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Characterizing the height profile of the flux of wind-eroded sediment

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Abstract Wind erosion causes severe environmental problems, such as aeolian desertification and dust storms, in arid and semiarid regions. Reliable prediction of the height profile of the wind-eroded sediment flux is crucial for estimation of transport rates, verification of computer models, understanding of particle-modified wind flows, and control of drifting sand. This study defined the basic height profile for the flux of wind-eroded sediment and the coefficients that characterize its equation. Nine grain-size populations of natural sand at different wind velocities were tested in a wind tunnel to measure the flux of sediment at different heights. The resulting flux profiles resemble a golf club with a small back-turn where the flux increases with increasing height within 20 mm above the surface. If the small back-turns are neglected, the flux profiles can be expressed by an exponential-decay function $q_r(z) = a e^{-bz_r}$, where $q_r(z)$ is the dimensionless relative flux of sediment at height z, which follows the exponential-decay law proposed by previous researchers for aeolian saltation. Three coefficients (a creep proportion, a relative decay rate, and an average saltation height) are proposed to characterize the height profile. Coefficients a and b in the above equation represent the creep proportion and relative decay rate as a function of height, respectively. Coefficient a varies widely, depending on grain size and wind velocity, but averages 0.09. It is suggested that the grain size and wind velocity must be specified when discussing creep proportion. Coefficients a and b are nearly linearly correlated and decrease as grain size and wind velocity increase. The average saltation height (the average height sediment particles can reach) was a function of grain size and wind velocity, and was well correlated with coefficients a and b.

Keywords Wind-eroded flux · Exponential-decay function · Creep proportion · Relative decay rate · Average saltation height

Introduction

Wind erosion is a common phenomenon in arid and semiarid regions, which are extensive and comprise about one-third of the world's area (Lal 1990; Sterk and Raats 1996). It is also active on planets, such as Mars and Venus, and satellite, such as Titan (Greeley and Iversen 1985). The wind erosion gives rise to soil degradation through the loss of fertile topsoil, damages crops by abrasion and burial in sand, reduces atmospheric visibility by the transport of sediments in the air, and creates problematic dune formation by deposition. The problems become even worse as the climate becomes drier and more loose surface soil is exposed to the wind as a result of increasing quantity of adverse human disturbance, such as land reclamation and over-grazing. This is demonstrated by the more frequent dust storms that occur in desertified arid and semi-arid zones.

Sediment movement with the wind has generally been described as being composed of three separate phases: creep, saltation, and suspension (Bagnold 1941; Pye and Tsoar 1990). During creep, a thin surface layer moves by rolling or sliding of particles due to shear stresses or particle impacts, but these particles do not leave the surface. Above the surface-creep zone, particles move by means of saltation. Saltating particles are lifted from the surface by the wind or by impacts from other particles, but are too heavy to be suspended in the air stream and fall back to the soil surface. Finally, in suspension, the lightest particles are suspended in the air stream and move along with it. According to the estimates of Chepil (1945), the proportions of particles in each of the three components range from 5 to 25% for surface creep, from 50 to 75% for saltation, and from 3 to 40% for suspension. It has been suggested that the diameters of the transported particles vary approximately from 500 to 2,000 μm for creep, vary from 50 to 1,000 μm for saltation, and are smaller than 100 μm for suspension (Hudson 1973; Shao 2000). The overlaps in these proportions and diameters indicate that certain particles may be moved by different transport modes, depending on the wind's force (i.e. speed). In blowing sediment, particles usually move in a mixture of all the three modes, depending on the factors that control movement for each combination of particle size and wind speed. Consequently, the movement of sediment particles with the wind is stochastic, with sediment particles following different trajectories and producing variation as a function of height in the horizontal flux. This flux is the product of the concentration of sediment particles in the air and the horizontal velocities of those particles (Anderson and Haff 1988; Liu and Dong 2004; Zheng et al. 2004).

Reliable prediction of height profiles within the flux of sediment is crucial for the estimation of transport rates, verification of computer models, and understanding of particle-modified wind flows and the vertical intensity of aeolian abrasion (Butterfield 1999). It is also the basis for controlling drifting sand because the problems caused by blown sand are related not only to the quantity transported, but also to the vertical distribution of the transported sediment (Wu 2003). Therefore, considerable efforts, including wind tunnel tests, field observations, numerical simulations, and theoretical analyses, have been devoted to characterizing the variation of flux of blown sediment as a function of height since Bagnold's (1936) pioneering work. In addition to the qualitative descriptions in early studies (e.g. Bagnold 1936; Chepil 1945), various theoretical and empirical equations have been developed in subsequent studies by using various methods to characterize the height profile of the flux of aeolian sediment (e.g. Ni et al. 2002).

Theoretical attempts had been performed to define the vertical distribution of the flux of wind-eroded sediment before conducting actual measurements. The mass of suspended flux was found to decay with height following a power function defined by turbulence diffusion theory (Liu 1960; Takeuchi 1980). Kawamura (1948, 1951) performed the first theoretical attempt to define the vertical flux profile of saltating particles. He assumed that saltating particles leave the surface vertically, and that the vertical component of a particle's velocity as it leaves the randomly oriented surface obeys Maxwell's law of statistical distribution. He therefore proposed that the vertical profile of the flux of saltating particles could be given by the following equation:

$$q(z) = N_0\left(\frac{\alpha U}{g}\right)e^{(z/\pi h_0)} \tag{1}$$

where $q(z)$ is the mass flux at height z, N_0 is the number of particles leaving the surface per unit area in a unit time, α is a constant equal to $3\pi\mu d$ (where μ is the viscosity coefficient of the air and d is the particle diameter), h_0 is the average maximum height of a particles' trajectory, and U is the wind velocity, g is acceleration due to gravitation. A similar method was used in subsequent research, but different results were obtained because researchers adopted different assumptions and distribution functions for the lift-off velocity of particles (Anderson and Hallet 1986; Zheng et al. 2004). As a result of differences in these assumptions and functions, most researchers have resorted to experimental studies designed to support the development of empirical equations.

Several researchers (e.g. Znamenskii 1958; Wu and Ling 1965) have used simple indices to characterize the vertical profile of the flux of wind-eroded sediment in their early experimental studies. Based on the measured flux of sediment at different heights within 100 mm above the surface in a wind tunnel, Znamenskii (1958) proposed a sediment flux profile index (the Znamenskii's index):

$$S = \frac{q_{\max}}{\bar{q}} \tag{2}$$

where S is the Znamenskii's index, $q_{\max}$ is the mass flux in the 0- to 10-mm layer, and $\bar{q}$ is the mean mass flux within 100 mm above the surface (with measurements taken in 10-mm increments). He found that S was a function of the total sand transport rate and wind speed. Similarly, Wu and Ling (1965) proposed another sediment flux profile index (the Wu–Ling index) based on their field measurements:

$$\lambda = \frac{q_{10-100}}{q_{0-10}} \quad (3)$$

where λ is the Wu–Ling index, q_{10-100} is the mass flux between 10 and 100 mm above the surface, and q_{0-10} is the mass flux within 10 mm above the surface. They related λ to its sedimentological significance and suggested that wind erosion occurs when $\lambda > 1$, deposition occurs when $\lambda < 1$, and erosion balance deposition when $\lambda = 1$.

Most researchers have characterized the vertical profile of flux of wind-eroded sediment by developing an empirical relationship between the flux of sediment and height based on measured results. This has been achieved by means of many wind tunnel tests (e.g. Williams 1964; White, 1982; Gerety and Slingerland 1983; Greeley et al. 1983; Sørensen 1985; Anderson and Hallet 1986; Dong et al. 2003), and field observations (e.g. Sharp 1964; Wu and Ling 1965; Nickling 1978, 1983; Takeuchi 1980; Rasmussen et al. 1985; Fryrear 1987; Vories and Fryrear 1991; Fryrear and Saleh 1993; Greeley et al. 1996). Many results have supported the use of an exponential-decay function expressed as follows:

$$q(z) = q_0 e^{-bz} \quad (4)$$

where $q(z)$ and q_0 are the mass flux at height z above the bed and at the bed surface ($z = 0$), respectively. The parameter b has a different significance for different authors (Nalpanis 1985; Nalpanis et al. 1993; Dong et al. 2003). However, the exponential-decay function has been questioned by some researchers (Dong and Liu 1997), who believed that the available measurement techniques made it difficult to obtain accurate results, especially in the near-surface layer (Rasmussen and Mikkelsen 1998; Goossens et al. 2000; Dong et al. 2004a).

Some researchers (e.g. Zingg 1953; Chepil and Woodruff 1957; Rasmussen and Mikkelsen 1998; Ni et al. 2002) have instead proposed power functions or modified power functions. For example, the fitting of 43 sets of wind tunnel data for five sand grain sizes in the 0.200- to 0.715-mm range led Zingg (1953) to propose the following function:

$$q(z) = \left(\frac{b}{z+c}\right)^{\frac{1}{n}} \quad (5)$$

where b and c are coefficients that may be constants for a single experiment, and n is the exponent of the power function. In wind tunnel tests, Ni et al. (2002) found that except for measurements within the lowest compartment of the tray array used to collect suspended particles, the vertical distribution of mass flux was well represented by an equation of the following form:

$$q(z) = a\left(\frac{1}{z} - \frac{1}{H}\right)^{n} \quad (6)$$

where a is a coefficient, H is the maximum height of the saltation layer, and n is the exponent.

Other researchers (e.g. Zobeck and Fryrear 1986; Fryrear 1987; Vories and Fryrear 1991; Stout and Zobeck 1996; Sterk and Raats 1996; Hasi 1997; Butterfield 1999) suggested a combined mass-flux model that combines a power function or modified power function with an exponential function to describe the vertical distribution of horizontal mass fluxes between the soil surface and a given height. For example, Sterk and Raats (1996) proposed:

$$q(z) = a\left(\frac{1+z}{\alpha}\right)^{-b} + ce^{(-z/\beta)} \quad (7)$$

where coefficients a, b, c, and β can be defined using empirical data. The first term at the right is considered to describe the suspension mass flux, whereas the second term is considered to represent the saltation flux, including creep (Fryrear and Saleh 1993).

Many factors, such as wind speed, particle size, and the characteristics of the underlying surface, influence the height profile of the flux of aeolian sediment (Zou et al. 1981, 1986; Ma 1988; Yin 1989; Han et al. 1995; Zou et al. 1995; Dong et al. 2004b). Addressing the significances of these factors in descriptions of flux profiles will provide more information about the drift of blowing sediment. The objectives of the present study were to use wind tunnel tests to investigate the flux profile for sediment blowing over a loose sandy surface, and to define the influence of wind velocity and particle size on the flux profile.

Materials and experimental methods

To characterize the height profiles of the flux of wind-eroded sediment, experiments were conducted in the wind tunnel at the Shapotou Desert Experimental Research Station, Key Laboratory of Desert and Desertification, Chinese Academy of Sciences. This blow-type non-circulating wind tunnel has a total length of 37 m and is composed of a power section (2.60 m long), expansion section (6.40 m long), stabilization section (1.50 m long, where drag screens and honeycombs are set to reduce large-scale eddying), compression section (2.50 m long), working section (21 m long), and diffusion section (3 m long). The maximum cross-sectional area in the expansion section is 2.4×1.2 m^2. The 21-m-long working section has a cross-sectional area of 1.2×1.2 m^2. Wind speed can be changed continuously from 2 to 30 m s^{-1}. The depth of the boundary layer can reach 0.4 to 0.5 m in the working section.

Natural sand from the Tengger Desert was screened into nine grain-size populations for the tests: 100 to 150, 150 to 200, 200 to 250, 250 to 400, 400 to 500, 500 to 560, 560 to 630, 630 to 800, and 800 to 1,000 μm. The prepared sand samples were placed in a 4-m-long, 0.8-m-wide, and 25-mm-deep tray, whose surface was placed level with the wind tunnel floor, and then were exposed to the required free-stream wind velocity (measured at the centerline height of the wind tunnel, 0.6 m above the tunnel floor, using a Pitot static probe) above the threshold for initiation of sand movement. The length of the sand tray was chosen to ensure a significant development of the saltation cloud.

The horizontal flux of sediment at 10-mm intervals to a total height of 0.60 m was measured using a WITSEG sampler (for more details of the sampler, refer to Dong et al. 2004a). In this experiment, the sampler was positioned 16 m downwind from the entrance of the working section of the wind tunnel, 50 mm downwind from the sand tray. The bottom of the lowest opening of the sampler was set flush with the tunnel floor. Because it took some time for the wind tunnel to reach a pre-set free-stream wind velocity, an automatic sliding lid was used to cover the sand tray until the required wind speed had been reached; this prevented the sand from being blown away prematurely. The wind tunnel was powered off when about 20 mm of the top sediment in the tray had been blown away. For each sample at each free-stream wind velocity, three repetitions were performed to obtain the mean values.

Results and discussion

Flux profiles

We focused mainly on the variation in the *pattern* of the sediment mass flux as a function of height rather than the actual magnitude of the flux. To do so, the measured mass fluxes at different heights and the height itself were normalized into dimensionless relative values using the following transformation:

$$q_r(z) = \frac{q_m(z)}{Q}, \quad z_r = \frac{z}{Z} \tag{8}$$

where $q_r(z)$ is the dimensionless relative flux, which represents the portion of the flux at height z divided by the total flux, $q_m(z)$ is the measured flux of sediment at height z, z_r is the dimensionless relative height (i.e. the height of the measurement divided by the total height), z is the actual height, Z is the maximum measurement height of the sampler (equal to 0.6 m in this study), and Q is the total flux of sediment, which equals the sum of the measured flux of sediment at all heights (Table 1).

Figure 1 shows the variation with height of the proportion of total flux at different free-stream wind velocities for different grain-size populations. In general, the flux of sediment decreases rapidly with increasing height. The curves are similar in all cases, with a shape that resembles a golf club. The back-turn of the curves in the lowest layer (below 20 mm) represents a section in which flux increases with an increase in height. We are not sure whether this phenomenon results from measurement error or the inherent nature of the flux profile. This feature of the curves has been ignored in most previous studies. For example, Yin's (1989) observations over a gravel gobi surface showed a flux increment section in the near-surface layer, and wind tunnel tests of Dong et al. (2004b) verified the existence of a flux-increment section in the near-surface layer over a gravel surface. Based on this evidence, the phenomenon is considered to be an inherent property of the flux profile over a gravel surface because saltation is more intense over a gravel surface, resulting in greater saltation height. Recently, numerical simulations of Zheng et al. (2004) revealed a stratified flux-profile pattern composed of a linear increment layer, a saltation layer, and a monotonic decrement layer. However, they did not specify the kind of ground surface their conclusions applied to. The reconstructed flux profiles of Dong et al. (2006) for aeolian sediments in a wind tunnel based on the particle-image velocimetry technique showed similar back-turns in the lowest layer. It is possible that the flux increases with increasing height in the lowest layer because the velocity of the saltating particles increases rapidly with increasing height within that layer. However, direct measurement to verify the flux increment section is difficult because it occurs very close to the surface.

The curves in Fig. 1 also show that the flux decreases more rapidly as wind velocity decreases. This is qualitatively in agreement with previously reported results (Znamenskii 1958; Wu and Ling 1965; Chen et al. 1996). Znamenskii's (1958) wind tunnel results showed that within the 100 mm nearest to the surface, the proportion of the total sediment decreased with increasing wind speed within the 0- to 10-mm layer, was independent of wind speed within the 10- to 40-mm layer, and increased with increasing wind speed in the layers above 40 mm. These results suggest that blown sediment is transported higher as the wind speed increases, similar to Bagnold's (1941) conclusion. The field observations of Wu and Ling (1965) in shifting dune fields in Xinjiang (China) indicated that an increase in the wind speed decreases the relative proportion of the sediment flow in the lower layers, but increases the proportion in the higher layers. The wind tunnel tests and field observations of dune sands of

Table 1 Correlation between normalized flux (flux at a given height divided by total flux) and normalized height (actual height divided by maximum measurement height)

Serial no.	d (μm)	U (m s^{-1})	Q (kg m^{-1} s^{-1})	a	b	r^2
01	100–150	8	0.023	0.2948	14.23	0.91
02		10	0.050	0.2120	10.54	0.92
03		12	0.114	0.1883	9.54	0.92
04		14	0.247	0.2028	10.45	0.95
05		16	0.442	0.1957	10.24	0.95
06		18	0.701	0.1722	9.00	0.89
07		20	1.077	0.1714	9.20	0.96
08		22	1.567	0.1533	8.29	0.95
09	150–200	8	0.028	0.2195	11.28	0.98
10		10	0.055	0.1660	8.75	0.98
11		12	0.126	0.1398	7.48	0.98
12		14	0.265	0.1382	7.47	0.98
13		16	0.438	0.1323	7.14	0.97
14		18	0.641	0.1250	6.77	0.96
15		20	0.934	0.1133	6.10	0.92
16		22	1.317	0.1095	5.87	0.92
17	200–250	10	0.050	0.1064	5.69	0.97
18		12	0.116	0.0877	4.93	0.98
19		14	0.221	0.0841	4.67	0.98
20		16	0.361	0.0831	4.67	0.99
21		18	0.512	0.0755	4.20	0.99
22		20	0.741	0.0750	4.19	0.98
23		22	1.074	0.0711	3.94	0.97
24	250–400	10	0.047	0.0846	4.57	0.97
25		12	0.091	0.0723	3.96	0.98
26		14	0.171	0.0684	3.79	0.98
27		16	0.293	0.0647	3.59	0.98
28		18	0.433	0.0633	3.53	0.98
29		20	0.664	0.0624	3.50	0.99
30		22	1.022	0.0574	3.18	0.98
31	400–500	10	0.040	0.0768	4.15	0.96
32		12	0.086	0.0654	3.61	0.98
33		14	0.152	0.0621	3.44	0.99
34		16	0.250	0.0588	3.26	0.98
35		18	0.383	0.0533	2.93	0.98
36		20	0.547	0.0499	2.71	0.97
37		22	0.709	0.0481	2.59	0.98
38	500–560	12	0.067	0.0563	3.05	0.97
39		14	0.117	0.0527	2.86	0.98
40		16	0.191	0.0481	2.59	0.97
41		18	0.307	0.0489	2.61	0.99
42		20	0.425	0.0436	2.30	0.98
43		22	0.593	0.0440	2.29	0.99
44	560–630	12	0.046	0.0487	2.52	0.97
45		14	0.097	0.0480	2.53	0.99
46		16	0.161	0.0437	2.30	0.97
47		18	0.269	0.0419	2.19	0.98
48		20	0.410	0.0407	2.10	0.98
49		22	0.473	0.0373	1.85	0.98
50	630–800	14	0.041	0.0372	1.83	0.95
51		16	0.122	0.0410	2.11	0.97
52		18	0.188	0.0418	2.17	0.98
53		20	0.294	0.0418	2.19	0.98
54		22	0.374	0.0421	2.21	0.99
55	800–1,000	16	0.038	0.0384	1.93	0.98
56		18	0.149	0.0388	1.96	0.99
57		20	0.208	0.0387	1.96	0.99
58		22	0.310	0.0385	1.95	0.99

d represents the particle diameter, U is the free-stream wind velocity, Q is the total flux of sediment, and r^2 is the correlation coefficient at $P = 0.05$. The fitted function was $q_r(z) = ae^{-bz}r$, where $q_r(z)$ is the dimensionless relative flux (the proportion of the total flux at height z), z_r is the dimensionless relative height, a and b are regression coefficients

Chen et al. (1996) in China's Taklimakan Desert demonstrated that as the wind speed increased, the higher flow layers experienced the most rapid increase in their mass flux rate.

The difference between the flux profile curves at different wind velocities narrows as grain size increases. The curves nearly merge, and the back-turns in the lowest layer nearly disappear, for the 630- to 800-μm

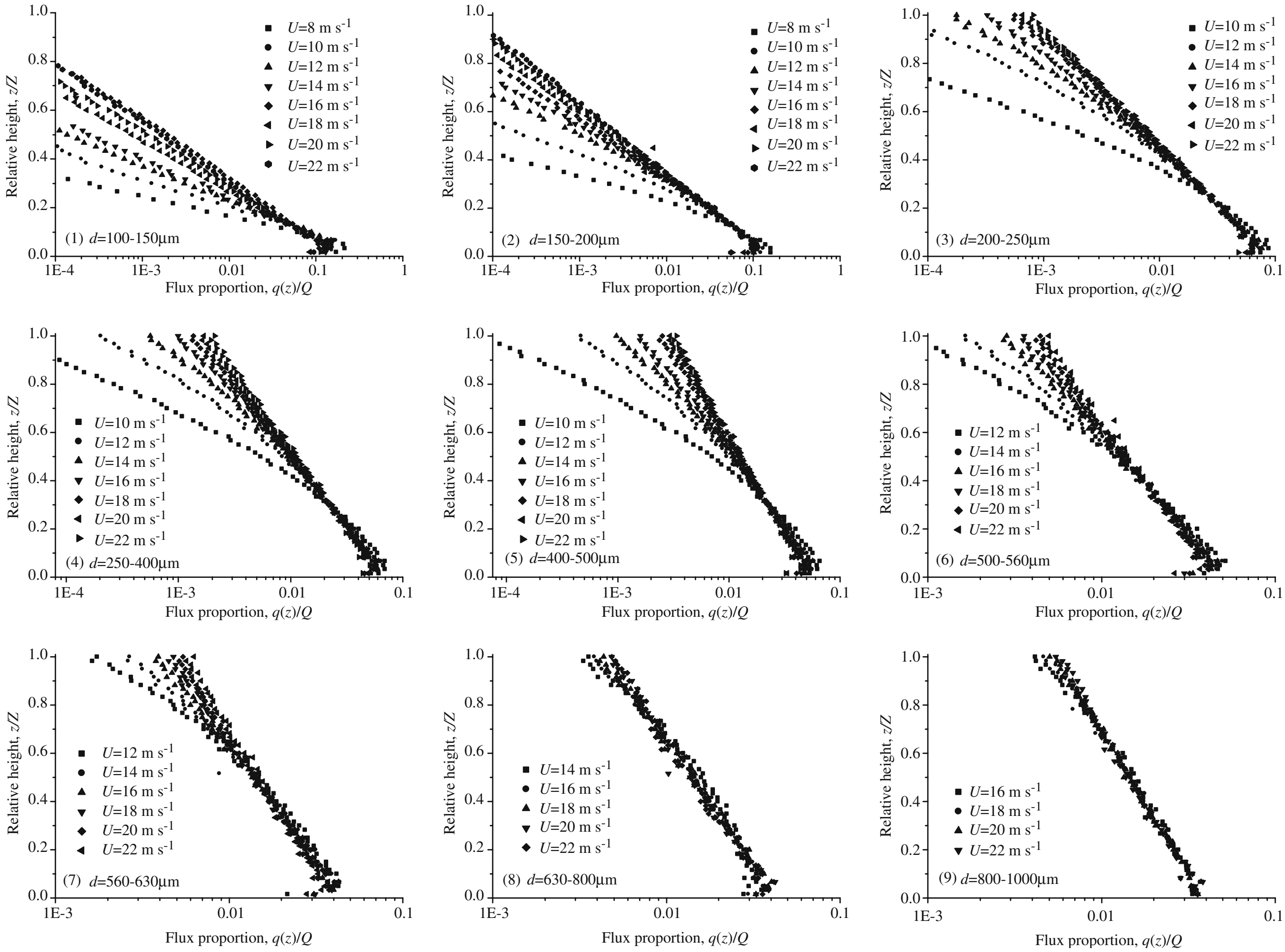

Fig. 1 The variation with height of the dimensionless flux of sediment for different grain-size populations and wind velocities

and 800- to 1,000-μm grain-size populations. This means that as the total sediment transport increases, the flux of sediment increases, with a relatively more constant proportion between heights for the 630- to 800-μm and 800- to 1,000-μm populations, whereas a relatively larger proportion is transported at greater heights for the smaller grain-size populations. The smaller the grain size, the greater the proportion that is added to greater heights above the surface as total transport increases, suggesting that the vertical movement of smaller particles is more sensitive than that of the larger particles to changes in wind speed.

If we neglect the back-turn in the flux profile very near the surface (because the flux increment layer is very thin compared with the total measurement height, at around 3% of the total), regression analyses reveal that for all grain-size populations and wind speeds, the dimensionless relative flux of sediment follows an exponential-decay function with respect to height:

$$q_{\mathrm{r}}(z) = a\mathrm{e}^{-bz_r} \tag{9}$$

where a and b are regression coefficients. Mathematically, a represents the proportion of the flux at the surface ($z = 0$), and thus represents the creep proportion; b defines the slope of the straight lines for $\ln[q_{\mathrm{r}}(z)] - z_{\mathrm{r}}$ and thus represents the relative decay rate of the dimensionless relative flux as a function of height. The coefficients a and b are listed in Table 1, in which the correlation coefficients between the relative flux of sediment and height are generally good or excellent. In more than 90% of the cases, the squared correlation coefficient (r^2) is greater than 0.95.

Combining Eqs. (8) and (9) produces the following equation:

$$q(z) = q_0\mathrm{e}^{-b'z} \tag{10}$$

where $q_0(z) = aQ$ is the flux of sediment at the surface ($z = 0$; i.e. the creep flux), and $b' = b/Z$. Eq. (10) is similar to Eq. (1) and has the same form as Eq. (4). Thus, the flux of wind-eroded sediment decays exponentially with height.

Creep proportion and relative decay rate

The height profile for the flux of wind-eroded sediment that we developed [Eqs. (9) and (10)] contains two important coefficients: a and b, which represent (respectively) the creep proportion and the relative decay rate. If these two coefficients are known, the flux profile can be reconstructed. This section discusses these two coefficients.

The creep proportion is important in estimating the total sediment transport. Because this flux is difficult to measure, the total sediment transport is usually estimated by measuring the saltating and suspended fluxes and assuming the creep proportion. To develop his sand transport equation, Bagnold (1941) proposed a creep proportion of 0.25 in his classic work. However, according to Chepil (1945), creep proportions vary from 0.05 to 0.25. In our tests, the flux of sediment above the sampler's measurement height was assumed to be negligible because the maximum flux percentage (for the 800- to 1,000-μm grain-size population at 22 m s^{-1} free-stream wind velocity) in the top flow layer (0.6 m above the surface) amounted to only 0.58% of the total. The total flux measured within the 0.6-m layer can thus be used to represent the total flux of sediment, and the creep proportion discussed in this section can be regarded as the proportion of the total flux of sediment accounted for by creep. Figure 2 shows that this proportion varies widely, ranging from about 0.04 to 0.29, depending on wind speed and grain size. This range is slightly wider than the range estimated by Chepil (1945), but very different from the single value of 0.25 proposed by Bagnold (1941). Thus, it is clear that we must specify the grain size and wind velocity when discussing the creep proportion. The overall average for the creep proportion was about 0.09, but this value generally decreased with the increasing grain size, suggesting that relatively more smaller grains are transported in the air than by the creep. The average creep proportions for the 100–150, 150–200, 200–250, 250–400, 400–500, 500–560, 560–630, 630–800, and 800–1,000 μm grain-size populations are 0.200, 0.140, 0.083, 0.068, 0.059, 0.049, 0.043, 0.041, and 0.039, respectively. The creep proportion tends to decrease with increasing wind speed, except for the 630- to 800-μm and 800- to 1,000-μm grain-size populations, for which the creep proportion remains nearly constant as wind velocity changes; this suggests that more grains are transported at higher levels as the wind velocity increases. The smaller the grain size, obviously more the creep proportion decreases with increasing wind velocity, which suggests that the vertical movement of smaller grains is more sensitive to changes in wind velocity. The creep proportion for the 560- to 630-μm grain-size population decreased from 0.049 at 12 m s^{-1} wind velocity to 0.037 at 22 m s^{-1} wind velocity (i.e. a relatively small decrease), whereas that for the 100- to 150-μm size population decreased from 0.29 at 8 m s^{-1} wind velocity to 0.15 at 22 m s^{-1} wind velocity (i.e. a relatively large decrease of nearly 50%).

Eq. (9) can be rewritten as follows:

$$\ln(q_r(z)) = a' - bz_r \tag{11}$$

where $a' = \ln(a)$, and b is the slope of the straight lines for $\ln[q_r(z)] - z$, and this coefficient defines the relative decay rate function for the flux of sediment as a function of height. Figure 3 shows that coefficient b in Eq. (9) decreases with increasing grain size and increasing wind speed, with a trend similar to that for coefficient a (the creep proportion). To some extent, the creep proportion and the relative decay rate reflect two aspects of the same phenomenon, because transport of more grains at greater heights decreases both the creep proportion and the relative decay rate. In fact, a and b are almost linearly related to each other, with a high correlation coefficient (Fig. 4).

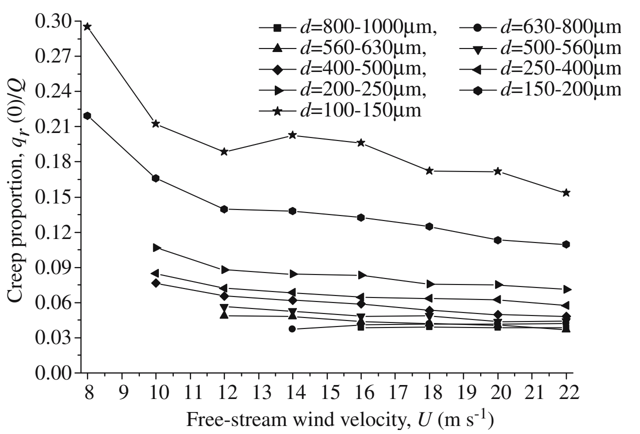

Fig. 2 The proportion of total flux accounted for by creep for different grain-size populations and different wind velocities

Average saltation height

The average saltation height is often used to characterize aeolian saltation clouds, but different methods have been used to define this height. Usually, the height at which 50% of the total flux of sediment from the top of the surface is defined as the average saltation height (Zingg 1953; Wu 2003). Zingg (1953) found that the average saltation height was determined by the grain size and "wind stress", which is proportional to wind speed, and proposed the following empirical relationship:

$$z_a = 7.7d^{3/2}\tau^{1/4} \tag{12}$$

where z_a is the average saltation height (in inches), d is the sand's grain size (in mm), and τ is the shear stress on

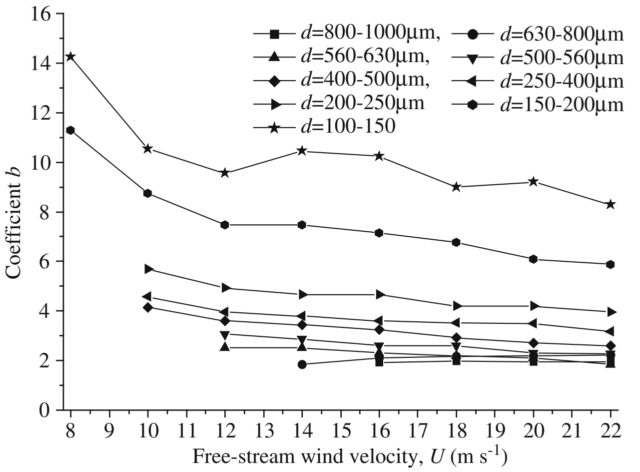

Fig. 3 Variation in coefficient b (the decay rate) as a function of grain size and wind velocity

the ground surface (lb in.$^{-2}$). Equation (12) indicates that the greater the grain size and wind velocity, the greater the average saltation height.

Equation (13) is used here to define the average saltation height on a more mathematically sound basis:

$$z_a = \frac{\int_0^\infty q(z)z\mathrm{d}z}{\int_0^\infty q(z)\mathrm{d}z}, \text{ or } z_m = \frac{\int_0^\infty q_r(z)z_r\mathrm{d}z_r}{\int_0^\infty q_r(z)\mathrm{d}z_r} \tag{13}$$

where z_m is the dimensionless relative average saltation height (i.e. the actual height expressed as a proportion of the total range of heights being studied). The average saltation heights obtained using Zingg's method and Eq. (13) are very well correlated (Fig. 5), but the latter method produces a greater average saltation height (30% greater on average). Figure 6 shows the dimensionless average saltation height calculated using Eq. (13). The average saltation height generally increases with increasing grain size. To develop a relationship between average saltation height and grain size for each grain-size population, the mean values of the average saltation height for all wind speeds were obtained using the combined data in Fig. 6. The mean average saltation heights for the 100–150, 150–200, 200–250, 250–400, 400–500, 500–560, 560–630, 630–800, and 800–1,000 µm grain-size populations were 0.10, 0.14, 0.21, 0.24, 0.27, 0.30, 0.33, 0.34, and 0.35, respectively. The relationship between mean average saltation height and the midpoint of the grain-size population can be expressed by:

$$z_m = 0.38 + 0.13\ln(d_m), \quad r^2 = 0.98 \tag{14}$$

where d_m is the midpoint of the grain-size population (in mm). Except for the 630- to 800-µm and 800- to 1,000-µm grain-size populations, the average saltation height increases with increasing wind velocity. The average saltation height shows the opposite trend compared to the creep proportion and relative decay rate with respect to grain size and wind velocity because increasing the average saltation height means that grains are transported to greater heights, thereby decreasing the creep proportion and the relative decay rate.

Figure 7 shows that the coefficients a and b are both closely related to the average saltation height, with a high correlation coefficient in both cases. The resulting regression equations are as follows:

$$a = \frac{0.022}{z_m} - 0.023, \quad r^2 = 0.999 \tag{15a}$$

$$b = \frac{5.54}{\sqrt{z_m}} - 7.49, \quad r^2 = 0.997 \tag{15b}$$

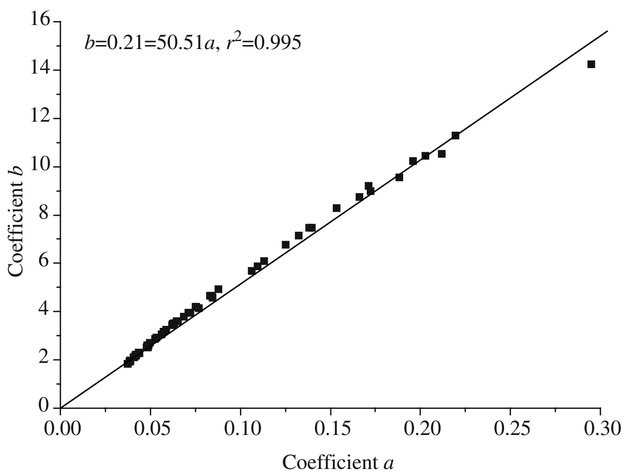

Fig. 4 The relationship between coefficients a (the creep proportion) and b (the decay rate)

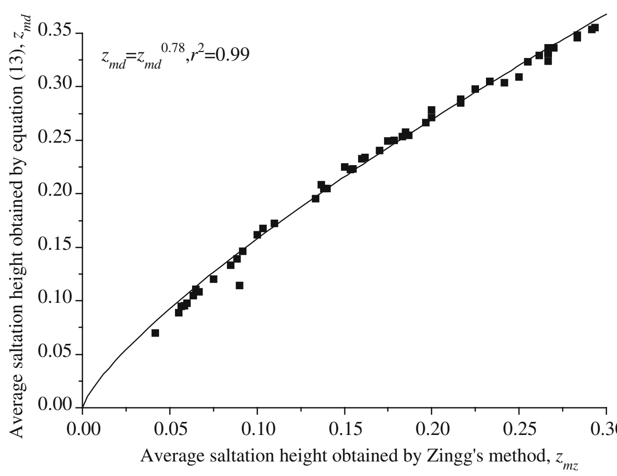

Fig. 5 Comparison of the average saltation heights obtained using Zingg's (1953) method and Eq. (13) from the present study

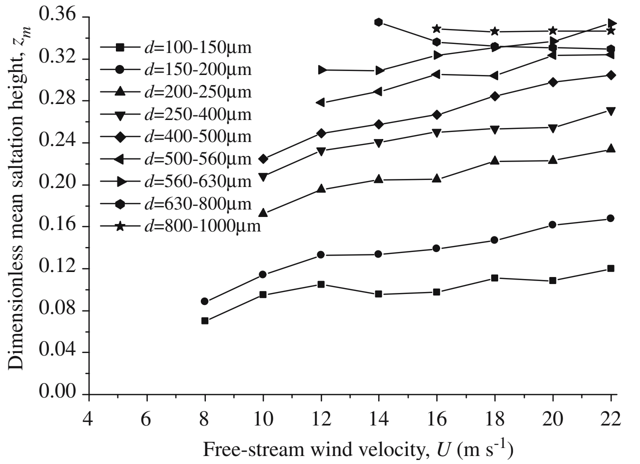

Fig. 6 The average saltation height obtained using Eq. (13) for different grain-size populations as a function of wind velocity

Based on this analysis, the average saltation height is clearly an important variable in characterizing the height profiles of the flux of wind-eroded sediment. This conclusion is similar to those of Kawamura (1948) and Nalpanis (1985), who suggested that the flux of sediment followed an exponential-decay curve with respect to height and related the proportionality coefficient before height z to the maximum height or other characteristic height of saltation.

Conclusions

In this study, a height profile for the flux of wind-eroded sediment has been developed based on detailed wind tunnel tests of different grain-size populations at different wind speeds. The resulting flux profiles on the $\ln[q_r(z)] - z_r$ plots have the shape of golf club, with small back-turns at the lowest level (within 20 mm above the surface) for which the flux increases with increasing height. If these back-turns are neglected, the flux profiles can be expressed as exponential-decay functions, in agreement with the exponential-decay law proposed by many previous researchers for the drift of wind-eroded sediment. However, the results indicate that the relative decay rate differs for different size populations and wind speeds. In general, the flux decreases more rapidly as the grain size and wind velocity decreases. The vertical movement of smaller grains is more sensitive to changes in wind velocity than is the case for larger grains. Given that the transport fraction in the top flow layer (0.6 m) in our tests was so small (less than 0.58% of the total transport), sand transport above 0.6 m could be neglected and the exponential-decay law should be a good descriptor of the flux profile of wind-eroded sediments in which saltation is the primary mode of grain movement. Whether the exponential-decay function remains valid for fine materials, such as dust, that move primarily in suspension needs further study.

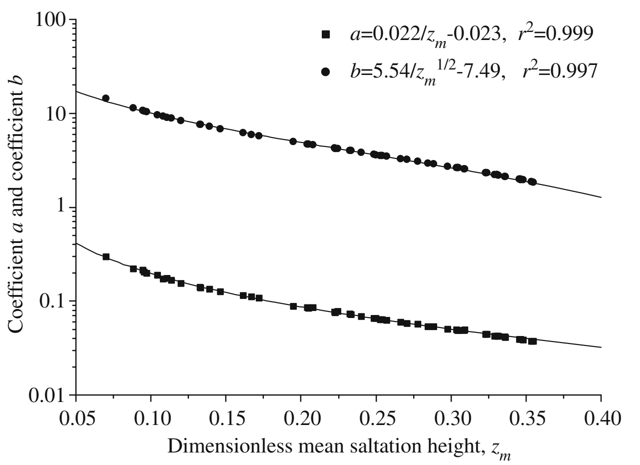

Fig. 7 The variation in average saltation height as a function of coefficients a and b

The exponential-decay function for the flux of wind-eroded sediment has two coefficients. Coefficient a represents the creep proportion and coefficient b defines the relative decay rate of the flux of sediment as a function of height. These coefficients, which reflect two aspects of the same phenomenon, are almost linearly correlated, as both define the relative proportion of the flux that is transported at different heights. The creep proportion varies widely as a function of grain size and wind velocity, with a range of 0.04 to 0.29 and an overall average of 0.09. This is close to the range estimated by Chepil (1945), but very different from Bagnold's fixed estimate of 0.25, which has been widely applied in developing aeolian transport models. It is suggested that the grain size and wind velocity must be specified when discussing the creep proportion.

The average saltation height has also been used to characterize the height profile of the flux of wind-eroded sediment. This height is strongly correlated with both coefficients (a and b) in the exponential-decay function for flux. The average saltation height is a function of grain size and wind velocity. In summary, the height profile of the flux of wind-eroded sediment can be characterized by three coefficients: the creep proportion, relative decay rate, and average saltation height.

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