The History and Nature of Wind Erosion in Deserts

Andrew S. Goudie

St. Cross College, University of Oxford, Oxford, OX1 3LZ, United Kingdom;
email: andrew.goudie@stx.ox.ac.uk

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Key Words

pans, dust, deflation, yardangs, stone pavements

Abstract

Recently, the importance of wind erosion has been reevaluated. Many low-angle surfaces, developed on susceptible materials, possess closed depressions (pans), and these often have a distinctive morphology and lunette dunes on their lee sides. It has also become apparent that in drylands there are extensive areas of stone pavements, some of which have been molded by deflation; however, it is now recognized that other horizontal sorting processes have played a role, especially movement of fines by overland flow. Vertical sorting associated with wetting and drying, salt heave, bioturbation, frost action, and dust accretion have also played a role. In hyperarid areas, especially with unidirectional winds, bedrock outcrops and old lake beds have been molded to give aerodynamic forms, called yardangs. The study of dust storms by analysis of climatological data and remote sensing has revealed the importance of deflation, especially in hyperarid areas with centripetal drainage. Deflation hot spots, such as the Bodélé Depression, have been identified. Analysis of ice and ocean cores and loess deposits has indicated that wind activity was greater during glacial phases than now. There is, however, great variability in the importance of wind at the present day, depending on the wind velocity characteristics of different deserts.

Pan: a closed desert topographic depression developed on low-angle surfaces as a result of processes such as deflation and salt weathering

Lunette: a crescent-shaped accumulation of eolian sediment that is deflated from a pan and then deposited as a positive topographic feature on its lee side

Deflation: the process by which the wind removes fine material from a surface

INTRODUCTION

Recently, the power of wind erosion in deserts has become more apparent and is discussed here in relation to the formation of such phenomena as closed depressions, inverted relief, stone pavements, yardangs, and dust storms. The review also examines the evidence for wind activity having been of greater intensity in the past—notably during the Pleistocene glacials. The review is not concerned with the theory of the wind erosion of soils—an area that has recently been reviewed by Shao (2000).

PANS

Pans, closed depressions, occur in the Pampas and Patagonia of Argentina; east Tierra del Fuego; the High Plains of the United States; the interior of southern Africa; east Mozambique; and extensive tracts of Australia, West Siberia, and Manchuria (Goudie & Wells 1995) (**Figure 1**). Some pan fields consist of tens of thousands of basins—for example, those of Western Australia (Boggs et al. 2006). They are a feature of low-angle surfaces on which the development of integrated drainage is limited. They result from a variety of processes, of which eolian excavation is probably the most important, and lunette dunes, composed of excavated material, frequently (though not invariably) occur on their lee sides (Sabin & Holliday 1995). On the southern High Plains, there are some 25,000 pans (playas), only 1100 of which have lunettes, but elsewhere this proportion is higher. In West Siberia, for example, of 8900 pans, 71.4% had lunettes (P. Kent, personal communication). Dating of lunettes by luminescence techniques has potential for establishing when deflation has been active (e.g., Lawson & Thomas 2002, Telfer & Thomas 2006).

Pans tend to occur in areas of semiaridity, as the following list of mean annual rainfalls (in mm) suggests:

Atbasar, Kazakhstan	320
Bahia Blanca, Argentina	550
Duolan, China	389
Katanning, W. Australia	480
Keetmanshoop, Namibia	143
Kimberley, South Africa	420
Lubbock, United States	450
Mildura, Australia	260
Mongu, Zambia	960
Rio Grande, Tierra del Fuego	330

In addition to a rainfall control on their distribution, there is a strong control exercised by surface materials. The West Siberian pans occur on Neogene and Quaternary materials, those of southern Africa on Ecca shales and Kalahari sands, those of the High Plains of the United States on Ogallala beds and various shales and sandstones, and those of the Yorke Peninsula (Australia) on Permian clays and sands. Elsewhere, pans have developed in dune fields, on coastal plains, and in deflated lake beds. That is, they only develop in classic and abundant form on materials that are relatively lightly lithified and that are fine-grained enough to be susceptible to deflation.

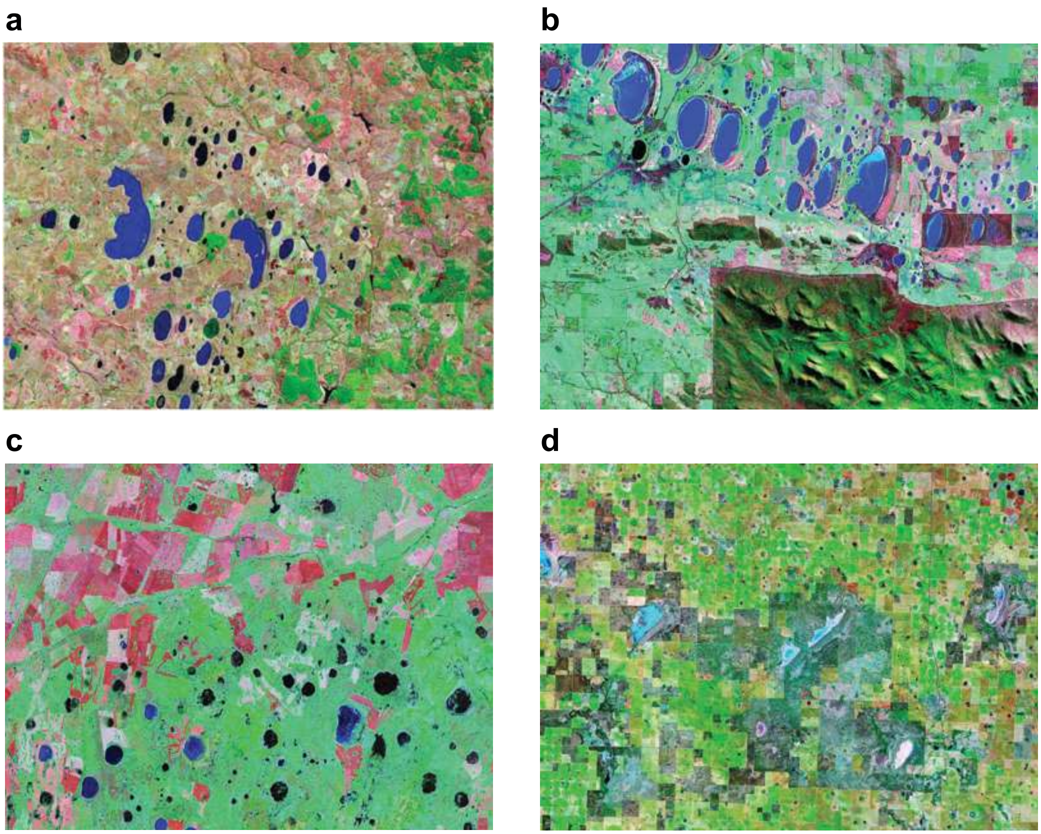


Figure 1

Landsat 7 images of pans (courtesy NASA Zulu): (*a*) Lake Chrissie area, South Africa (the largest pan is approximately 5.5 km long); (*b*) Stirling Range, Western Australia (the largest pan is approximately 3.5 km long); (*c*) West Siberian Plain, Kazakhstan (the largest pan is approximately 2 km long); (*d*) Llano Estacado, High Plains, United States (the largest pan is approximately 5 km long).

That pans are at least in part eolian is indicated by their distinctive morphology, their orientation with regard to prevailing wind directions, the presence of lunettes (composed in part of sediment deflated from pan floors) on their lee sides, and observations on the ground and from space of plumes blowing from their surfaces. That said, other processes may contribute to the development of closed depressions, including solution and animal excavation (Gustavson et al. 1995). Moreover, they have undergone periods of modification during humid phases (Holliday et al. 1996, Holliday 1997).

Goudie (1999b) developed a model of pan development that recognized the variety of formative influences, and classified them into various categories. First of all, there is the predisposing condition of low precipitation, which means that vegetation cover is limited and that deflation can occur. Low precipitation also means that animals concentrate at pans, causing trampling and overgrazing, which also promote deflation. Aridity also promotes salt accumulation so that salt weathering can attack the fine-grained bedrock in which the depression lies, producing rock flour that can be evacuated by wind (Goudie & Viles 1997). These are accentuating processes, which enlarge hollows, whether they are formed by other initial formative processes, such as solution of carbonate and gypsum beds or tectonics. It is also important if pans are to develop that the initial surface depression is not obliterated by the action of integrated or effective fluvial systems. Nonrestraining conditions that limit fluvial activity are low-angle slopes, episodic desiccation and dune encroachment, the presence of dolerite intrusions, and tectonic disturbance. In addition, it is important that pans do not lie in areas of active sand accumulation, which might cause infilling of an existing hollow, although pans can and do develop in interdune depressions.

Yardang: a primarily wind-abraded ridge with an aerodynamic form that can range in size from the centimeter to the kilometer scale

YARDANGS

Yardangs are wind-abraded ridges of cohesive material (Hedin 1903). They result from a number of processes, including wind abrasion, deflation, fluvial incision, desiccation cracking, weathering, and mass movements (McCauley et al. 1977, Laity 1994, Goudie 1999b). The presence of polished, fluted, and sand-blasted slopes, and the undercutting of the steep windward face and lateral slopes (Peel 1966) indicates the importance of abrasion. Their streamlined shapes are also suggestive of the role of wind abrasion. Deflation is probably important on softer sediments, such as desiccated lake beds. Fluvial incision, caused by occasional intense rain storms, may play a role in the early stages, creating initial depressions along which wind can be channeled, but excessive fluvial erosion would eventually lead to their obliteration. Once established, the steep sides of yardangs may encourage mass movements, whereas in playa environments, salt weathering and wetting and drying may prepare material for wind evacuation.

Yardangs range in scale from microyardangs (small centimeter-scale ridges), through mesoyardangs (forms that are meters in height and length) to megayardangs (features that may be tens of meters high and some kilometers long) (Cooke et al. 1993, pp. 296–97; Halimov & Fezer 1989). These megayardangs are ridge and swale features of regional extent (Mainguet 1972) (**Figure 2**), and their curving trajectories in the Sahara closely mirror the trajectories of the winter trade winds, both being largely N to S over Egypt, and almost E to W in the lee of Tibesti.

The significance of yardangs is now being reevaluated, largely because analysis of satellite images has shown that megayardangs are extensive in the coastal deserts of Peru and Chile, the high Andes, the Sahara and Libyan Deserts, the northern and southern Namib, Saudi Arabia (Vincent & Kattan 2006), the Lut Desert of Iran, the Seistan Basin of Afghanistan, and the Taklamakan and Turpan basins of Asia.

Goudie (2006) attempted to analyze the factors that determine the global distribution of megayardangs and came up with the following relationships:

1. Large yardangs occur in hyperarid areas where rainfall totals are less than 50 mm per annum, whereas pans, as we have seen, become more significant where

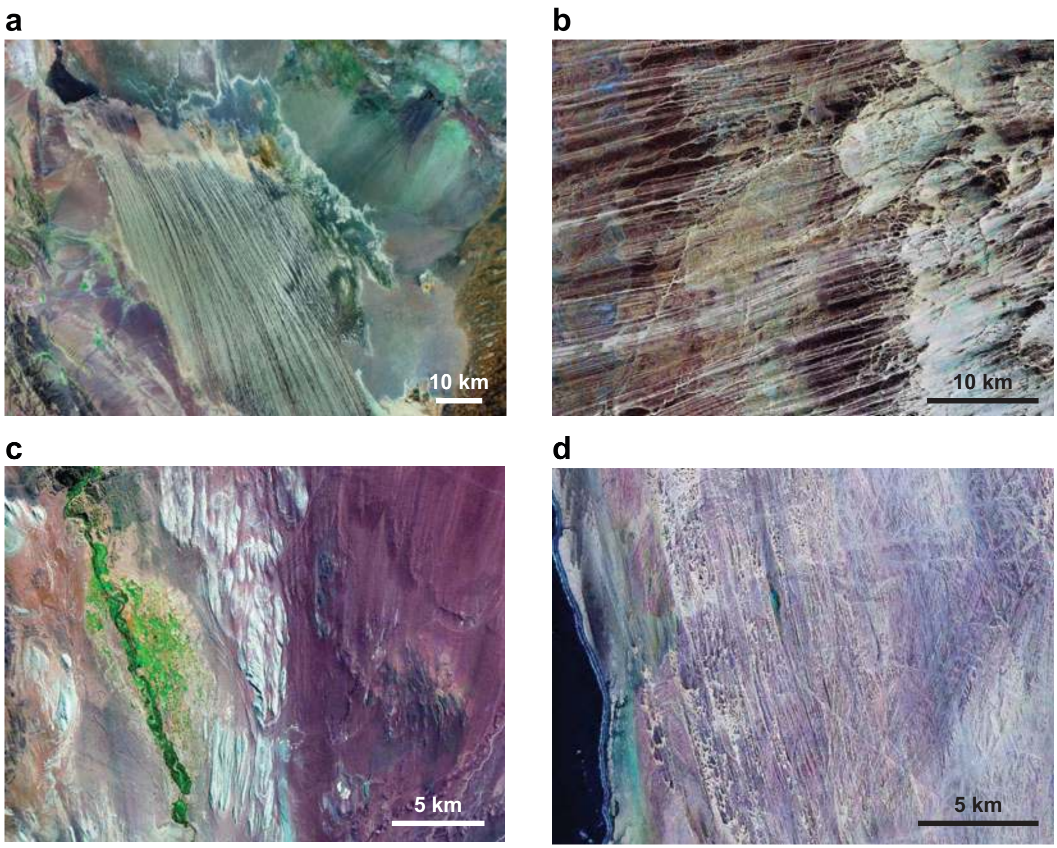


Figure 2

Landsat 7 images of megayardangs (courtesy NASA Zulu): (*a*) Lut Desert, Iran; (*b*) central Sahara; (*c*) Peruvian Desert; (*d*) northern Namib, Namibia.

rainfall is between 150 and 500 mm per year. Large yardangs occur in dry areas where deflation is at a maximum, vegetation cover is minimal, and sand abrasion occurs.

2. Yardangs do not occur in sites of active dune accumulation (e.g., sedimentary basins), although they do occur in former pluvial lake depressions. Basins are areas of sand sea development rather than eolian erosion. Yardangs do not occur in areas with massive alluvial fans, in truly mountainous areas, or in areas with integrated drainage.

3. Megayardangs occur in areas with unidirectional or narrow bimodal wind directions, as is made evident by their frequent association with barchans (e.g., Namib, Peru, Chile, Egypt), a dune form that only occurs where winds are relatively constant in direction. It is only with such constant wind directions that forms can develop parallel to the prevailing wind. They sometimes occur

upwind of sand seas, in areas where sand transport occurs (e.g., the Lut, N. and S. Namibia, Saudi Arabia).

4. Megayardangs occur in relatively homogeneous rocks without complex structures (e.g., sandstones), but with jointing along which incision can occur. They do not solely occur in "soft rocks," as has so often been asserted, although chalk landscapes in the Libyan Desert have been dramatically scoured and molded into mound and trough topography. Indeed, small lake deposits only lead to small forms, although the large spreads of diatomite and other lake deposits in the Bodélé depression have features approximately 10 m high (Bristow et al. 2007).
5. There is little evidence as to the age and rate of formation of megayardangs developed in hard rocks, although there is evidence that smaller yardangs can be excavated to depths of some meters in mid-Holocene lake and swamp deposits (Embabi 1999, Goudie 1999b, Goudie et al. 1999).

Stone pavement: also called desert pavement, it is an armored surface composed of a thin mosaic of rock fragments that is set in or on a matrix of finer material

Megayardangs may be persistent features that have been shaped over millions of years, not least by high-velocity glacial age trade winds (Rea 1994). The Atacama and the Namib, for example, are deserts that originated in pre-Pleistocene times, possibly in the Miocene or earlier (Goudie 2002), so there has been an extended time for yardang formation. However, some yardangs, like those of parts of S. Namibia, may be relict features, currently undergoing weathering rather than abrasion, because of past changes in sand flow associated with sea-level changes (Corbett 1993).

INVERTED RELIEF

Deflation of desert surfaces can cause relief inversion to occur. This is the case, for example, with old gravel channels, which, being resistant to deflation, may be left upstanding as sinuous ridges (Oviatt et al. 2003). Raised channels of this type are widespread in Arabia (Maizels 1987). Equally, valley calcretes, because of their ability to create resistant low-lying deposits, may be left upstanding during periods of increased eolian activity (McClaren 2004).

STONE (DESERT) PAVEMENTS

Another major landform that has been attributed to wind erosion is the stone (or desert) pavement. These are armored surfaces composed of a mosaic of fragments, usually only one or two stones thick, set on or in matrices of finer material. They occur on debris mantles, pediments, alluvial terraces, and fans. Pavements are formed by a range of processes that cause coarse particle concentration at the surface: (*a*) the classic mechanism of deflation of fine material, (*b*) removal of fines by surface runoff and/or creep, and (*c*) processes causing upward migration of coarse particles to the surface. In addition, it has become increasingly clear that pavements may evolve in close association not only with eolian erosion but also with dust deposition and soil-profile differentiation caused by weathering.

Deflation. Pavements have usually been explained by deflation of fine material, which leaves a surface lag of coarse particles. Loose fines can indeed be removed by wind, as the frequency of dust storms, together with experimental work, has demonstrated (Symmons & Hemming 1968). However, arguments can be advanced against the ubiquitous application of the deflation hypothesis. Although desert winds are probably strong enough to move most loose fine material, in semiarid areas vegetation may reduce winds to below the requisite threshold velocities, and vegetable litter may protect loose fines. Even more important is the fact that loose fine material may be bound at the surface into a thin biological or rain splash crust. Deflation will occur only if this carapace is broken by, for example, prior wind abrasion, animal activities, or vehicular traffic.

Water sorting. Experiments show some pavements are often composed, at least in part, of coarse particles that remain after finer materials have been dislodged and removed by raindrop erosion and running water (Wainwright et al. 1995). At cleared pavement sites in the Mojave, Cooke (1970) showed that surface fines and some buried coarse particles were removed by surface runoff and collected in sediment traps down slope. Plainly, the role of sheetfloods should not be ignored as a horizontal transport mechanism (Williams & Zimbelman 1994).

Upward clast migration. The relative scarcity of coarse particles in the upper soil profile suggest that stones may have moved upward to the surface by cycles of freezing and thawing, wetting and drying, or salt heave. Indeed, laboratory experiments have shown that alternate freezing and thawing of saturated mixed sediments causes coarse particles to migrate toward the ground surface (Mackay 1984). Although sorting by freeze/thaw action may seem improbable in warm deserts today, it cannot be ruled out, especially in high-altitude deserts. It may also have been more effective during cooler, moister periods of the Quaternary. A more effective and widespread upward migration mechanism is wetting and drying of the soil (McFadden et al. 1987), and various experiments have shown this to be a feasible mechanism (Cooke 1970). Subsidiary mechanisms of upward migration may be salt heave (Horta 1985, Searl & Rankin 1993) and the activity of soil fauna. Whether or not bioturbation causes pavement formation or disruption is still a matter of debate. On the one hand, churning and burrowing may bring fine material to the surface where it can be deflated, whereas on the other hand, the process may cause coarse particles to sink and for homogenization to occur. With higher rainfall, for example, during pluvials, it is probable that pavement disruption predominates.

The role of dust deposition. McFadden et al. (1987), followed by Anderson et al. (2002),Valentine & Harrington (2006), and Meadows et al. (2007), have argued that in the Mojave, pavements are born at the surface, and that the individual pavement stones were never buried. The pavements in this area were said to evolve as follows: First, volcanic rocks are broken up by weathering into clasts, a process accelerated by the ingress of eolian silt into cracks and its subsequent volume changes with wetting and drying. Such clasts move by colluvial and alluvial processes from higher to lower areas that are already filled with silt and clay. Deposition of eolian material continues. The eolian mantle is altered pedogenically by, for example, leaching (Ikeda et al. 1998), to create a vesicular horizon. Finally, the clasts are maintained at the surface, while eolian deposition and pedogenesis continue. In this explanation, therefore, no eolian deflation or upward migration is necessary, although it is probable that both may occur from time to time. Indeed, many of the characteristics of pavement clasts, including their degree of splitting and their grain sizes, may be determined by weathering (Amit & Gerson 1986, Al-Farraj & Harvey 2000).

Dust storm: a dust-raising event that reduces horizontal visibility to 1000 m or less

Therefore, deflation may create pavements, although its operation is reduced by their presence. However, other processes contribute to their development, including surface runoff, weathering, heave, and dust accretion. They cannot, therefore, be thought of solely as products of eolian action.

DUST STORMS

Dust storms are a manifestation of wind activity in deserts (Goudie & Middleton 2006) (**Figure 3**). The Total Ozone Mapping Spectrometer (TOMS) has proved to be an effective instrument for detecting atmospheric dust (Herman et al. 1997, Prospero et al. 2002, Washington et al. 2003), but we also have global maps of aerosol optical thickness (a measure of aerosol column concentration) derived from satellites such as the Advanced Very High Resolution Radiometer (AVHRR) and the Moderate Imaging Spectroradiometer (MODIS) (see, for example, Chin et al. 2004, Ginoux et al. 2004, Yu et al. 2003). Different sensors do not always provide the same picture of dust storm activity, and Brooks et al. (2005) point to the difference in results obtained from TOMS compared with those derived as an infrared difference dust index (IDDI) from METEOSAT.

TOMS can be used to derive a semi-quantitative aerosol index (AI), values for which are linearly proportional to aerosol optical thickness. The world pattern of annual mean AI values has certain clear features. First, the largest area with high values is a zone that extends from the western coast of Africa through the Sahara to Arabia and southwest Asia (**Figure 4**). In addition, there is a large zone with high AI values in central Asia—the Tarim Basin. Australia has a relatively small zone, located in the Lake Eyre basin, whereas southern Africa has two zones, Mkgadikgadi and Etosha. In Latin America, there is only one easily identifiable zone—one of the great closed basins of the Bolivina Altiplano, the Salar de Uyuni. North America has only one relatively small zone with high values—the Great Basin.

The importance of these different dust hot spots can be gauged by looking not only at their areal extents but also at their relative AI values. **Table 1** lists the latter. This again brings out the dominance of the Sahara in particular and of the Old World deserts in general. The Southern Hemisphere and the Americas both have relatively low AI values. Those for the Bodélé Depression of the Sahara are around four times greater than those recorded for either the Great Basin of the United States or the Salar de Uyuni of Bolivia.

TOMS data has demonstrated the primacy of the Sahara, and has highlighted the importance of some other parts of the world's drylands. One characteristic

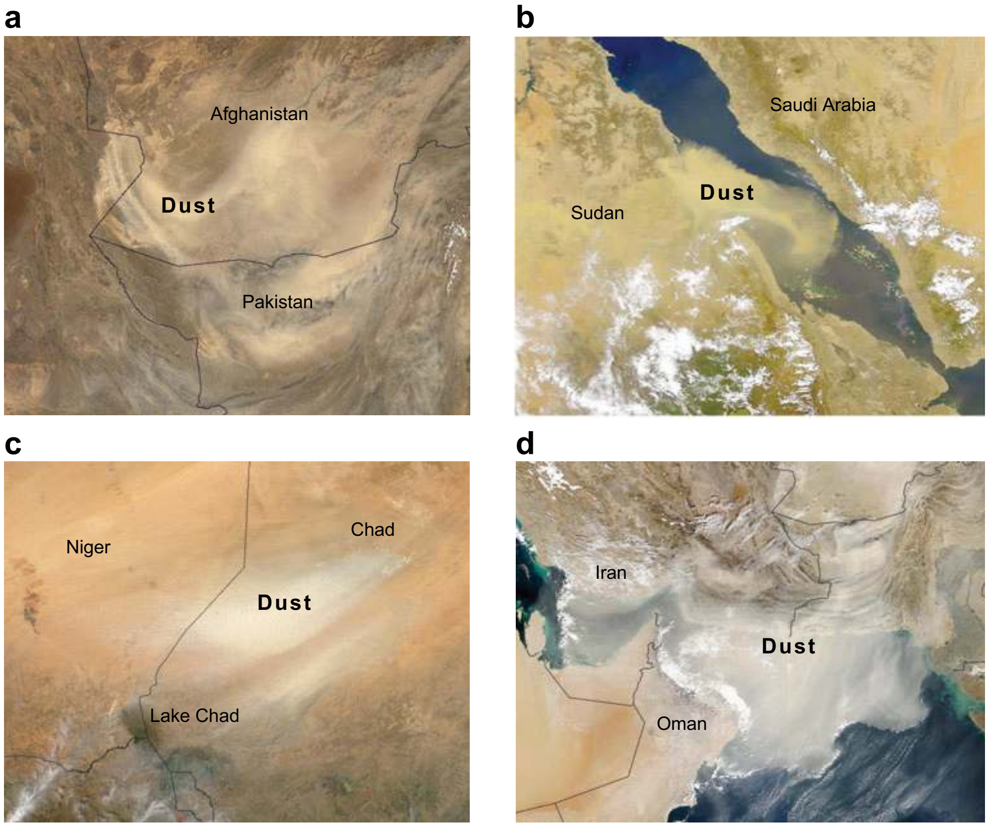


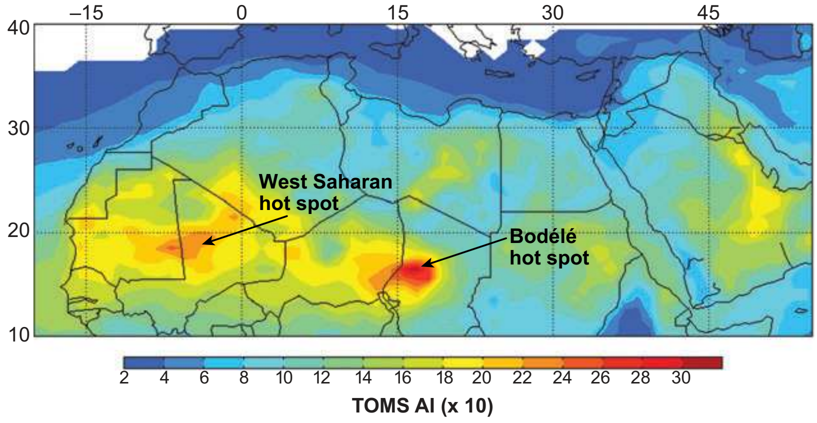
Figure 3

Satellite images of dust plumes: (*a*) Seistan Basin, Afghanistan (MODIS, June 16, 2004); (*b*) Tokar Delta and Red Sea (MODIS, September 1, 2004); (*c*) Bodélé Basin (MODIS, January 5, 2005); (*d*) Gulf of Oman and Makran Coast (SEAWIFS, December 13, 2003).

that emerges is the importance of large basins of internal drainage as dust sources (Engelstädter 2001) (Bodélé, Taoudenni, Tarim, Seistan, Eyre, Etosha, Mkgadikgadi, Etosha, Uyuni, and the Great Salt Lake), although there will be great variability in the importance of individual basins, depending on their surface state (e.g., whether they are wet or dry, compacted or puffy). Also, many sources are associated with deep and extensive alluvial deposits (Prospero et al. 2002) or with extensive piedmont alluvial fans (Wang et al. 2006). However, sand dune systems and sandy deserts are not good sources of fine dust. Furthermore, TOMS data indicate that major dust sources are very arid. The prime global source, Bodélé, has a mean annual rainfall of 17 mm, whereas the west Sahara source has between 5 and 100 mm. In Arabia, dust storms are most prevalent where the mean annual rainfall is less than 100 mm (Goudie & Middleton 2001), and the Taklamakan source in China has large areas with less than 25 mm.

Figure 4

Mean TOMS AI (aerosol index values) for 1980–1992 across northern Africa and into Arabia.



The main sources for Saharan dust used to be unclear (Herrmann et al. 1999, Goudie & Middleton 2001), but they include Bodélé in Niger and Chad; an area that comprises southern Mauritania, northern Mali, and central-southern Algeria; southern Morocco and western Algeria; the fringes of the Mediterranean Sea in Libya and Egypt (Koren et al. 2003); and northern Sudan (Brooks & Legrand 2000, Brooks et al. 2005). Bodélé is the world's largest dust source because of the existence of the topographically controlled low-level jet, channeled between the Tibesti and Ennedi Mountains, which deflates diatomites and other materials from a desiccated lake floor (Engelstädter et al. 2006, Koren et al. 2006). Some of the Saharan dust is carried thousands of kilometers to the Americas, Europe, and the Near East. Much is also moved by the NE trades over Nigeria and the Guinea zone to give the Harmattan haze (Breuning-Madsen & Awadzi 2005).

Estimates of global dust emissions show a large range (Prospero 1996a), of between 1000 and 3000 millions of tons per year (Cakmur et al. 2006). The largest source is the

Table 1 Maximum mean aerosol index (AI) values for major global dust sources as determined from TOMS

Location	AI value	Average annual rainfall (mm)
Bodélé Depression of 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 (United States)	>5	400

Table 2 Estimates of the source strength of the Sahara

Author(s)	Annual quantity (millions of tons year^{-1})
Jaenicke (1979)	260
Schütz et al. (1981)	260
Prospero (1996a,b)	170
Swap et al. (1996)	130–460
D'Almeida (1986)	630–710
Marticorena & Bergametti (1996)	586–665
Callot et al. (2000)	760
Ginoux et al. 2004)	1400
Miller et al. (2004)	517

Sahara, which probably contributes approximately half of the global total. Estimates of its annual emissions are shown in **Table 2**. They show a wide range that may reflect differences in modeling procedures, the timescales considered, and the areal extent of the source.

QUATERNARY DUST LOADINGS

At certain times during the Quaternary, such as the Last Glacial Maximum (LGM) at approximately 18–20 kyr ago (Mahowald et al. 1999), the world was very dusty, as indicated by extensive deposits of loess and the presence of large amounts of dust in ocean, ice, lake, and peat bog core sediments and in speleothems. Mahowald et al. (2006) have suggested that dust deposition rates at the LGM were 2.1–3.3 times the present levels.

The enhanced dustiness, especially during glacials, may relate to a larger sediment source (e.g., areas of glacial outwash), changes in wind characteristics both in proximity to ice caps and in the trade-wind zone (Ruddiman 1997), and the expansion of low-latitude deserts. However, it would be simplistic to attribute all cases of higher dust activity to greater aridity in source regions, for as Nilson & Lehmkuhl (2001) point out, this is but one factor, albeit important. Also important are changes in the trajectories of the major dust transporting winds, changes in the strength of winds in source regions, the balance between wet and dry deposition (which may determine the distance of dust transport), the degree of exposure of continental shelves in response to sea-level changes, and the presence of suitable vegetation to trap dust on land (Werner et al. 2002).

There is evidence that the strength of the trades may have intensified in the Pleistocene as a whole, and also during particular phases of the Pleistocene. In the late Pliocene (3.2 to 2.1 Ma), there may have been an increase in atmospheric circulation driven by a steeper pole-equator temperature gradient owing to the development of the bipolar cryosphere (Marlow et al. 2000). Within the Pleistocene, analysis of ocean cores has shown variability in the grain size characteristics of eolian inputs to the oceans which may be explained by variations in wind velocities

(Sarnthein & Koopmann 1980). In addition, studies of pollen, diatom, and phytolith influx have shown differences that can be explained in terms of wind velocity changes (Hooghiemstra 1989). Moreover, changes in upwelling intensity and oceanic productivity, established by analyses of benthic foraminifera, have been linked to changes in wind intensity (Loubere 2000).

Studies based on this evidence have indicated that NE trade wind velocities were higher during glacials, probably because of an intensified atmospheric circulation caused by an increased temperature gradient between the North Pole and the Equator owing to the presence of an extended Northern Hemisphere ice cap (Stein 1985, Ruddiman 1997, Kim et al. 2003). However, work off Namibia suggests that the SE trades were also intensified during glacials compared with interglacials (Stuut et al. 2002). Off NW Africa the highest velocities may have occurred during the last deglaciation rather than at the times of maximum ice (Moreno et al. 2001). Moreno & Canals (2004) attribute this to the lowering of North Atlantic sea surface temperatures during deglaciation because of glacial water releases. This, in turn, strengthened the North Atlantic high-pressure system, caused a high temperature difference between land and sea, and enhanced the trade wind system.

On the basis of cores from the Arabian Sea, Sirocko et al. (1991) suggested that dust additions were approximately 60% higher during glacials than in postglacial times, although there was a clear spike of enhanced dust activity at approximately 4000 years BP associated with a severe arid phase. Jung et al. (2004) also report on Holocene dust trends in the Arabian Sea, and suggest that dry, dusty conditions were established by 3.8 kyr BP.

Approximately 18,000 years ago, the amount of dust transported from the Sahara into the Atlantic was augmented by a factor of 2.5 (Tetzlaff et al. 1989, p. 198) and Australia contributed three times more dust to the SW Pacific (Hess & McTainsh 1999). Increased dust loadings may have stimulated increases in planktonic productivity on the South Australian continental margin (Gingele & De Deckker 2005). Dust fluxes appear generally to have been 2–4 times higher than at present (Grousset et al. 1998). By contrast, they appear to have been very low during the African Humid Period (AHP). From 14.8 to 5.5 ka, the mass flux off Cape Blanc was reduced by 47% (DeMenocal et al. 2000). This is confirmed by analyses of the mineral magnetics record from Lake Bosumtwi (Ghana), which suggests a high dust flux during the Last Glacial Period and a great reduction during the AHP (Peck et al. 2004).

Another source of information on rates of dust accretion is the record preserved in cores retrieved either from the polar ice caps or from high-altitude ice domes at lower latitudes. Thompson & Mosley-Thompson (1981) were among the first to point to the great differences in microparticle concentrations between the Late Glacial and the Postglacial. Briat et al. (1982) maintained that at Dome C there was an increase in microparticle concentrations by a factor of 10 to 20 during the last glacial stage, and they explain this by a large input of continental dust. The Dunde Ice Core from high Asia (Thompson et al. 1990) also shows high dust loadings in the Late Glacial and a sudden fall off at the transition to the Holocene. The Little Ice Age, however, was also a time of relatively high dust activity (Yang et al. 2006). Yancheva et al. (2007) have shown how Chinese dust loadings in the Pleistocene and Holocene have responded to changes in the position of the Intertropical Convergence Zone and the strength of the winter monsoon. Over the past 1500 years, changes in the deposition of eolian material in the Aral Sea can be related to changes in the strength and frequency of winds, which in turn are associated with changes in the intensity of the Siberian high pressure system over central Asia (Sorrel et al. 2007).

Within the last glacial, dust activity both in Europe and Greenland appears to have varied in response to millennial scale climatic events (Dansgaard-Oeschger events and Bond cycles) (Rousseau et al. 2002). In the Epica core from Antarctica the dust flux rose by a factor of ~25, ~20, and ~12 in glacials [oxygen isotope stages (OIS) 2, 4, and 6, respectivly] compared with interglacials (the Holocene and OIS 5.5). Delmonte et al. (2004) found in Antarctic cores that during the LGM dust concentrations were between 730 and 854 ppb, whereas during the Antarctic Cold Reversal (14.5–12.2 kyr BP) they had fallen to 25–46 ppb, and from 12.1 to 10 kyr BP they were between 7 and 18 ppb. Isotopic studies suggest that most dust was derived from Argentina (Iriondo 2000). In the case of Greenland, a major source of dust in cold phases was east Asia (Svensson et al. 2000).

In the Pleistocene, dust was deposited as loess. Rates of loess accumulation range between 22 and 4000 mm 1000 year^{-1}. Pye (1987, p. 265) believed that at the LGM loess was probably accumulating at between 500 and 3000 mm 1000 year^{-1}, and suggests that "dust-blowing on this scale was possibly unparalleled in previous Earth history." Pye also argued that rates increased through the Quaternary, with rates in China, central Asia, and Europe averaging approximately 20–60 mm 1000 year^{-1} during Matuyama time and about 90–260 mm 1000 year^{-1} during the Brunhes epoch (post 0.78 Ma ago).

For China, Kohfeld & Harrison (2003) indicate that in glacials (e.g., OIS 2) eolian mass accumulation rates were approximately 310g m^{-2} year^{-1} compared to 65 g m^{-2} year^{-1} for an interglacial (e.g., OIS 5)—a 4.8-fold increase. For Europe, Frechen et al. (2003) found large regional differences in accumulation rates, but suggested that along the Rhine and in Eastern Europe, they were from 800–3200 g m^{-2} years^{-1} in OIS 2. Loess accumulation rates over the United States during the LGM were also high, being approximately 3000 g m^{-2} year^{-1} for midcontinental North America (Bettis et al. 2003). From 18,000 to 14,000 years ago, rates of accumulation in Nebraska were remarkable, ranging from 11,500 g m^{-2} year^{-1} to 3500 g m^{-2} year^{-1} (Roberts et al. 2003).

THE SPATIAL VARIABILITY OF WIND POWER

The significance of wind erosion varies hugely between and within deserts because of the great variability that exists in wind power. This becomes evident when one examines data for annual sand drift potentials based on the method of Fryberger (1979) (**Table 3**). These range from 11 vector units to 1401 vector units.

Fryberger (1979) classified locations with values greater than 400 as high-energy environments, those between 200 and 400 as intermediate-energy environments, and those with less than 200 as low-energy environments. On this basis it appears, for

Table 3 Annual drift potentials

Vector units	Location	Source
1401	El-Khanka, Egypt	Moursy et al. (2002)
1155	Akron, Colorado, USA	Muhs et al. (1996)
1090	Hutchinson, Kansas, USA	Arbogast (1996)
750	Northern Great Plains, USA	Muhs & Wolfe (1999)
655	Sabha, Libya	Linsenbarth (1991)
655	Wadi Araba, Jordan	Saqaa and Atallah (2004)
588	Ghadamis, Libya	Linsenbarth (1991)
540	Bahrain	Fryberger (1980)
518	Walvis Bay, Namibia	Fryberger (1980)
489	An Nafud, Saudi Arabia	Fryberger (1979)
468	Kufra, Libya	Linsenbarth (1991)
414	Jalu, Libya	Linsenbarth (1991)
405	Qatar	Embabi & Ashour (1993)
392	El Centro, California	Muhs et al. (1995)
391	Simpson Desert, Australia	Fryberger (1979)
384	Mauritania	Fryberger (1979)
366	Karakum/Kyzlkum	Fryberger (1979)
362	Hun, Libya	Linsenbarth (1991)
354	Kuwait	Al-Awadhi et al. (2005)
321	California (Algodones)	Muhs et al. (2003)
293	Algeria	Fryberger (1979)
270	Al Jaghbub (Libya)	Linsenbarth (1991)
266	Dunhuang (China)	T. Wang et al. (2005)
237	Namib	Fryberger (1979)
203	Blythe, southern California	Muhs et al. (1995)
201	Rub Al Khali (Saudi Arabia)	Fryberger (1979)
191	Kalahari	Fryberger (1979)
160	Badain Jaran, China	Z. Wang et al. (2005)
147	Israeli Coast (Ashdod)	Tsoar and Blumberg (2002)
139	Mali	Fryberger (1979)
127	Gobi, China	Fryberger (1979)
116	Ghat, Libya	Linsenbarth (1991)
114	Indio, southern California	Muhs et al. (1995)
86	Tazirbu, Libya	Linsenbarth (1991)
82	Thar Desert (India)	Fryberger (1979)
75	Northern Kalahari	Thomas (1984)
41	Taklimakan, China	T. Wang et al. (2002), Z. Wang et al. (2005)
37	Mu Us, China	Z. Wang et al. (2005)
11	Tengger, China	Ha Si (2002)

example, that the Great Plains of the United States is a high-energy environment, as are parts of the Libyan Desert, whereas the Kalahari, the Thar, and many of the Chinese deserts are relatively low-energy environments. That said, there is often great variability in wind energy within deserts, in part because of topographic control. Values in the Taklimakan, for example, range between 2 and 112 (Wang et al. 2002, Wang et al. 2005), and for northern China as a whole between 0.9 and 1052 (Wang et al. 2006). Those in Libya range between 86 and 655 (Linsenbarth 1991). There is also a substantial degree of temporal variability (Bullard et al. 1996). Saqqa & Atallah (2004), for instance, analyzed drift potentials in the Wadi Araba (Jordan) for each of 23 years, and found that they ranged from 141 to 1070.

However, the significance of wind erosion is also controlled by small-scale events with high vertical velocities and gustiness, something that is hidden in the general wind velocity characteristics discussed above.

CONCLUSIONS

1. Remote sensing imagery has shown the widespread importance of desert depressions, yardangs, and dust storms in drylands.

2. Although pans, yardangs, and stone pavements have complex histories and a range of formative factors, wind plays a major role in their development.

3. The importance of features produced by deflation and/or wind abrasion varies according to a range of environmental factors. Pans occur preferentially on weak, fine-grained rocks, or sediments; on plains; and in areas with more rainfall than is the case for yardangs. These occur in hyperarid areas with unidirectional winds but develop on a range of rock types. Dust storms are a feature of old lake basins and alluvial spreads in basinal areas, especially where high wind velocities are produced by topographic enhancement. The world's largest sources are in the Sahara (especially Bodélé) and overall dust emissions from deserts are probably on the order of 1000–3000 million tons.

4. Wind erosion and trade wind velocities may have been considerably more important than today in dry, cold phases of the Pleistocene, and elevated wind activity levels are made evident in ocean and ice cores, but also in thick loess deposits that accreted rapidly.

5. There is a great range of wind power variation both within and between deserts, and from year to year.

SUMMARY POINTS

1. Many low-angle surfaces developed on susceptible materials, including fine-grained sedimentary rocks and lacustrine deposits, and possess pans, many of which have lunettes on their lee sides.

2. Stone pavements, which have classically been attributed to deflation, may also be the result of other horizontal sorting processes (e.g., sheet floods) and a range of vertical sorting processes (e.g., dust accretion, frost and salt heave, wetting and drying).
3. In very dry areas with unidirectional winds, bedrock and old lakebeds may be molded into various types of yardang.
4. Dust storms provide evidence of the power of deflation, and key areas of dust storm activity include closed depressions, especially in the Northern Hemisphere.
5. In the Pleistocene glacials, wind activity was at times greater than now, as is made evident by sediment preserved in loess and in ice and ocean cores.
6. The study of current drift potentials indicates that the importance of wind erosion varies greatly between different desert areas.

DISCLOSURE STATEMENT

The authors are not aware of any biases that might be perceived as affecting the objectivity of this review.

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