Chapter One
Wind and Sand

Part I of this book (in which this is the first chapter) has a two-dimensional spatial frame. The principal consequence is that the wind is assumed to come from one direction only.

The chapter is about the lifting and carrying of sand by the wind. As to winds at this scale, only their velocity and small-scale patterns of turbulence are relevant. As to sand, the primary interest here is in its grain size (shape in a secondary concern). In the most widely accepted taxonomy for the size of sedimentary particles, grains of 'sand' have diameters of between 0.625 mm and 2 mm (Wentworth, 1922, RL). A global survey, by its own admission limited, but probably representative, found that most dune sands were in Wentworth's 'fine sand' category (0.10–0.40 mm) (Ahlbrandt, 1979). A quick scan of recent papers confirms this. Another assumption here is that sand is composed only of quartz. Dunes built of smaller and coarser particles, and of other minerals, are described in Chapter 6. Because changes in the shape and surface texture of windblown sand grains have long histories, they are issues for Chapter 4.

Wind versus Bed

The mechanical energy spent when two bodies (such as the wind and the surface) slide past each other is termed 'shear'. Shear on a surface over which a wind passes is denoted by τ_0.

Dunes: Dynamics, Morphology, History, First Edition. Andrew Warren.

The Law of the Wall

Because, until very recently, the shear of the wind on a surface could not be measured, it had to be approached through theory. The first model of the relationship grew out of the work of Ludwig Prandtl and Theodore von Kármán, working first at Göttingen, and came to be known as the 'Law of the Wall', where the word 'wall' was chosen with wind tunnels in mind. For dunes, 'bed' (as in the 'bed of a stream') is more appropriate than 'wall' and is adopted here. Despite some serious revision (shortly), the Law is still a good introduction to how the wind shears a surface.

The Law is built on two observations (or, as will be seen, simplifying assumptions):

- The velocity of the wind increases upward from the bed, because friction on the bed retards the wind, and this retardation is transferred, with weakening effect, to the wind at greater heights. Figure 1.1a shows the velocity of a wind at successive heights above the bed of a wind tunnel, both on arithmetic scales. In Figure 1.1b, the data and the velocity scale are the same as in Figure 1.1a, but the height scale is now logarithmic. The Law declares that on a semi-logarithmic plot, like this, the data fall on a straight line, as they do in this and many other observations in wind tunnels. The slope of the straight line in Figure 1.1b is a function of the strength of the wind that it represents: steep slopes represent gentle winds (low velocities even at some height); gentler slopes represent faster, more powerful winds (high velocities near the bed). This relationship comes into the argument again shortly.
- Figure 1.1b also shows that the straight lines depicting winds of different strength reach the same focal point on the vertical (height) axis. Those who built the Law took this to imply that there was a very thin layer of air, just above the bed, at the same height for all winds, where the air was stationary or moving very slowly. The focal point is higher on rougher beds, which is why it is termed the 'roughness height' or 'roughness length' (shorthand, 'z_0'). Because few measuring instruments, even today, can penetrate this layer, z_0 has to be estimated. A common estimate is ~1 mm over a smooth, stable sandy surface on Earth. A newer formulation of z_0 is discussed very shortly.

In order for the Law to apply to fluids of differing density, a parameter, shear velocity (or friction velocity), or u_*, was introduced:

$$u_* = \sqrt{\tau_0 / \rho_a} \, ,$$

where τ_0 is the shear force on the bed, and ρ_a is the density of the fluid; this is the 'Prandtl–Kármán equation'.

Because the dimensions of u_* are those of velocity (m s^{-1}), it is termed the 'shear velocity'. The equation is applicable to thin air on the Tibetan plateau, or on

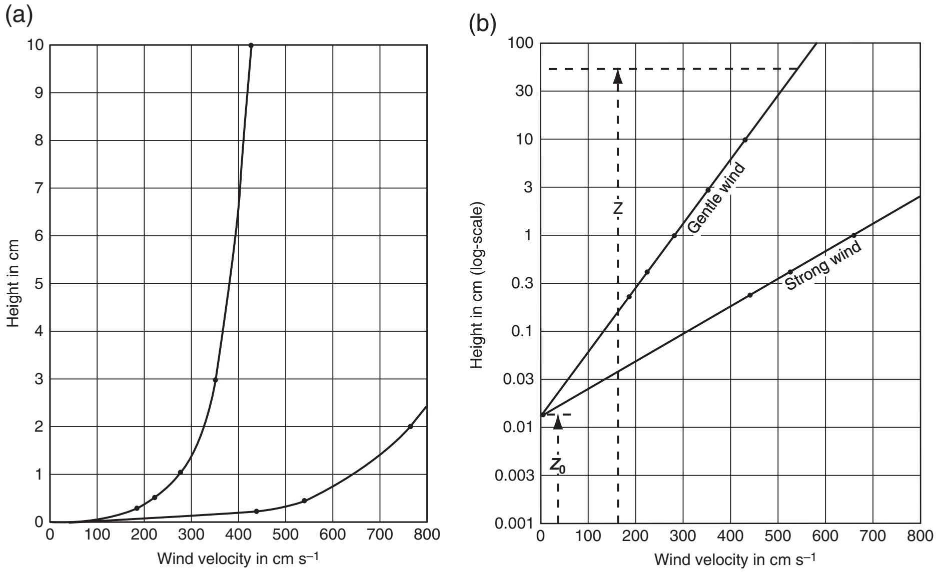

Figure 1.1 Velocity profile of the wind over a smooth, flat bed in a wind tunnel. In (a), both axes are arithmetic; in (b), the velocity (on the abscissa) is still arithmetic, but the height scale (on the ordinate) is now logarithmic. The data are Bagnold's (1941, pp. 48–49). The notations are explained in the text. Plots like this are known as 'Prandtl curves'. Redrawn after Bagnold (1941).

Mars; to dense atmospheres at very low temperatures, as near the Poles on Earth, or on Venus; or to any denser fluid, including water. The shorthand, 'u_*', is used in many places in this and later chapters.

The next step in the building of the Law was to relate u_* to measurements of the slope of the velocity/log height plot (Figure 1.1b). This step built on the observation (earlier) that the slopes of the lines are related to the strength of the wind. More precisely,

$$u_z / u_* = 1/\kappa \ln\left(z / z_0\right),$$

where u_z is wind velocity at height z, and κ is Kármán's constant.

There are various estimates of κ, from theory and experiment; most are ~0.40.

Improving the wind/bed model

For all its simplicity and clarity, the Law is now only the first step in understanding wind shear on a solid surface. Three of its essential components have been found to be oversimplifications.

The velocity profile

Many plots of the velocity of flow against height over a bed do not fit the logarithmic profile. This is shown by attempts to fit straight lines to measurements of the wind at different heights above a beach, which produced a wide range of values of z_0 (Bauer *et al.*, 1992). Similar problems have been found with measurements in a wind tunnel (Butterfield, 1999a). The implication is that the way in which u_* and z_0 in the Law were derived was somewhat arbitrary.

There are many reasons for divergence from the semi-logarithmic curve. Three apply to common situations on dunes.

- When sand is in motion, there is a change in the slope in the velocity/height profile near the top of the cloud of bouncing sand. This point has been dubbed 'Bagnold's kink', after its discoverer (Bagnold, 1941, p. 59) (McEwan, 1993).
- Above a heated surface, which is the norm in deserts during the day, the air is unstable, and the wind-velocity profile, below about 0.5 m from the bed, is not semi-logarithmic. Errors of up to 15 times the value of u_* may occur if the stability condition of the atmosphere is not allowed for (Frank and Kocurek, 1994). The reason for this kind of deviation in the wind-velocity profile is that heat then joins shear as a driver of vertical mixing, and ensures that there is less change in velocity with height than in neutral conditions. The sand-flow rate in these conditions reaches equilibrium more quickly than in neutral conditions (Lu Ping and Dong ZhiBao, 2008). In very stable conditions, as during cold nights in some deserts and, more distinctly, in extreme cold at high latitudes or altitudes, the profile also deviates from the logarithmic. Mosaics of surfaces with different responses to heating, also the norm in deserts, create even more complex wind-velocity profiles as the wind repeatedly passes over surfaces of different roughness (Butler *et al.*, 2005).
- The height distribution of velocity above a sloping or curved bed is less easy to accommodate in estimates of shear. This is an issue in the explanation for flow over dunes (Chapter 3).

There are now many alternatives to the semi-logarithmic model of the relation between wind velocity and height, but most are useful only in wind tunnels, and few have been used by geomorphologists (Bauer *et al.*, 2004).

The roughness height (z_0)

This was the second element in the Law to be questioned. As explained earlier, the definition of z_0 in the Law is as arbitrary as that of u_*.

The most commonly used estimate of z_0 in a cloud of bouncing sand is now Owen's model (1964):

$$z_0 = \alpha\left(u_*^2 / 2g\right),$$

where g, as ever, is the gravitational constant.

Owen proposed that α=~2; others have found different values (Rasmussen *et al.*, 1996). Owen's model assumes that the cloud of salting sand is a form of roughness. Unlike z_0 in the Law, z_0 in Owen's model is not at the same height in all winds. Owen also discovered that shear on a bed under a cloud of saltating sand was the same as the shear anywhere within the saltating cloud. In other words, the saltating cloud is a self-organising system, the first of a number of self-organising systems that shape dunes, as will be explained later, especially in Chapter 4. As with the height/velocity curve, there are now several other formulations of z_0 for situations in which sand is in movement (detail in Bauer *et al.*, 2004).

Microturbulence

Turbulence is one of the main ways in which energy is transferred from the wind to the bed. In the scale limits of this chapter, it is only small-scale turbulence that is relevant. At this scale, turbulence is structured into 'burst–sweep' sequences. A burst is a downward spurt of air that replaces the air that has just been removed by a sweep, which is a slower upward ejection from the bed. On a loose sandy bed, the leading edge of a burst may dislodge sand, which is then taken up by a sweep. Burst–sweep sequences are responsible for most entrainment, even when, as is probably common, they are effective for only about 20% of the time (Sterk *et al.*, 1998). Measurements on a sandy, eroding field in Burkina Faso revealed that the burst–sweep sequence (at that site and on that occasion) had a downwind dimension of 0.25 m (Leenders *et al.*, 2005).

Turbulence at this scale can be measured by the Reynolds stresses, which describe the variation of velocity in three dimensions. Velocity in the windward direction is denoted 'u'; in the vertical (up or down) 'w'; and in the lateral (sideways in either direction) 'v'. u is positive downwind; w is positive upward; v would take the discussion beyond the two-dimensional frame of this chapter; its role in determining the two-dimension form of dunes has anyway hardly been explored. In a burst (towards the bed), u' is positive (wind-directional flow faster than the mean), and w' is negative (flow more downward than the mean). In a sweep (movement away from the bed), u' is negative (wind-directional flow in the sweep being slower than the mean), and w' is positive (upward flow at a faster rate than the mean). The prime symbol (′) denotes a fluctuating variable.

The best measure of shear on the bed

This is the most important issue raised by all these reservations about the Law. Shear velocity (u_*) is, by definition, a description of the mean wind, which is seen in the relationship between u_* and the transport rate at progressively smaller averaging intervals (Namikas *et al.*, 2003). Thus, any study of entrainment at a small scale must acknowledge burst–sweep sequences. One alternative for measuring turbulent flow is the 'Reynolds Shear Stress', which combines stresses in the forward and vertical dimensions: $-\overline{u'w'}$ (the overbar denotes the mean). When multiplied by the air density (ρ) this gives a force, $\rho\left(-\overline{u'w'}\right)$.

The value of u_* has also been questioned even at larger scales. The solution could be as simple as deriving u_* from a wind profile measured down to about 0.05 m of the bed (Bauer *et al.*, 2004), or measuring the free-stream velocity 'well above' the bed (Schönfeldt and von Löwis, 2003), both of which solutions are empirical rather than theoretical. Developments in measuring sand transport, including shear or force balances, may deliver more theoretically acceptable measures of shear (Gillies *et al.*, 2000).

Lift-Off

A particle of sand starts to move (is 'entrained') when the forces that hold it down are exceeded by those that might rip it away.

Holding down by gravity

The gravitational force is defined thus:

$$g\left(\rho_{\mathrm{p}} - \rho_{\mathrm{a}}\right)d^3,$$

where g is the acceleration owing to gravity; ρ_{p} is particle density; ρ_{a} is the density of the air (or other fluid); and d is the particle diameter.

In other words, where the densities (ρ_{p} and ρ_{a}) are constant (when all the sand is of the same mineral, say quartz, and the air density does not change, as in many situations), particles of greater size (d), are held down by a greater gravitational force. The model can accommodate the behaviour of sands that are denser than quartz (say, of magnetite) or less dense (say, of diatomite), and of fluids with different density. Chapter 12 includes a discussion of the effects of the differences in gravity, temperature and air density on the lifting and carrying of sand by the wind on Mars, Venus and Titan.

Holding down by cohesion

Cohesion derives from several of the characteristics of particles. First, finer particles pack more closely, which means that they touch each other in more places and are thus more coherent. Second, rough particles touch each other in many more places than do smooth ones and so also cohere better. Third, platy shapes, as of many fine particles (particularly clays) allow much more contact than do rounded shapes, if packing is parallel (as it usually is for clays). Fourth, physicochemical bonds, known as London–van der Waals forces, increase cohesion between clay particles of some mineralogies (many clays) but are weaker between particles of some common rock minerals, such as quartz (Cornelis and Gabriels, 2004).

The fifth and sixth forms of cohesion come from water held between particles. The fifth is the meniscus force, which is strongest where a meniscus has a small angle of contact with a particle. This is the case where the voids between particles are small, as they are between fine particles. The strength of this force also depends on the roughness, roundness and surface properties of the particles, and on contaminants in the water.

None of the cohesive forces is as strong as the sixth form of cohesion. It depends on water held (adsorbed) on the surfaces of particles. The amount of water held in this way increases with relative atmospheric humidity, but, contrary to intuition, the static threshold (shortly) peaks at a relative humidity of 35–40%, below and above which value, entrainment is easier (Ravi and D'Orico, 2005). All of these properties are difficult to measure, and their combined effect is a major challenge to modellers and experimentalists (McKenna-Neuman and Sanderson, 2008). The relation between moisture and movement is discussed again later in the section on the dynamic behaviour of moisture in an eroding bed.

In sum, fine particles are more coherent than coarse, other things being equal, which they often are. Cohesion, of all sorts, is a function of $(1/d)^3$, where d is the diameter of the particles. This is the start of an explanation for why dunes are sandy: fine particles cohere too well to be easily moved by the wind (with some exceptions, later).

Raising by lift

Shear moves particles on a loose bed by two kinds of 'aerodynamic entrainment'. The first, lift, occurs because fast flow is accompanied by low pressure, following Bernoulli's equation. Flow over a bed of particles is faster than the velocity of the surrounding fluid in two situations: first, where there is a difference in pressure between the slow flow very near the bed, and the faster flow just above it, the slope of the velocity/height curve being steepest there; and second, where the wind is accelerated over a protruding particle.

Lift is more effective on rough than on smooth beds, on moving than static particles, and in sudden changes in velocity, as caused by turbulence. In some circumstances, lift may also be augmented by thermal diffusion from a heated surface, or by electrostatic forces arising from friction between the wind and the sediment (Rasmussen *et al.*, 2009). Lift alone entrains few particles, but it lightens the task of the other forces.

Raising by drag

Drag is usually stronger than lift. 'Surface drag' is caused by friction between the wind and the bed. It causes both rolling and sliding. 'Form drag' is caused by the difference in pressure between the windward and lee sides of a protruding particle, especially in high turbulence. Because it is greatest on top of the particle, it causes rolling. Both contribute to 'aerodynamic moment', which is the force on a particle that is dependent on its projection above the surface: particles that project more (bigger or longer ones) are toppled and therefore entrained more easily. Drag can help to eject particles when they collide or are dragged over projections. The magnitude of both forms of drag is proportional to $u_*^2 d^2$ (*d* being the particle diameter). Drag, like lift, is effective only very close to the bed, and raises few grains on its own. When the particles are clear of the bed, many acquire spin, which may contribute up to 24% of lift (the Magnus effect) (Huang Ning *et al.*, 2010).

Raising by bombardment

When particles are lifted into the wind, they pick up momentum from the wind in their trajectory above the surface and take some back to the bed on their return. This is 'bombardment', and, when sand is in movement, it is more powerful than any of the other sand-raising mechanisms. It both dislodges loose grains and breaks up aggregates (pellets) and crusts (both later). Once sand is lifted in sufficient amounts, further entrainment is almost wholly by bombardment.

Thresholds

This section adds to, but does not yet complete, the explanation for why dunes are built of sand. The wind can lift grains if it has sufficient power, which is to say (following the line of reasoning earlier), it is fast enough. The critical condition, when sand begins to blow, is the threshold velocity (u_t), or, more generally, the threshold shear velocity (u_{*t}).

In Bagnold's (1941, pp. 85–90) terms, the *static* (or fluid) threshold is the wind velocity at which grains start to move under the influence of lift and drag alone; and the dynamic (or impact) threshold is passed when particles are bombarding the bed. The dynamic threshold occurs at a lower velocity than the static threshold, because of bombardment. A much later model shows that in most places on Earth, the velocity at the dynamic threshold is ~0.96 (the ratio is different on Mars, Venus and Titan, as described in Chapter 12; Almeida *et al.*, 2008a).

Bagnold (1941, pp. 85–90) built the first mathematical model of the static threshold, which described the balance between the lifting and retaining forces on a particle. Subsequent theorising was reviewed by Cornelis and colleagues (2004b; Figure 1.2), who developed a model of their own, which is simpler and more testable than some earlier versions.

Thresholds have been found to be much more complicated than this, in theory, in wind tunnels, and in the field. Even in dry sand (moist sand is discussed shortly), surface conditions, such as roughness, the grain-size mix, and other factors, each produce their own thresholds, and these may vary in time and space

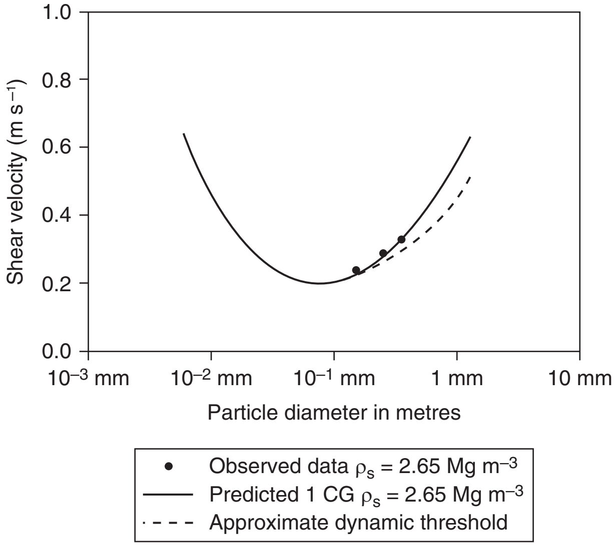

Figure 1.2 Threshold curve for the start of motion (the static threshold) of particles of the density of quartz (observed and modelled data) (Cornelis and Gabriels, 2004). The curve for the dynamic threshold is added and very approximate. Reprinted with permission from John Wiley & Sons.

(Baas, 2007). The following list, therefore, is far from exhaustive. Different thresholds occur when: (1) saltation reaches the intensity at which it can move sand in reptation (shortly); (2) bouncing grains are powerful enough to disperse clods, pellets, and crusts, of varying cohesiveness (Hu ChunXiang *et al.*, 2002) (also shortly); (3) ripples appear (Chapter 2); and (4) ripples move from a 'subcritical' shape (with gentle lee slopes), to a 'supercritical' shape, in which there are small slip faces (Hoyle and Woods, 1997).

Even within one of these groups of threshold, there is a range of behaviour. In a wind tunnel, the static threshold occurs at a spread of velocities, from that at which particles begin to rock back and forth; start rolling; or are lifted from the bed; or, at a larger scale, between the point at which there are only a few flurries of movement in response to bursting turbulence and the point at which the whole bed is mobile. If the wind is accelerated in the wake of even quite small roughness elements, local entrainment occurs at velocities lower than the ambient threshold (Sutton and McKenna Neuman, 2008). The difference between early, sporadic movement and the movement of the whole bed produces static thresholds with a wide spectrum of values. It would be better to choose a probability density of values, although this is seldom done (Williams *et al.*, 1994).

In the field, thresholds are yet more elusive. Measurement is more complex (both of the wind and of blowing sand): winds are gustier; sediments have more sizes and densities; controls like moisture or crusting can limit the supply of loose particles (later); there is far greater spatial variability in all these controls; and features like sand streamers (shortly) complicate measurements of sediment movement. In one field experiment, sand was seen to move at a velocity below a theoretical threshold (calculated from grain size and wind speed), almost certainly because of high instantaneous wind velocities (Rasmussen and Sørensen, 1999). In another experiment, now on a wet beach, it was found that each size and size-mixture of sand, and each set of environmental variables, had its own threshold (Wiggs *et al.*, 2004a). Many of the problems involved in measuring thresholds in the field may be overcome by using terrestrial laser scanning, which can quickly and non-invasively measure surface topography, moisture content of the surface, and perhaps even the sizes of grains in saltation (shortly; Nield and Wiggs, 2011). A recent study, based on measurements in the field, discovered yet another cause of variability: the way in which the threshold is calculated from data (Barchyn and Hugenholtz, 2011).

Grain size

This section continues, but does not, even yet, complete, the explanation for why most dunes are sandy. The relationship between the static threshold and the grain size of sand is shown by a simple experiment: trays, containing grains of a succession of sizes of particle, each with only one size, are exposed, one by one, to increasing wind speeds in a wind tunnel, and the threshold velocity (u_t) or threshold shear velocity (u_{*t}) at which each size of particle begins to move is plotted against its size. The results of such an experiment are shown in Figure 1.2.

The most important (and most obvious) characteristic of the curve in Figure 1.2 persists, whatever the definition of the static threshold and by whatever means it is verified or theorised: there is a minimum value of the velocity at a grain size of ~0.07 to ~0.1 mm (somewhere between medium and fine sand on the Wentworth scale, earlier). Both values are different in higher or lower pressures and temperatures, as at altitude, near the poles, or on Mars, Venus or Titan (Figure 12.1).

The increase in the threshold on the right-hand side of Figure 1.2 is what one might expect intuitively, and by the earlier explanation for the effect of gravity: bigger, heavier grains need stronger winds to move them than do smaller, lighter ones. But, even for coarse grains, there are complications, the main one being when there is a mixture of sizes. A wind-tunnel study found that the greater the spread of particle sizes in a mixture, the greater was the difference between the static and dynamic threshold (Nickling, 1988). Coarse particles are usually the first to move from a mixture with finer grains, probably because they stick up further into the wind (Martz and Li, 1997). The finer grains in a mixture have a higher threshold of movement than in a one-size sandy bed, because of the sheltering effect of the larger grains (Komar, 1987, RL).

The increase in the threshold on the left-hand side of Figure 1.2 is, of course, because of the increasing effects of the cohesive forces on finer grains (earlier). In short: the finer the particles, the faster the wind must be to move them. This is true in the conditions of the experiment, and in many situations in nature, but does not hold where a bed of fine particles is being bombarded by saltating sand, in which case clods and crusts (shortly) are broken down to dust and raised by the wind.

The slope of the bed

This has a small effect on the threshold values for most sands and in most field situations. If the slope is ≥15°, a slope that is seldom exceeded on the windward slopes of desert dunes (perhaps for this reason, Chapter 3), the threshold is not materially affected. There is, however, a distinct slope effect with coarse sands: field measurements (as reported by Bagnold, 1941, p. 220) and modelling have found that particles coarser than 0.23 mm in diameter (fine sand) are unable to climb a 20° slope normal to the wind at common wind speeds (Tsoar *et al.*, 1996). Although this slope is steeper than many of the windward slopes of dunes, and the sand is not as coarse as many, this may explain why there are accumulations of coarse sand at the base of many dunes. A study of grain-size variations along transects over dunes in the Taklamkan found very distinctly coarser sands on the lower slopes (Wang Xunming *et al.*, 2002c).

The dynamics of water content

It must be true (given earlier arguments) that the amount of water held in a sediment has a strong effect on cohesion, and therefore increases the threshold of movement, but it is difficult, perhaps impossible, to measure the full effect. This is because moisture content in the field and even in the wind tunnel is not constant beneath a surface subject to a wind strong enough to raise particles, because the winds also take moisture.

The various models of the interactions of moisture and the speed of the wind have been reviewed by Cornelis and Gabriels (2003b; Figure 1.3), who developed their own model and tested it in a wind tunnel. This curve would probably be different for different grain sizes of sand. For example, the coarse pores in a coarse sand would lose moisture more quickly, and the capillary replacement of water from lower in the sediment would be slower.

Research in wind tunnels delivers far smaller water contents at the threshold than do field experiments. In one field experiment on a beach of well-sorted coarse sand, the threshold lay between 4% and 6%, which is much greater than has been found in experiments in wind tunnels for sand of the same size (Wiggs *et al.*, 2004a). The reasons are not hard to find. In the field, there is much greater variation in water content, caused by varying inputs of rainwater or dew, and greater variation in output by upward movement (by capillarity) from groundwater; and varying output by evaporation, itself accelerated by the wind, insolation,

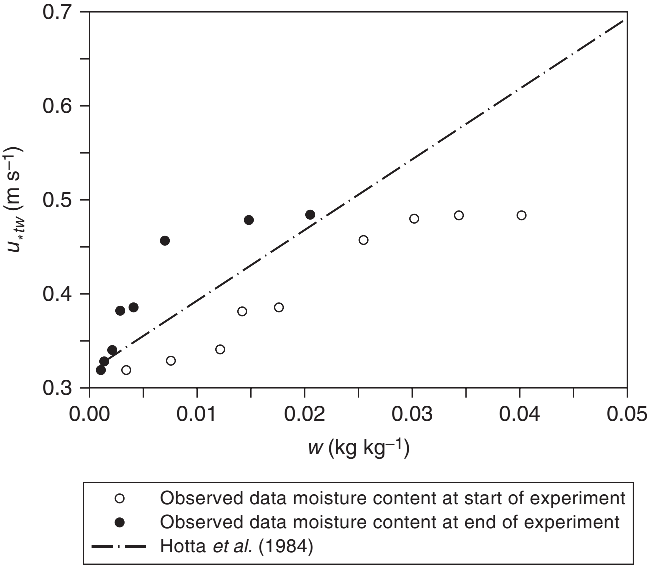

Figure 1.3 Static threshold for grain motion (u_{*tw}) related to the gravimetric water content of the sediment at the start and end of an experiment. Only the results predicted by the model of Hotta *et al.* (1985) (as the best fit of many models) have been retained (Cornelis and Gabriels, 2003). Reprinted with permission from John Wiley & Sons.

high temperatures, drainage under the influence of gravity and capillary rise. The threshold over wet sands varies much more rapidly than that over finer wet sediments, because the higher porosity of sands allows them both to imbibe water more quickly and, if there is not a shallow water table, to drain more quickly. Sands on the surface also dry out more quickly because capillarity in the large pores between sand grains is too weak to bring in water laterally or from below to replace water lost to evaporation. Thresholds may be passed suddenly if water is suddenly lost from the equally sized pores between well-sorted sand. In some circumstances a wet surface can even be hard enough for saltons to bounce off it, thus increasing the intensity of saltation (shortly; McKenna-Neuman and Maljaars Scott 1998).

For these reasons, the threshold in sands may have a daily cycle, according to variations in temperature (Stout, 2003). On a wind-eroding beach, moisture content may also vary with the state of the tide, sea spray and the content of hygroscopic salts. In a set of field experiments on a beach, the threshold was closely correlated with the moisture content at the upwind end of the beach but not at sites further downwind, where bouncing sand drove entrainment (Davidson-Arnott *et al.*, 2008). Erosion and transport create their own spatial and temporal patterns on moist surfaces. If the removal of a layer of dry sediment exposes wet sand, entrainment is suddenly checked, and does not pick up again until the new surface dries. The result is the pulsing of erosion (Pease *et al.*, 2002).

Complexity increases at extremes. At very low temperatures, sand may be released only if the rate of sublimation of ice is sufficient, and this rate itself depends on a number of factors including the temperature, local wind speed, air humidity and ice content of the sediment (Speirs *et al.*, 2008a). In cold wind-blown environments, the patchy growth or thawing of needle ice may withhold or release sediment to the wind, as in parts of South Island, New Zealand, and on some Chinese dry lakes (McGowan, 1997; Mu GuiJin *et al.*, 2002). Some simple statements, notwithstanding, can be made with a modicum of confidence: there is no particle movement at soil water contents over 75%; and the inhibiting effect of moisture is less in coarse than in fine sand (Cornelis *et al.*, 2004b; Ishizuka *et al.*, 2005).

Crusts

Crusts (~2–3 mm thick) are very common on dunes, even in some very dry climates, and can seriously hamper entrainment. There are two types. Abiotic crusts are created by the impact of raindrops, or after evaporation (Ishizuka *et al.*, 2008). Both these processes can produce crusts, even on a bed of glass beads. There are stronger abiotic crusts in the presence of salts, such as gypsum, and where dust has been added to the surface (Scheidt *et al.*, 2010). Abiotic crusts offer less resistance to bombardment than do most biotic crusts (McKenna-Neuman and Sanderson, 2008).

Biotic crusts are formed by animalcules, mosses, lichens, liverworts, fungi and cyanobacteria, whose organic filaments or glues bind the sand (Malam-Issa *et al.*, 2001). On beaches, organisms may aggregate patches of sand, according to subtle ecological differences in moisture and disturbance (Maxwell and McKenna-Neuman, 1994). In the dry, seasonally very cold, Gurbantunggut Desert in far north-western China, where the mean annual rainfall (MAR) is only 70–150 mm, biotic crusts cover 30.5% of the crests of dunes, 48.5% of the upper slopes, 55.5% of middle and lower slopes and 75.5% of interdune areas (Chen YaNing *et al.*, 2007). On dunes in the Negev in Israel (MAR 95 mm, and a much milder winter), morning dew is sufficient to sustain an algal crust, which offers very effective protection against entrainment (Kidron *et al.*, 2009).

The strength of biotic crusts varies from species to species (Godinho and Bhosle, 2009). They are both thicker and more flexible than abiotic crusts and seldom crack or fall apart. If cracks do appear, their frayed edges are the most vulnerable to entrainment by bombardment (Langston and McKenna-Neuman, 2005). Biotic crusts develop at a rate dependent on the environment and the crust-forming species. The rate of development accelerates only after a few years, and some crusts do not mature for ~20 years and can therefore fully develop on already very stable (and undisturbed) sites (Belnap *et al.*, 2009); but disturbed and therefore rougher crusts may and raise the threshold of entrainment.

Pellets

Pellets are aggregates of sand size. Windblown pellets have been reported (or inferred) from: Devonian deposits in Scotland; Late Pleistocene sand dunes in Indiana; Late Pleistocene loess; agricultural fields; Mars (Chapter 12); and many other situations (Rogers and Astin, 1991; Mason *et al.*, 2003a; Kilibarda *et al.*, 2008; Colazo and Buschiazzo, 2010). Their binding agents include micro-organisms, clays, salts and electrostatic forces. In sufficient concentration, sand-sized pellets can be built into dunes (Chapter 4), but most aggregates never reach dunes because they break down quickly in transport and are then dispersed as dust. The abrasion of pellets is faster where most of the saltation load is unaggregated sand (Hagen, 1984). In Western Australia, much stronger pellets, known as 'spherites', have been reported from near-coastal dune sands (Killigrew and Glassford, 1976). They are held together by various cements. The most robust survive in many environments.

Sand in Motion

Once raised, particles travel in four ways, which, in order of increasing velocity, are creep (and related near-surface activity), reptation, saltation and suspension (Figure 1.4). Reptating particles are reptons; saltating ones, saltons; suspended ones, dust. This account begins with saltation because it drives all the others.

Saltation

Saltation is the leaping of wind-driven grains. In the controlled environment of a wind tunnel, and with rounded sand, most saltons are ejected at an angle between 50° and 80° (forwards) from the surface (Li Wanqing and Zhou Youhe, 2007;

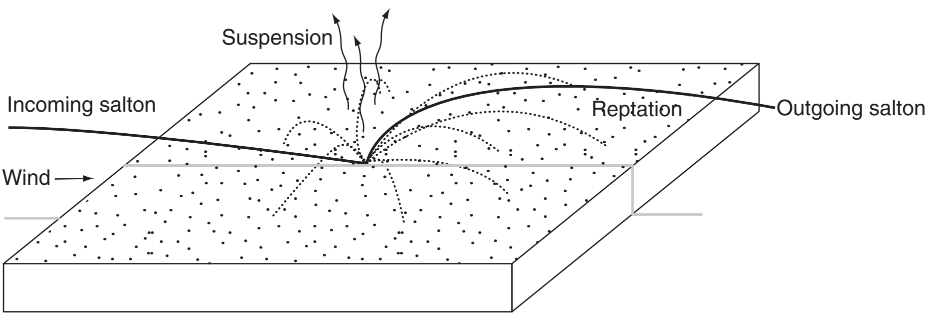

Figure 1.4 Modes of grain transport in the wind (partly based on data from high-speed filming by Mitha *et al.*, 1986). The terms are explained in the text.

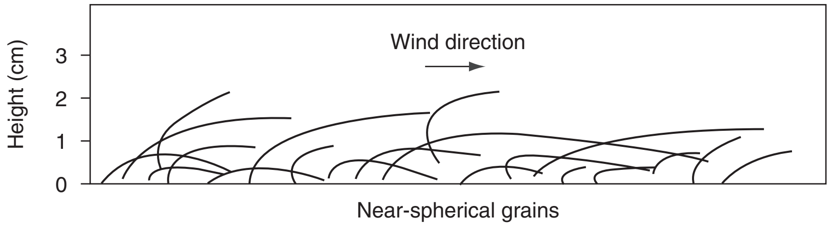

Figure 1.5 Trajectories of near-spherical saltons (Willetts, 1983). Reprinted with permission from John Wiley & Sons.

Figure 1.5). The modal lift-off angle is ~20° to ~40°. No take-off angles less than 20° have been observed. Take-off angles are greater with less rounded grains (Willetts and Rice, 1985). Ejection occurs at a higher velocity where saltons meet an upward-sloping surface, as on the windward side of a ripple or a dune (Willetts and Rice, 1989). A few saltons leap back into the wind, most of them having rebounded from the upwind side of a protrusion; there are more backward leaps at higher wind speeds (Dong Zhibao *et al.*, 2002d). The time of flight is ~50 ms, and a hop on Earth is ~50 cm long (much greater on Mars; Chapter 12). Saltons spend a large part of their flight near the top of their trajectory, which explains the everyday observation that saltating grains seem to 'float' above the surface of a beach or a sandy desert. Angles of return for common sizes of sand are between 9° and 15° to a flat surface (Rice *et al.*, 1996a).

In wind-tunnel experiments, launch velocity does not alter much with u_*, which may be because increasing u_* drives more grains in saltation, rather than increasing the velocity of individuals (Namikas, 2003; Rasmussen and Sørensen, 2008). When they reach the highest point of their trajectory (~80 mm above the bed), the velocity of 0.242 mm and 0.320 mm diameter saltons reaches that of the wind. As they descend through the slower wind nearer the bed, the ratio of their velocity to that of the wind around them steadily increases, until it reaches ~2 when they meet the bed, all irrespective of grain size. Because of the added impetus given by collisions, the forward speed may exceed the wind speed. The denser the cloud and the faster the wind, the more collisions there are, and these widen the distribution of the velocities of the grains. Finer particles have less chance of collision because they travel high in the wind, where there are fewer grains in motion (Dong Zhibao *et al.*, 2005). Mixtures of sizes of sands have little effect on the characteristics of saltation, such as ejection speeds, ejection angles and the mass flux profile (Xing Mao *et al.*, 2011).

In general, smaller particles travel faster, but a few large saltons leap high, perhaps because of their greater momentum, and because their lower specific surface area suffers proportionally less drag (Jensen and Sørensen, 1986). In a sand storm in which fine particles of the order of 0.1 mm rose only a few centimetres, a few with diameters of 1.0 mm reached 1.5–2 m above the surface (Sharp, 1964). In a severe windstorm in the San Joaquin Valley in California, in which wind speeds at

10 m above the surface reached 53 m s^{-1}, particles of 23 mm diameter were imbedded in a telegraph pole 0.8 m above ground (Sakamoto-Arnold, 1981).

Almost all observations of saltons have been in wind-tunnel experiments. The few that have been made in the field show more complex behaviour, for several reasons, principally the variations in roughness, and the hardness of the surface and the coverage of crusts (Stout and Zobeck, 1996c).

Streamers and other medium-scale patterns of saltating sand

Streamers of bouncing sand, so familiar on windy beaches and dunes, may be related not to the burst–sweep process but to the downward propagation of larger, higher turbulent structures. In one experiment, streamers on a fairly uniformly dry sandy surface were ~2 m wide and spaced at ~1 m. These dimensions appear to be independent of the velocity of the wind and perhaps also of the nature of the bed. Streamers intertwine and bifurcate, in ways that are more complex in higher winds. They may have a dominant role in initiating and maintaining the movement of sand, implying that there is more yet to learn about sediment transport by the wind, but their study presents some major challenges, and they have been little researched (Baas, 2008).

There are other horizontal patterns of transport on Sahelian fields, and doubtless in many other similar situations. They are created by roughness elements such as bushes (Visser *et al.*, 2004b). Another pattern, on a dry lake in western Queensland, is also likely to be common. It was generated by variations in the availability of sediment. The horizontal variability of supply was greater at low wind speeds than at high speeds, when many sources yielded sediment (Chappell *et al.*, 2003c).

Reptation

Reptation (or 'impact creep') is the 'splashing' or low hopping of grains that have been dislodged by descending saltons (Figure 1.4). Reptons hop just once. They have much less momentum than saltons, so that, when they return to the surface, they neither rebound nor disturb others, although they may roll a few millimetres. The size distributions of reptons and saltons have substantial overlap, and grains continually pass between the two modes of travel, but the velocity distribution of reptons is heavily skewed towards small velocities, with a long tail of faster ones, whereas that for saltons follows a peaked Gaussian distribution (Anderson, 1987b). The velocity of reptons is only weakly dependent on u_* but strongly related to their size (Schwämmle and Herrmann, 2005a).

Reptons absorb more of the energy of the saltons that disturb them than is taken by outgoing saltons, and at any one time most of the grains in motion are reptons (Anderson *et al.*, 1991; Rice *et al.*, 1995). The transport rate in reptation, being powered by saltation, scales with u_*, but its contribution to the overall transport rate declines as the overall transport rate increases. Because they are not driven directly by the wind, reptons respond to the effect of slope, where saltons cannot. Chapter 4 shows how this behaviour may play a major role in controlling the shape of transverse dunes.

Creep

Before the discovery of reptation (as it is now understood), the term 'creep' was applied to all near-surface movement (as by Bagnold, 1941, pp. 33–35). 'Creep' is now reserved for two types of slow surface movement, which, like reptation, are caused wholly by bombardment (rather than wind shear). The first is the rolling of coarse particles driven by the impact of saltons (unlike the small leaps of reptating grains); the second is rolling under gravity, as into craters created by the saltation impacts, or down the lee sides of ripples (Chapter 2). Most particles in creep are coarser than those in saltation or reptation, but the size of creeping grains is also a function of the prevailing mix of grain sizes and of u_*. Data from fast-shutter images show that with u_* at 0.48 m s^{-1}, 0.355–0.6-mm-diameter sand grains creep at ~0.005 m s^{-1}. Grains of the same size begin their journey together but rapidly disperse, as some move faster than others; in one experiment, dispersal was complete within 3 min (Willetts and Rice, 1985b). Some creeping grains (like some reptons and saltons) are buried for long periods of time, many in ripples (Barndorff-Nielsen *et al.*, 1982). Partly because of differences in sampling methods, partly because of the use of different sizes of sand and of different values of u_* in experiments, and partly because of the occasional lumping of reptation with creep, estimates of creep as a proportion of the total wind load have varied from 6.5 to 50% (Wang Zhenting and Zheng Xiaojing, 2004). In Wang and Zheng's model, the proportion of the load in creep is a high proportion of the total flux rate at low u_* but rapidly declines at higher u_*.

Other near-surface activity

A closer look finds that yet more activity occurs near the surface. The first is the tunnelling beneath the surface of high-velocity saltons on their return (Willetts and Rice, 1985b). The second is a compressional-dilational wave that radiates from the point of impact of a salton, shaking the sand to a depth of about five grain diameters. Models show that the shaking raises large particles by rotating and ratcheting them against smaller ones, which lifts rougher particles more quickly. All this activity may explain the very thin layer of exceptionally well-sorted coarse sand on an erosional surface (Sarre and Chancey, 1990). Third, bombardment may elevate some grains to positions where they are more vulnerable to dislodgement (Iversen *et al.*, 1987). Countering the destabilising processes, bombardment sometimes consolidates and partly immobilises the surface.

Suspension

Suspension depends on both u_* and the fall velocity of a particle, u_f, which is a function of the balance between the weight of the particle and the drag of the air upon it (Stoke's Law). The vertical velocity in turbulence near the ground is directly related to u_*, so that, as a rule of thumb, if $u_f < u_*$, particles stay aloft. Observations at quite a spread of values of u_* show a sharp transition to suspension or an intermediate behaviour, 'modified saltation', where grain size is less than ~0.1 mm (Nalpanis, 1985). There is therefore a fairly clear and common distinction between the behaviour of particles of silt and clay size (or dust), which can be held aloft at many common wind speeds, and that of sands, which rarely go into suspension (except in the lee of dunes, Chapter 3). This completes the explanation for why dunes are sandy.

The vertical distribution of load and grain size

Measurements of the movement of a mixture of sizes of sand using a Phase Doppler system have now shown that the maximum flux is just above the surface (Figure 1.6); the peak height increases with, and is more marked in, stronger winds. Above this near-bed convexity, the profile adopts a form of curve more like Bagnold's (1941, p. 63; Dong Zhibao *et al.*, 2006a), which showed a near-bed plateau below a smooth curve. The near-surface peak is probably the mean height of reptating grains (Ni JinRen *et al.*, 2003a).

The top of the saltation 'cloud' proper (in which there are many more particles in motion) is between 14 and 15 cm above the surface. In most cases, with the exceptions noted earlier, the modal size of grains in transport declines smoothly with height above the bed. Different values of u_*, different grain sizes and grain-size mixtures, and other variables affect the vertical distribution (Dong Zhibao and Qian GuangQiang, 2007b).

The saturation length

As well as responding to the speed of the wind, the quantity of sand carried responds to a change in the character of the surface over which it blows. The response to a change from a hard, compact surface to a loose sandy surface is of special interest because it probably determines the minimum size of a dune, and may be one of the main determinants of the shape of a mature dune (Chapter 3). The adjustment to a loose surface is the outcome of changes in many different

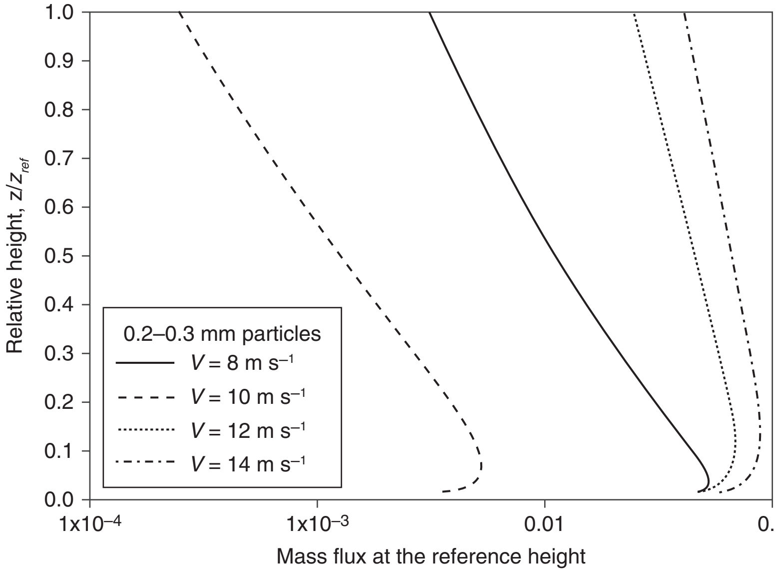

Figure 1.6 Vertical distribution of dimensionless sand flux, for 0.2–0.3 mm particles, at different free-stream velocities (Dong Zhibao *et al.*, 2006a, which contains data for many more particle sizes).

processes, such as: the length needed to complete the release of new grains into the wind; their acceleration to a new steady velocity; the change in hop length of saltons on the new, loose surface; and the decelerating effect on the wind of the increase in the load of sand it is carrying. It is the slowest of these processes that determines the ultimate distance to overall adjustment.

Bagnold (1941, pp. 180–183) discovered what is now known as the saturation length in his wind-tunnel experiments. More recently, the phenomenon has been studied, also in a wind tunnel, by Andreotti and colleagues (2010). As shown on Figure 1.7, they found two phases in the overall response: first, an exponential increase in the load as new grains are released, which they labelled '$L_{1/4}$'. $L_{1/4}$ becomes rapidly shorter at higher wind speeds and is negligible in high winds; second, a stage of steadily increasing load towards a new plateau, over a distance that they labelled 'L_{sat}' or the 'saturation length'. In their experiment, L_{sat} was ~1.5 m long. This is well short of Bagnold's result, a difference they attribute to Bagnold's system for measuring sand flux with spring balances, compared with the spatial resolution of 10 cm for their own measurements of sand flux. L_{sat} is not sensitive either to changes in wind speed or to whether the approaching wind does or does not carry sand. Andreotti and colleagues noted that:

$$L_{sat} \approx 4.4\rho_s / \rho_f d,$$

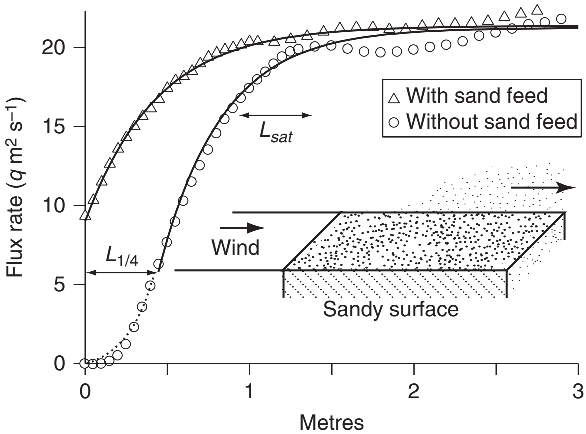

Figure 1.7 Saturation length or L_{sat} (the distance taken for the load carried by the wind to adjust it passes from a hard, cohesive to a sandy surface). The terms $L_{1/4}$ and L_{sat} are explained in the text (redrawn from Andreotti *et al.*, 2010).

where ρ_s and ρ_f are the densities of the sand and of the air, and d is the sand grain diameter

As can be seen on Figure 1.7, Andreotti and colleagues' experiment revealed a minor fluctuation immediately after the first peak in the curve of L_{sat}. Bagnold's finding (1941, p. 182), which was more or less confirmed by Spies and McEwan (2000) and Arnold (2002), was that there were more distinct and more fluctuations beyond a first peak, which continued for ~7 m downwind of the upwind edge of the patch of sand. The explanations for the fluctuations offered by McEwan and Arnold were: (1) the equilibrium between the cloud of saltating sand and the wind above it takes time to propagate up through the saltating curtain (Arnold, 2002); (2) saltation trajectories shorten as the number of grains in the wind increases, and this brings down the transport rate (Almeida *et al.*, 2007), which could be because; (3) the buildup in the rate of collisions between saltating particles, which disperses the available energy, as shown later by Dong Zhibao and colleagues (2005).

The fetch effect

This is a longer spatial and temporal adjustment between the wind and its load. The adjustment occurs on surfaces, such as agricultural fields, dry-lake beds or tidal beaches, where the sediment on the surface contains a significant content of silt and clay, or is of variable wetness, both being situations where the surface releases sand slowly and which therefore delays the point of maximum load, which is not achieved until the 'fetch' distance (Delgado-Fernandez, 2010). There is no fetch effect where there is an adequate supply of sediment, as on a dry beach, or in most desert conditions.

The response of a loose bed to erosion by the wind

Because the fine particles on a bed of unaggregated particles are removed before the coarse ones, the surface of the eroding bed coarsens (Bagnold and Barndorff-Nielsen, 1980), which raises the threshold of movement and reduces the erodibility of the bed. The same is true where an eroding surface is covered with loosely cemented aggregates or a crust. In that case also, the fine particles are the first to be released, at least until the crust or pellets have disintegrated, although this point is seldom reached within a single wind storm (Stout and Zobeck, 1996).

The Transport Rate

The transport rate is commonly denoted q (or Q), defined as the mass of sediment passing through a plane perpendicular to the wind, of unit width and of infinite height, per unit time (in kg (m-width)$^{-1}$ s^{-1}; or m^3 (m-width)$^{-1}$ s^{-1}).

Because of the difficulties in measuring the rate of blowing sand, effort has been focused on the prediction of the transport rate from wind data (which are easier to collect, and are often collected routinely for other purposes). For all the effort, however, there is still no universally accepted relationship. This is partly because of the evolution of ideas about the importance of u_* (earlier), but some pessimists believe that the transport rate is inherently indeterminate, given the number of variables involved (Bauer *et al.*, 1996; Smith and Stutz, 1997). Indeterminacy can be kept to a minimum in a wind tunnel and is manageable where a surface in the field is dry and level, and where there is a copious supply of well-sorted quartz sand of appropriate size, but these requirements are rarely met. In the field, it is common to find variations in: grain size, hardness of the surface (shortly), slope and curvature of slope, microtopography, wetness, crusting and other controls on the availability of sand.

The modelling of the relationship between the speed of the wind and its load has been approached in four ways (Ni JinRen *et al.*, 2004): those based on (1) relations between the paths of saltating grains and the wind (Bagnold's approach, 1941, pp. 64–71); (2) the relations between the concentration of grains and their trajectories (Owen's approach, earlier); (3) linkages between trajectories with grain-surface collisions and the corresponding adaptation of the wind; finally, there has been (4) wholly empirical curve-fitting to experimental observations from either wind-tunnel or field measurements (for example, Liu Xianwan *et al.*, 2006). All of these approaches have relied heavily on data from experiments in wind tunnels.

Most of the models relate the transport rate (q) and shear velocity (u_*) in the general form:

$$q = \alpha\left(u_* - u_{*_t}\right)^b,$$

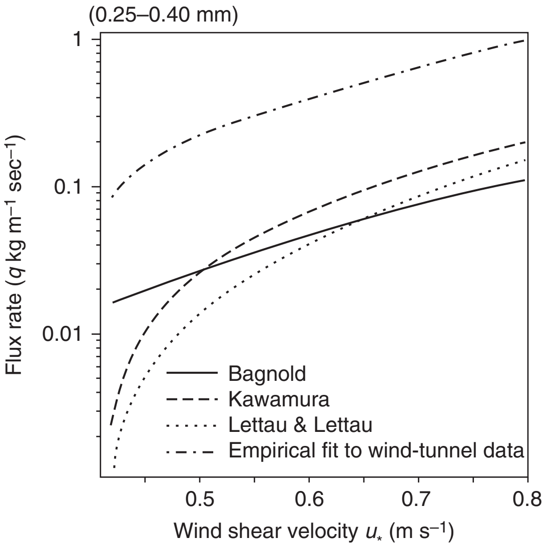

Figure 1.8 Comparison of the performance of selected transport formulae for particles 0.25–0.40 mm (medium sand) in size (Liu XianWan *et al.*, 2006).

where the constant α=~2; u_* is shear velocity; u_{*_t} is threshold shear velocity; and b is an exponent (Almeida *et al.*, 2007).

Comparisons of the predictions of various models show that they diverge in their predictions, some wildly. Figure 1.8 shows divergence of the relationship of wind speed and sand transport for one grain size (curves for many more grain sizes are given in Liu Xianwan *et al.*, 2006). The divergences are wider where the experiments have modelled extreme conditions, as for very fine and very coarse sands, or at very high or very low wind speeds. A field test on a windy Irish beach found most of the models gave poor results, probably because they could not account for surface moisture. Bagnold's (1941, p. 67) and Zingg's (1953c) models, both aging, were the best of a bad bunch (Sherman *et al.*, 1998).

The model that has been most widely used in studies of dunes (and especially in calculating the directional variability of sand movement, Chapter 4), is Lettau and Lettau's (1978):

$$q = C\left(\rho_a / g\right) u_*^3 \left(1 - u_t / u_z\right) t^{-1},$$

where q is the discharge rate of sand in grams (m-width)$^{-1}$; C is an empirical constant related to grain size, commonly ~6.5; ρ_a is the density of the air; u_t is the impact threshold velocity; u_z is wind velocity at height z; and t is a specified time period.

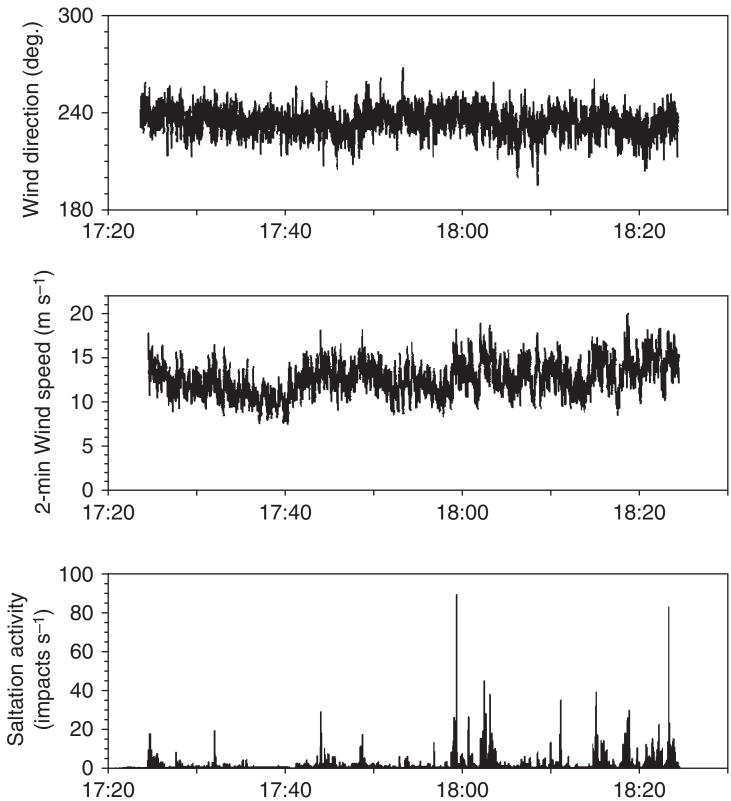

Figure 1.9 Variation in the speed and direction of the wind in the field resulting in a very variable rate of transport, labelled 'intermittent saltation' (Stout and Zobeck, 1997). Reprinted with permission from John Wiley & Sons.

New transport models continue to be proposed. One of the latest is designed to be used for small fluxes (Almeida *et al.*, 2006).

At the small scale, the transport rate has been shown to be remarkably responsive to changes in wind speed. If the frequency of variation in velocity is ~1 Hz, the transport rate tracks the speed of the wind, but it does not follow higher frequency variations (Butterfield, 1999a). In the field, wind speeds and directions usually fluctuate wildly so that sand may be moving only part of the time. In one field experiment, saltation was active for only 26% of the time (Stout and Zobeck, 1997; Figure 1.9).

Shapes, densities and mixtures of size

At low u_*, platy grains (as of shell sands) have higher transport rates than more rounded ones, but at low u_*, platy grains have markedly lower transport rates, size, density and sorting being held constant (Willetts *et al.*, 1982) (Figure 1.10).

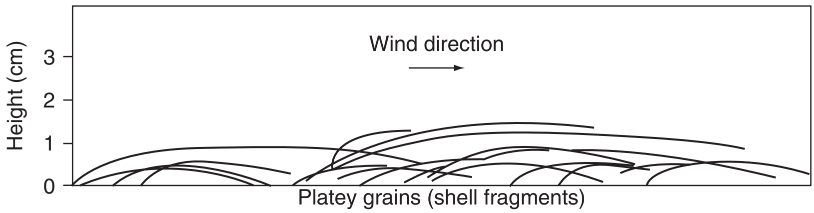

Figure 1.10 Trajectories of platy saltons (Willetts *et al.*, 1982). Reprinted with permission from John Wiley & Sons.

There are many more dimensions to the shape of particles, each having an effect on transport in the wind. When shapes are complex, as are the shapes of the minute bivalves of the genus *Mya*, the wind, unaccountably, sorts left- from right-curving shells (Cadée, 1992). Denser sands (as in those composed of magnetite) have a markedly lower transport rate at low u_* and a somewhat lower rate at high u_* (size, sorting and shape being held constant; Williams, 1964). In some size mixtures, large saltating grains dislodge smaller ones, in which case, the transport rate of the coarse fraction is higher than that of the mean for the mixed size (Iversen and Rasmussen, 1999b). The long-term development of roundness in windblown sands is discussed in Chapter 9.

Hard surfaces

Bagnold (1941, p. 71) observed that saltation trajectories are higher and longer over hard surfaces, as of rock, or over a surface strewn with pebbles ('desert pavement'), than on sandy surfaces. Hardness is measured by the coefficient of restitution, being the ratio of the velocity of an object before it hits the surface to its velocity on rebound. At the same wind speed, and given a sufficient supply of sand, therefore, more sand is in movement over hard surfaces than over loose sand. There is also a more distinct peak height of flux over the harder surfaces, which increases in height downwind of the windward edge of a patch of gravel (Dong Zhibao and Qian Guangqiang, 2007b).

Rough surfaces

The transport rate is also dependent on the roughness of a surface, as shown in a field experiment at a contrast in roughness between an alluvial plain and a very rough lava field in the Mojave Desert by Greeley and Iversen (1987). When the wind first crossed from the alluvium to the lava field, there was a surge in stress (and hence sand-carrying capacity), after which the stress settled down, but to a higher level than over the smooth alluvium. There was another adjustment where the wind passed back from the rough to the smooth surface. Here, too, the change was at first abrupt, both in shear stress and in transport capacity, but stress and carrying capacity again slowly picked up. The transport rate is further enhanced by the greater air turbulence over the rough surface.

Moisture, temperature and humidity

For all the inherent, some perhaps insuperable, problems of studying the details of the effect of moisture on thresholds (earlier), there have been some constructive studies of its effect on transport rates. Many have shown, for example, that when shear velocities are well above the threshold on a dry surface, the moistness of the surface has little effect on the transport rate. This is partly because of the increase in the drying capacity of the wind. On a Dutch beach, sand flow did not reach near the rate predicted by transport equations from wind-speed measurements, until relative humidity fell below 85% (Arens, 1996b). Moisture is thought to be the main reason that few measurements of sand flux over a wet beach agree with the predictions of mathematical models (Bauer *et al.*, 2009).

Rain

Intense wind-driven rain increases the transport rate by splashing up particles into the path of the wind and by lengthening saltation trajectories (by, on average, three times) (Erpul *et al.*, 2004b). The effect is strongest on saturated surfaces, as on beaches exposed by a fall in the tide, or wet, bare upland peats (Foulds and Warburton, 2007a). In these cases, driving rain substantially enhances sand transport. The effectiveness of the rain-driven process depends on the size of the raindrops, the slope of the surface, the angle at which the raindrops meet the surface (raindrops driven by a wind are more effective than vertically falling rain; Erpul *et al.*, 2005) and the grain size of the sand (Furbish *et al.*, 2007).

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