Chapter 20

Dust

Joanna E. Bullard and Ian Livingstone

Introduction

In aeolian systems there is a fundamental difference between the behaviour of coarse sediments (sands) and fine sediments (dusts). Sand-sized material (63–2000 μm) travels predominantly by saltation, reptation and creep within the lowest levels of the atmospheric boundary layer (<3 m above the surface) and travels short distances. In contrast, dust-sized material, generally defined as <63 μm, is transported in suspension at a wide range of heights above the surface and can rapidly travel considerable distances. Dust plumes disperse as they travel away from source diffusing the concentration of sediment. Consequently, although there are clearly definable sources for dust emissions, these finer particles can be transported around the globe and their deposits can be both far removed from their origins and extensive. Every year up to three billion tonnes of dust are released into the atmosphere from the Earth's surface. The spatial and temporal patterns of these dust emissions are often closely controlled by desert geomorphology, and in turn have an impact both directly and indirectly on the desert landscape and further afield.

Some of the research on dust has been driven by environmental concerns which have included issues associated with health (as in the 'Dust Bowl' of the 1930s in the Great Plains of the USA) and with climate change because dust in the atmosphere affects radiation budgets and may, through iron fertilisation, enhance carbon dioxide uptake in the oceans. Mineral dust in the atmosphere can cause cooling (via scattering) or warming (via absorption) and its impact depends on factors such as mineralogy, chemical composition and particle size, each of which is strongly-correlated with dust source characteristics, as well as the position of the particles within atmospheric layers. There has also been considerable interest in the movement of dust in geomorphological systems. These impacts include, for example: deflation of dust-sized material leading to the formation or enhancement of topographic lows such as playas; ventifact and yardang formation; stone pavement formation; duricrust and rock varnish development. And the movement of dust has an impact on soils, for the deflation of dust-sized material leads to a loss of soil nutrients, usually concentrated at the surface, and a reduction in soil moisture-holding capacity at source, but can enhance soil nutrients in areas of deposition.

Characteristics of Dust Particles

Airborne dust is derived from a range of sources, both natural and anthropogenic, and the former constitutes 89% of global emissions (Satheesh and Moorthy 2005). It is becoming increasingly important to be able to distinguish rates of natural dust entrainment and its impacts from human-induced entrainment (such as mechanised agriculture and off-road vehicle use) or anthropogenically-generated dusts (such as industrial emissions), due to the impact of the latter on the physical and chemical properties of the atmosphere. This chapter focuses principally on

J.E. Bullard (✉)
Department of Geography, Loughborough University of Technology, Loughborough, LE11 3TU, UK
e-mail: J.E.Bullard@lboro.ac.uk

A.J. Parsons, A.D. Abrahams (eds.), *Geomorphology of Desert Environments*, 2nd ed.,
DOI 10.1007/978-1-4020-5719-9_20, © Springer Science+Business Media B.V. 2009

natural dust. The dominant material comprising dust is SiO_2, usually in the form of quartz, but it can also include feldspars, calcite, halite and organic material such as pollen and diatoms.

Most sedimentological definitions of dust include all particles below a maximum size in the range 60–70 μm and consequently include the both the silt (4–63 μm) and clay (<4 μm) classes of the Udden-Wentworth grain-size scale as well as occasionally part of the very fine sand class (63–125 μm) (Wentworth 1922). In terms of aeolian geomorphology, the important distinction between 'sand' and 'dust' is the way in which the particles are transported. This difference in behaviour (and hence the precise size boundary) is governed by the balance between the forces holding particles aloft in the atmosphere and those pulling them towards the ground surface.

If the rate at which gravitational forces are pulling the particle towards the surface (its terminal velocity – w_t) is the same as, or less than the mean velocity at which the air parcel within which the particle is contained is moving upwards by atmospheric turbulence (the Lagrangian vertical velocity), then the particle is carried in suspension. In a thermally neutral atmospheric surface layer, the Lagrangian vertical velocity is κu_*, where κ is von Karman's constant (0.4) and u_* is the friction velocity (Hunt and Weber 1979). Particles travel in suspension when $\mid w_t \mid / \kappa u_* << 1$ and travel by saltation or creep when $\mid w_t \mid / \kappa u_* >> 1$. The critical size boundary between 'sand' and 'dust' (d_1) can be calculated by solving $\mid w_t \mid /(d_1) = \alpha_d \kappa u_*$. The value of α_d makes a significant difference to the position of the boundary (Fig. 20.1) but for dust entrainment studies is commonly 0.5, which means that once dust particles are lifted from the surface they are likely to remain suspended for some time (Shao et al. 1996, Shao 2000). This balance between upward and downward forces means that larger particles can be considered to be dust particles in stronger and more turbulent airflows whilst in weaker or less turbulent air flows the same sized particle might travel in saltation or by creep. It may also help to explain why particles far larger than 63 μm can occasionally be found in dust deposits located far from their source – for example the >75 μm quartz particle transported in a dust storm from Asia to Hawaii (Betzer et al. 1988).

In reality, when close to the sediment source, any body of sediment transported by the wind is likely to comprise a mix of saltating, creeping and suspended

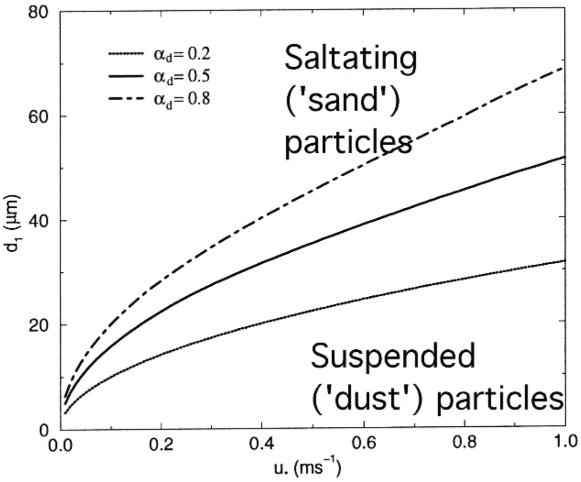

Fig. 20.1 The impact of variations in α_d on the size boundary between saltating (sand) particles and particles travelling in suspension (dust) (modified from Shao 2000)

particles and also to include particles travelling in a mode part way between pure suspension and pure saltation in what has been called modified suspension (e.g. Nalpanis 1985) where particle trajectories are influenced by both their inertia and settling velocity (see Nickling and McKenna Neuman, this volume). With increasing distance from the source, coarser particles will be deposited and only true suspended particles will remain in the air mass. Pye (1987) suggested that particles in the size range 20–70 μm will travel in short-term suspension or by modified saltation (for example moving up to 30 km from source) whilst those <20 μm are likely to travel in long-term suspension (travelling thousands of km from source). Tegen and Fung (1994) estimate that the majority of contemporary dust emissions are in the size range 1–25 μm (Table 20.1).

The forces acting to transport or stabilise particles are discussed in chapter 17 and include cohesive, gravitational and aerodynamic forces. The relative importance of each of these is dependent upon the size of the particles: below 10 μm cohesive forces tend to dominate, for particles in the range 10–300 μm

Table 20.1 Estimates of global dust emissions by size (Tegen and Fung 1994)

Particle-size range	Global dust source strength
0.5–1 μm	390 Mt a^{-1}
1–25 μm	1960 Mt a^{-1}
25–50 μm	650 Mt a^{-1}

aerodynamic forces dominate, and >300 μm gravitational force dominates. For this reason the majority of wind-transported sediments are in the 10–300 μm category. However, finer particles can be transported because the dominance of cohesive forces means that there is a tendency for fine particles to aggregate together rather than to act as individual grains. Aggregation can occur when fine silts and clays attach themselves to coarser sand-sized particles as individual adhesions or to form thin layers or coatings around them. In soils with a high clay content dust-sized particles are usually aggregated together. Under low wind speeds aggregates and dust-coated grains behave like coarser particles – the size of the aggregate. In higher wind speeds, aggregates break up, releasing the individual dust particles into suspension (Gomes et al. 1990). It is very difficult to quantify the degree of aggregation of particles, or the transport-stable size of aggregates for different wind speeds because methods for establishing particle size distributions tend to break-up the aggregates. An approximation can be obtained by examining the difference between minimally-dispersed and fully-dispersed sample treatments – for example Shao (2000) demonstrated a shift in modal particle size from ≈100 μm to ≈4 μm following fully-dispersed analysis of clay soil (Fig. 20.2).

In addition to size, the shape of dust particles can be important in terms of entrainment, transport and deposition, and is increasingly recognised as a key factor affecting the radiative properties of mineral dust particles in the atmosphere (e.g. Kalashnikova and Sokolik 2004, Meloni et al. 2005). The terminal velocity (w_t) of dust depends on the particle's mass and shape and these factors can affect the size of particles that can be transported in suspension. Spherical particles are streamlined and consequently have higher settling velocities for a given mass than platy particles. Natural dust particles are rarely perfectly spherical and it is more common for particles to be platy or 'flattened' (Baba and Komar 1981, Le Roux 2002). The shape of a particle is linked to lithology but also to the mechanism by which it was formed – for example crushing or grinding results in blade-shaped particles (Assallay et al. 1998). For clay-sized sediments the effective particle shape is determined by aggregation. As the number of particles comprising the aggregate increases, its overall shape will become more irregular and there will be a concomitant decrease in density (Goossens 2005). Wind tunnel experiments have shown marked differences between measured and predicted rates of dust deposition depending on whether or not grain shape has been taken in to account. For example, using sub-angular to sub-rounded dust particles with a Corey Shape Factor of 0.5–0.9 (blocky to near-spherical), Goossens (2005) demonstrated that predicted mass deposition flux was closest to the measured rates when grain shape was taken into account, although the accuracy of the prediction varied considerably for individual sediment size fractions.

Mechanisms of Dust Production

Whilst the direct release of particles from fine-grained rocks might be a locally-important source of silt-sized material, the mean grain size of quartz in igneous and metamorphic rocks is around 700 μm compared with

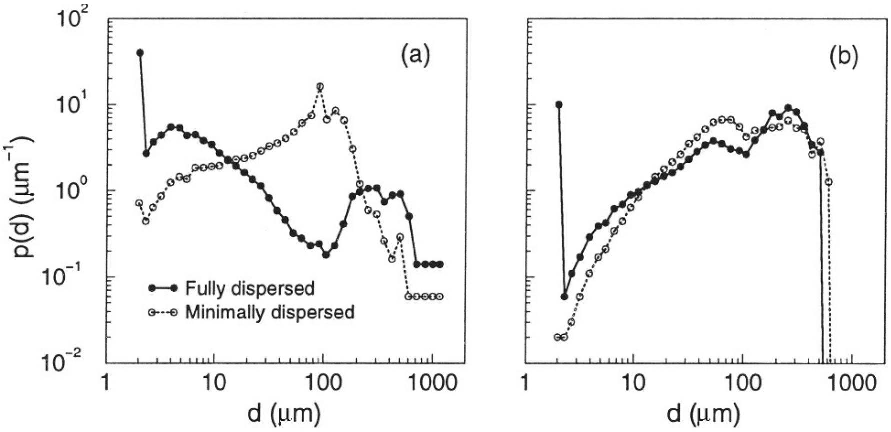

Fig. 20.2 The difference between fully-dispersed and minimally-dispersed particle-size analysis may indicate the presence of aggregation in clay soils (**a**) and results in a shift in modal grain size from ≈100 μm to ≈4 μm. The two treatments have very little impact on the particle-size distributions of sandy soils (**b**) in which few aggregates would be expected (redrawn from Shao (2000) using unpublished data from McTainsh)

a mean size of detrital quartz of just 60 μm. Particles therefore need to undergo processes that can reduce their size by about 90% before they can be carried as suspended dusts. This size reduction can be achieved by the fracturing of whole particles into smaller grains, the removal of corners and protuberances from larger grains, usually as a result of physical stresses and impacts, or the gradual reduction of the grain size by rubbing or grinding against larger particles or more resistant surfaces. These processes are not mutually exclusive, for example the components of a fractured particle may subsequently undergo rounding or grinding (Fig. 20.3).

The main geomorphological processes proposed for silt-sized particle production are physical and chemical weathering, glacial grinding, fluvial comminution and aeolian abrasion, whilst chemical weathering is thought to be the dominant primary source of particles $<4\,\mu m$ in diameter (Pye 1987). Some of these mechanisms are far more important in deserts than others, and it is these desert processes that we concentrate on here.

Weathering Processes

In desert environments salt weathering appears to be one of the most effective mechanisms of rock breakdown and has been demonstrated to produce both silt- and clay-sized particles. The process can be effective on both rock blocks and on sand grains. For example, field experiments by Goudie et al. (1997) in the Namib Desert found that on disintegration oolitic limestone blocks in contact with the ground surface produced fine particles with a bimodal size distribution. All the samples reported had a well-defined mode at 32–63 μm and a second, more variable mode in the medium-coarse sand range. The mean diameter of the ooids in the limestone was 689 μm (range 150–1500 μm) suggesting particle disintegration rather

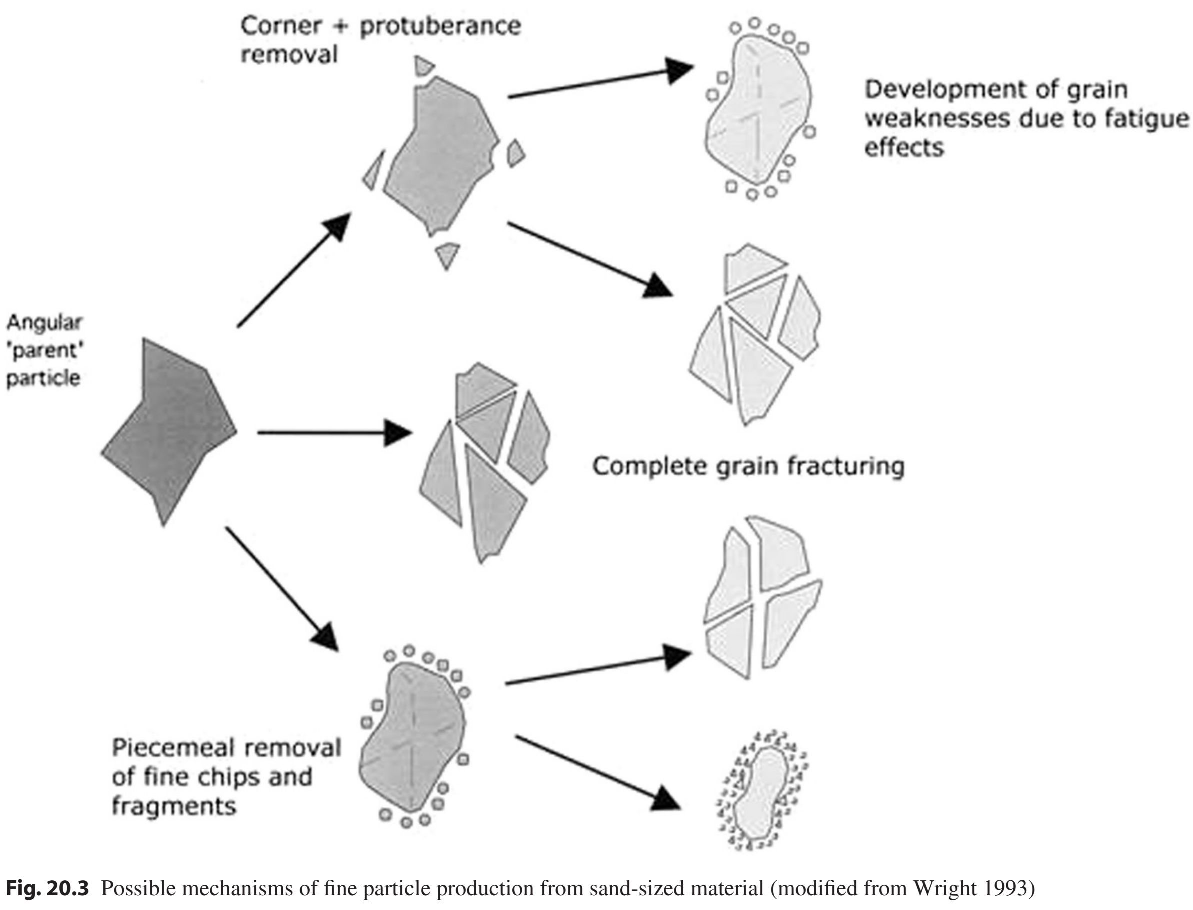

Fig. 20.3 Possible mechanisms of fine particle production from sand-sized material (modified from Wright 1993)

than simply the release of ooids by weathering. Viles and Goudie (2007) also reported a mean grain size of 32–63 μm for weathered schists in the Namib with >30% of the debris released from these rocks being finer than 63 μm diameter. The impact of salt weathering on quartz sand grains examined experimentally by Goudie et al. (1970) and Pye and Sperling (1983) has been shown to produce angular silt-sized particles with the shape of the resulting fine particles a function of both the lithology of the parent grain and also the weathering process.

Similarly, frost weathering has been widely demonstrated as capable of causing rock disintegration and the formation of silt-sized material in the laboratory (e.g. Lautridou and Ozouf 1982, Wright et al. 1998), and field observations in contemporary cold climates point to the importance of frost action in silt formation (e.g. Pye and Paine 1984). Whilst frost weathering is not generally considered to be as effective as salt weathering for fine particle production (Wright et al. 1998), the two processes can act in concert in cold, arid regions, however local environmental conditions are likely to determine whether their combined actions increase or decrease the rate of weathering (Pye 1987).

Aeolian Abrasion

Aeolian transport of sand involves both grain-to-ground impacts and mid-air grain-to-grain collisions which together provide considerable potential for particle breakage and fracturing and consequently for the production of fine material (Kuenen 1960, Whalley et al. 1982, 1987, Wright et al. 1998). Controlled experiments on selected size fractions of sediments show positive relationships between rates of aeolian abrasion and increasing particle size, angularity and surface roughness. Angular particles initially yield high amounts of fine material as corners and protuberances are removed, followed by a decrease in the production of fine particles as the grains become more rounded (Kuenen 1960, Whalley et al. 1987, Wright et al. 1998). However the extent to which the samples used are representative of sediments in desert environments or these experiments simulate the impacts of natural saltation is debatable. For example, Wright (2001a) noted that aeolian abrasion of artificial or freshly-crushed quartz grains with a very angular initial particle shape resulted in higher fine particle production than is likely in natural conditions where particles are generally sub-angular to sub-rounded (Goudie and Watson 1991) and Bullard et al. (2007) found no relationship between particle size or sorting and fine particle production during experiments using natural dune sands.

Other Dust Sources

As discussed above, clay minerals <10 μm may aggregate to form larger sand-sized particles or may adhere to sand grains forming a surface coating. These aggregates can be broken down during aeolian activity. Gillette and Walker (1977), for example, found that sandblasting by saltation removed clay platelets from the surfaces of quartz grains and separated aggregates. There has been some debate about the extent to which clay coatings can be removed by aeolian abrasion (Walker 1979, Gomes et al. 1990, Wopfner and Twidale 2001). Experiments by Evans and Tokar (2000) to determine the robustness of iron-rich coatings on natural sands showed that after one week of dry abrasion in a rock tumbler about 50% of the grains lost their clay coating entirely and the remainder of the particles retained only 5–20% of their coatings (by surface area). These results did not correspond to field data, however, as clay coatings were found to persist for more than two and a half years in a coastal environment (Evans and Tokar 2000). Other experiments (Bullard et al. 2007) indicated that for sub-rounded to sub-angular grains, the quantity of clay coating on a grain surface is the most important determinant of the amount of material <63 μm produced by aeolian abrasion. The removal of iron-rich clay coatings has received attention in the literature due to the impact of the process on the release of bio-available iron oxides to the atmosphere. Whilst it is difficult to study in the field, the removal of iron-rich red-coloured clay coatings can cause visible changes in soils over very short time periods (Shao et al. 1993a, McTainsh 1985) suggesting that this may be an important process.

Dust-sized particles can also be generated by the breakdown of organic materials. For example diatomaceous material deposited during wet phases can form extensive deposits when lakes desiccate – flakes of the diatomite entrained by the wind undergo a 'self-abrasion' process in which they break up into fine dust-sized particles on collision with each other or the ground surface (Tegen et al. 2006). This process is similar to the break-up of particles by aeolian abrasion as the diatomite flakes are single particles rather than aggregates.

Relative Importance of the Mechanisms

The relative importance of each of the mechanisms described above is a function of macro- and micro-climatic conditions, geomorphological setting and lithology (Bullard et al. 2007, Viles and Goudie 2000, Smith et al. 2000, Wright 2001b, 2007). Wright et al. (1998) compared the relative efficiency of five mechanisms of fine-particle production during a range of experiments (Table 20.2), however the magnitude and frequency characteristics of the different processes make it difficult to draw conclusions. For example, the amount of material produced by weathering is low but the fact that it is a ubiquitous process may mean weathering is potentially the most important geomorphological process in silt formation (Wright et al. 1998). Fluvial tumbling and aeolian abrasion are efficient producers of silt but fluvial activity is usually spatially and temporally confined and aeolian activity is also sporadic in most desert environments. It is likely that over long periods of time, several different mechanisms of fine particle production will take place: for example in the production of the silts comprising desert loess deposits (Wright 2001b, Fig. 20.4). In contemporary environments, the presence of specific agents of weathering may determine the relative importance of different mechanisms, for example salt weathering is an important contemporary process operating in the Namib due to the presence of salts and frequent fog in the environment. Similarly, dune sands that have developed a coating of clay-sized materials may be a significant source of dust during periods of sustained aeolian activity. The effectiveness of a particular dust-production process and the precise particle-size distribution of the resulting material is also related to the character of the source materials and the presence of defects and susceptibilities in the particles (Assallay et al. 1998, Kumar et al. 2006).

Table 20.2 A comparison of the relative effectiveness of silt-producing mechanisms (calculation of the theoretical maximum amount of silt produced from 1 kg of the original sample) (Wright et al. 1998)

Run type	Amount of fine ($<63\,\mu$m) material produced (g kg^{-1})	Run duration
Glacial grinding	47.4	24 h
Aeolian abrasion	287	96 h
Fluvial tumbling: spheres	900	32 h
Salt weathering: Na_2SO_4	41.6	40 cycles
Frost weathering	0.44	360 cycles

Global Dust Emissions

Globally most atmospheric dust is derived from the deflation of surface material in arid and semi-arid areas including North Africa, the Middle East and Asia, in the northern hemisphere, and Australia and southern Africa, in the southern hemisphere. Within these dryland zones, ground meteorological observations and satellite remote sensing data have enabled the identification of the main global dust sources and transport paths.

Remote Sensing of Dust Sources, Transport and Sinks

Detailed analyses of global dust sources by Prospero et al. (2002) using data from the Total Ozone Monitoring Spectrometer (TOMS) mounted on the Nimbus 7 satellite launched in September 1978 (Herman et al. 1997) demonstrated that all major contemporary dust sources are identified with topographic lows in areas with an annual rainfall of usually less than 250 mm (Fig. 20.5). These areas are predominantly located within internally draining basins with seasonally active rivers, streams and playas. The implication is that fine alluvial material deposited within these channels and on the flood plains following flood events and playa deposits developed during more humid conditions are the major global sources of dust. This is the cornerstone of the 'inland-basins hypothesis' (Bullard

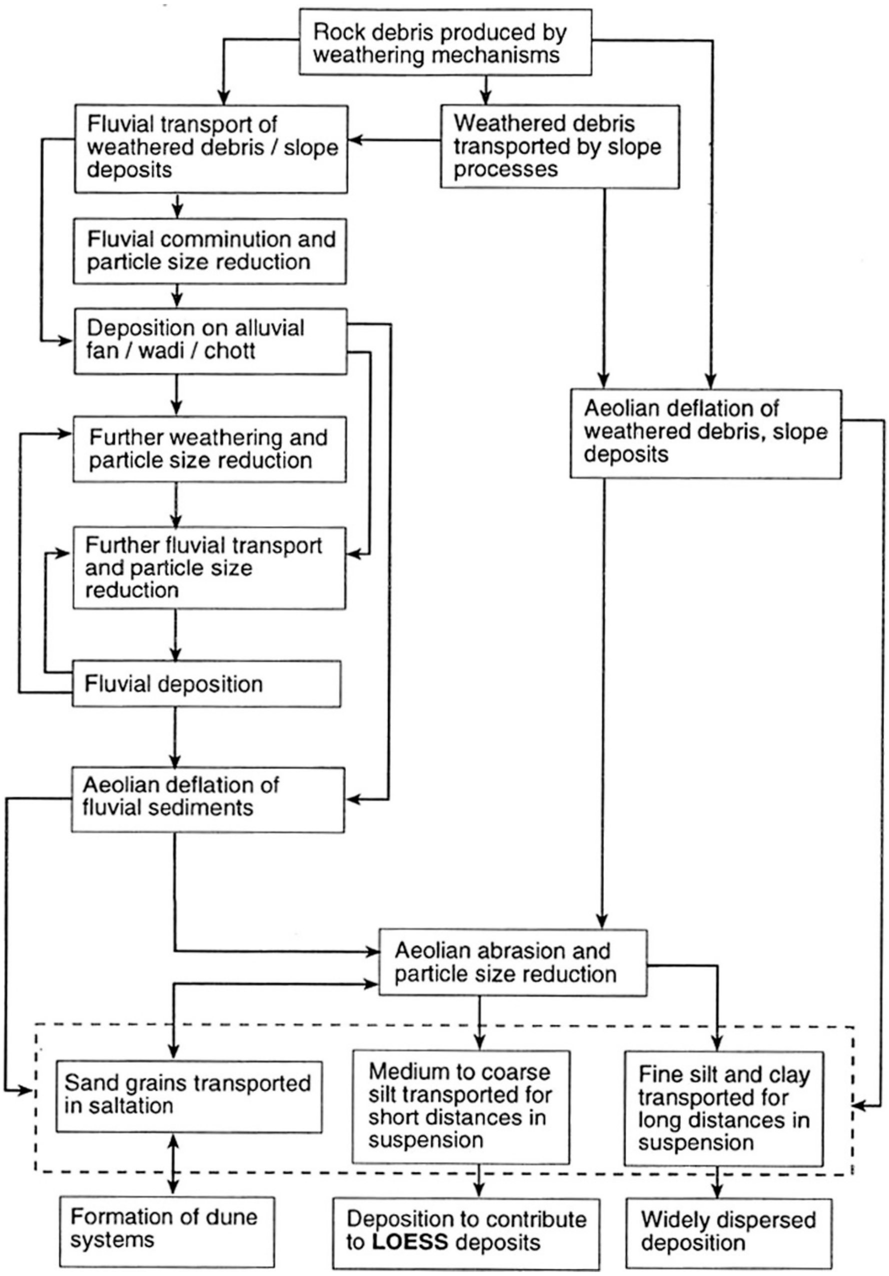

Fig. 20.4 Possible mechanisms for the production of fine particles in a hot desert environment (after Wright 2001b)

and McTainsh 2003). In some cases, deep alluvial deposits laid down during the Pleistocene are also dust sources (Prospero et al. 2002).

Satellite-based sensors have revolutionised the mapping of dust emissions and have proved an excellent source of data for ascertaining spatial patterns of dust emissions at the global scale, seasonal temporal trends, and likely zones of deposition (Prospero et al. 2002, Washington et al. 2003). No single satellite data source can identify all dust emissions. The frequently used Total Ozone Mapping Spectrometer Aerosol Index (TOMS-AI), for example, can be affected by cloud, dust mineralogy and chemical composition, and is more sensitive to middle and upper troposphere aerosols than lower level (<1–2 km) aerosols. This sensitivity difference can result in an under-representation of dust events at some locations (Washington et al. 2003), and of some types, such as haboob-style dust storms that are confined to the lower 1 km of the atmosphere (Miller et al. 2008).

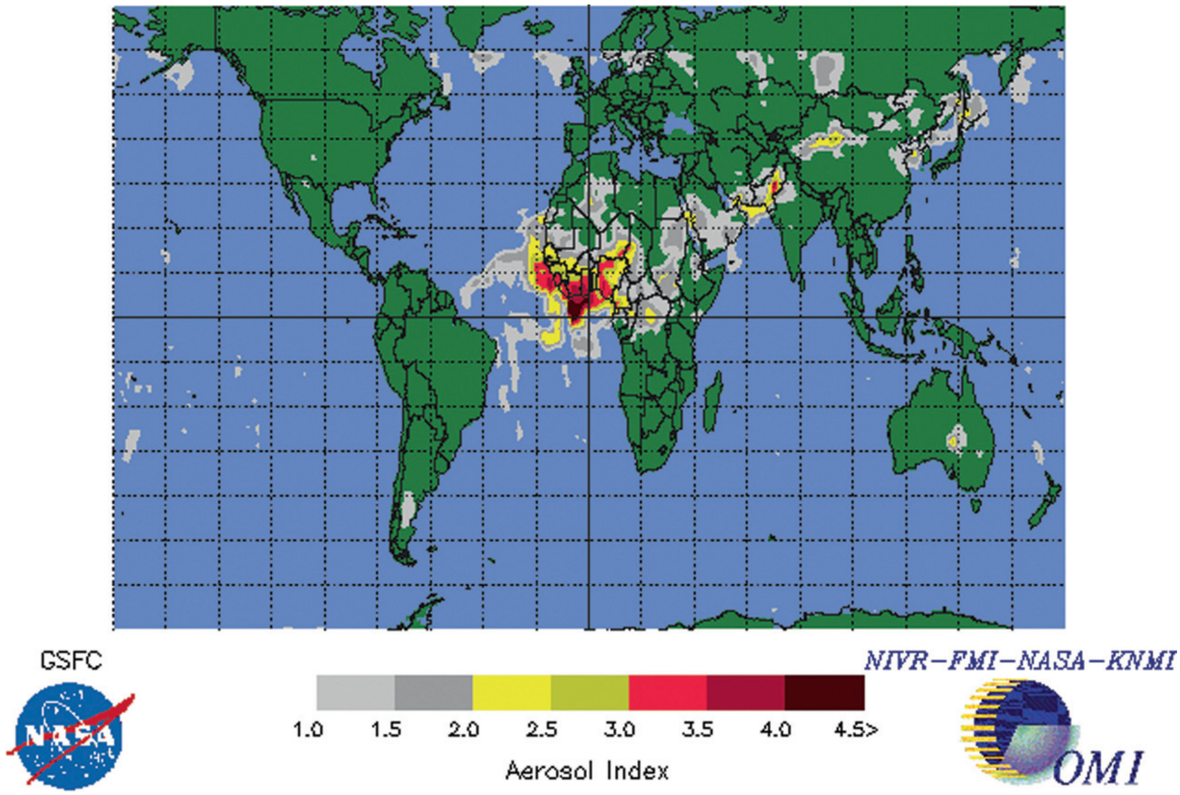

Fig. 20.5 A composite image showing the Ozone Monitoring Instrument aerosol index (TOMS-AI) on 22 February 2008 (reproduced courtesy of the Ozone Processing Team at NASA's Goddard Space Flight Center)

However, when a range of complementary satellite sensors is used, such as TOMS, MODIS (Moderate Resolution Imaging Spectroradiometer) (e.g. Kaufman et al. 2005), GOES (Geostationary Operational Environmental Satellite), SeaWIFS (Sea-viewing Wide Field of View Sensor) and LiDAR (Laser Imaging Detection and Ranging) (e.g. Berthier et al. 2006), each of which has strengths and limitations, then characterisation of dust emissions can be very precise (e.g. Rivera et al. 2006, Miller et al. 2008).

Global Dust Models

As dust is increasingly recognised as an important component of the atmosphere capable of driving or affecting global climate change, a number of attempts have been made to model global emissions. Models vary but generally include a wind erosion or erodibility component coupled with a global climate model. 'Bulk mobilization' models (Zender et al. 2003a) assume a simple relationship between dust emission and wind speed for particular sizes of emitted dust (e.g. Tegen and Fung 1994, Perlwitz et al. 2001) whilst more complex models include detailed specification of the physical processes acting in the emission zones, such as saltation mass flux (e.g. Shao 2001). Different models predict varying amounts of dust with the majority suggesting total global dust emissions in the range 1000–2000 Tg a^{-1} (Table 20.3) but considerably higher and lower estimates have been reported due largely to the different underlying assumptions incorporated into the models (Zender et al. 2004). Table 20.3 also demonstrates the impact of different model parameters on the relative proportions of dust attributed to different regions of the globe, with for example estimates for Australia ranging from 37–148 Tg a^{-1} and for Arabia from 43–415 Tg a^{-1}.

Researchers attempting to model global and continental-scale dust emissions increasingly recognise the need, not only to acknowledge surface heterogeneity in their calculations (Sokolik et al. 2001), but also to obtain high-quality land erodibility and

Table 20.3 Comparison of the regional annual mean dust flux (Tg a^{-1}) from selected global dust models. Numbers in brackets indicate the percentages for the annual mean global emission flux predicted by each model (from Tanaka and Chiba 2006)

	Africa		Asia			America			
	North	South	Arabia	Central	East	North	South	Australia	Global
Ginoux et al. (2004)	1430 (69.0)	63 (3.4)	221 (11.8)	140 (7.5)	214 (11.4)	2 (0.1)	44 (2.3)	106 (5.7)	1877
Luo et al. (2003)	1114 (67.0)		119 (7.2)		54 (3.2)			132 (8.0)	1654
Miller et al. (2004)	517 (51.0)		43 (4.2)	163 (16.0)	50 (4.9)	53 (5.2)		148 (15.0)	1019
Tanaka and Chiba (2006)	1087 (57.9)	63 (3.4)	221 (11.8)	140 (7.5)	214 (11.4)	2 (0.1)	44 (2.3)	106 (5.7)	1988
Werner et al. (2002)	693 (65.0)		101 (9.5)	96 (9.0)				52 (4.9)	1060
Zender et al. (2003a)	980 (66.0)		415 (28)			8 (0.5)	35 (2.3)	37 (2.5)	1490

dust concentration data at suitable resolutions. Research using global models has demonstrated that it is unrealistic to assume that soil erodibility is uniform and factors such as topography, geomorphology and hydrology have to be accounted for to explain global patterns of dust emissions. For example, Zender et al. (2003b) found that a model based on the assumption of uniform soil erodibility could explain none of the spatial distribution of dust emissions in Australia and was also a poor predictor of spatial patterns in North Africa and the Arabian Peninsula. However the inclusion of geomorphic erodibility improved their model considerably.

Sub-Continental Scale Dust Emissions

Although different models predict different absolute mean annual dust emissions, Table 20.3 and Fig. 20.4 indicate clear conclusions that can be drawn about the relative importance of different geographical areas. The largest global dust source is the Sahara in north Africa, followed by Arabia and southwest and central Asia. At present, Australia is the largest contributor of dust in the southern hemisphere and southern Africa and the Americas all contribute significantly lower quantities of dust to the global system. The TOMS-AI can be used to identify dust 'hot spots' within these zones more precisely, to the extent of identifying specific geomorphological regions (Table 20.4). For example the 37,000 km^2 Mkgadikgadi basin ephemeral lake complex is the largest dust source in Botswana, the south-central area of the 1.14 million km^2 Lake Eyre Basin is the most significant dust source in Australia and the Salar de Uyuni, possibly the world's largest salt flat extending 10,582 km^2 across the Bolivian altiplano in South America, is also a clearly definable dust source.

Whilst Table 20.4 emphasises the importance of topographic basins in areas of low rainfall as major sources of dust, these are rarely the only two parameters contributing to high dust emissions. For example, the Earth's single largest source of dust is the Bodélé Depression, estimated to emit 640–780 Tg a^{-1} (Ginoux et al. 2001). The Bodélé Depression, on the southern edge of the Sahara Desert, is part of the bed of Lake Chad. Currently the lake itself is approximately 1350 km^2, having shrunk from 25,000 km^2 in 1963 and up to 400,000 km^2 at its maximum extent in the mid-Holocene (Lake Mega-Chad) (Coe and Foley 2001, Ghienne et al. 2002). The recession of the water has left an extensive area of desiccated, highly erodible alluvial and diatomaceous material. The status of the Bodélé Depression as the dustiest place on Earth (Giles 2005) results from the coincidence of several

Table 20.4 Maximum mean aerosol index (AI) values for major global dust sources determined by TOMS (Goudie and Middleton 2006). The higher the value of the index, the greater the proportion of absorbing aerosols in the atmosphere

Location	AI value	Average annual rainfall (mm)
Bodélé Depression, south central Sahara	>30	17
West Sahara in Mali and Mauritania	>24	5–100
Arabia (Southern Oman/Saudi border)	>21	<100
Eastern Sahara (Libya)	>15	22
Southwest Asia (Makran coast)	>12	98
Taklamakan/Tarim Basin	>11	<25
Etosha Pan (Namibia)	>11	435–530
Lake Eyre Basin (Australia)	>11	150–200
Mkgadikgadi Basin (Botswana)	>8	460
Salar de Uyuni (Bolivia)	>7	178
Great Basin of the USA	>5	400

variables that combine to give consistent high winds and sediment supply. First, the area is a topographic low in which considerable quantities of sediment have collected. These sediments include thick layers of diatomite formed from the shells of freshwater diatoms which lived in the lake and which correspond with the areas of maximum dust deflation. High levels of dust production occur during the transport of saltating diatomite flakes which disintegrate during collision with one another and on impact with the ground surface (Giles 2005). The topographic characteristics of the area also promote strong winds with low directional variability as the winds are funnelled by the Tibesti Mountains to the north and by the Ennedi Mountains (Koren and Kaufman 2004, Washington and Todd 2005). Washington et al. (2006) suggested that enhancement of these topographically-generated winds (the Bodélé Low Level Jet) during drier, windier conditions in the past may have helped to create the depression, thus promoting the topographic and environmental conditions that make the area such an intense contemporary dust source.

At the local scale ($<10\,\text{km}^2$) there have been many high-resolution studies of the susceptibility of different surfaces to aeolian erosion. These studies have provided detailed information concerning critical threshold wind velocities for dust entrainment and the influences of particle size, surface roughness, soil moisture content and biocrust development on soil erodibility (see Nickling and McKenna Neuman, this volume). The most significant gap in our current understanding of dust emissions is at the meso-scale or sub-drainage basin scale (10^4–$10^6\,\text{km}^2$). Most inland basins are patchworks of soil and land types that yield both different quantities and different types of dust. For example, using measured erosion rates, Breshears et al. (2003) projected annual erosion mass flux to be $4.5\,\text{g m}^{-2}\,\text{a}^{-1}$ from semi-arid grassland compared with $14.3\,\text{g m}^{-2}\,\text{a}^{-1}$ in semi-arid shrubland. In southern Nevada and California, Reheis and Kihl (1995) found that playa and alluvial surfaces produced the same amount of dust per unit area (9–14 $\text{g m}^{-2}\,\text{a}^{-1}$) but the greater areal extent of alluvial surfaces meant they contributed a higher total volume of dust. The composition of dust from the two sources also differs: the playas producing dust much richer in soluble salts and carbonate compared with the majority of alluvial sources.

Whilst the spatial distribution of dust sources is becoming well known, and some dust sources are very consistent, the majority of dust sources are intermittent in terms of temporal productivity. The causes of this intermittency are variable, and a key current area of research lies in determining what causes the ‘switching on’ or ‘switching off’ of different dust sources. For some dust events, the triggers of dust emissions are relatively easy to discern. For example, in a supply-limited system such as in parts of the Mojave Desert, USA, if the supply of fine materials to dust source areas is not maintained (for example by regular input from flood events) the magnitude and frequency of dust events diminishes (e.g. Clarke and Rendell 1998). Where currently active ephemeral streams are major sources of dust, a temporal relationship between fluvial events and dust sources can be discerned. However this relationship can be complicated if the associated increase in soil moisture content affects other parts of the system, such as vegetation cover, soil salt content or surface crust development (Offer and Goossens 2001). McTainsh et al. (1999) reported that under conditions of low rainfall (150 mm in 1994) in the Channel Country of Australia, dust flux was higher from sand dunes than from clay pan surfaces. However subsequent increases in rainfall caused improved vegetation cover on the dunes which reduced dust emissions. Higher rainfall (370 mm in 1997) maintained the non-erodible vegetation cover on the dunes, but resulted in a pulse of sediment being delivered by flood waters to an adjacent clay pan facilitating a series of dust emission events. Similarly, Bryant (2003) and Mahowald et al. (2003) explored the interactions amongst ephemeral lakes, flood inundation, land use and dust emissions in the Etosha Basin, Namibia and found that although there was some indication that dust emissions increased following post-flood desiccation of the pan surface, the relationship between the two variables was not a simple one. Even where sediments are present they are not always available for deflation, for example due to vegetation protecting the land surface from erosion. Bullard et al. (2008) found an increase in the frequency of dust emissions from the Simpson desert, Australia, following widespread fires that removed vegetation cover, and Rostagno (2007) recorded rapid removal of 90% of the clay and silt fractions from soils during dust storms originating from rangelands following fires.

Whilst flood events in dryland areas can be an important source of dust-sized material, the delivery of coarser, sand-sized material to dust-emission areas can be important in controlling the extent to which the fine material can be raised into the atmosphere. Several field studies have highlighted the importance of sand-sized sediment in releasing dust into the atmosphere – the impact of the sand particles on the fine, often consolidated or compacted clays and silts, causes dust-sized material to be ejected into the airstream. (e.g. Nickling et al. 1999, Shao et al. 1993b, Nickling and Gillies 1993). This process of 'sandblasting' not only triggers the release of dust but can, by aeolian abrasion of the coarse saltators, simultaneously result in dust production. Grini and Zender (2004) found that predictions of dust emission were significantly improved when saltation and sandblasting were factored into a dust emission model.

Meteorological Conditions Promoting Dust Events

Dust emissions have a distinct spatial distribution and, as discussed above, are temporally discontinuous. Whilst variables such as sediment supply are important, there are also certain meteorological conditions that promote the entrainment and transport of dust. Nickling and McKenna Neuman (this volume) discuss thresholds of entrainment for fine particles; this section will focus on broader meteorological parameters.

Dust storms are defined as 'the result of turbulent winds raising large quantities of dust into the air and reducing visibility to less than 1000 m' (McTainsh and Pitblado 1987, p. 416). Dust storms vary in extent and duration, and can include both local and long-distance dust transport. There are four main types of weather systems commonly associated with dust storm generation.

Dust Storms Associated with Cold Fronts

The passage of frontal depressions is one of the most widespread causes of dust storms (Pye 1987, Livingstone and Warren 1996), especially in the Middle East, USA and Australia. These low-pressure fronts have steep pressure gradients that generate strong winds and are often accompanied by an increase in relative humidity and a fall in temperature. As a cold front moves displacing warm air, increasing instability and vertical air movements increase turbulence which is conducive to the raising of dust. The passage of a cold front is commonly marked by a sharp change in wind direction. Dust is raised along the cold front and can be lifted to high levels in the atmosphere and transported long distances. Widespread dust haze, such as that associated with the harmattan in north Africa can be caused by frontal activity. At lower levels (typically <2000 m) cold fronts can generate belt dust storms hundreds of kilometres long and tens of kilometres wide that can travel at speeds of over 20–30 ms^{-1}.

One of the best documented recent dust storms associated with a cold front occurred in Australia on 22–23 October 2002 (Chan et al. 2005, Leslie and Speer 2006, McTainsh et al. 2005, Shao et al. 2007). The storm was generated by a cold front moving eastwards from the Great Australian Bight. On 22 October 2002 moderate-strong pre-frontal north westerly winds affected parts of southeast Australia reducing visibility in some areas to <500 m (McTainsh et al. 2005). As the front travelled across the continent, large quantities of dust were entrained from rangelands and the central deserts and at its maximum extent the dust plume was 2400 km long (north-south), 400 km across and 1.5–2.5 km in height (Fig. 20.6). The estimated total dust load was 3.35–4.85 Mt and dust in the atmosphere increased

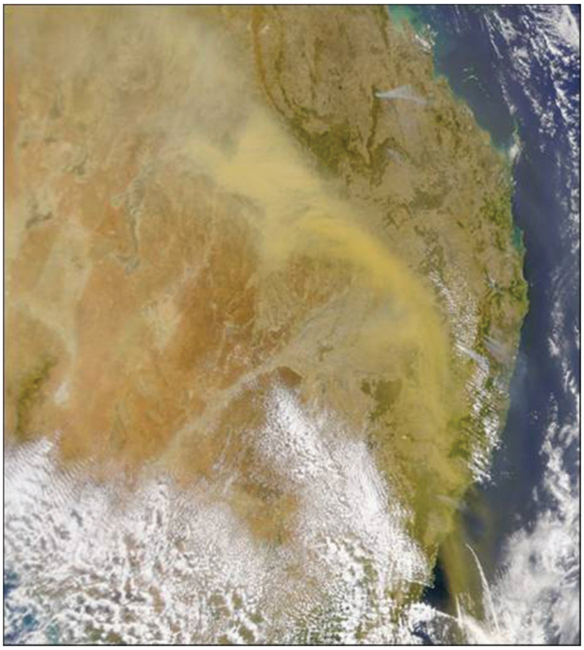

Fig. 20.6 SeaWIFS satellite image showing the dust transported by pre-frontal north westerly winds in southeast Australia on 22 October 2002

Fig. 20.7 Dust storm at Big Spring, Texas, USA, on 16 June 1997. Photo taken by and reproduced by kind permission of Weinan Chen

the average concentration of particles $<10\,\mu$m (PM_{10} load) in Brisbane and Mackay on the east coast to 161 and 475 μg m^{-3}, respectively, (greatly exceeding the recommended 50 μg m^{-3} air quality standard) (Chan et al. 2005). Average maximum wind speeds associated with the event were 32.4 km h^{-1} (9.0 ms^{-1}) which is lower than some comparable events. The magnitude of the storm is thought to have been exacerbated by six months of preceding drought conditions and high temperatures that reduced vegetation cover. Whilst the desert regions were the strongest dust source, fine particles were also entrained from grazing and farmlands (Shao et al. 2007).

Dust Storms Associated with Thundery Conditions

Dust storms are often associated with strong downdrafts of cooled air that descend steeply from cumulonimbus clouds and thunderstorm cells. These downdrafts can cause wind speeds as high as 50 ms^{-1} (Goudie and Middleton 2006) which raise dust causing localised dust storms along the storm line. This type of event, known in north Africa as a 'haboob', is typically characterised by a dense wall of dust which moves across the landscape. Haboob-type storms in north Africa (particularly Sudan) are associated with northward movement of the intertropical front bringing moisture inland from the central Atlantic and are most common from May to September, although they can occur at any time. They also occur in Israel (Offer and Goossens 2001), across the Arabian peninsula (Miller et al. 2008), in the southwest USA (Brazel and Nickling 1986), and in Australia (Leslie and Speer 2006). Chen and Fryrear (2002) measured wind speeds of 1.8–2 ms^{-1} rising to 2.5–5.5 ms^{-1} at the leading edge and 13 ms^{-1} at the rear of a haboob dust storm at Big Spring, Texas on 16 June 1997 (all velocities measured at 10 m above the ground surface) (Fig. 20.7). Dust concentration (integrated at 27 heights up to 1.567 m) was 84,960 kg km^{-1} h^{-1} (of which 21% was in the size range 10–20 μm). The mean diameter of the dust particles entrained by the event decreased with height above the surface from 33.4 μm (0.41 m) to 23 μm (1.567 m) and the dust became more poorly sorted with height (Chen and Fryrear 2002).

Dust Storms Associated with Major Depressions

In the northern hemisphere spring, off the coast of north Africa there is an extreme contrast between the surface temperatures of the sea and the desert air. This contrast leads to the generation of large, synoptic-scale depressions which travel eastwards over the Sahara and south-central Mediterranean. Passage of these depressions is associated in most cases with strong, hot, dry winds which entrain sediments and can lead to severe dust storms as well as widespread dust haze across the region.

Dust Storms Associated with Mountain Winds

Air that has undergone orographic lifting cools and dries as a result of condensation of water vapour. This cool, dry air becomes dynamically-heated as it descends from lower to higher atmospheric pressures creating strong, warm, dry, turbulent airflow known as föhn (or foehn) winds. Winds associated with mountain ranges are common but do not all trigger dust storms due to sediment supply limitations. Common dust-raising föhn winds include the Santa Ana winds of southern California and Berg winds of southern Africa. Soderberg and Compton (2007) report a berg wind event in Cape Town that occurred on 19–20 May 2002 in which visibility was reduced by wind-blown dust. On the 20 May, PM_{10} levels at Khayelitsha averaged

101 μg m^{-3} over the 24 period (classed as 'very high') and hourly records at stations including Table View exceeded 200 μg $^{-3}$ (City of Cape Town Air Quality Monitoring Network 2002). In this area dust storms are an important source of fine particles and nutrients (e.g. K, Al, Ca) for soil and vegetation (Soderberg and Compton 2007).

Small-Scale Dust Events

One of the most localised types of dust-raising event is the 'dust devil'. Dust devils form as air at the ground surface is heated and starts to rise rapidly through overlying cool air. The air column may start to rotate, gaining increasing wind speeds which enable entrainment of sediments from the surface that then rise with the hot air through the air column. Dust devils form on surfaces of low gradient that are susceptible to heating, such as bare soils, although vegetation can be present (Mattsson et al. 1983) and their development can be exacerbated by local topography. For example Hess and Spillane (1990) reported dust devils forming parallel to topographic ridges. The requirement for surface heating means that dust devils are most frequent during the hottest part of the day (late morning or early afternoon) and during spring and summer seasons (Wigner and Peterson 1987, Oke et al. 2007a). Dust devils range from <0.5 m to >100 m in diameter, although most are <10 m across. They are usually at least five times higher than they are wide (Hess and Spillane 1990) and exceptionally can reach heights of over 1000 m on Earth, although over 80% lie in the height range 3–300 m (Balme and Greeley 2006). The height of a dust devil is determined not only by atmospheric conditions but also by the characteristics of the dust particles incorporated within it. Oke et al. (2007b) found that over 80% of particles transported in dust devils in Australia (known locally as willy-willies) were $\leq 63\,\mu$m in diameter and suggest that this preferential entrainment of fine particles could have implications for soil surface texture, vegetation sustainability and hence geomorphology. Until recently there had been little geomorphological research on dust devils when compared with that concerning other dust-event phenomena, but the discovery that dust devils on Mars are significant enough in terms of magnitude and frequency to affect surface albedo and possibly regional-scale atmospheric circulation (e.g. Michaels 2006) has led to more studies on Earth which might act as useful analogues for other planets (e.g. Towner et al. 2004).

Dust Deposition

Dust Deposition in the Oceans

Of course, once entrained dust can travel long distances in suspension. Grousset et al. (2003), for example, reported that dust deflated from the Taklamakan Desert in western China had travelled over 20,000 km east across the Pacific Ocean, North America and the Atlantic to be deposited in western Europe. Similarly, Tanaka (2005) reported a two-day dust event recorded at meteorological stations in Japan that had its origins in dust storms in north Africa and the Middle East several days earlier, and he concluded that the Sahara and Arabian deserts were potentially important sources of dust in east Asia.

As a consequence of this propensity for long-distance travel, much of the dust generated in the desert areas is exported and deposited in the oceans. The deep-sea drilling programmes have provided much evidence in the form of sedimentary cores from the ocean floors of export of aeolian material from deserts, early examples of the recognition of aeolian dust in ocean cores including the work of Chester et al. (1971), Duce (1980) and Prospero et al. (1981). This dust has been seen as an important palaeoenvironmental proxy: accumulation rates indicating dust fluxes and particle sizes indicating wind strength. There is good evidence, for instance, of increased dust input into ocean cores at the time of the last glacial maximum around 18,000 years ago, notably off the west coast of north Africa (Sarnthein and Koopman 1980, Tetzlaff and Peters 1986) and into the Pacific off the east coast of Australia (Hesse and McTainsh 1999). In both cases these cores point to increased dust production in the deserts, probably associated with lowered precipitation rates, but in some instances they may also reflect greater wind energy, although Hesse and McTainsh (1999) suggested that this was not the case in Australia.

Dust deposited in oceans has negligible geomorphological impact but there are a range of consequences of this deposition. There may for instance be a significant link between aeolian dust and biological productivity, for example in the Pacific (Yuan and Zhang 2006), Indian (Piketh et al. 2000) and southern Oceans although there are still considerable uncertainties concerning this relationship (Jickells et al. 2005). Additionally, the geochemical properties of deposited dust have been seen as a potential tracer for dust sources that could find application in palaeoenvironmental studies of dust transport and deposition (e.g. Moreno et al. 2006). Dust deposited in the oceans can therefore be used as a tracer for global atmospheric circulation (Biscaye 1997, Prospero 1999, Grousset and Biscaye 2005, Tanaka 2005) and this potential has meant that dust transport and deposition has been the subject of modelling (e.g. Tanaka and Chiba 2006).

Contemporary Terrestrial Dust Deposition Records

At least part of the difficulty of establishing the importance of aeolian transport of dust has been the lack of effective monitoring. While dust flux may be monitored remotely (see the section on remote sensing above), it is much more difficult to measure the amounts of dust being deposited. Where attempts have been made, the characteristics and mass of dust trapped, and hence how representative the sample is of actual dust deposition, is highly dependent upon the type of trap used. A number of recent studies have compared the performance of different methods for ascertaining dust flux and rates of dust deposition. There is a fundamental difference between horizontal transport flux and vertical deposition of dust and most traps measure effectively one or the other, but not both. Relationships between the two variables can be determined (e.g. Goossens 2008) but these vary with grain size. Goossens and Offer (2000) evaluated the efficiency of five horizontal dust flux samplers and found this varied with wind speed. The semi-isokinetic, self-orienting Big Spring Number Eight (BSNE) sampler (Fryrear 1986), which is widely used in field studies, was recommended because its trap efficiency is largely independent of wind speed, however it is better suited to coarser particles than very fine particles and was not tested at high wind speeds. A wide range of studies have explored the relative efficiencies of dust deposition samplers (Goossens 2005, Goossens 2007, Goossens and Rajot, 2008, Sow et al. 2006), the big practical difficulty to be overcome being developing equipment that can prevent the re-entrainment of dust deposited on the trap. Common approaches include using vessels (which vary widely in terms of their aerodynamics) containing marbles or water to prevent re-entrainment, or traps with high roughness values such as astroturf (O'Hara et al. 2006).

Dust deposition varies both spatially and temporally and the types of traps used should be tailored to the research questions being asked, but will also be dependent on available resources and how frequently data can be retrieved. Recently, a number of projects have been established to trap airborne dust and estimate deposition rates both spatially and over long time scales. The US Geological Survey (USGS), for instance, has established the CLIM-MET network of meteorological stations in the south-west USA which include a number of instruments for monitoring both dust transport and deposition. From their data they have been able to piece together a picture of dust sources, fluxes and deposition rates in the Mojave Desert (e.g. Reynolds et al. 2006). (Data from the CLIM-MET project are publically available from: http://esp.cr.usgs.gov/info/sw/clim-met/index.html). In North Africa O'Hara et al. (2006) monitored dust deposition in Libya over an annual cycle and found values in the range 38.6–311.0 $\mathrm{g\,m^{-2}\,a^{-1}}$, with a mean of 129.1 $\mathrm{g\,m^{-2}\,a^{-1}}$. Dust activity in the McMurdo valleys (Antarctica) showed high inter-annual variability ($<1\,\mathrm{g\,m^{-2}\,a^{-1}}$ to $>200\,\mathrm{g\,m^{-2}\,a^{-1}}$) from 1999 to 2005 with rates of dust deposition strongly influenced by the availability of material (Lancaster 2002). These studies have all used different methods for quantifying dust deposition, but one possibility is that sampler comparison studies, such as those cited above, could provide a framework for comparing studies conducted using different techniques.

Factors Promoting Dust Deposition

The nature of the transport of dust in the wind means that it is often carried large distances and spread over large areas, as a consequence of which it only rarely builds distinguishable deposits and arguably never builds distinctive landforms. Frequently this transport is away from the terrestrial source and out over the oceans where it may be incorporated into marine sediments. On land it is often spread so thinly that it is indistinguishable from other surface materials. Particularly in areas where weathering rates are sufficient, the dust will be quickly incorporated

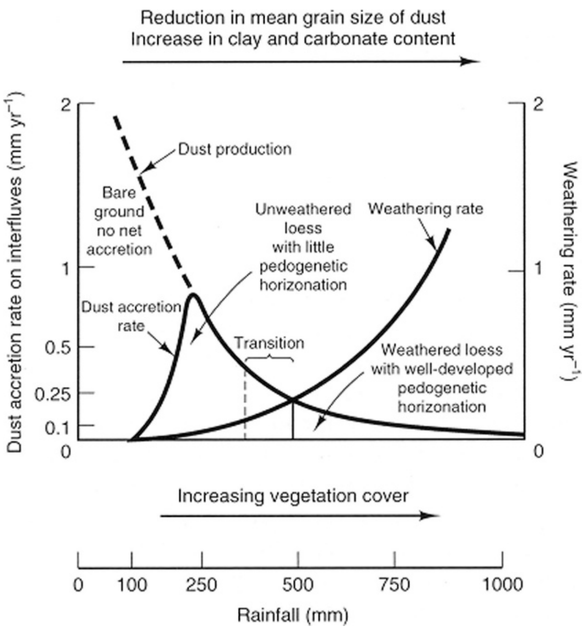

Fig. 20.8 Schematic model of dust accretion in relation to mean annual rainfall (from Pye and Tsoar 1987)

into soils. If deposited where there is little vegetation or other surface materials to trap the dust, it will be re-entrained. There are therefore rather particular conditions that need to be satisfied before aeolian dust deposits accumulate sufficiently to be recognisable. Where the terrestrial dust deposits are recognisable they are often called 'loess'.

Pye and Tsoar (1987) summarised the conditions for terrestrial dust deposition by plotting dust accretion as a function of rainfall (Fig. 20.8). Where rainfall is low there may be plenty of dust generated by all the mechanisms discussed above (and represented by the dashed line on Fig. 20.8) but it is too easily re-entrained to accumulate. Increasing vegetation cover generated by greater rainfall acts as an effective trap to the dust that does fall to the surface and helps to prevent re-entrainment. Increasing precipitation is also often the mechanism for deposition by washing the dust out of the atmosphere. However, more moisture also intensifies the weathering rate and this means that deposited dust is quickly incorporated into soils. Although airborne desert dust does reach temperate areas, it does so in small quantities and is soon 'lost' into the soil. It may also be redistributed by runoff processes generated by the rainfall. Consequently, the schematic model of Pye and Tsoar showed that dust most readily accumulated in areas of around 250 mm a^{-1}. Drier than this and the dust is easily re-entrained; wetter and it is 'lost' into the soil.

Airborne dust does also accumulate in combination with other surface materials. McFadden et al. (1987) suggested that dust was being trapped by coarser material in the Mojave Desert, in their example on the surface of a lava flow. Once it reached the surface they argued it was washed into the interstices between fragments of the lava to create a stone pavement with finer material concentrated below the surface under an armour of coarser material. Stone pavements have been described as forming by a number of mechanisms, including aeolian deflation (Cooke 1970, Chapter 19) but this mechanism of trapping of fine material carried by suspension in the air has been reported from a number of locations (e.g. Gerson and Amit 1987, Goossens 1995, Li et al. 2005).

The trapping of wind-blown dusts by rough surfaces can also lead to distinctive localised deposits. For example, in northern Chile dust streaks have formed on uneven salar surfaces (Stoertz and Ericksen 1974). Figure 20.9 shows dust deposits extending from the base of a hillside on the southwest of the Salar de Pintados. Gullies in the hillside funnel the wind which entrains fine sediments and deposits them forming streaks with an average length of 1.5 km, although the longest extend over 3 km on to the salar and have a

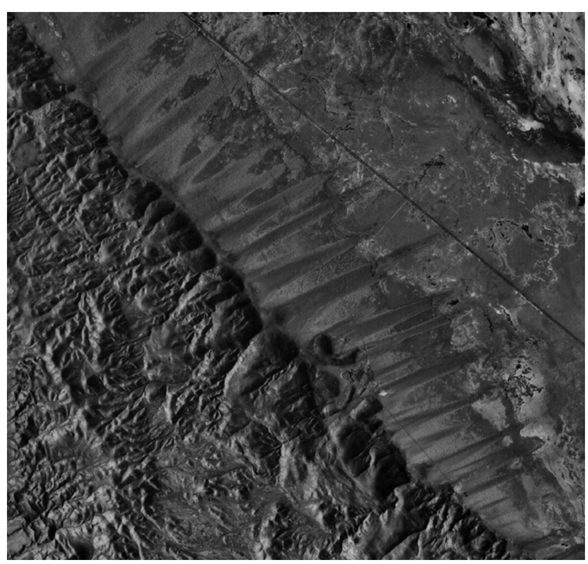

Fig. 20.9 Dust streaks on the surface of the Salar de Pintados, north Chile. Aerial photograph extract reproduced by kind permission of Servicio Aero Fotogrametrico, Chile

maximum width of 350 m. The dust particles, which are predominantly silt-sized aggregates, are trapped in pockets on the salar surface and the streaks are clearly visible because the dusts are dark grey in colour and contrast with the pale salt crust of the salar.

Loess

Where recognisable deposits of wind-blown dust accumulate they are termed 'loess'. There is a huge literature reflecting the disputes over the recognition of loess. For a long time loess was defined by its size, and the best known of the world's loess deposits were associated with glacial outwash – often specifically linked to greater glacial extents at the time of the last global glaciation – and the source of dust-sized material in the world's deserts was contested. Even as recently as 1990, Smalley and Derbyshire felt able to argue that 'most of what was called desert loess can now be seen as mountain loess' (Smalley and Derbyshire 1990, p.301). Their argument was that loess was a Quaternary phenomenon because only glaciers were able to grind rock to the size fraction required for loess deposition. Their concession was that in some low-latitude mountain environments it might be possible for cold-climate conditions to produce loess-sized material by periglacial weathering processes. As Yaalon (1991) pointed out in his rebuttal to Smalley and Derbyshire, other sedimentary deposits are not distinguished by their place of origin; we do not talk, for instance, of igneous sand dunes or plutonic alluvium. But in addition, as we have seen, deserts do provide an excellent environment for generating silt-sized material to be carried by the wind, and Yaalon (1991) could not see how loess at the margins of deserts could be called 'mountain loess'. Even in 1995, Smalley was reluctant to admit that deserts might provide the material for loess. He pointed to the mountains as the source for the loess of central Asia and north China. He did admit that Africa provided him with something of a problem and failed to mention loess deposits in Arabia or Australia. Assallay et al. (1996) suggested that the particles in the Libyan deposits were too large for them to be termed 'loess' although Coudé-Gaussen (1987) attributed large particle size elsewhere in north Africa to proximity to the source. Sarnthein and Koopman (1980) reported that there were deposits of terrigenous quartz silt in the size range characteristic of loess in the Atlantic Ocean off the West African coast. There is now no doubt that deserts produce the basic material for loess deposits.

Notwithstanding the lengthy debates about the origin, provenance, mineralogy and size of loess, loess can most straightforwardly be defined as 'a terrestrial clastic sediment, composed predominantly of silt-size particles, which is formed essentially by the accumulation of wind-blown dust' (Pye 1995, p. 653). The mineralogy reflects the mineralogy of wind-blown dust discussed above: it is predominantly quartz with components of feldspar, mica, carbonates and clay minerals. Despite some arguments that true loess must be exclusively silt-sized, the modal size is nonetheless 20–40 μm, and many loess deposits at desert margins include a significant sand component. Arbitrarily Pye (1987) termed loess with more than 20% sand-sized material as 'sandy loess', an example being the peri-desert loess in Tunisia which has a mode of 63 μm (Coudé-Gaussen and Rognon 1988). In part this may be because much loess associated with deserts is relatively close to its source. Smith et al. (2002) noted studies (Morales 1979) in which dusts blowing out of the Sahara were described as having modal particle sizes predominantly less than 10 μm.

The conditions that lead to loess accumulation (summarised by Smith et al. 2002) are: first, atmospheric and ground conditions in the source area that are conducive to deflation of material that can be carried in suspension; second, a specific combination of atmospheric and ground conditions that encourage the preferential deposition of silts from all the particle sizes being carried by the wind; and third, prolonged duration of these conditions, during which time there is minimal reworking of deposits or additions of material from other sources via different transport mechanisms.

Given these conditions and the model of Pye and Tsoar (Fig. 20.8) we might expect that material is being entrained and exported from the arid cores of the deserts, and deposited once moisture and vegetation cover are great enough to trap the material in the semi-arid areas at the desert margins. These deposits are commonly termed 'peri-desert loess'. Until the 1980s these peri-desert loesses were frequently omitted from or poorly represented on world maps of loess deposits partly because the deposits were not recognised as wind-blown dust. In Australia, for example, Butler (e.g. 1956) felt unable to use the term 'loess' for deposits that he recognised as wind blown and instead used the local term ‘parna’ for what he described as ‘wind-blown clays’. Consequently ‘these so-called parna soils were relegated to the position of a Southern Hemisphere curiosity, rather than representing early examples of desert loess’ (Hesse and McTainsh 2003), a situation addressed by Haberlah (2007).

The recognition in the 1970s and 1980s that deserts provided efficient machines for generating suspendible material and the desert margins were ideal environments to trap that material led to a flurry of reports of peri-desert loess from around the world. The thickest loess deposits are in China, where up to 300 m of loess has been deposited over the past 2.5 million years and at least some of these deposits are thought to have a desert origin. Other peri-desert loesses are much thinner than the Chinese loess or than their high-latitude counterparts but they are now reported from the margins of all the world’s major deserts (Table 20.5). At least part of the debate about peri-desert loess has been concerned with whether the deposited material was generated in the desert or whether the desert was merely an intermediate stopping-off point on the way from origin to loess deposit. Either way, of course, the desert environment has played an important part in the creation of peri-desert loess deposits (Figs. 20.10 and 20.11).

The largest accumulations of loess on the planet are in China (Fig. 20.12) where they cover 440,000 km^2 (Liu 1985) and appear to have built up over at least the last 22 million years according to Guo and co-workers (Guo et al. 2002, Hao et al. 2008). One view has been that the three northwestern inland basins (the Junggar Basin, the Tarim Basin and the Qaidam Basin) have been important source areas for the loess. In contrast, Sun (2002) argued that the gobi (stony desert) in southern Mongolia and the adjoining gobi and sand deserts (the Badain Jaran Desert, Tengger Desert, Ulan Buh Desert, Hobq Desert and Mu Us Desert) in China, rather than the three inland basins, were the dominant source areas. However, although Sun regarded these gobi and sand deserts as the main source regions, he suggested that they served as dust and silt holding areas rather than dominant producers and that the Gobi Altay, Hangayn and Qilian mountains may have been the initial sources. It appears that the material of the right size is created by a variety of processes in the mountains but is entrained by the wind from the desert basins and then deposited as loess on the Chinese loess plateau. The considerable thickness of the Chinese loess may therefore be the consequence of a particular combination of rapid uplift of the Tibetan Plateau and surrounding mountain ranges, high rates of sediment production and supply to adjacent basins, a strong northwesterly and westerly wind regime, and the existence of effective dust traps downwind of the source regions (Pye 1995). The Libyan deserts have also been regarded as ‘mixing points’ for dust before being re-entrained (O’Hara et al. 2006).

Table 20.5 Examples of peri-desert loess locations

Location	Reference
Sahara (northern margin) ~ Matmata Plateau, Tunisia	Coudé-Gaussen and Rognon 1988, Dearing et al. 2001
Sahara (northern margin) ~ Jebel Gharbi, Libya	Giraudi 2005
Sahara (southern margin) Nigeria	McTainsh 1987
Namib Desert, Namibia	Blümel 1982, Brunotte and Sander 2000
Arabia UAE	Goudie et al. 2000, Edgell 2006
Australia	Butler 1956, Hesse et al. 2003
Peru	Eitel et al. 2005
Yemen	Nettleton and Chadwick 1996
Negev Desert, Israel	Ginzbourg and Yaalon 1963, Yaalon and Dan 1974, Yaalon and Ganor 1975

Elsewhere conditions at desert margins do not favour the development of such thick loess deposits. For example, Breuning-Madsen and Awadzi (2005) argued that deposition rates of Harmattan dust in Ghana simply were not great enough to develop loess deposits. There may be a number of reasons for this, but clearly for thick loess deposits to develop there needs to be an effective trap downwind and nearby a considerable source of dust. In particular, Pye (1995) and Mason et al. (1999) have drawn attention to the importance of topographic effects on the distribution of loess. Loess tends to accumulate in basins and at the foot of mountains, both because of aerodynamic effects and because of post-depositional reworking by water and gravity.

Loess in the Palaeoenvironmental Record

The importance of the deposition of airborne dust in the palaeoclimatic record was recognised by the establishment of the DIRTMAP project (e.g. Kohfeld

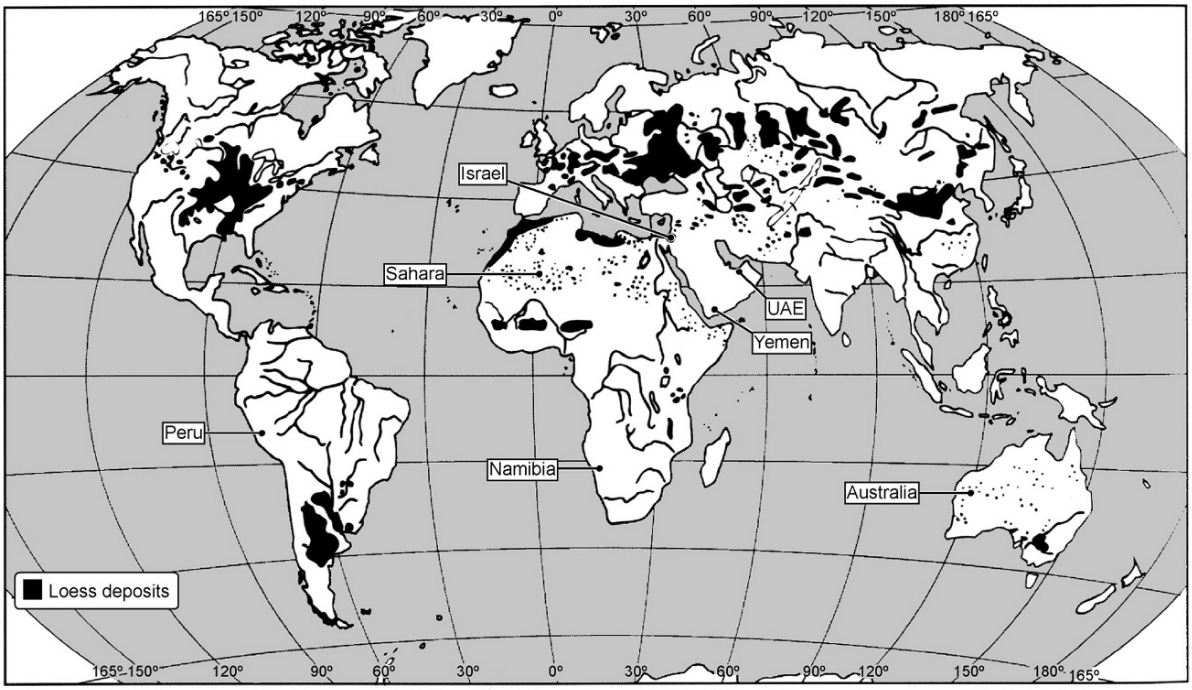

Fig. 20.10 World map showing major loess locations (after Livingstone and Warren 1996) and the major deserts

Fig. 20.11 Loess deposits in (**a**) China (courtesy Frank Eckardt) and (**b**) Tunisia

and Harrison 2001) which created a database of records of dust deposition both from contemporary monitoring and in the sedimentary record. Where loess has been deposited it has usually accumulated over a considerable period of time and therefore provides a very important terrestrial record of past environments (Sun et al. 2000). Rates of loess deposition appear to have varied quite considerably. Derbyshire (2003) suggested that 'some marine and terrestrial sedimentary records suggest that rates of aeolian deposition during the Last Glacial Maximum (LGM) may have been up to 10 times greater than those of the present (see summary by Kohfeld and Harrison 2001)', and Hesse and McTainsh (2003) reported that the dust flux from Australia to the oceans was at least three times greater at the LGM than in the Holocene. Pye (1987) estimated that loess was accumulating at 0.5–3.0 mm a^{-1} at the LGM. Global climate changes have affected not only the dust fluxes and rates of dust deposition, but also changed the location of sources and sinks. In China, for instance, Derbyshire (2003) noted that during the LGM, China's dryland margins

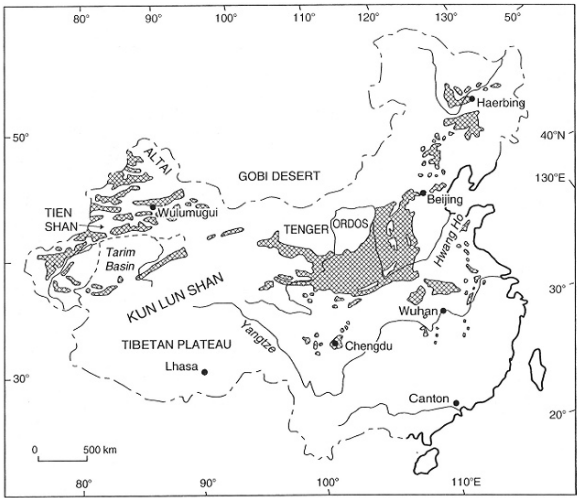

Fig. 20.12 Map of loess deposits in China (from Goudie and Middleton, 2006)

advanced several hundred kilometres south and east of those of the Holocene climatic optimum, affecting the winter monsoon intensity and influencing loess accumulation rates (Ding et al. 1999, Porter 2001).

This record of dust deposition has become particularly important especially as luminescence dating techniques have developed since the 1980s. Luminescence dating allows the dating of minerals such as quartz and feldspar, both important constituents of loess, and has therefore removed the reliance on finds of organic material required for radiocarbon dating. There is now a plethora of luminescence dates that have been provided for the loess in China, but dated sections have also been provided for other peri-desert areas, such as the Matmata Plateau in Tunisia (Dearing et al. 2001) and bordering the northern Atacama Desert in Peru (Eitel et al. 2005). Some of the issues associated with luminescence dating of loess were discussed by Singhvi et al. (2001).

As noted above, peri-desert loess is frequently coarser than its temperate and higher latitude counterparts, but its particle size is far from homogeneous. Chinese dust is typically bimodal representing a far-travelled, poorly sorted, fine component and a more localised, well-sorted component (Sun et al. 2004). The variability of particle size has been seen as a useful indicator of past environmental conditions.

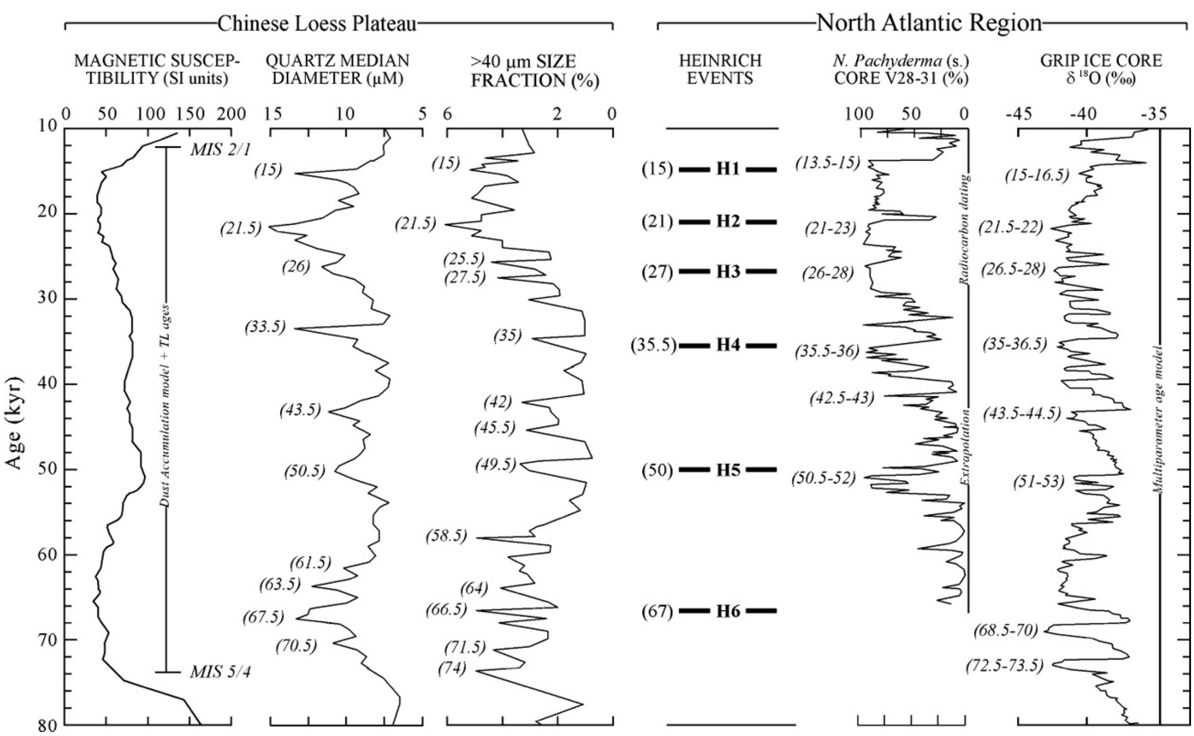

Fig. 20.13 Variation through a profile in the Chinese Loess Plateau of particle size and magnetic susceptibility plotted alongside variations in oxygen isotope ratios for the North Atlantic (from Porter and An 1995)

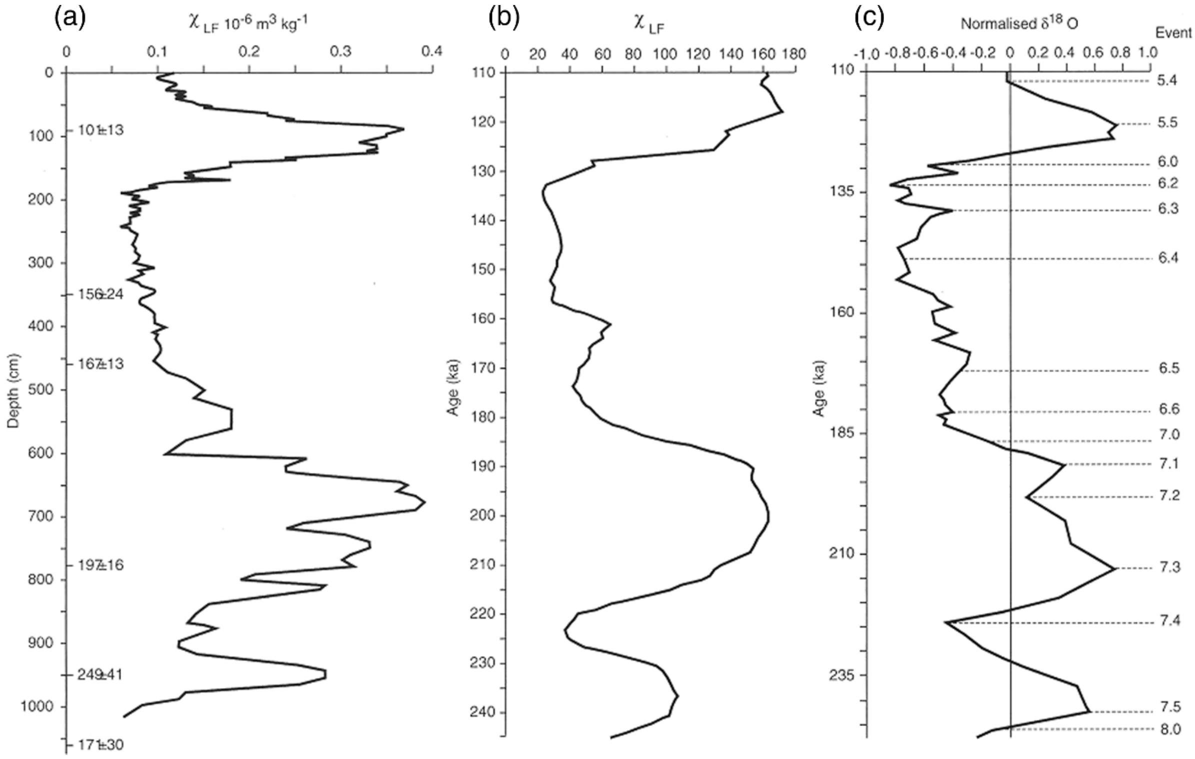

Fig. 20.14 Magnetic susceptibility records from the loess on the Matmata Plateau plotted against depth (**a**) with a similar record from the Chinese loess (Kukla et al. 1988) (**b**), compared with the oxygen isotope record (**c**) (Martinson et al. 1987) (from Dearing et al. 2001)

Porter and An (1995), for example, suggested that particle size (particularly as measured by the percentage >40 μm) was a proxy for the rate of deposition of the Chinese loess and therefore coarser particles indicated times when the capacity of winds to carry dust increased. They linked this measure of particle-size variability to oxygen isotope stages and were able to propose that windier periods occurred during colder periods in the marine record (linked to Heinrich iceberg-discharge events in the north Atlantic) (Fig. 20.13). Similar links have been made by, for example, Nugteren et al. (2004) and Sun and Huang (2006), who related particle-size variation in the northwestern Chinese Loess Plateau through the last interglacial to half-precessional cycles in insolation. Feng and Wang (2006), however, suggested that some caution was required because a number of factors contributed to the particle size distribution and it might not be possible to use particle size as a direct proxy for climate, particularly the strength of the winter monsoon.

Much of the effort in loess studies has concentrated on the recognition of palaeosols in the loess sequences. Generally the palaeosol has been seen as an indicator of surface stability and zero or reduced dust input, probably associated with greater humidity, whereas un-weathered loess without apparent soil development has been viewed as evidence of increased dust input and associated with periods of aridity. Often the palaeosols have been recognised using measures of magnetic susceptibility, increased susceptibility being taken as an indication of secondary mineralisation by pedogenic processes. However, Kemp (1999, 2001) was keen to urge caution when using palaeosols as palaeoenvironmental indicators particularly because of problems associated with reworking or syn-depositional alteration of palaeosols. Frequently, continuous deposition has been the basis for developing the Quaternary climate record from the Chinese loess. The assumption has been that loess sequences represent the whole record and that either the loess is accumulating or the ground surface is stable and a soil is developing. The reality is likely to be much more complicated with episodes of erosion when loess has been removed and the possibility of episodes when loess deposition is contemporaneous with soil formation. For example, work by Stevens et al. (2006), based on luminescence dates from sample points closely-spaced within a section, suggested that Chinese loess deposition has been episodic rather than continuous. Notwithstanding Kemp's concerns, there are a large number of studies of loess palaeosol sequences using mineral magnetic measures from the Chinese loess (see the reviews by Porter 2001 and Liu et al. 2007). There are few from other peri-desert loess areas although Dearing et al. (2001) provided magnetic data in combination with luminescence dates to place development of the Matmata Plateau loess in Tunisia alongside the oxygen isotope record in stages 8–5 (Fig. 20.14).

Conclusion

Plainly the past couple of decades have seen a very considerable increase in the interest in mineral dust generated in deserts. Although desert geomorphologists have long been aware of dust generation in deserts, the monitoring and accurate recording of dust entrainment, transport and deposition has been a relatively recent phenomenon. In part this has been made possible by improving technology. Unlike other geomorphological processes, the movement of dust often does not leave clear landform evidence of erosion and frequently the deposition of even quite large amounts of sediment is spread as thin mantles over large areas. Remote sensing has helped greatly in monitoring the transport of dust, such that we now have a much clearer impression of the global budget for dust transport. In addition the technology used in field studies has been developed to allow good measurements of dust fluxes at ground level to be ascertained. Added to these studies, the recent past has seen a widespread recognition that the dust generated within the world's deserts is frequently exported by the wind and deposited at their margins, either as recognisable loess deposits or as dust-sized additions to semi-arid soils. Some of this work on dust fluxes has been driven by a desire to comprehend more fully the role of suspended dust (of which deserts are one source) in the global atmospheric system, but there has also been a general recognition that we need to develop a fuller understanding of the aeolian transport of dust if we are to explain the movement of clastic material in deserts.

References

Assallay, A.M., Rogers, C.D.F. and Smalley, I.J. 1996. Engineering properties of loess in Libya. *Journal of Arid Environments* **32**, 373–386.

Assallay, A.M., Rogers, C.D.F., Smalley, I.J. and Jefferson, I.F. 1998. Silt: 2-62 mu m, 9-4 phi. *Earth-Science Reviews* **45**, 61–88.

Baba, J. and Komar, P.D. 1981. Measurements and analysis of settling velocities of natural quartz sand grains. *Journal of Sedimentary Petrology* **51**, 631–640.

Balme, M. and Greeley, R. 2006. Dust devils on Earth and Mars. *Reviews of Geophysics* **44** (3), RG3003, doi:10.1029/2005RG000188.

Berthier, S., Chazette, P., Couvert, P., Pelon, J., Dulac, F., Thieuleux, F., Moulin, C. and Pain, T. 2006. Desert dust aerosol columnar properties over ocean and continental Africa from LIDAR in-Space Technology Experiment (LITE) and Meteosat synergy. *Journal of Geophysical Research – Atmospheres* **111** (D21), D21202.

Betzer, P.R., Carder, K.L., Duce, R.A., Merrill, J.T., Tindale, N.W., Uematsu, M., Costello, D.K., Young, R.W., Feely, R.A., Breland, J.A., Berstein, R.E. and Greco, A.M. 1988. Long-range transport of giant mineral aerosol particles. *Nature* **336**, 568–571.

Biscaye, P.E. 1997. Asian provenance of glacial dust (stage 2) in the Greenland Ice Sheet Project 2 Ice Core, Summit, Greenland. *Journal of Geophysical Research – Oceans* **102**, 26765.

Brazel, A.J. and Nickling, W.G. 1986. The relationship of weather types of dust storm generation in Arizona (1955–1980). *Journal of Climatology* **6**, 255–275.

Breshears, D.D., Whicker, J.J., Johansen, M.P. and Pinder, J.E. 2003. Wind and water erosion and transport in semi-arid shurbland, grassland and forest ecosystems: quantifying dominance of horizontal wind-driven transport. *Earth Surface Processes and Landforms* **28**, 1189–1209.

Breuning-Madsen, H. and Awadzi, T.W. 2005. Harmattan dust deposition and particle size in Ghana. *Catena* **63**, 23–38.

Bryant, R.G. 2003. Monitoring hydrological controls on dust emissions: preliminary observations from Etosha Pan, Namibia. *Geographical Journal* **169**, 131–141.

Bullard, J.E., Baddock, M.C., McTainsh, G.H. and Leys, J.F. 2008. Sub-basin scale dust source geomorphology detected using MODIS. *Geophysical Research Letters*, **35**, 15, L15404.

Bullard, J.E. and McTainsh, G.H. 2003. Aeolian-fluvial interactions in dryland environments: scales, concepts and Australia case study. *Progress in Physical Geography* **27**, 471–501.

Bullard, J.E., McTainsh, G.H. and Pudmenzky, C. 2007. Factors affecting the rate and nature of fine particle production by aeolian abrasion. *Sedimentology* **54**, 1169–1182.

Butler, B.E. 1956. Parna – an aeolian clay. *Australian Journal of Science* **18**, 145–151.
Chan, Y.-C., McTainsh, G., Leys, J., McGowan, H. and Tews, K. 2005. Influence of the 23 October 2002 dust storm on the air quality of four Australian cities. *Water, Air and Soil Pollution* **164**, 329–348.
Chen, W. and Fryrear, D.W. 2002. Sedimentary characteristics of a haboob dust storm. *Atmospheric Research* **61**, 75–85.
Chester, R., Elderfield, H. and Griffin, J.J. 1971. Dust transported in north-east and south-east trade winds in Atlantic Ocean. *Nature* **233**, 474–476.
City of Cape Town Air Quality Monitoring Network. 2002. http://www.capetown.gov.za/airqual/episodes/default.asp?yr=2002 Last accessed 5 December 2006.
Clarke, M.L. and Rendell, H.M. 1998. Climate change impacts on sand supply and the formation of desert sand dunes in the southwest USA. *Journal of Arid Environments* **39**, 517–531.
Coe, M.T. and Foley, J.A. 2001. Human and natural impacts on the water resources of the Lake Chad basin. *Journal of Geophysical Research* **106**, 3349–3356.
Cooke, R.U. 1970. Stone pavements in deserts. *Annals of the American Association of Geographers* **60**, 560–577.
Coudé-Gaussen, G. 1987. The perisaharan loess: sedimentological characterization and paleoclimatical significance. *GeoJournal* **15**, 177–183.
Coudé-Gaussen, G. and Rognon, P. 1988. The Upper Pleistocene loess of southern Tunisia: a statement. *Earth Surface Processes and Landforms* **13**, 137–151.
Dearing, J.A., Livingstone, I.P., Bateman, M.D. and White, K.H. 2001. Palaeoclimate records from OIS 8.0-5.4 recorded in loess-palaeosol sequences on the Matmata Plateau, southern Tunisia, based on mineral magnetism and new luminescence dating. *Quaternary International* **76–77**, 43–56.
Derbyshire, E. 2003. Loess, and the Dust Indicators and Records of Terrestrial and Marine Palaeoenvironments (DIRTMAP) database. *Quaternary Science Reviews* **22**, 1813–1819.
Ding, Z.L., Sun, J.M., Rutter, N.W., Rokosh, D. and Liu, T.S. 1999. Changes in sand content of loess deposits along a north-south transect of the Chinese Loess Plateau and the implications for desert variations. *Quaternary Research* **52**, 56–62.
Duce, R.A. 1980. Long-range atmospheric transport of soil dust from Asia to the tropical north Pacific – temporal variability. *Science* **209**, 1522.
Edgell, H.S. 2006. *Arabian Deserts: nature, origin and evolution*, Berlin: Springer.
Eitel, B., Hecht, S., Machtle, B. and Schukraft, G. 2005. Geoarchaeological evidence from desert loess in the nazca-palpa region, southern Peru: Palaeoenvironmental changes and their impact on pre-columbian cultures. *Archaeometry* **47**, 137–158.
Evans, J.E. and Tokar Jr, F.J. 2000. Use of SEM/EDS and X-ray diffraction analyses for sand transport studies, Lake Eyrie, Ohio. *Journal of Coastal Research* **16**, 926–933.
Feng, Z.D. and Wang, H.B. 2006. Geographic variations in particle size distribution of the last interglacial pedocomplex S1 across the Chinese Loess Plateau: their chronological and pedogenic implications. *Catena* **65**, 315–328.
Fryrear, D.W. 1986. A field dust sampler. *Journal of Soil and Water Conservation* **41**, 117–120.
Gerson, R. and Amit, R. 1987. Rates and modes of dust accretion and deposition in an arid region-the Negev, Israel. In: Desert Sediments: Ancient and Modern, Spec. Publ. Geol. Soc.
Ghienne, J.-F., Schuster, M., Bernard, A., Duringer, P. and Brunet, M. 2002. The Holocene giant Lake Chad revealed by digital elevation models. *Quaternary International* **87**, 81–85.
Giles, J. 2005. The dustiest place on earth. *Nature* **434**, 816–819.
Gillette, D.A. and Walker, T.R. 1977. Characteristics of airborne particles produced by wind erosion of sandy soil, High Plains of West Texas. *Soil Science* **123**, 97–110.
Ginoux, P., Chin, M., Tegen, I., Prospero, J.M., Holben, B., Dubovik, O. and Lin, S.J. 2001. Sources and distributions of dust aerosols simulated with the GOCART model. *Journal of Geophysical Research – Atmospheres* **106** (D17), 20255–20273.
Ginoux, P., Prospero, J.M., Torres, O., and Chin, M. 2004. Long-term simulation of global dust distribution with the GOCART model: correlation with the North Atlantic Oscillation. *Environmental Modelling and Software*, **19**, 113–128.
Ginzbourg, D., Yaalon, D.M. 1963. Petrography and origin of the loess in the Be'er Sheva basin. *Israel Journal of Earth Sciences* **12**, 68–70.
Giraudi, C. 2005. Eolian sand in peridesert northwestern Libya and implications for Late Pleistocene and Holocene Sahara Expansions. *Palaeogeography Palaeoclimatology Palaeoecology* **218**, 159–171.
Gomes, L., Bergametti, G., Coudé-Gaussen, G. and Rognon, P. 1990. Submicron desert dusts: a sandblasting process. *Journal of Geophysical Research* **95**, 13927–13935.
Goossens, D. 1995. Field experiments of aeolian dust accumulation on rock fragment substrata *Sedimentology* **42**, 391–402.
Goossens, D. 2005. Quantification of the dry aeolian deposition of dust on horizontal surfaces: an experimental comparison of theory and measurements. *Sedimentology* **52**, 859–873.
Goossens, D. 2007. Bias in grain size distribution of deposited atmospheric dust due to the collection of particles in sediment catchers. *Catena* **70**, 16–24.
Goossens, D. 2008. Relationships between horizontal transport flux and vertical deposition flux during dry deposition of atmospheric dust particles. *Journal of Geophysical Research-Earth Surface* **113** (F2), art.no. F02S13.
Goossens, D. and Offer, Z.Y. 2000. Wind tunnel and field calibration of six aeolian dust samplers. *Atmospheric Environment* **34**, 1043–1057.
Goossens, D. and Rajot, J.L. 2008. Techniques to measure the dry aeolian deposition of dust in arid and semi-arid landscapes: a comparative study in west Niger. *Earth Surface Processes and Landforms* **33**, 178–195.
Goudie, A.S., Cooke, R.U. and Evans, I. 1970. Experimental investigation of rock weathering by salts. *Area* **4**, 42–48.
Goudie, A.S. and Middleton, N.J. 2006. *Desert Dust in the Global System*. Berlin: Springer.
Goudie, A.S. and Watson, A. 1991. The shape of desert sand dune grains. *Journal of Arid Environments* **4**, 185–190.
Goudie, A.S., Viles, H.A. and Parker, A.G. 1997. Monitoring of rapid salt weathering in the central Namib desert using limestone blocks. *Journal of Arid Environments* **37**, 581–598.

Goudie, A.S., Parker, A.G., Bull, P.A., White, K. and Al-Farraj, A. 2000. Desert loess in Ras Al Khaimah, United Arab Emirates. *Journal of Arid Environments* **46**, 123–135.

Grini, A. and Zender, C.S. 2004. Roles of saltation, sandblasting, and wind speed variability on mineral dust aerosol size distribution during the Puerto Rican Dust Experiment (PRIDE). *Journal of Geophysical Research* **109**, D07202, doi:10.1029/2003JD004233.

Grousset, F.E. and Biscaye, P.E. 2005. Tracing dust sources and transport patterns using Sr, Nd and Pb isotopes. *Chemical Geology* **222**, 149–167.

Grousset, F.E., Ginoux, P., Bory, A. and Biscaye, P.E. 2003. Case study of a Chinese dust plume reaching the French Alps. *Geophysical Research Letters* **30**, Art. No. 1277.

Guo, Z.T., Ruddiman, W.F., Hao, Q.Z., Wu, H.B., Qiao, Y.S., Zhu, R.X., Peng, S.Z., Wei, J.J., Yuan, B.Y., Liu, T.S. 2002. Onset of Asian desertification by 22 Myr ago inferred from loess deposits in China. *Nature* **416**, 159–163.

Haberlah, D. 2007. A call for Australian loess. *Area* **39** 224–229.

Hao, Q.Z., Oldfield, F., Bloemendal, J., Guo, Z.T. 2008. The magnetic properties of loess and paleosol samples from the Chinese Loess Plateau spanning the last 22 million years. *Palaeogeography Palaeoclimatology Palaeoecology* **260**, 389–404.

Herman, J.R., Bhartia, P.K., Torres, O., Hsu, C., Seftor, C. and Celarier, E. 1997. Global distribution of UV-absorbing aerosols from Nimbus-7/TOMS data. *Journal of Geophysical Research* **102**, 16,911–16,922.

Hesse, P.P. and McTainsh, G.H. 1999. Last glacial maximum to Early Holocene wind strength in the mid-latitudes of the Southern Hemisphere from aeolian dust in the Tasman Sea. *Quaternary Research* **52**, 343–349.

Hesse, P.P. and McTainsh, G.H. 2003. Australian dust deposits: modern processes and the Quaternary record. *Quaternary Science Reviews* **22**, 2007–2035.

Hesse, P.P., Humphreys, G.S., Smith, B.L., Campbell, J. and Peterson, E.K. 2003. Age of loess deposits in the Central Tablelands of New South Wales. *Australian Journal of Soil Research* **41**, 1115–1131.

Hess, G.D. and Spillane, K.T. 1990. Characteristics of dust devils in Australia. *Journal of Applied Meteorology* **27**, 305–317.

Hunt, J.C.R. and Weber, A.H. 1979. A Lagrangian statistical analysis of diffusion from a ground-level source in a turbulent boundary layer. *Quarterly Journal of the Royal Meteorological Society* **105**, 423–443.

Jickells, T.D., An, Z.S., Andersen, K.K., Baker, A.R., Bergametti, G., Brooks, N., Cao, J.J., Boyd, P.W., Duce, R.A., Hunter, K.A., Kawahata, H., Kubilay, N., laRoche, J., Liss, P.S., Mahowald, N., Prospero, J.M., Ridgwell, A.J., Tegen, I. and Torres, R. 2005. Global iron connections between desert dust, ocean biogeochemistry and climate. *Science* **308**, doi.10.1126.

Kalashnikova, O.V. and Sokolik, I.N. 2004. Modeling the radiative properties of nonspherical soil-derived mineral aerosols. *Journal of Quantitative Spectroscopy and Radiative Transfer* **87**, 137–166.

Kaufman, Y.J., Koren, I., Remer, L.A., Tanre, D., Ginoux, P. and Fan, S. 2005. Dust transport and deposition observed from the Terra-Moderate Resolution Imaging Spectroradiometer (MODIS) spacecraft over the Atlantic Ocean, *Journal of Geophysical Research – Atmospheres* **110** (D10), D10S12.

Kemp, R.A. 1999. Micromorphology of loess-paleosol sequences: a record of paleoenvironmental change. *Catena* **35**, 179–196.

Kemp, R.A. 2001. Pedogenic modification of loess: significance for palaeoclimatic reconstructions. *Earth-Science Reviews* **54**, 145–156.

Kohfeld, K.E. and Harrison, S.P. 2001. DIRTMAP: the geological record of dust. *Earth-Science Reviews* **54**, 81–114.

Koren, I. and Kaufman, Y.J. 2004. Direct wind measurements of Saharan dust events from Terra and Aqua satellites. *Geophysical Research Letters* **31**, L06122, doi:10.1029/2003GL019338.

Kuenen, P.H. 1960. Experimental abrasion 4: aeolian action. *Journal of Geology* **68**, 427–449.

Kukla, G., Heller, F., Ming, L.X., Chun, X.T., Sheng, L.T. and Sheng, A.Z. 1988. Pleistocene climaes in China dated by magnetic susceptibility. *Geology* **16**, 811–814.

Kumar, R., Jefferson, I.F., O'Hara-Dhand, K. and Smalley, I.J. 2006. Controls on quartz silt formation by crystalline defects. *Naturwissenschaften* **93**, 185–188.

Lancaster, N. 2002. Flux of eolian sediment in the McMurdo Dry Valleys, Antarctica: a preliminary assessment. *Arctic, Antarctic and Alpine Research* **34**, 3, 318–323.

Lautridou, J.-P. and Ozouf, J.C. 1982. Experimental frost shattering: fifteen years of research at the Centre de Geomorphologie du CNRS. *Progress in Physical Geography* **6**, 215–232.

Le Roux, J.P. 2002. Shape entropy and settling velocity of natural grains. *Journal of Sedimentary Research* **73**, 363–366.

Leslie, L.M. and Speer, M.S. 2006. Modelling dust transport over central eastern Australia. *Meteorological Applications* **13**, 141–167.

Liu, T.S. 1985. *Loess and the Environment*. Beijing: China Ocean Press.

Liu, Q.S., Deng, C.L., Torrent, J. and Zhu, R.X. 2007. Review of recent developments in mineral magnetism of the Chinese loess. *Quaternary Science* **26** 368–385.

Li, X.Y., Liu, L.Y., Gao, S.Y., Shi, P.J., Zou, X.Y., Hasi, E. and Yan, P. 2005. Aeolian dust accumulation by rock fragment substrata: influence of number and textural composition of pebble layers on dust accumulation, *Soil and Tillage Research* **84**, 139–144.

Livingstone, I. and Warren, A. 1996. *Aeolian Geomorphology: an introduction*. London: Longman.

Luo, C., Mahowald, N.M. and del Corral, J. 2003. Sensitivity study of meteorological parameters on mineral aerosol mobilization, transport and distribution. *Journal of Geophysical Research* **108** (D15), 4447. Doi:10.1029/2003JD003483.

McFadden, L.D., Wells, S.G. and Jercinovich, M.J. 1987. Influences of eolian and pedogenic processes on the origin and evolution of desert pavements. *Geology* **15**, 504–508.

McTainsh, G.H. 1985. Dust processes in Australia and West Africa: a comparison. *Search* **16**, 104–106.

McTainsh, G. 1987. Desert loess in northern Nigeria. *Zeitschrift für Geomorphologie* **31**, 145–165.

McTainsh, G.H. and Pitblado, J.R. 1987. Dust storms and related phenomena measured from meteorological records in Australia. *Earth Surface Processes and Landforms* **12**, 415–424.

McTainsh, G.H., Leys, J.F. and Nickling, W.G. 1999. Wind erodibility of arid lands in the Channel Country of western Queensland, Australia. *Zeitschrift für Geomorphologie NF Supplementband* **116**, 113–130.

McTainsh, G., Chan, Y.-C, McGowan, H., Leys, J. and Tews, K. 2005. The 23rd October 2002 dust storm in eastern Australia: characteristics and meteorological conditions. *Atmospheric Environment* **39**, 1227–1236.

Mahowald, N.M., Bryant, R.G., del Corral, J. and Steinberger, L. 2003. Ephemeral lakes and desert dust sources. *Geophysical Research Letters* **30**, art.no. 1074.

Martinson, D.G., Pisias, N.G., Hays, J.D., Imbrie, J., Moore, Jr. T.C. and Shackleton, N.J. 1987. Age dating and the orbital theory of the ice ages: development of a high-resolution 0 to 3,000,000-year chronostratigraphy. *Quaternary Research* **27**, 1–29.

Mason, J.A., Nater, E.A., Zanner, C.W. and Bell, J.C. 1999. A new model of topographic effects on the distribution of loess. *Geomorphology* **28**, 223–236.

Mattsson, J.O., Nihlén, T. and Yue, W. 1993. Observations of dust devils in a semi-aird district of southern Tunisia. *Weather* **48**, 359–363.

Meloni, D., di Sarra, A., Di Iorio, T. and Fiocco, G. 2005. Influence of the vertical profile of Saharan dust on the visible direct radiative forcing. *Journal of Quantitative Spectroscopy and Radiative Transfer* **93**, 397–413.

Michaels, T.I. 2006. Numerical modeling of Mars dust devils: albedo track generation. *Geophysical Research Letters* **33**, art no. L19S08.

Miller, R.L., Tegen, I. and Perlwitz, J. 2004. Surface radiative forcing by soil dust aerosols and the hydrologic cycle. *Journal of Geophysical Research* **106** (D16), 18193. Doi:10.1029/2003JD004085.

Miller, S.D., Kuciauskas, A.P., Liu, M., Ji, Q., Reid, J.S., Breed, D.W., Walker, A.L., and Al Mandoos, A. 2008. Haboob dust storms of the southern Arabian Peninsula. *Journal of Geophysical Research* **113** (D1), art.no. D01202.

Morales, C. (ed.) 1979. *Saharan Dust: mobilization, transport, deposition*. New York: Wiley.

Moreno, T., Querol, X., Castillo, S., Alastuey, A., Cuevas, E., Herrmann, L., Mounkaila, M., Elvira, J. and Gibbons, W. 2006. Geochemical variations in aeolian mineral particles from the Sahara-Sahel Dust Corridor. *Chemosphere* **65**, 261–270.

Nalpanis, P. 1985. Saltating and suspended particles over flat and sloping surfaces. II. Experiments and numerical simulations. In: Barndorff-Nielsen, O.E., Møller, J.T., Romer-Rasnussen, K., Willetts, B.B. (eds.) *Proceedings of an International Workshop on the Physics of Blown Sand*. Department of Theoretical Statistics, Institute of Mathematics, University of Aarhus, Memoir 8, 1, 37–66.

Nettleton, W.D. and Chadwick, O.A. 1996. Late Quaternary, redeposited loess-soil developmental sequences, south Yemen. *Geoderma* **70**, 21–36.

Nickling, W.G. and Gillies, J.A. 1993. Dust emission and transport in Mali, Africa. *Sedimentology* **40**, 859–868.

Nickling, W.G., McTainsh, G.H. and Leys, J.F. 1999. Dust emissions from the Channel Country of western Queensland, Australia. *Zeitschrift für Geomorphologie NF Supplementband* **116**, 1–17.

Nugteren, G., Vandenberghe, J., Van Huissteden, J.K. and An, Z. 2004. A Quaternary climate record based on grain size analysis from the Luochuan loess section on the Chinese Loess Plateau, China. *Global and Planetary Change* **41**, 167–183.

Offer, Z.Y. and Goossens, D. 2001. Ten years of aeolian dust dynamics in a desert region (Negev Desert, Israel): analysis of airborne dust concentration, dust accumulation and high-magnitude dust events. *Journal of Arid Environments* **47**, 211–249.

O'Hara, S.L., Clarke, M.L. and Eltrash, M.S. 2006. Field measurements of desert dust deposition, *Atmospheric Environment* **40**, 3881–3897.

Oke, A.M.C., Dunkerley, D. and Tapper, N.J. 2007a. Willy-willies in the Australian landscape: the role of key meteorological variables and surface conditions in defining frequency and spatial characteristics. *Journal of Arid Environments* **71**, 201–215.

Oke, A.M.C., Dunkerley, D. and Tapper, N.J. 2007b. Willy-willies in the Australian landscape: sediment transport characteristics. *Journal of Arid Environments* **71**, 216–228.

Perlwitz, J., Tegen, I. and Miller, R.L. 2001. Interactive soil dust aerosol model in the GISS GCM: 1. Sensitivity of the soil dust cycle to radiative properties of soil dust aerosols. *Journal of Geophysical Research* **106** (18), 167–18,192.

Piketh, S.J., Tyson, P.D., Steffen, W. 2000. Aeolian transport from southern Africa and iron fertilization of marine biota in the South Indian Ocean. South African Journal of Science **96**, 244–246.

Porter, S.C. 2001. Chinese loess record of monsoon climate during the last glacial-interglacial cycle. *Earth-Science Reviews* **54**, 115–128.

Porter, S.C. and An, Z. 1995. Correlation between climate events in the North Atlantic and China during the last glaciation. *Nature* **375**, 305–308.

Prospero, J.M. 1999. Long-term measurements of the transport of African mineral dust to the southeastern United States: implications for regional air quality. *Journal of Geophysical Research – Atmospheres* **104**, 15917.

Prospero, J.M., Glaccum, R.A. and Nees, R.T. 1981. Atmospheric transport of soil dust from Africa to South-America. *Nature* **289**, 570–572.

Prospero, J.M., Ginoux, P., Torres, O., Nicholson, S.E. and Gill, T.E. 2002. Environmental characterization of global sources of atmospheric soil dust identified with the Nimbus-7 Total Ozone Mapping Spectrometer (TOMS) absorbing aerosol product. *Reviews of Geophysics* **40**, doi: 10.1029/2000RG000095.

Pye, K. 1987. *Aeolian dust and dust deposits*. London: Academic Press.

Pye, K. 1995. The nature, origin and accumulation of loess. *Quaternary Science Reviews* **14**, 653–667.

Pye, K. and Paine, A.D.M. 1984. Nature and source of aeolian deposits near the summit of Ben Arkle, northwest Scotland. *Geologie en Mijnbouw* **63**, 13–18.

Pye, K. and Sperling, C.H.B. 1983. Experimental investigation of silt formation by static breakage processes: the effect of temperature, moisture and salt on quartz dune sand and granitic regolith. *Sedimentology* **30**, 49–62.

Pye, K. and Tsoar, H. 1987. The mechanics and geological implications of dust transport and deposition in deserts with particular reference to loess formation and dune sand diagenesis in the northern Negev, Israel, In: Frostick, L. and Reid I. (eds.) *Desert Sediments: Ancient and Modern*. *Geological Society Special Publication* **35**, 139–156.

Reheis, M.C. and Kihl, R. 1995. Dust deposition in southern Nevada and California, 1984–1989: relations to climate, source area and source lithology. *Journal of Geophysical Research* **100** (D5), 8893–8918.

Reynolds, R.L., Reheis, M., Yount, J. and Lamothe, P. 2006. Composition of aeolian dust in natural traps on isolated surfaces of the central Mojave Desert – Insights to mixing, sources, and nutrient inputs. *Journal of Arid Environments* **66**, 42–61.

Rivera Rivera, N.I., Bleiweiss, M.P., Hand, J.L. and Gill, T.E. 2006. Characterization of dust storm sources in southwestern US and northwestern Mexico using remote sensing imagery. 14th Conference on Satellite Meteorology and Oceonography, American Meteorological Society, February 2006, 17pp.

Rostagno, C.M. 2007. Fire and wind erosion in the northeastern portion of Patagonia. Abstract. Multidisplinary workshop on southern South American dust, Puerto Madryn, Argentina, October 3–5, 2007.

Sarnthein, M. and Koopman, B. 1980. Late Quaternary deep-sea record of northwest African dust supply and wind circulation. *Palaeoecology of Africa* **12**, 239–253.

Satheesh, S.K. and Moorthy, K.K. 2005. Radiative effects of natural aerosols: a review. *Atmospheric Environment* **39**, 2089–2110.

Shao, Y. 2000. *Physics and Modelling of Wind Erosion*. Dordrecht: Kluwer Academic.

Shao, Y. 2001. A model for mineral dust erosion. *Journal of Geophysical Research* **106**, 20, 239–20,254.

Shao, Y., McTainsh, G.H., Leys, J.F. and Raupach, M.R. 1993a. Efficiency of sediment samplers for wind erosion measurement. *Australian Journal of Soil Research* **31**, 519–532.

Shao, Y., Raupach, M.R. and Findlater, P.A. 1993b. Effect of saltation bombardment on the entrainment of dust by wind. *Journal of Geophysical Research* **98**, 12719–12726.

Shao, Y., Raupach, M.R. and Leys, J.F. 1996. A model for predicting aeolian sand drift and dust entrainment on scales from paddock to region. *Australian Journal of Soil Research* **34**, 309–342.

Shao, Y., Leys, J.F., McTainsh, G.H. and Tews, K. 2007. Numerical simulation of a dust event in Australia. *Journal of Geophysical Research-Atmospheres* **112**, D8, D08207.

Singhvi, A.K., Bluszcz, A., Bateman, M.D. and Rao, M.S. 2001. Luminescence dating of loess-palaeosol sequences and coversands: methodological aspects and palaeoclimatic implications. *Earth-Science Reviews* **54**, 193–211.

Smalley, I. 1995. Making the material: the formation of silt-sized primary mineral particles for loess deposits. *Quaternary Science Reviews* **14**, 645–651.

Smalley, I. and Derbyshire, E. 1990. The definition of 'ice-sheet' and 'mountain' loess. *Area* **22**, 300–301.

Smith, B.J., Warke, P.A. and Moses, C.A. 2000. Limestone weathering in a contemporary arid environment: a case study from southern Tunisia. *Earth Surface Processes and Landforms* **25**, 1343–1354.

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

Sokolik, I.N., Winker, D.M., Bergametti, G., Gillette, D.A., Carmichael, G., Kaufman, Y.J., Gomes, L., Schuetz, L. and Penner, J.E. 2001. Introduction to special section: outstanding problems in quantifying the radiative impacts of mineral dust. *Journal of Geophysical Research – Atmospheres* **106** (D16), 18015–18027.

Soderberg, K. and Compton, J.S. 2007. Dust as a nutrient source for fynbos ecosystems, South Africa. *Ecosystems* **10**, 550–561.

Sow, M., Goossens, D. and Rajot, J.L. 2006. Calibration of the MDCO dust collector and of four versions of the inverted Frisbee dust deposition sampler. *Geomorphology* **82**, 360–375.

Stevens, T., Armitage, S.J., Lu, H.Y. and Thomas, D.S.G. 2006. Sedimentation and diagenesis of Chinese loess: implications for the preservation of continuous, high-resolution climate records. *Geology* **34**, 849–852.

Stoertz, G. and Ericksen, G. 1974. Geology of salars in northern Chile. *USGS Professional Paper*, **811**, 65pp.

Sun, D.H., Bloemendal, J., Rea, D.K., An, Z.S., Vandenberghe, J., Lu, H.Y., Su, R.X. and Liu, T.S. 2004. Bimodal grain-size distribution of Chinese loess, and its palaeoclimatic implications. *Catena* **55**, 325–340.

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

Sun, J. M. and Huang, X.G. 2006. Half-precessional cycles recorded in Chinese loess: response to low-latitude insolation forcing during the Last Interglaciation. *Quaternary Science Reviews* **25**, 1065–1072.

Sun, J., Kohfeld, K.E. and Harrison, S.P. 2000. *Records of aeolian dust deposition on the Chinese Loess Plateau during the Late Quaternary*. Jena, Germany: Max Planck Institute for Biogeochemistry.

Tanaka, T.Y. 2005. Possible transcontinental dust transport from North Africa and the Middle East to East Asia. *Atmospheric Environment* **39**, 3901.

Tanaka, T.Y. and Chiba, M. 2006. A numerical study of the contributions of dust source regions to the global dust budget. *Global Planetary Change* **52**, 88–104.

Tegen, I. and Fung, I. 1994. Modeling of mineral dust in the atmosphere: sources, transport and optical thickness. *Journal of Geophysical Research – Atmospheres* **99** (D16), 22897–22914.

Tegen, I., Heinold, B., Todd, M., Helmert, J., Washington, R. and Dubovik, O. 2006. Modelling soil dust aerosol in the Bodélé depression during the BoDEx campaign. *Atmospheric Chemistry and Physics Discussions* **6**, 4171–4211.

Tetzlaff, G. and Peters, M. 1986. Deep-sea sediments in the eastern equatorial Atlantic of the African coast and meteorological flow patterns over the Sahel. *Geologische Rundschau* **75**, 71–79.

Towner, M.C., Ringrose, R.J., Patel, M.R. Balme, M., Metzger, S.M., Greeley, R. and Zarnecki, J.C. 2004. A close encounter with a terrestrial dust devil. *Lunar and Planetary Science* **XXXV**, 1259.

Viles, H.A. and Goudie, A.S. 2000. Weathering, geomorphology and climatic variability in the central Namib desert. In: McLaren, S.J. and Kniveton, D.R. (eds.) *Linking Climate Change to Land Surface Change*. Amsterdam: Kluwer, 65–82.

Viles, H.A. and Goudie, A.S. 2007. Rapid salt weathering in the coastal Namib Desert: implications for landscape development. *Geomorphology* **85**, 49–62.

Walker, T.R. 1979. Red colour in eolian sand. In: E.D. McKee (ed.) *A Study of Global Sand Seas*. USGS Professional Paper **1052**, 62–81.

Washington, R. and Todd, M.C. 2005. Atmospheric controls on mineral dust emission from the Bodélé Depression, Chad: the role of the low level jet. *Geophysical Research Letters* **32**, L17701, doi:10.1029/2005GL023597.

Washington, R., Todd, M., Middleton, N.J. and Goudie, A.S. 2003. Dust-storm source areas determined by the Total Ozone Monitoring Spectrometer and surface observations. *Annals of the Association of American Geographers* **93**, 297–313.

Washington, R., Todd, M.C., Lizcano, G., Tegen, I., Flamant, C., Koren, I., Ginoux, P., Engelstaedter, S., Britsow, C.S., Zender, C.S., Goudie, A.S., Warren, A. and Prospero, J.M. 2006. Links between topography, wind, deflation, lakes and dust; the case of the Bodélé Depression, Chad. *Geophysical Research Letters* **33**, L09401, doi:10.1029/2006GL025827.

Wentworth, C.K. 1922. A scale of grade and class terms for clastic sediments. *Journal of Geology* **30**, 377–392.

Werner, M., Tegen, I., Harrison, S.P., Kohfeld, K.E., Prentice, I.C., Balkanski, Y., Rodhe, H. and Roelandt, C. 2002. Seasonal and interannual variability of the mineral dust cycle under present and glacial climate conditions. *Journal of Geophysical Research* **107** (D24), 4744. Doi: 10.1029/2002JD002365.

Whalley, W.B., Marshall, J.R. and Smith, B.J. 1982. Origin of desert loess from some experimental observations. *Nature* **300**, 433–435.

Whalley, W.B., Smith, B.J., McAlister, J.J. and Edwards, A.J. 1987. Aeolian abrasion of quartz particles and the production of silt-size fragments: preliminary results. In: L. Frostick and I. Reid (eds.) *Desert Sediments: ancient and modern*. Geological Society Special Publication **35**, 129–138.

Wigner, K.A. and Peterson, R.E. 1987. Synoptic climatology of blowing dust on the Texas South Plains. *Journal of Arid Environments* **13**, 199–209.

Wopfner, H. and Twidale, C.R. 2001. Australian desert dunes: wind rift or depositional origin? *Australian Journal of Earth Sciences* **48**, 239–244.

Wright, J.S. 1993. Non-glacial origins of loess-sized quartz silt. Unpublished PhD thesis, 1993, The Queen's University of Belfast, 502 pp.

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

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

Wright, J.S. 2007. An overview of the role of weathering in the production of quartz silt. *Sedimentary Geology* **202**, 337–351.

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

Yaalon, D.H. 1991. Mountain loess is not a suitable term. *Area* **23**, 255–256.

Yaalon, D.H. and Dan, J. 1974. Accumulation and distribution of loess-derived deposits in the semi-arid desert-fringe areas of Israel. *Zeitschrift für Geomorphologie Supplementband* **20**, 27–32.

Yaalon, D.H. and Ganor, E. 1975. Rate of aeolian dust accretion in the Mediterranean and desert fringe environments of Israel. 19th International Congress on Sedimentology Theme 2, 169–174.

Yuan, W. and Zhang, J. 2006. High correlations between Asian dust events and biological productivity in the western North Pacific. *Geophysical Research Letters* **33**, L07603.

Zender, C.S., Bian, H. and Newman, D. 2003a. Mineral Dust Entrainment and Deposition (DEAD) model: description and 1990s dust climatology. *Journal of Geophysical Research* **108** (D17), 4543, doi: L10.1029/2002/JD002775.

Zender, C.S., Newman, D. and Torres, O. 2003b. Spatial heterogeneity in aeolian erodibility: uniform, topographic, geomorphic and hydrologic hypoetheses. *Journal of Geophysical Research* **108**, doi: 10.1029/ 2002JD003039.

Zender, C.S., Miller, R.L. and Tegen, I. 2004. Quantifying mineral dust mass budgets: terminology, constraints and current estimates. *Eos* **85** (48), 30 November 2004.