movement as a function of sediment size and wind speed. (Inspired by Figure 2 of Owen, 1964.)

9 As the size of particles in saltation decreases toward the silt range, the decrease in particle mass means an increasing effect of turbulence on particle trajectories. Wind speed is important in this respect as well, inasmuch as the characteristic magnitude of velocity fluctuations from eddy to eddy increases with mean wind speed. At sufficiently fine particle sizes, and for sufficiently strong winds, the particles are carried in suspension rather than in saltation; see Figure 11-1 (in the same spirit as Figure 10-3 in Chapter 10), showing in cartoon form the regions of distinctive modes of eolian particle movement as a function of sediment size and wind speed. What is known about the transition from saltation to suspension is described in a later section of this chapter.

10 A distinction is commonly made between saltation, whereby particles take ballistic excursions well above the bed, and ***surface creep*** (also called ***impact creep*** or ***reptation***), whereby particles are moved for short distances without losing contact with the bed surface. Particles that are too large to be moved in saltation (but not so large as to be immovable by the given wind) characteristically engage in surface creep. Particles of sizes susceptible to saltation can also move as creep, however, if a saltation impact is sufficient only to impart slight movement to a given particle on the bed surface. Even in very well sorted sediments, surface creep as well as saltation is an important mode of transport.

Observing Saltation

11 The very best way to appreciate saltation is to observe it for yourself. Imagine yourself out on the surface of a sand dune on a windy day. If you get your eye level down to within a few decimeters of the surface—you risk getting sand in your eyes, ears, nose, and mouth—and sight horizontally across the wind, you see a blurry layer of saltating sand, with concentration tailing off upward for as much as a meter above the surface. You are seeing the characteristic saltation cloud. Unfortunately, your eye cannot easily follow the trajectories of individual particles.

12 To see saltation trajectories clearly, you need to build your own wind tunnel (Figure 11-2). That's not a difficult matter, even if you are on a limited budget and have no more space than an ordinarily large spare room. A classic "Bagnold" wind tunnel consists of a horizontal rectangular duct, wider than high and with a flared entrance, emptying into a large box equipped with a fan mounted high in the wall opposite the downwind end of the duct. If you fabricate the roof of the duct in the form of several removable segments, it is easy to gain access to the sand bed. Because fans with continuously variable speed are not easy to obtain or arrange, it would be helpful to mount an adjustable louver just outside the fan, in order to set the wind velocity in the duct to any desired value. Lay in a planar bed of medium sand in the duct, turn on the fan, and gradually increase the wind velocity until saltation is established. The only significant difference between saltation in your wind tunnel and the saltation you observed on the sand dune is that the range of eddy sizes in the duct is much smaller, the consequence being that the wind is not nearly as gusty: the saltation is much closer to being steady (unchanging with time).

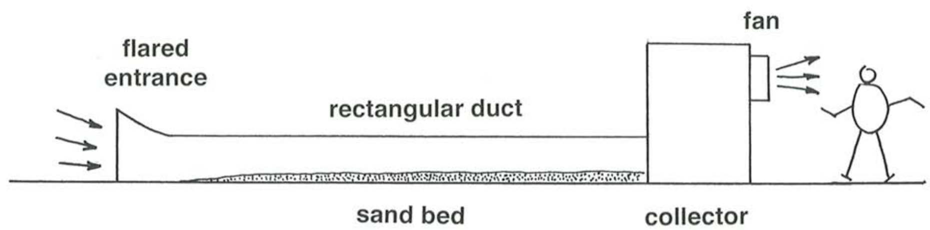

Figure 11-2. A simple but effective wind tunnel.

13 To see saltation trajectories (Figure 11-3), cut a thin slit along the centerline of the tunnel roof, not far from the downwind end, and mount a strong light source above the slit, with a second slit between the strobe and the roof, for good collimation. With that arrangement you can illuminate a thin streamwise slice of the flow. Trajectories of saltating particles that move in this illuminated slice show up well as curving bright streaks. It would be even better to use a stroboscope as the light source. Then the trajectories show up as series of closely spaced illuminated dots. If the concentration of saltating particle is not too high, so that individual trajectories can be discriminated, then by making sufficiently careful measurements of a photographic image you could compute velocities and accelerations of individual saltating particles along their trajectories.

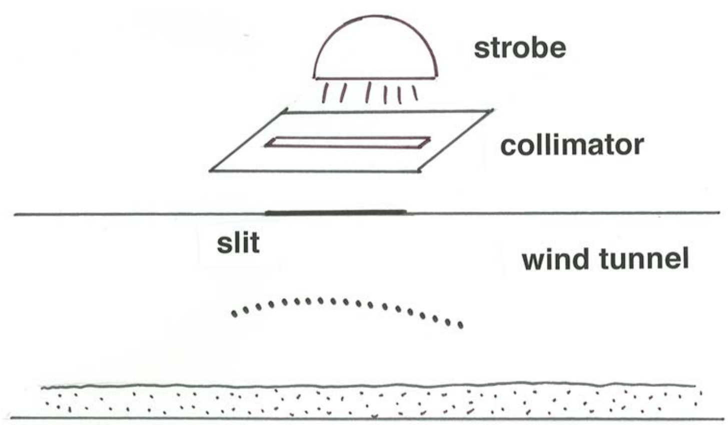

Figure 11-3. A lighting arrangement to see saltation trajectories.

Saltation Trajectories

14 The general nature of the trajectories of saltating particles is known from early descriptions by many investigators, most notably Bagnold (1941) and Zingg (1953), but also in several later studies. Figure 12-4, from Maegley (1976), is a representation of a typical saltation trajectory from the early literature on saltation. After launch, the subsequent path of the particle is the outcome of the constant downward force of gravity (that is, the weight of the particle), on the one hand, and the fluid drag force occasioned by the motion of the particle relative to the surrounding air, which evolves as the particle traverses its path.

15 Although some authors have described the saltation trajectory as parabolic, what is immediately apparent about the trajectory in Figure 11-4 is that it is *asymmetrical*: the angle of takeoff is much larger than the angle of impact. You could make two other significant observations about saltation trajectories: they are convex upward all along their courses, from takeoff to landing; and their length, from takeoff to landing, is much greater than the maximum height they reach.

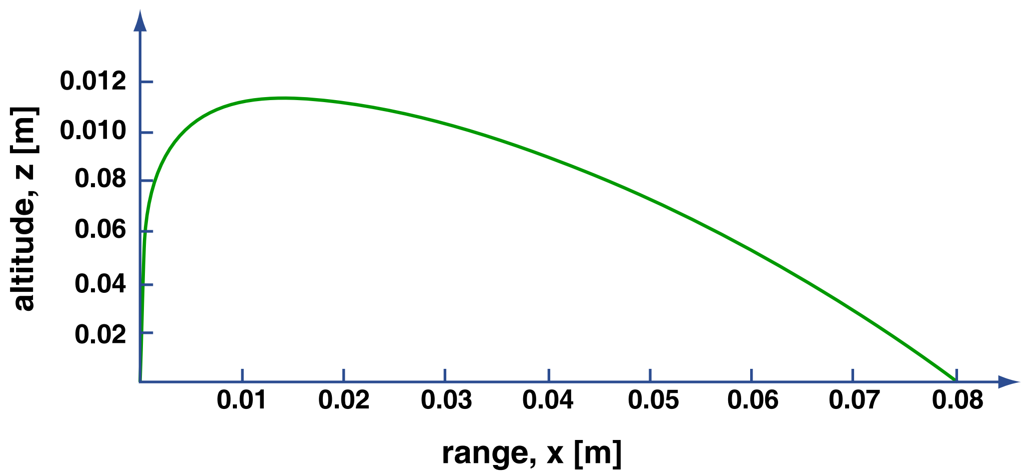

Figure by MIT OpenCourseWare.

Figure 11-4. A typical saltation trajectory. (From Maegley, 1976.)

16 You can gain some qualitative insight into the asymmetry and upward convexity of saltation trajectories noted above by means of simple thought experiment on particle accelerations. Suppose that a particle is somehow launched into the air at some representative angle, say forty to fifty degrees, at some initial speed (Figure 11-5). If the medium is a vacuum, you know from elementary physics that the trajectory of the particle, from takeoff to landing, would be a perfect parabola (Figure 11-5A). If the medium is air at rest, then the height of the trajectory would be slightly smaller, because air drag adds to the downward force of gravity and makes the vertical component of deceleration during ascent smaller. Air drag also acts to decrease the horizontal component of velocity throughout the course of the trajectory, so the descent of the particle is at a steeper angle than the ascent (Figure 11-5B).

17 Now suppose that the particle is launched at the same angle and initial speed into a wind stream. There are two cases to consider: (1) the initial speed of the particle is less than that of the wind stream, and (2) the initial speed of the particle is greater than the wind speed. (Here we assume, for simplicity, that because of the logarithmic shape of the velocity profile the particle traverses only a very thin layer of low wind velocity in the immediate proximity to the surface and thereafter finds itself, for most of its path, in a region in which the wind velocity is nearly constant with height. This does not do damage to our first-order thought experiment.)

18 If the initial speed of the particle is *greater* than the wind speed, the wind causes horizontal deceleration, just as in the case of launch into still air. Qualitatively, the shape of the trajectory is the same as in the case of launch into still air (Figure 11-5B). If, however, the initial speed of the particle is *smaller* than the wind speed, the wind causes horizontal acceleration, and the steepness of the ascending part of the trajectory is smaller (Figure 11-5C). The steepness and the shape of the descending part of the trajectory depends on the relative importance of the downward pull of gravity and the remaining horizontal acceleration, but in any case the downward path is less steep than the ascending part. What we can conclude from this simple exercise is that, by comparison of Figure 11-5C with Figure 11-4, in typical saltation the particle is launched into the wind with a smaller horizontal component of velocity than the speed of the wind in the region well above the surface.

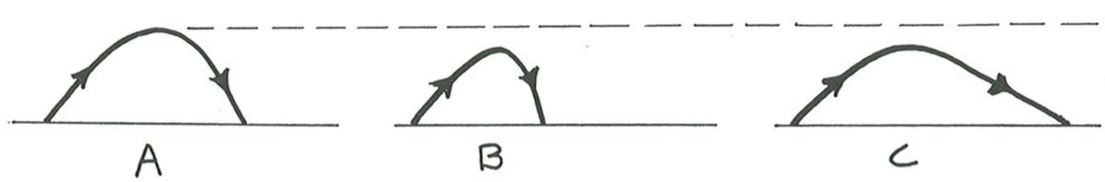

Figure 11-5. Qualitative trajectories of particles launched at a fixed angle from a horizontal surface: **A)** in a vacuum; **B)** into air at rest; **C)** into a wind stream with speed greater than the initial horizontal component of particle velocity.

19 Following the early observations of saltation trajectories by Bagnold (1941) and Chepil (1945), many authors have assumed that the particles typically leave the bed at a steep, nearly vertical angle. Careful measurements of frequency distribution of takeoff angles by White and Schulz (1977), by use of the technique described above for viewing saltation trajectories in a wind tunnel, together with high-speed cinematography, showed that the average takeoff angle was 50°, and less than 10% of the particles observed had takeoff angles of more than 80° (Figure 11-6A). A notable feature of the distribution shown in Figure 11-6A is that the distribution is strongly skewed: the mode lies in the range 20–40°, and the distribution tails off steadily toward steeper angles, but no angles less than 20° were measured. White and Schulz also found that the average angle of impact at the end of a saltation trajectory was 14° (Figure 11-6B), and the distribution was much more nearly symmetrical.

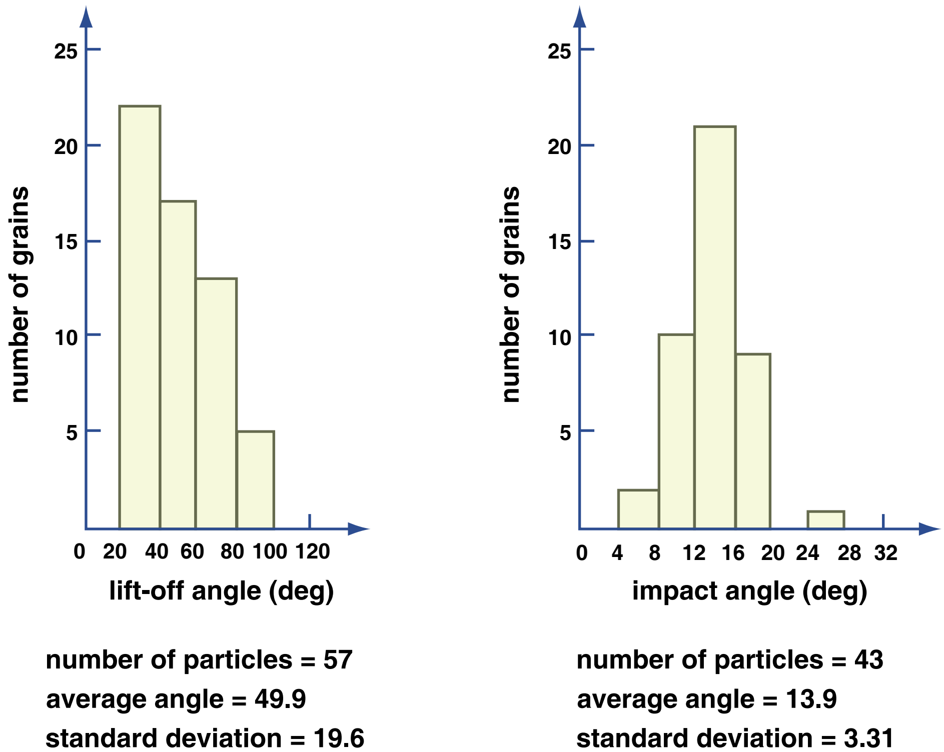

Figure by MIT OpenCourseWare.

Figure 11-6. Frequency distribution of **A)** takeoff angle and **B)** impact angle for 0.5 mm glass spheres saltating in a wind tunnel (From White and Schulz, 1977.)

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20 The results obtained by White and Schulz might be questioned because they were obtained from single-size glass spheres. More recent studies have found lower launch angles. Willetts and Rice (1985), using natural sands, measured average takeoff angles of 52–54° for particles ejected from rest by impacts of already saltating particles but considerably smaller average angles of 21–33° for rebounds of already saltating particles. Nalpanis et al. (1993) measured takeoff angles of 35–41°, for natural sands, and Nishimura and Hunt (2000) measured even lower takeoff angles of 21–25° for ice spheres and for spherical mustard seeds. If large and immovable particles are present on the bed surface, finer particles in saltation are observed to rebound from them upon impact at sometimes very steep angles, in some cases even with a component in the direction opposite to the wind.

Saltation Lengths

21 Why are the lengths of saltation trajectories so much greater than the heights? It was noted at the beginning of this chapter that the relative inertia of sand particles in air is extremely large, but nonetheless the air at all times exerts a drag force on the particles, because there is always a difference between the velocity of the particle and the velocity of the wind. Only for very fine dust particles in suspension does this velocity difference become negligible.

22 White and Schulz (1977) also measured takeoff speeds and impact speeds of saltating particles (Figure 12-7). Takeoff speeds averaged about 70 cm/s, not much more than the friction velocity u_*—but keep in mind that such a value of u_* corresponds to wind speeds of several meters per second only some centimeters above the bed. Suppose that a sand particle is launched vertically into a wind stream at such a speed. The rising particle almost immediately encounters much higher wind speeds. At any given instant, the velocity of the particle relative to the air is the vector difference between the velocity of the particle relative to the ground and the horizontal velocity of the wind relative to the ground (Figure 11-8). In the initial, rising part of the trajectory, this vector velocity is directed upward and upwind.

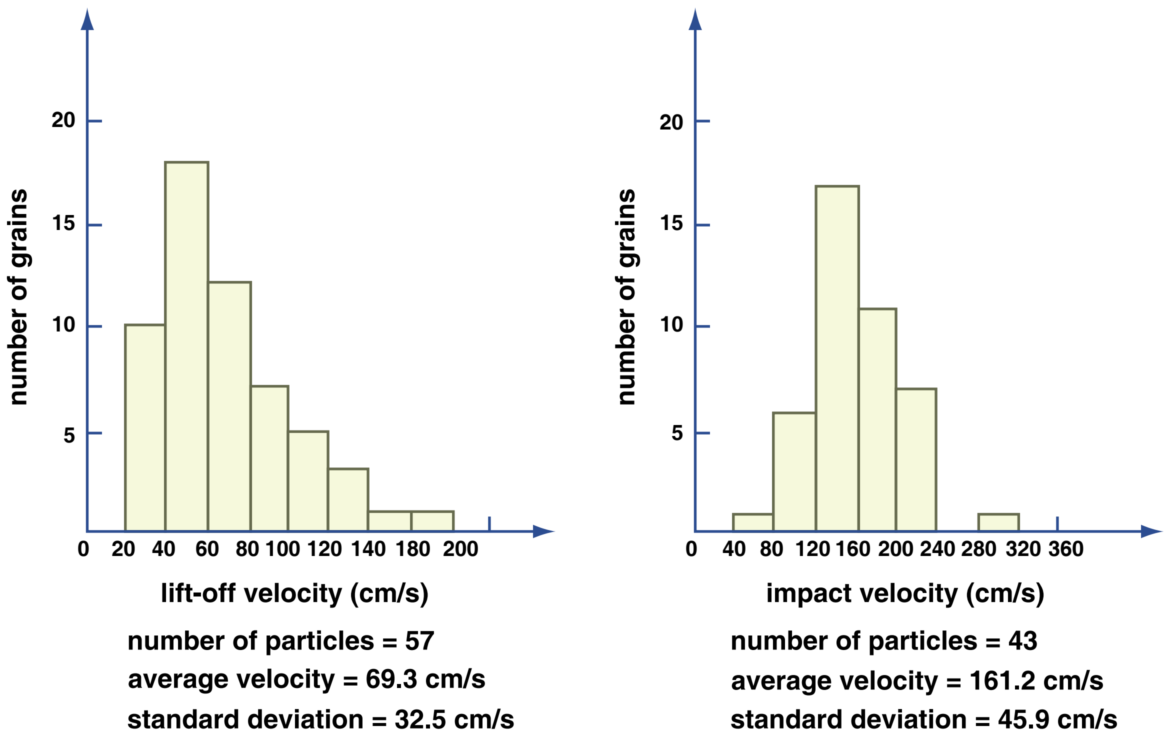

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Figure 11-7. Frequency distribution of **A)** takeoff speed and **B)** impact speed for 0.5 mm glass spheres saltating in a wind tunnel (From White and Schulz, 1977.)

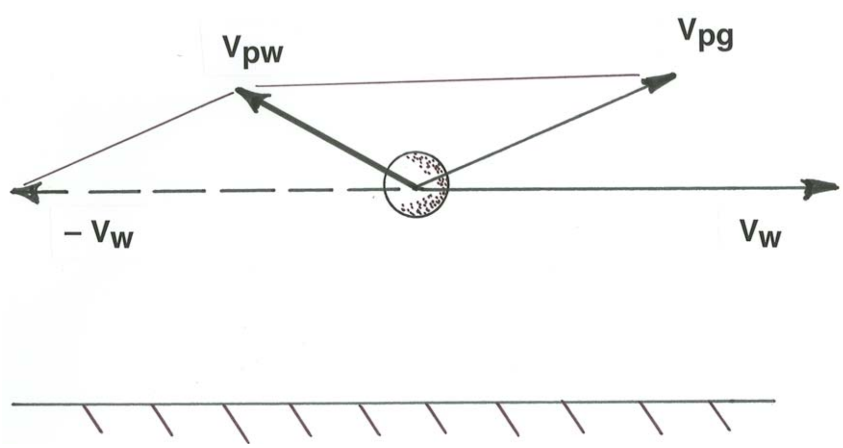

Figure 11-8. The speed of a saltating particle relative to the surrounding air. V_w = the velocity of the wind; V_{pg} = the velocity of the particle relative to the ground; V_{pw} = the velocity of the particle relative to the wind.

23 Bagnold (1941) supposed that the importance of the effect of particle speed relative to the air can be characterized by the ratio of fluid drag force to particle weight, a quantity he termed the ***susceptibility*** (although that useful term has not subsequently propagated itself through the literature on saltation). Figure 11-9, from Bagnold (1941), shows the susceptibility of several sand sizes as a function of wind speed. You can see from Figure 11-9 that for relative speeds of several meters per second the susceptibility of sizes between 0.3 mm and 1.0 mm—which largely span the range of sizes of saltating particle—lies between about one and ten: the air drag is greater than the particle weight, but not far greater. The implication is that the air drag does not much affect the details of the saltation trajectory but is important in determining the overall course of the trajectory. If the fluid drag were much less, the saltation length would be reduced. A further implication then seems to be that for saltation trajectories on Mars, where the density ratio ρ_s/ρ is even greater than on Earth, saltation height should be greater, relative to saltation length, than on Earth.

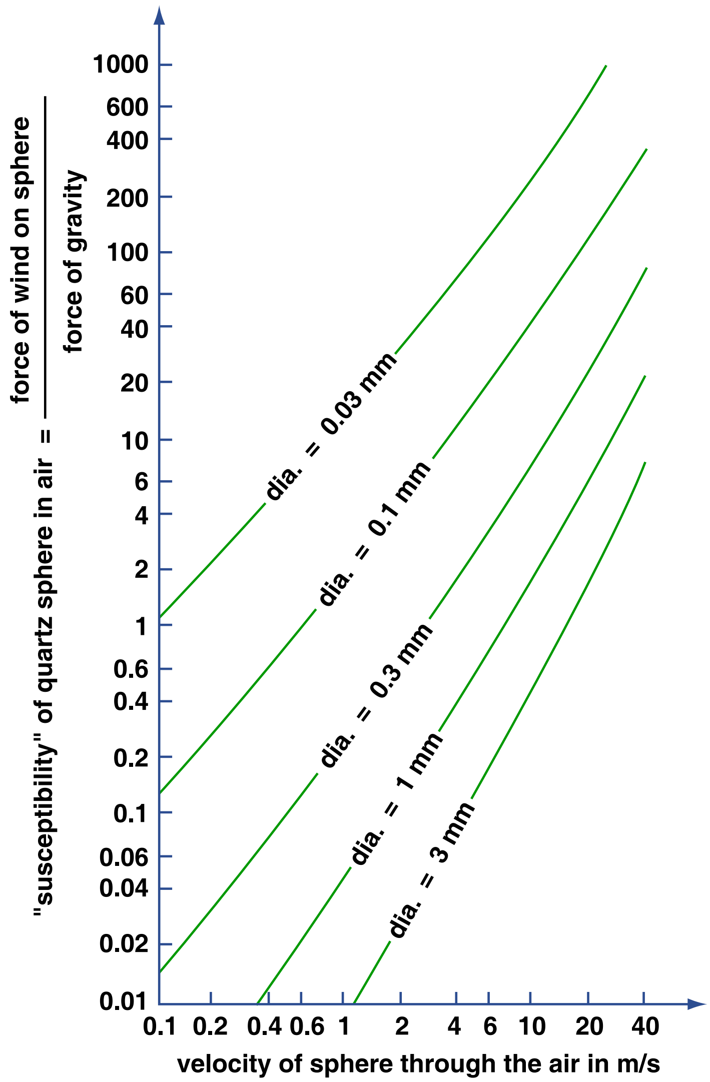

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Figure 11-9. The susceptibility of two particle sizes as a function of wind speed. (From Bagnold, 1941.)

24 The average impact speeds of about 160 cm/s measured by White and Schulz (Fig. 11-7) are much less than the wind speed at heights traversed by the particles near the tops of their saltation trajectories. Given that wind speeds are greater than that down to heights of only a few centimeters, those values of impact speed tell us that the wind has not nearly "finished the job" of accelerating the particle to the prevailing wind speed before the particle descends to splash down again onto the bed.

25 The foregoing material is only the briefest qualitative introduction to saltation trajectories. Several authors, beginning with Bagnold, have developed methods for computing saltation trajectories; see, for example, Owen (1964) and White and Schulz (1977). As Bagnold notes, it is essentially the same problem as the practical computation of the trajectories of cannonballs and artillery shells. The basic computational problem is that neither the velocity nor the fluid drag on the particle can be assumed independently: the two evolve together.

Saltation Heights, and the Magnus (Robins) Effect

26 It seems to be a common belief that the near-surface zone of saltation (the saltation cloud) has a well-defined upper limit. This might in part be because of the statement in Bagnold's influential 1941 book that the saltation cloud has "a clearly marked upper surface" (p. 10). Also, Owen, in his classic 1964 paper, illustrates a series of saltation trajectories all with the same shape, height, and length (his Figure 1), which a casual reader might assume was intended to represent real saltation—but Owen in fact took care to point out that the figure was meant only to illustrate the simplifying assumptions he made in his study, and that the saltation "in reality must be endowed with a certain randomness" (Owen, 1964, p. 226).

27 It is clear, from later observational studies, that for a given sand and wind there is a considerable variation in the height to which saltating particles rise. This shown perhaps most clearly by results of measurements of sand transport rate as a function of height above the bed. Using beds of moderately well sorted sand, both Zingg (1953) and Williams (1964) found that the sediment transport rate, per unit width across the wind and for unit height above the bed, varied as a negative exponential function of the height above the bed. Several later studies have shown similar results. (For more on sand transport rates in saltation, see the later section.)

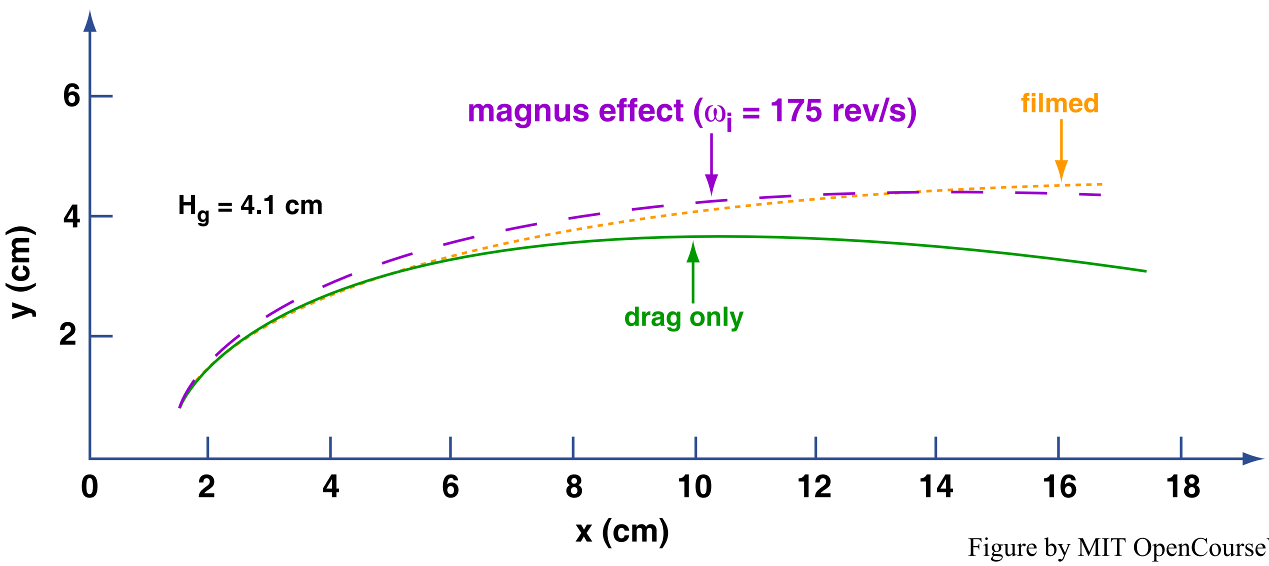

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Figure 11-10. Trajectories of a saltating glass sphere calculated for the case of drag only (non-rotating sphere; dashed curve) and drag plus lift (a sphere with a rotation rate of 275 per second; semi-dashed curve) compared with the observed trajectory (solid curve). (From White and Schulz, 1977.)

28 If you go back to what you learned in Physics I, you can easily compute the theoretical height to which a saltating particle would rise in the contrary-to-fact case of no air drag on the particle. The particle has some initial kinetic energy, $mv^2/2$, where m is the mass of the particle and v is the initial speed of the particle. As the particle rises, against the pull of gravity, its kinetic energy is converted to potential energy of height, *mgh*, where *g* is the acceleration due to gravity and *h* is height above the bed. To find the maximum height of rise, at the top of the parabolic trajectory, set the kinetic energy equal to the potential energy and solve for *h*: $h = v^2/2g$.

29 The value of the no-air-drag result is that it serves as a standard for comparison of actual saltation trajectories. In light of what was said in the earlier sections on saltation trajectories, we might conclude that real trajectories should always have a lesser maximum height of rise, owing to air drag. We would, however, be mistaken: experiments (e.g., by White and Schulz, 1977) slow clearly that saltation heights are even greater than the no-air-drag value (Figure 11-10). The reason seems to lie in the spin of the saltating particles.

30 As observed early on by Chepil (1945), particles in saltation have spectacularly high spin rates of hundreds of revolutions per second. The spinning must somehow be imparted to the particles at, and/or soon after, takeoff into the wind stream. Spinning generates a lift force that acts while the particle is in flight. This effect of spinning is generally called the Magnus effect for cylinders and the Robins effect for spheres (Figure 11-11). Rotation of the particle changes the streamlines so that they are no longer symmetrical about the particle: streamlines are closer together above the particle, implying that velocities are greater there than they are below the particle (Figure 11-11). From the Bernoulli equation (Chapter 3) it follows that the pressure is less above the particle than below, and the particle experiences a lift force. The variation in lift coefficient with rate of spinning is known, so the lift force can be calculated. White and Schulz (1977) could account for the observed saltation trajectories only by taking this effect into account. For most observed trajectories the rate of spinning could not be observed directly, but a good fit of observed trajectories to theoretical calculations could be made by assuming a rate of spin of several hundred revolutions per second. This is known from photographic studies to be about the right value for the spin.

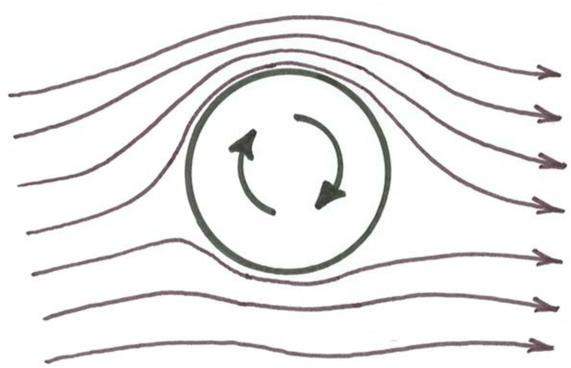


Figure 11-11. Vertical streamwise cross section through a spinning sphere immersed in a flowing fluid, to illustrate the Robins effect. See text for explanation.

Threshold of Motion for Eolian Sand Transport, and the Question of the Forces That Cause Saltation

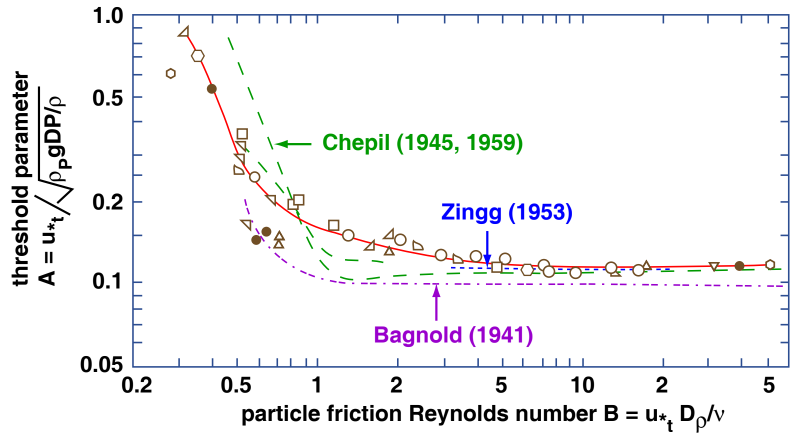
31 Clearly, no particles at rest on a broad horizontal surface of sand are set in motion until the wind reaches a certain strength. At wind speeds below the threshold for movement, the forces on the sand particles are the same as was discussed in Chapter 9 for water flows, because the fluid dynamics of the wind very near the ground is the same for air as for water. As in water flows, the nature of the fluid forces on the bed-surface particles—pressure forces and viscous forces, which can be resolved into a drag component, parallel to the bed, and a lift component normal to the bed—are a function of the particle Reynolds number. In fact, much of what is known about lift and drag forces as a function of particle Reynolds number has been learned from experiments in wind tunnels, beginning with Einstein and El-Samni (1949) and Chepil (1958, 1961).

32 The difficulties in defining the onset (or even the existence) of a definite threshold flow strength as discussed in Chapter 9 for sediment under water flows exist for sediment under air flows as well, although with certain important differences. As you saw in Chapter 9, in water flows the sediment transport rate in the range of flow strengths for which the threshold might be located is wide, and the mode of movement (bed load) is the same over that range. In contrast, in air flows a different mode of sediment movement—saltation—sets in soon after movement begins, and transport rates increase far more rapidly once sediment movement begins than in water flows.

33 As the wind speed increases, particles are set in motion by the fluid forces. Beginning with Bagnold, this has been called the ***fluid threshold*** or the ***aerodynamic threshold***. Soon after particle motion starts—in just a few seconds—saltation sets in, in a kind of cascade whereby the concentration of saltating particles increases rapidly to its equilibrium state. Then, if the wind speed decreases, the saltation eventually ceases. The condition of cessation of saltation is called the ***impact threshold***. One of the first-order facts about saltation is that the fluid threshold is at a wind speed less than the fluid threshold, as first remarked by Bagnold (1941) and confirmed observationally many times since. There is thus a strong hysteresis effect in saltation.

34 There has been a long-standing controversy about whether bed particles hop and roll for a brief time before cascading into fully developed saltation, as first proposed by Bagnold (1941), or whether they vibrate in place, in response to the rapidly fluctuating fluid forces they feel, before finally being launched into movement above the bed surface, as reported by soil scientists studying entrainment of soil particles by the wind. The consensus seems to be that, in the case of sand particles, the sand particles undergo some brief movement as bed load for a brief time before saltation develops.

35 Observations of movement threshold under air flows have been made since the early days of the modern era of research on sand movement by the wind. Following on early studies by Bagnold (1941), Chepil (1945, 1959), and Zingg (1952, 1953), Iversen et al. (1976a) made extensive observations of eolian thresholds by use of sediments of varying size, density; their results (Figure 11-12) show a nearly constant value of threshold Shields parameter for boundary Reynolds numbers down to about five, and then increasing threshold Shields parameter with further decrease in boundary Reynolds number. As mentioned in Chapter 9, the Shields parameter for threshold under air is somewhat greater than for under water, for the same values of boundary Reynolds number.



material	density, gm/cm^3	diameter, µm
⬡ instant tea	0.21	719
◺ silica gel	0.89	17; 169
□ nut shell	1.1	40 to 359
✧ clover seed	1.3	1290
▿ sugar	1.59	393
◸ glass	2.42	31 to 48
○ glass	2.5	38 to 586
◺ sand	2.65	526
▷ aluminum	2.7	36 to 204
△ glass	3.99	55 to 519
⌂ copper oxide	6.0	10
▽ bronze	7.8	616
● copper	8.94	12; 37
○ lead	11.35	8; 720

Figure by MIT OpenCourseWare.

Figure 11-12. Plot of threshold Shields parameter against boundary Reynolds number for observations of threshold conditions for a number of sediments under air. From Iversen et al. (1976a); their threshold parameter A is the same as the Shields parameter except that ρ_s is used in the denominator instead of $(\rho_s - \rho)$ in the variable $(\rho_s - \rho)g$, called γ' in these notes.

36 Nickling (1988) devised an experiment in which particles newly set into motion at near-threshold conditions were observed by means of a horizontal laser beam directed horizontally across the flow one millimeter above an originally intact planar sand bed. Sediments with a range of size and sorting were used. Nickling's results showed (Figure 11-13) that for the relatively poorly sorted sediments there is a range of flow strengths (as measured by the shear velocity) for which small number of particles are moved before flow strengths become great enough for saltation to begin, whereupon the number of particles in motion increases sharply. For the relatively well-sorted sediments, however, that range of flow strengths effectively vanishes: saltation begins immediately upon attainment of motion brought about by the fluid forces.

37 Most studies of threshold of eolian transport have been made in wind tunnels, in which nearly steady winds can be arranged. In the field, observations of threshold are far more difficult, in large part because winds across natural sand surface are much gustier, owing to the much larger scale of eddies in the lower atmosphere. In small wind tunnels, fluctuations in bed shear stress with time at a point are short relative to the time scales of saltation of individual particles, whereas in the field they are typically far longer. Such considerations point toward a later section of this chapter, on saltation in unsteady winds.

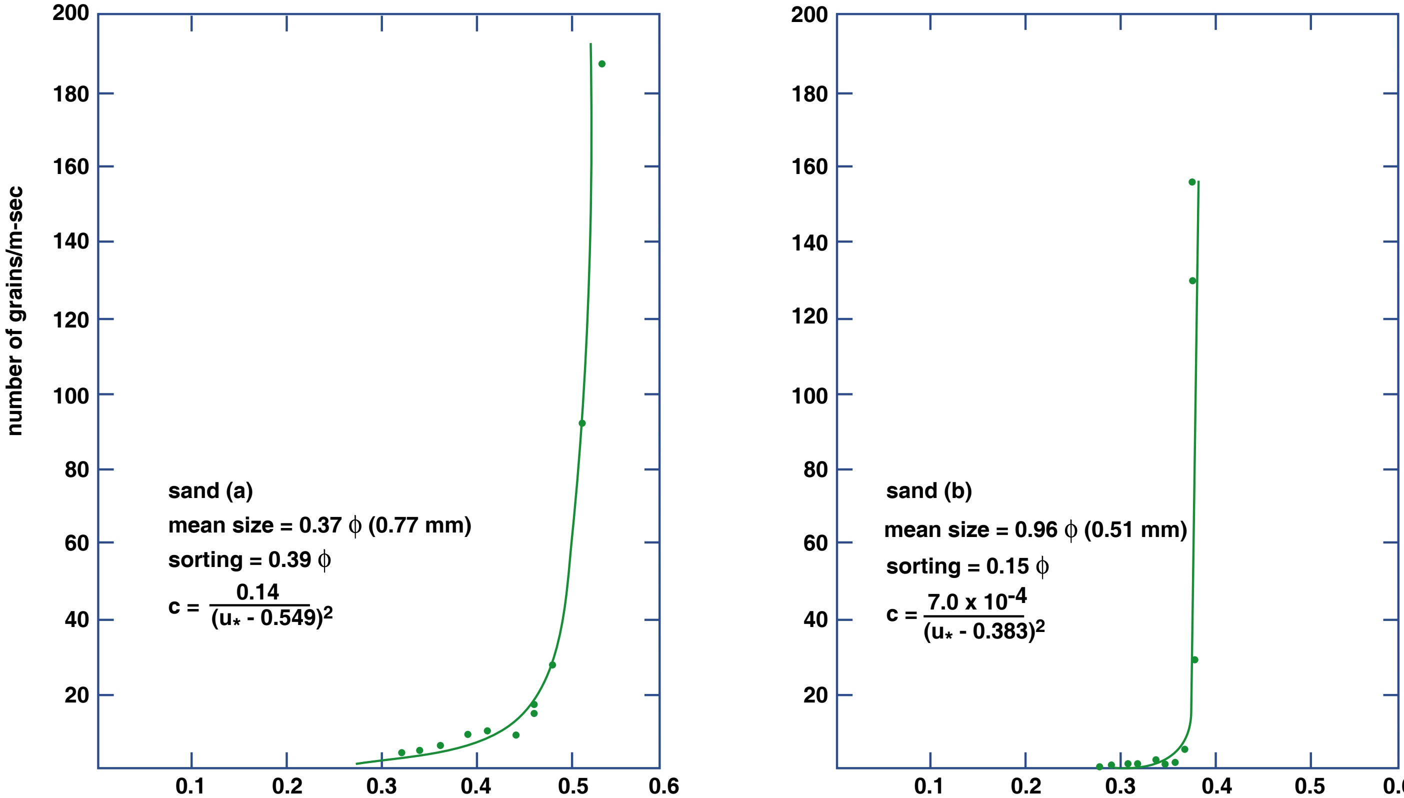


Figure by MIT OpenCourseWare.

Figure 11-13. Plots of numbers of particles in motion, per unit time and per unit width normal to the wind, versus shear velocity, for two sediments: **A)** a relatively poorly sorted sand, with mean size 0.77 mm and with a sorting value of 0.39 phi units, and **B)** a relatively well sorted sand, with mean size 0.51 mm and sorting of 0.15 phi units. (From Nickling, 1988.)

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38 The forces that cause a particle to be launched into a saltation jump in the wind have been controversial. There are two candidates: aerodynamic forces of lift and drag, and impacts by other saltating particles as they splash down onto the bed. (Of course, the two could, and probably do, act in concert; the question is which is the more important.) The moderate to large takeoff angles of saltating particles do not in themselves indicate the relative importance of the two kinds of forces: it might be supposed that strong aerodynamic lift forces should be responsible for steep takeoff angles, but it is clear also that similarly steep angles can be the result of rebounds upon splashdown. The controversy dates back to the early days of the modern era of study of eolian sand movement: Chepil (1945, 1961) considered aerodynamic forces to be dominant, whereas Bagnold (1941) believed saltation impacts to be principally responsible for saltation takeoff.

39 It seems clear that the presence of the saltating particles extracts momentum from the wind within the saltation layer, as discussed in a later section, so the fluid shear stress on the bed must be much less than would be the case with the same sand bed and with the same overlying wind but with the bed particle immovable. Owen (1964) went so far as to hypothesize that the shear stress exerted by the wind on the sand bed is just sufficient to maintain the surface particles in a mobile state. The implication of that hypothesis is that the aerodynamic forces of lift and drag should be much less important in maintaining saltation than rebound of particles, as well as mobilization of other particles, at the point of collision.

40 Theoretical models of continuous saltation, beginning with Tsuchiya (1969, 1970) and Reizes (1978), demonstrate that saltation can continue once started, without the necessity of any fluid lift or drag forces acting on particles resting on the bed, but they do not lead to any predictions about particle trajectories that can distinguish this hypothesis conclusively from the fluid-force hypothesis.

The Effect of Saltation on the Velocity Profile of the Wind

41 You have seen that the air exerts a drag force on the saltating particles as they rise from the bed. Conversely, the equal and opposite force exerted by the particles on the wind tends to slow the wind. Given the commonly substantial concentration of particles in the saltation layer, you should expect the structure of the wind in the saltation layer to be different from that in the absence of saltation. At first thought, you might assume that there is a kind of symmetry at work here: perhaps the particles tend accordingly to speed up the wind as they descend from the tops of their trajectories down into region of lower wind speed. If, however, our earlier deduction to the effect that the particles have not yet been fully accelerated by the wind even when they reach the ends of their trajectories is true, then the saltating particles must be responsible for a net decrease in wind velocity. You will see below that this is indeed the case.

42 As with so many aspects of eolian saltation, Bagnold was the first to give systematic attention to the effect of saltation on wind velocity. Bagnold (1941), and many later researchers, have measured wind-velocity profiles in the presence of saltation. Figure 11-14, taken directly from Bagnold’s book, shows actual measurements.

43 Recall from Chapter 4 that the air speed over a fixed rough bed varies logarithmically with height above the bed, according to the law of the wall for rough boundaries (Equation 4.33, reproduced here as Equation 11.1, for your convenience):

$$\frac{\overline{u}}{u_*} = A \ln \frac{y}{y_0} \tag{11.1}$$

where y_o, the roughness length, is nothing more than a convenience variable to put the law of the wall as expressed in the form of Equation 4.41 into a neater form. The roughness length y_o has the property that when the profile expressed by Equation 11.1 is extrapolated downward, its intercept with the $\overline{u}/u_*$ axis (nominally, zero wind velocity) is at a value of y, the height above the bed, of $D/30$ for close-packed granular roughness—but in reality Equation 11.1 ceases to hold at heights above the bed not much greater than the particle diameter, as discussed in Chapter 4.

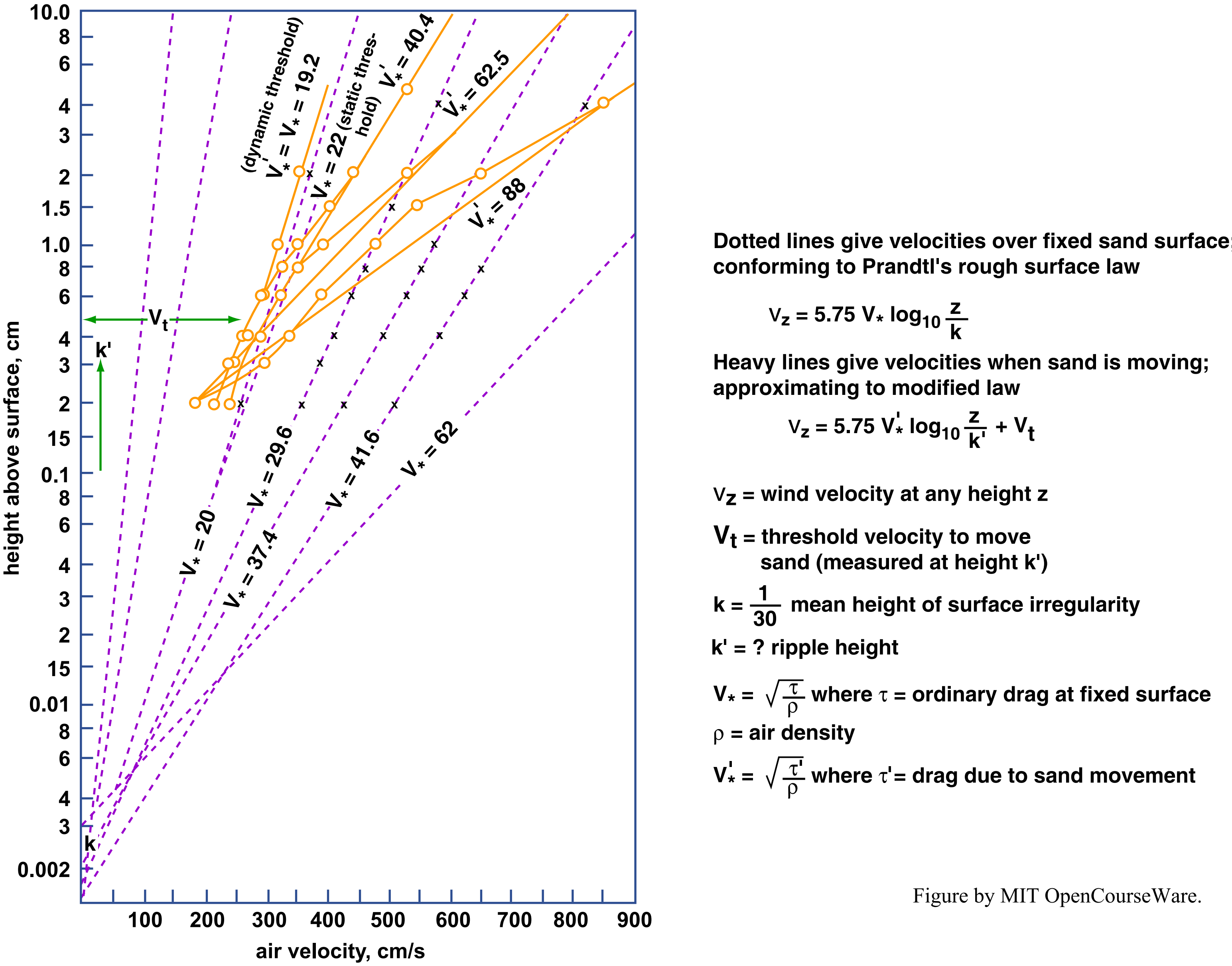

Figure 11-14. Profiles of wind velocity in saltation. (From Bagnold, 1941.)

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44 In a dimensional plot of wind speed u against height y above the sand bed, if u_* is changed, the slope of the velocity profile varies, but the intercept y_o does not, according to Equation 11.1 (see Figure 11-15, an idealization of Figure 11-14). What Figure 11-15 shows is that when a saltation layer is present the profile of wind speed in the region above the saltation layer is still logarithmic, but with a significant modification: profiles for different shear velocities no longer converge on the point (0, y_o) located on the y axis (where $u = 0$) but, approximately, on a point (u_o', y_o'), where u_o' is not equal to zero and y_o' is much larger than y_o. The effect on the velocity profile above the saltation layer is the same as if the roughness of the bed had been increased—as if, in Equation 4-41 or 4-42 the size of the roughness elements, *D*, had been increased. The saltation layer thus adds resistance to the wind, as we deduced at the beginning of this section.

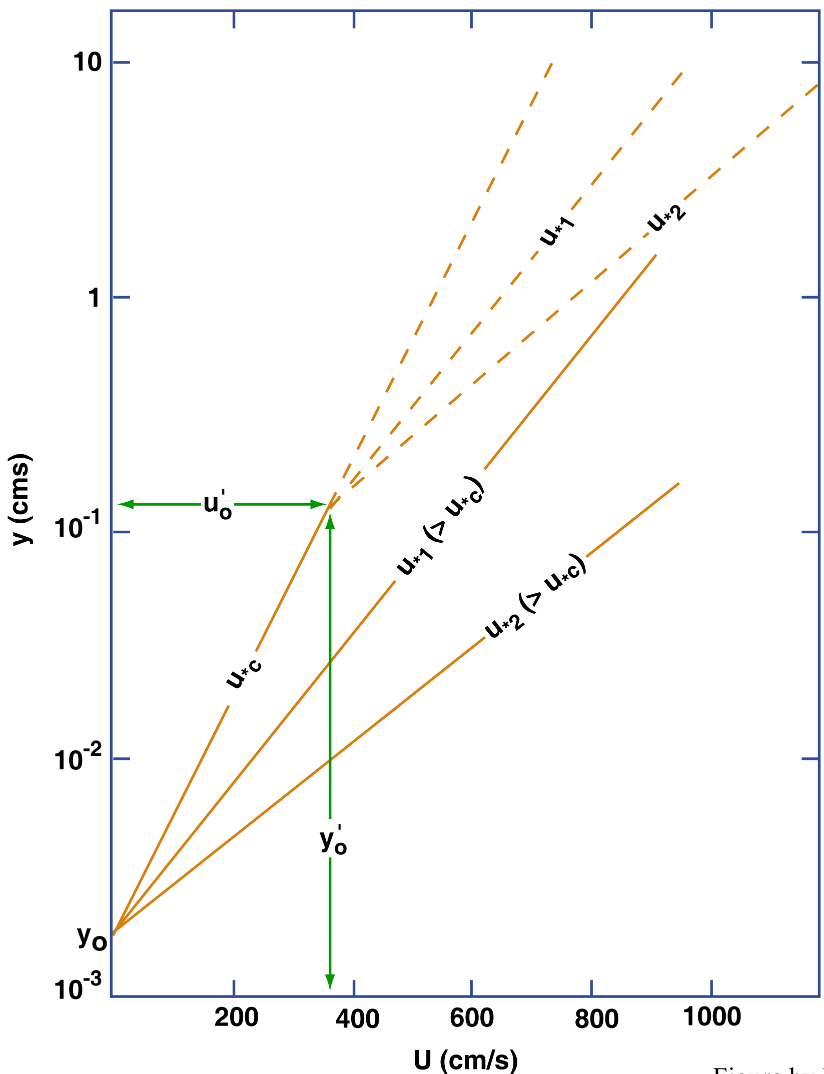

Figure by MIT OpenCourseWare.

Figure 11-15. Idealized plot of vertical distribution of wind velocities in saltation. Solid lines show profiles observed where the particles are fixed to the bed; dashed lines show profiles observed where particles are saltating over a planar bed of loose sand. (Figure by G.V. Middleton.)

45 The question then arises: how low does the wind speed become, deep in the saltation layer, just above the tops of the bed particles? Owen (1964) offered the following hypothesis, noted in an earlier section: the shear stress exerted by the wind on the sand bed is just sufficient to maintain the surface particles in a mobile state—which is much lower than would be the case with the same sand bed and with the same overlying wind but with the bed particle immovable.

Jump-Distance Distribution

46 The downwind distance traversed by saltating particles ranges from very short, perhaps of the order of a few millimeters (the minimum saltation distance is partly a matter of semantics, hinging upon one's view of the transition from particle movement in surface creep to particle movement in saltation) to very long, as much as several meters in strong winds under which the saltation layer extends upward by more than a meter. When the flights of a large number of saltating particles in a uniform wind are considered, there is some well defined probability distribution of jump distances.

47 Measuring the jump-distance frequency distribution is not straightforward. Direct measurement of jump distances, by means of tracking trajectories photographically, is likely to be biased toward the longer trajectories, owing to the greater particle concentrations at lower levels, which tend to obscure the individual trajectories, and the slower particle speeds, which makes measurements of speeds from photographic images more difficult. The few attempts at measurement have exploited the indirect method of measuring the catch of particles in long bed-level traps of various designs (Kawamura, 1951; Horikawa and Shen, 1960; Belly, 1964).

48 It is not difficult to show that the jump-distance distribution is related to the distribution of catch in a horizontal sand trap by

$$f(\eta) = \frac{1}{G_O} \frac{dG}{dx} \tag{11.2}$$

(Kawamura, 1951), where $f(\eta)$ is the frequency distribution of saltation jump distances η, G is the saltation catch (mass per unit area and unit time) in a horizontal trap with leading edge at $x = 0$ and extending downwind in the positive x direction, and G_O is the total mass launched into saltation from a unit area in unit time.

49 The few measurements of jump-distance distribution show three significant features:

- The frequency of jump distances increases monotonically with decreasing jump distance, apparently right up to the transition to surface creep; in other words, the maximum of the curve is at very small, or even zero, jump distance.
- The mean jump distance is significantly greater than the spacing of the wind ripples over which the saltation takes place.
- There is no well-defined maximum jump distance, as is to be expected, given the gradually decreasing concentration of saltating particles with height, but the frequency of jump distances several meters long is not negligible.

50 Mass-balance considerations in the context of jump-distance distributions are enlightening. Think about saltation that is uniform, in the sense that the picture of saltation is exactly the same at every point along the wind direction. Uniform saltation is very closely approximated where a sand-moving wind blows steadily over a level sand surface of great extent. In uniform saltation, the mass of particles launched from a small unit area of the bed must be equal to the mass of particles arriving onto that area—and, more specifically, the jump-distance distributions of both the incoming and outgoing particles must be identical, or the saltation would not be uniform. This is a demanding requirement, because each incoming particle gives rise to zero, one, or more outgoing particle motions with jump distances not likely to be identical to its own. Nature somehow manages to adjust the jump-distance distribution of outgoing particles to be the same as that of the incoming particles. There must be a self-regulating mechanism at work: if not enough downwind transport is engendered from the unit area by the incoming particles, the intensity of saltation falls off downwind until what leaves matches what arrives, and if the incoming particles cause an even greater transport rate out of the area, the saltation transport rate increases until the rate becomes uniform. This transformation of incoming saltation to outgoing saltation can be described in terms of what Werner (1990) calls the ***splash function***. The following makes these matters more concrete.

51 In eolian saltation the mass of moving particles that make contact with a small reference area on the bed includes particles launched into saltation from a range of distances upwind, from only a fraction of a particle diameter, in the case of the surface creep, to as much as a few meters, in the case of the highest-flying particles in saltation. With x as incoming jump length, let the function $g_{in}(x)$ represent the jump-distance distribution of this incoming mass of particles, expressed as mass per unit bed area per unit time. Similarly, with y as outgoing jump length the function $g_{out}(y)$ represents the corresponding jump-distance distribution of the outgoing mass of particles. In uniform saltation, incoming and outgoing mass must be the same for any given jump length, so g_{out} and g_{in} are identical distributions. Mathematically this can be expressed as

$$\int_0^\infty g(x)F(x,y)dx = g(y) \qquad (11.3)$$

Where $F(x, y)$ is the splash function of Werner (1990). Equation 11.3 is an integral equation—one that contains an integral. A function with the form of F in Equation 11.3 is said to be the *kernel* of the equation. In this case, a mathematician might call F a *self-replicating kernel function*, because it has the remarkable property of transforming the other factor in the integral on the left, $g(x)$, into an identical function, $g(y)$, on the right.

52 The requirement, mentioned above, that in uniform saltation the jump-distance distributions become adjusted so that the incoming and outgoing jump-distance distributions, g_{in} and g_{out}, are identical and a function of the wind strength can be expressed in the context of Equation 11.3 as follows. For each value of incoming jump distance x, the splash function acts on the incoming mass of saltating particles to give a contribution to the mass distribution of outgoing jump distance, and the sum of all of these contributions is the outgoing mass distribution of jump distances.

53 What can we say, qualitatively, about the nature of the splash function F?

• The momentum of incoming particles, and therefore their ability to set particles in motion at any given outgoing jump distance, increases with increasing incoming jump distance, so F should be a monotonically increasing function of x at constant y for all y, including $y = 0$.

• The mass of particles set in motion by arrival of particles with a given jump distance *x* should be greater for smaller outgoing jump distances than for larger, so *F* should be a monotonically decreasing function of *y* for constant *x*.

• Incoming particles with very small jump distances can give rise to only a narrow range of jump distances, and therefore relatively small momentum, not much larger than their own, whereas incoming particles with very large jump distances, and therefore relatively large momentum, can give rise to a wide range of outgoing jump distances from very small to even larger than their own, so the overall rate of decrease of *F* with increasing *y* at constant *x* should be sharpest for very small *x* and become gentler with increasing *x*.

• *F* must approach zero as *x* approaches zero, because the mass of particles mobilized must go to zero as the incoming jump distance, and therefore the momentum of the incoming particles, goes to zero.

Figure 11-16 shows, qualitatively, what the splash function *F* might actually look like.

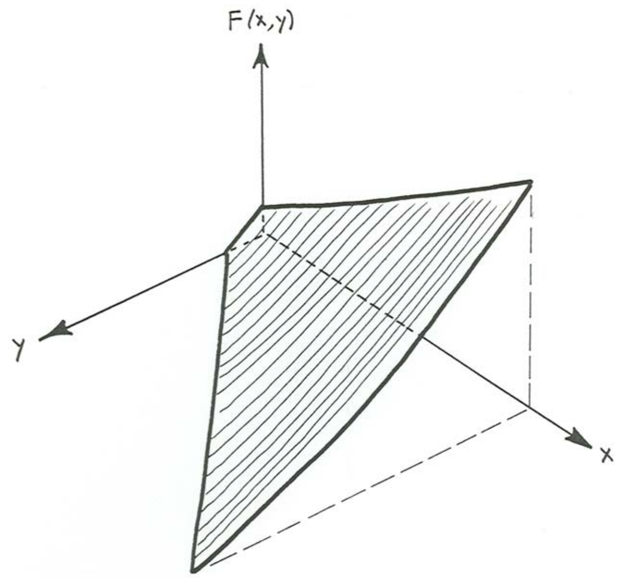

Figure 11-16. A qualitative representation of the splash function.

Saltation Transport Rates

54 It was mentioned in the section on saltation heights that the concentration of saltating particles tails off gradually upward. This is known from sampling to measure the transport rate of saltating particles. Such measurement is simple in principle but somewhat troublesome in actual practice. The common procedure is to install, on a vertical shaft or frame in the sand, a series of particle-catching devices, which are uncovered for a fixed time and then the mass of particles caught in each is measured. A curve of catch versus height is plotted, and the total transport rate is the integral of that curve from the bed to a level above the highest saltation heights. Once the transport at any given level is known, the concentration of the saltating particles at that level can be found if the time-average wind speed is measured at the same level at the same time, inasmuch as the transport rate must equal the concentration times the speed of passage of the parcel of air that contains the particles. Systematic measurements of transport rate date from the time of Bagnold (1941); see also the early and widely cited work of Williams (1964).

55 One practical problem is that any such catching devices, no matter how well designed, inevitably disturb the passing wind to some extent, and even aside from that, measurements near the sand bed, where the mass flux of particle is greatest, is difficult to arrange. In recent years, high-resolution measurements using non-intrusive optical sensors have been developed (e.g., Butterfield, 1999), thus mitigating some of the problems. Another problem is that it is not easy to measure the transport rate of sediment moved as surface creep.

56 A more general problem, however, has to do with what is actually being measured. The wind is gusty on natural sand surfaces. Even on a broad, horizontal sand-coved plain, the large-scale eddy structure in the lowermost atmosphere means that the saltation catch varies with time on periods of seconds to many minutes. The problem is exacerbated on the upwind flanks of sand dunes, owing to the strong wake produced by an upwind dune. A catch averaged over many minutes may be very different from an "instantaneous" measurement, taken over a number of seconds. This problem could be circumvented in a wind tunnel, but the tunnel would have to be large enough that the saltation profile is fully developed vertically even in very strong winds. Few wind tunnels are of such a size.

Saltation in Unsteady Winds

57 In recent years, increasing attention has been given to how the saltation cloud adjusts to changing wind conditions, given that winds in the outdoors are characteristically highly variable, on time scales of minutes to hours. The problem can be posed as follows. A surface of loose sand lies susceptible to saltation. A strong gust of wind initiates saltation. How do the conditions of saltation respond? The saltating cloud responds rapidly. The response of the saltation to the changing wind speed has been studied in wind tunnels and in the field (e.g., Butterfield, 1991, 1998) (Figure 11-17), and several numerical models have been developed to account for the observations (e.g., Anderson and Haff, 1991; McEwan and Willetts, 1991; Spies and McEwan, 2000; Spies et al., 2000). In Figure 11-18, from numerical simulations by Spies and McEwan (2000), you can see how the transport rate develops in time and space: at a given time after onset of the wind, the transport rate reaches a maximum near the upstream edge of the sand bed, and the maximum in transport rate moves downstream with time.

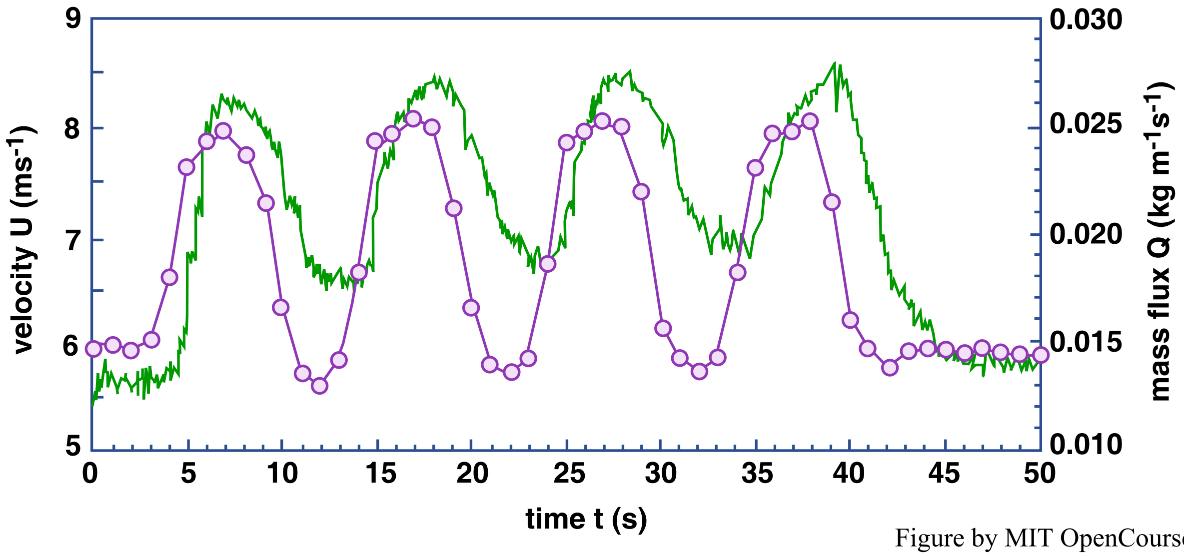

Figure by MIT OpenCourseWare.

Figure 11-17. Synchronized measurements of transport rate (grams per centimeter width per second) and shear velocity (meters per second) versus time for a sinusoidally varying wind velocity. The open squares are for wind velocity, and the heavy curve is for transport rate. (From Butterfield, 1998.)

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58 One significant aspect of the response of saltation to a sudden increase in wind speed, from below threshold to well above, is that the saltation transport rate first increases but then decreases somewhat before settling into equilibrium with the wind. The reason is easy to understand: it takes some time for the effect of theft of fluid momentum on the part of the saltating particles to develop—so there is a brief period of time during which the aerodynamic forces on bed particles has not decreased significantly, while the impact forces exerted by saltating particles on the bed have already become significant. As the wind adjusts in such a way as to exert a smaller bed shear stress (see the earlier section), the saltation cloud settles down to a state of somewhat less vigorous saltation. There is thus a transient maximum in saltation transport at the outset of a transport event. Spies et al. (2000) have done numerical simulations of this effect (Figure 11-19).

The Transition from Saltation to Suspension

59 You learned way back in Chapter 3 that the characteristic fluctuations in velocity in a turbulent flow are a certain small percentage of the mean velocity. Because of that, the characteristic vertical fluctuating velocity in near-surface winds should increase with wind speed. If those vertical velocities are sufficiently large, even saltating sand particles are affected in the trajectories by the fluctuations. Likewise, in a wind with a given speed, the effect of the velocity fluctuations on particle trajectories increases with decreasing particle size.

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Please see: Spies, P. J., and I. K. McEwan. "Equilibration of Saltation."
Earth Surface Processes and Landforms 25 (2000): 437-453.

Figure 11-18. Results of numerical simulations to show how saltation transport rate develops in time and space after initiation of a steady wind.

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60 The transition from classic saltation trajectories to trajectories that are non-negligibly affected by turbulence is an area of study in eolian sedimentation that has less attention than the study of saltation. A distinction needs to be made here between (1) fine particles (usually referred to in the eolian literature as *dust*), which are raised either directly by the wind or indirectly by the impact of saltating larger particles on exposed surfaces of sediment or bedrock, and which go directly into true suspension even at wind speeds for which vertical fluctuating turbulent velocities are much lower than the settling velocities of the coarser saltating particles, and (2) sand particles moved by winds so strong that the vertical fluctuating velocities become comparable to the settling velocities of the particles, causing particle trajectories to show at least some influence of turbulence. Nishimura and Hunt (2000) found, in a wind-tunnel study of particle trajectories, that the transition from saltation to suspension begins to be noticeable when the shear velocity is still as low as one-tenth the particle settling velocity. As wind speeds increase beyond that, particle trajectories show greater and greater irregularity due to interaction with turbulent eddies (Figure 11-20).

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Please see: Spies, P. J., I. K. McEwan, and G. R. Butterfield. "One-dimensional Transitional Behaviour in Saltation." *Earth Surface Processes and Landforms* 25 (2000): 505-518.

Figure 11-19. Simulated transport rate as a function of time for saltation in a wind tunnel. The initial shear velocity was 0.37 m/s, and the shear velocity one steady-state saltation had developed was 0.55 m/s.

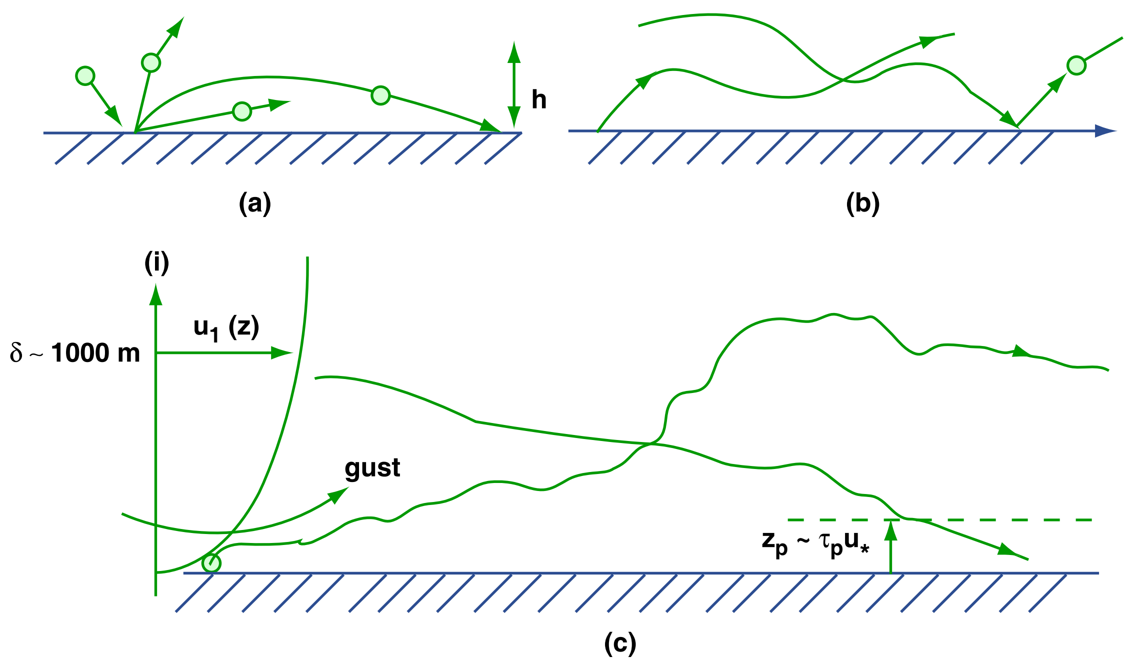

Figure by MIT OpenCourseWare.

Figure 11-20. Cartoon of the transition from saltation to suspension. **A)** Saltating particles are unaffected by fluid turbulence; **B)** saltating particles are slightly affected by fluid turbulence; **C)** particle trajectories are strongly affected by fluid turbulence. (From Nishimura and Hunt, 2000.)

Models of Eolian Saltation

61 After the early work of Reizes (1978), and concurrently with the development and elaboration of the concept of the splash function by Werner and co-workers, the focus of studies of eolian saltation began to shift toward modeling of eolian sediment transport as a unified phenomenon with saltation dynamics as the basis (e.g., Anderson and Hallet, 1986; Ungar and Haff, 1987; Anderson and Haff, 1988; Werner and Haff, 1988; Werner, 1990; Haff and Anderson, 1993). As time has gone on since the late 1980s, with the development of ever greater computing power, numerical models of eolian transport have become more and more able to simulate the physics of saltation and the consequences for eolian sediment flux.

62 Models at first aimed at simulating saltation transport in steady and fully developed winds, of the kind that can be produced without difficulty in a long wind tunnel (e.g., McEwan and Willets, 1991, 1993a, 1993b; Willetts, 1998). More recent models have moved on to simulation of unsteady winds—for example, a saltation event in which a sudden strong wind gust generates a cloud of saltating particles, which develops in time and with downwind distance, as described in an earlier section (e.g., Spies and McEwan, 2000; Spies et al, 2000).

Sand Movement on Mars and Venus

63 Look back at Figure 8-5, in Chapter 8, to remind yourself that the case of sand transport by wind on the Earth's surface is only one point in the wide range of density ratios for which solid particles are transported by fluid flows. The density ratio for sand movement on Mars (if we assume that the mineral particles available on the Martian surface are not greatly different in density from those on the surface of the Earth) lies even farther to the right along the ρ_s/ρ axis than the density ratio for eolian sediment transport on Earth. In contrast, the Venus case lies not much farther to the right than the case of transport of ultra-heavy minerals (gold being the obviously important example) by water flows on the Earth's surface! It seems fair to say that the great bulk of the research so far on transport of loose particulate sediment on Mars and Venus has come from the research group headed by R. Greeley, and especially on the part of J.D. Iversen and of B.R. White (Greeley et al., 1974; Greeley et al., 1976; Iversen et al., 1975; Iversen et al. 1976a; Iversen et al. 1976b; Iversen et al. 1976c; White, 1979; Iversen and White, 1982; White et al., 1987) as well as more recent contributions (e.g., Fenton and Bandfield, 2003; Bourke et al., 2004). Much of the data and conclusions from the work of Greeley's group is presented in the book by Greeley and Iversen (1985). The emphasis in these notes is on eolian sand movement on Mars, in light of the spectacular recent advances in our understanding, and the much enhanced interest, that have arisen from the Rover results.

64 A first-order and seemingly unassailable deduction we can make at the outset is that saltation should be the dominant mode of movement of sand-size particles on Mars—because the relative inertia of the particles is even greater than in eolian transport on Earth. In the case of Venus, for which the density ratio is greater than for sand in water on Earth by not much more than one order of magnitude, particle trajectories are much more likely to be affected by the turbulence in the wind than is the case for saltation on Mars.

65 Look back to the discussion of the effect of density ratio on thresholds, in Chapter 9, to see that in terms of the Shields diagram, in which the threshold for sediment motion is expressed in terms of the Shields parameter and the particle Reynolds number, the difference between dimensionless threshold for mineral particles in water and for mineral particles in air is not entirely clear (to me, at least). Given the great differences in atmospheric density between Earth and Mars, as well as the difference in gravity, you should expect that when expressed in dimensional terms the thresholds should be quite different. Figure 11-21 shows a comparison of motion thresholds expressed in terms of the shear velocity of the wind.

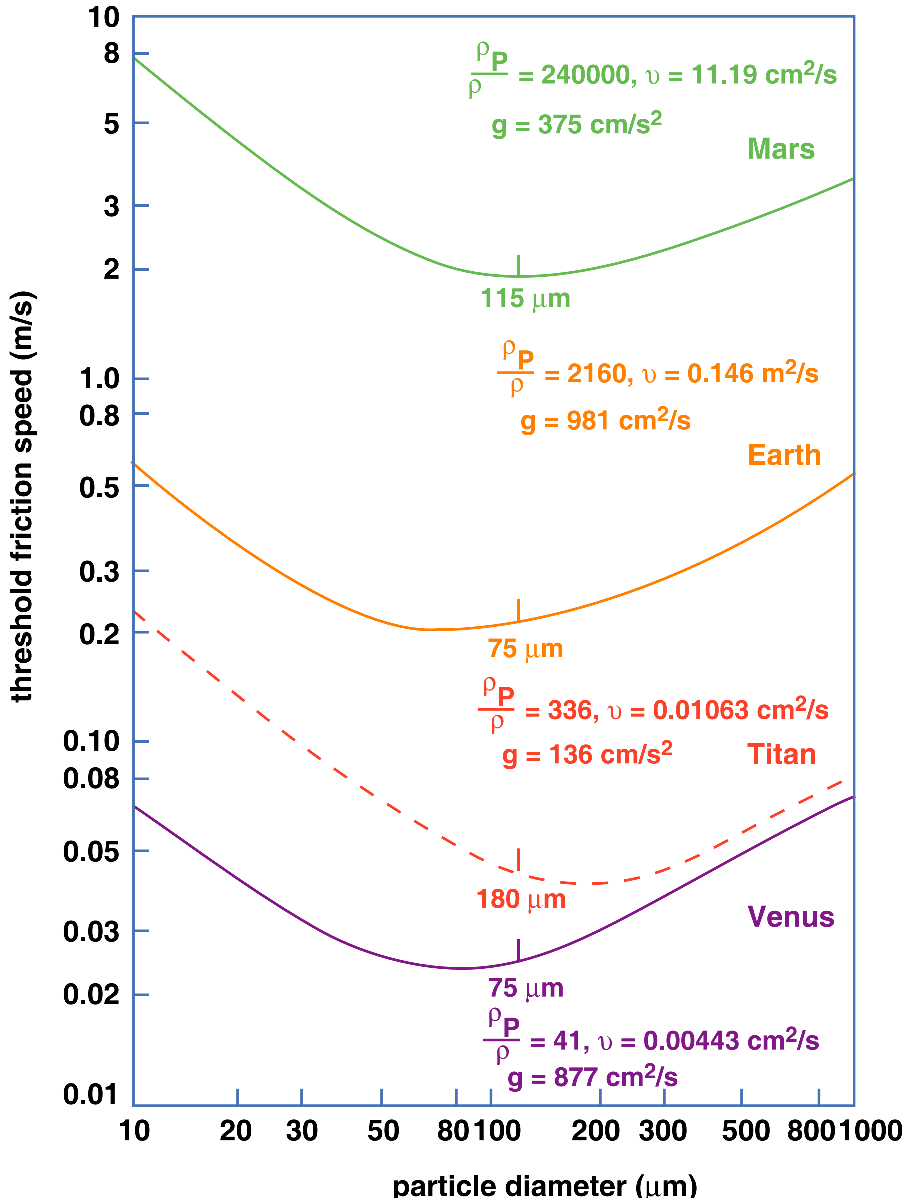

Figure by MIT OpenCourseWare.

Figure 11-21. Predicted threshold shear velocity versus particle diameter for Earth, Mars, and Venus. (From Greeley and Iversen, 1985.)

66 It seems clear that saltation jump heights and lengths must be much greater on Mars than on Earth, owing to the greater wind speeds and lesser gravity. Another significant deduction we can make is that because of the much greater wind speeds on Mars, together with the even greater relative inertia of the particles, the destructive effects of impacts of saltating mineral particles on rock surfaces should be even greater on Mars than on Earth.

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