Chapter 19

Landforms, Landscapes, and Processes of Aeolian Erosion

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Introduction

Aeolian erosion develops through two principal processes: deflation (removal of loosened material and its transport as fine grains in atmospheric suspension) and abrasion (mechanical wear of coherent material). In a vegetation-free environment, the relative significance of each of these processes is a function of surface material properties, the availability of abrasive particles, and climate. The resulting landforms include ventifacts, ridge and swale systems, yardangs, desert depressions (pans), and inverted relief. Dust is an important by-product of some forms of erosional activity.

The significance of wind erosion as a geomorphological process has been long debated (Goudie 1989). In recent years, remote sensing images of terrestrial deserts and rover and satellite imagery of Mars (Greeley et al. 2002, Bridges et al. 2004) have stimulated new research, providing information on the extent of large-scale erosion systems, surface textures of Martian rocks, and the timing, frequency, and size of sand and dust storms associated with erosion. On Mars, ventifacts and yardangs are important proxies of climate, wind regime, sediment type and transportation, and thus consistent interpretations of geomorphic process are critical.

Despite recent advances, much remains to be learned about landforms of wind erosion. With few exceptions, detailed environmental analyses and short- or long-term process measurements are hampered by remoteness. Consequently, landform ages, processes and rates of formation, and evolutionary history remain poorly understood.

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Ventifacts

Published work on ventifacts dates back to the mid-19th century. They were first described by Blake (1855) from the Salton Sea region, California. Faceting of stones by wind abrasion was documented by Travers (1870), and further discussion on wind erosion was provided by Gilbert (1875). Evans (1911) proposed that "ventifact" encompass the multiple and compound terms (wind-grooved stones, wind-faceted stones, etc.) then in use. By 1931, Bryan had published a bibliography of 258 titles on ventifacts, including a useful discussion on terminology, and had provided an overview of prevailing theories concerning ventifact formation.

The study of ventifacts has spawned a number of controversies, particularly because many are fossil (relict of earlier surface and climatic conditions) and are thus not subject to field-based process studies. An area of early debate concerned whether faceted pebbles are shaped by mono-directional winds, winds from two opposing directions, or winds from variable directions. In addition, the factors that control the final shape of ventifacts were in contention. Woodworth (1894) suggested that a rock of moderate size will have a facet cut at right angles to the wind, with new facets formed as a result of accidental overturning or rotation. However, researchers in Europe proposed a final form primarily conditioned by (a) the shape of the original base (a square base would yield a pyramid of four faces), (b) winds that split along

A.J. Parsons, A.D. Abrahams (eds.), *Geomorphology of Desert Environments*, 2nd ed.,
DOI 10.1007/978-1-4020-5719-9_19,

the ground and impinge upon the rock from variable directions, or (c) the original size and shape of the rock. Woodworth's ideas (1894) were later supported by Sharp (1949, 1964, 1980) and appear to represent the present majority opinion. In the literature of the past few decades, differences of opinion have appeared with regard to the relative importance of saltating and suspended grains in ventifact formation and the orientation of the abraded face relative to the wind.

The term ventifact, although widely applied, is ill-defined, describing as it does wind eroded forms of varying size, form, and material composition. Ventifacts range in size from small pebbles to large boulders. Although facets are often regarded as fundamental features of ventifacts (Sharp and Malin 1984), they are often absent from large boulders and rock outcrops, where the presence of pits, flutes, and grooves may be the principal indicators of wind erosion. On lava flows and bedrock outcrops, eroded semi-planar surfaces may develop and linear grooves form on playa surfaces subject to saltating sand grains. The material that forms ventifacts may not be stone at all: over short time periods, playa (Williams and Greeley 1981) or river bed ventifacts may form in cohesive sediments.

Ventifacts are valuable as evidence of past climatic conditions (Figs. 19.1 and 19.2). The presence of wind-worn stones in ancient formations (Precambrian, Cambrian, Devonian, Permian, Jurassic, Triassic, Cretaceous, Tertiary, and Pleistocene) has been used as evidence of more arid conditions (Bryan 1931). Palaeoventifacts are indicative of intense wind activity, but they do not necessarily signify warm desert conditions, as they also form in temperate regions along ocean or lake margins, and in periglacial settings. Aspects of ventifact form (the orientation of facets, grooves, flutes, and pits) enable a determination of palaeowind direction, and the presence of ventifacts provides clues to the relative abundance of abrasive materials and vegetation cover in earlier environments. Although it is recognized that many ventifacts are no longer forming ("relict" or "fossil" forms), the actual age of most ventifacts is unknown.

Fig. 19.2 In this photograph taken in the Bristol Mountains, California, fossil ventifacts are located on the lower two-thirds of the hill, whereas actively-forming ventifacts are found near the crest and summit. Wind speed and sand transport increases towards the crest of a hill, and active sand is more common here

Fig. 19.1 Abraded granite boulder approximately 2.7 m high located on a hill crest near June Lake, California, east of the Sierra Nevada. Pitting covers much of the high-angle windward face, transitional to fluting as the wind approaches at an oblique angle on the left side of the ventifact. The fluted abrasion-hardened rind stands out in relief (detached by about 25 cm) relative to the pitted surface

Environments of Ventifact Formation

Ventifacts form in environments characterized by (a) a supply of an abrasive material, (b) lack of complete vegetative cover, (c) strong winds, (d) a topographic situation that allows the free sweep of wind or that locally accelerates wind, and (e) ground surface stability. Ventifacts are found in desert, periglacial, and coastal environments on Earth and are probably even more widespread on Mars. Many are fossil in nature, as a change in any of the aforementioned environmental factors will cause the diminishment or cessation of wind erosion. As the processes of ventifact formation in coastal and periglacial settings are similar to those in deserts, all environments will be briefly reviewed, and the research results incorporated into the discussion.

Ventifacts occur in numerous present-day and palaeo-periglacial environments, including North America (Blackwelder 1929, Powers 1936, McKenna-Neuman and Gilbert 1986, Bach 1995, Dorn 1995), Antarctica (Lindsay 1973, Miotke 1982, Pickard 1982, Hall 1989), Europe (Schlyter 1989, Christiansen 2004), Iceland (Antevs 1928, Greeley and Iversen 1985, Mountney and Russell 2004), Greenland, the Falkland Islands (Clark and Wilson 1992), and higher elevations in South America (Czajka 1972). These ventifacts range in size from small faceted forms to boulders of considerable size: palaeoventifacts along the eastern base of the Sierra Nevada, California exceed a metre in height (Blackwelder 1929) (Fig. 19.1) and Powers (1936) described 60-cm-long grooves in crystalline boulders strewn on the sandy Lake Wisconsin plain. Clark and Wilson (1992), Mountney and Russell (2004), McKenna-Neuman (1993) and Tremblay (1961) emphasize that most erosion occurs in the presence of sand (associated with sand sheets or sand dunes). In Antarctic dry valleys, very high wind speeds (up to 300 km h^{-1}) allow very coarse material, including medium sand, granules, and pebbles, to abrade rocks (Hall 1989), such that the yearly abrasion rates average a few millimetres and ventifacts form within a few decades or, at most, a few centuries (Miotke 1982). McKenna-Neuman (1990) reported aeolian transport of coarse granules up to 4 m in height in an arctic proglacial setting in Canada.

Ventifacts occur in coastal locations where winds are strong and drifts of sand provide an effective abrading agent (Travers 1870, King 1936, Wentworth and Dickey 1935, Bishop and Mildenhall 1994, Knight and Burningham 2003). Few ventifacts have been reported for moist temperate climates outside of the coastal zone. In Nova Scotia, Canada, they occur as a result of highly permeable sediments that result in locally dry surface conditions (Hickox 1959).

Ventifacts appear to be fairly common in warm deserts and semi-arid environments. However, their spatial distribution is poorly known because such small features cannot be identified from aerial imagery. Our knowledge of ventifact sites thereby depends on fortuitous discovery by individuals who recognize these features and furthermore publish their observations. High altitude photography is sometimes useful in locating ventifacts, because their occurrence is often coincident with deposits of active and stabilized sand, with wind streaks, and with regions where the air is locally accelerated, either by topographic constrictions or by passing over a hill (Laity 1987).

In appearance, there is no evident difference between ventifacts formed in desert, coastal, or periglacial environments, suggesting that the proceses of formation are broadly similar. Most researchers have assumed that the abrading agent in deserts is wind-blown sand (Fig. 19.3) (Blake 1855, Blackwelder 1929, Maxson 1940, Sharp 1964, 1980, Smith 1967, 1984, Greeley and Iversen 1986, Laity 1987, 1992, 1995, Bridges et al. 2004), but few have provided specific information as to grain size, shape, or lithology. Sediment analyses by Sharp (1964), Greeley and Iversen (1986), and Laity (1995) suggest that fine to medium-sized sand dominates in ventifact formation.

Fig. 19.3 Actively forming ventifacts surrounded by aeolian sand at the hillslope summit shown in Fig. 19.2. The rocks are dense basalts, with very little vesicularity. Multiple wind directions are represented by several facets, but there are few lineations developed

Morphology of Ventifacts

Lithology and Size of the Original Rock

Ventifacts develop in a wide range of rock types, ranging from exceptionally hard quartzite to softer limestone and marble. Most lithologies develop characteristic faceting, pitting, fluting, or grooves. Rocks such as schist, gneiss, and ignimbrites are particularly prone to linear erosion features and to etching (Sharp 1949, Lancaster 1984). Dense, hard, and homogeneous rocks (such as cherts or basalts of low vesicularity) are less likely to develop pits, flutes, or grooves, as they lack vesicles or soft inhomogeneities that abrasion can exploit (Fig. 19.3). Polishing and faceting of rock surfaces have been observed on all lithologies.

In deserts, ventifact preservation is affected by weathering processes that modify or destroy surface features. Limestones and dolomites are particularly susceptible to solution, even in arid environments. In a periglacial setting in Wyoming, Sharp (1949) found no evidence of erosion in rocks of this type. Laity (unpublished data) has observed rillenkarren that cross perpendicular to grooves on fossil ventifacts of the Panamint Valley, California. Coarse-grained granites are particularly susceptible to granular disintegration and exfoliation in deserts, and tend to preserve evidence of erosion less well than other materials (Lancaster 1984, Laity 1992). Likewise schists and pegmatites of the Namib Desert (Selby 1977, Lancaster 1984) and andesite in the Mojave Desert (Laity 1992) were found to weather too rapidly to preserve significant evidence of abrasion.

The ultimate form of the ventifact is strongly controlled by the size of the original material. If the rock is small, with a diameter of only a few centimetres, the ventifacts are more likely to be the classic faceted types (with planar faces and smooth surfaces) that lack lineations (Schoewe 1932, Wentworth and Dickey 1935, King 1936, Needham 1937, Maxson 1940, Glennie 1970, Czajka 1972, Whitney and Dietrich 1973, Babikir and Jackson 1985, Nero 1988). Schoewe (1932) suggested that the ratio of the height of a faceted ventifact to the height of the sand-laden current does not exceed 1:8. Thus, if the zone of saltation extends to 64 cm, the diameter of the largest fragment faceted would be 8 cm. Maxson (1940) noted that faceted ventifacts in Death Valley, California, did not exceed a height of 8 cm, and that larger fragments were striated and polished, but not faceted.

Fig. 19.4 Wind-eroded rock formed in a bidirectional flow regime on a hill crest in the Little Cowhole Mountains, Mojave Desert, California. Wind directions are from the north and south, producing two grooved facets and a central keel. The keel is perpendicular to the wind flow. Material surrounding the ventifacts is active aeolian sand

Given sufficient time for erosion, moderate-sized boulders often develop a semi-planar face, or two, if the wind is bidirectional (Fig. 19.4): these faces are invariably fluted or grooved on the windward exposure (Powers 1936, Sharp 1949, Czajka 1972, Laity 1987). Rocks with three facets are occasionally observed in regions with three distinct wind directions. Owing to their great mass, facets usually do not develop on boulders exceeding 1 m in diameter, although considerable bevelling may occur. Grooves, pitting, and fluting have been observed up to heights of several metres on rocks (Fig. 19.1). Powers (1936) noted such features on a crystalline boulder 4.5 m in diameter. Outcrop ledges may show evidence of planation and be pitted or grooved.

Erosional Forms

Ventifact surfaces are characterized by a wide range of erosional features, that vary both in form and scale, resulting from differences in rock type, face angle, weathering history, sediment supply, and wind speed. To date, there has been little examination of the causes of these features.

Some degree of rock heterogeneity seems important in the development of specific features. For example, pits are best developed in materials that have vesicles or a softer matrix, such as coarse grained granites, volcanic tuffs, and basalts. Flutes and grooves also appear to develop from initial vesicles or weaker spots within rocks. In 1993, I placed 6 large blocks of modelling foam of different densities, but a homogeneous texture, in an area of active abrasion in the Little Cowhole Mountains. The blocks ranged from 50 g to 325 g in weight, and were placed at a 45° angle to the horizontal, more-or-less parallel to the slope. Thirteen years later, the foams have all been strongly abraded, with the low density foam almost completely eroded away. However, over this period, erosion was essentially uniform across the face of the targets and no lineations formed. By contrast, styrofoam blocks (that are heterogeneous in texture) and plaster targets with small air vesicles developed lineations within a few months.

Although the various erosional forms that occur on ventifacts are discussed separately below, it is important to note that some forms are transitional in nature. For example, a pit may be seen to be elongating into an incipient flute. In soft materials such as tuffs, a case-hardened fluted surface may be undermined by subsequent erosional episodes, to produce a surface that is deeply eroded and almost fretted. Additional images and discussion of ventifact features may be found in Viles and Bourke (2007).

Smoothing and Polishing of Rock Surfaces

Smoothing and polishing of rock surfaces is perhaps the most common feature reported for ventifacts. Polish occurs both on smooth facets and within flutes and grooves (Maxson 1940). In periglacial settings, ventifact surfaces may exhibit a high gloss that exceeds that of glacial polish (Tremblay 1961). Rock coatings may develop after abrasion ceases, including the development of a silica glaze on periglacial ventifacts and desert varnish in drylands (Dorn 1995).

Facets

King (1936) and Sharp (1949) used the term facet to describe a relatively plane surface cut at right angles to the wind, regardless of the original shape of the stone, and the term face to describe the original surface of the rock fragment. Facets commonly join along a sharp ridge or keel (Figs. 19.3 and 19.4), and the number of keels (kante) is used to describe the stones as einkante, zweikanter, dreikanter (one-, two-, three-ridged), etc. (Bryan 1931). Much effort has been expended on the morphological classification of ventifacts (Bryan 1931, King 1936, Czajka 1972), particularly for small faceted forms.

Multiple facets developed on ventifacts have been attributed to (a) the original shape of the stone, (b) splitting of the wind around the rock, (c) winds from different directions, and (d) shifting of ventifacts owing to undermining by wind scour, and overturning by wind, frost action, rain wash, and animal disturbance. In cold environments, frost shifting may be a particularly effective mechanism, as discussed by Sharp (1949) and Lindsay (1973).

Pits

A pitted surface is one indented by closed depressions, often of irregular shape (Fig. 19.5). According to Whitney (1978, 1979), McCauley et al. (1979), and Garvin (1982), the wind is capable of producing pits on the surface of dense, homogeneous stones. However, Sharp and Malin (1984) emphasized that pitting is not an inevitable by-product of aeolian erosion, as material such as chert and limestone may develop facets, flutes, and grooves, but no pits. It is probably easier for the wind to modify pre-existing pits (such as vesicles in basalts or tuffs) by enlargement or integration, or to

Fig. 19.5 Pits developed on high-angle windward face in tuff. Soft inclusions are preferentially eroded, and abrasion in tuff yields pits in a great range in shapes. In basalts, pits tend to be rounder and more regular in form

erode softer minerals, as is evident in pitting of coarse-grained granites (Fig. 19.1).

Pits occur on surfaces that are inclined at high angles to the wind (55–90°) and thereby indicate the windward side of boulders. As the angle between the face and the wind decreases, a transition from pits to deep flutes with overhanging ends occurs (Sharp 1949).

Flutes

In form, flutes are open at one end and closed at the other, and broadly U-shaped in cross-section (Figs. 19.6, 19.7, 19.8, and 19.9). They may appear as "arrowheads" that point in a downwind direction. Flutes form independently of material hardness, composition, or rock structure on surfaces that are nearly horizontal or inclined at low angles (less than 40°) to the wind: flutes become shorter and deeper as the inclination of the surface steepens (Maxson 1940, Sharp 1949). Figure 19.6 indicates how the scale of the flute may increase in height up the ventifact. Occasionally, smaller flutes will begin to develop within larger ones as a second cycle of erosion begins (Fig. 19.9).

Flute development by aeolian processes is not understood. Maxson (1940) proposed that flutes are cut beneath vortices of fine sand, and Whitney (1978) argued that vortex pits may coalesce into flute pits and pit chains. However, most flutes exhibit smooth interior surfaces that lack pits.

Fig. 19.6 Ventifact in sandy terrain near Silver Lake, California. The windward face has been bevelled and fluted. The scale of the flutes increases from the base of the rock to the upper face

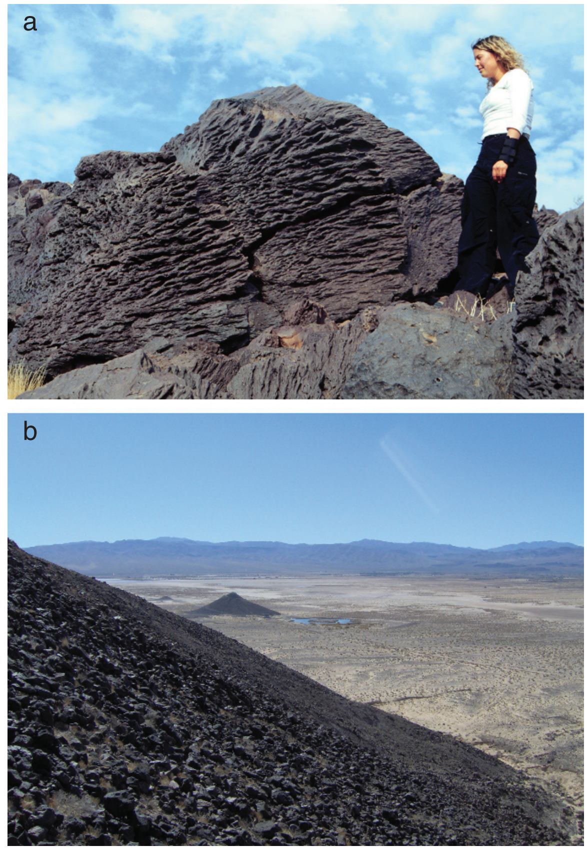

Fig. 19.7 **(a, b)** Intensely pitted and fluted basaltic ventifact, approximately 1.8 m in height, located near a hill crest in the Cady Mountains, Mojave Desert, California. The flutes and grooves radiate outward from a central area. The elevation of the ventifact, relative to the surrounding sand-covered plain is shown in Fig. 19.7b. Ventifacts occur across the entire slope, with feature scale and intensity increasing with altitude

Scallops

U-shaped or scalloped erosional features are open at one end and closed at the other and have similar length to width ratios. They lack the "arrowhead" form of flutes. They appear to be rare on terrestrial ventifacts, but are occasionally observed on Martian rocks.

Grooves

Grooves are longer than flutes and open at both ends. They are best developed on surfaces gently inclined or parallel to the wind. However, they are sometimes seen on the vertical sides of rocks, particularly where the wind has accelerated between

Fig. 19.8 Fossil ventifact with exceptionally large flutes located on a hill that rises ~100 m above the Mojave River Sink, Mojave Desert, California. The largest flutes are 10–17 cm wide, up to 60 cm in length, and 7 cm in depth. The side illustrated here faces into the westerly winds. Part of the upper boulder surface has detached and fallen to the base of the ventifact. The fossil ventifacts in this area are heavily varnished, but those near the crest show signs of reactivation (varnish abrasion) and sand is present in some of the flutes

Fig. 19.9 Closeup of flutes shown in Fig. 19.8. A secondary cycle of new flute generation can be seen within some of the larger forms. Litre-sized water bottle for scale

Fig. 19.10 Helical forms developed in marble. These features occur at scales up to 20 cm in length and several centimetres in width and depth. The abrading agent is sand, some of which can be seen trapped within the helical forms. Fine lineations (striae) also cross the lower left face of the ventifact

adjacent large boulders. Like flutes, they may cut indiscriminately across mineral grains and rock structures.

Ventifact groove and flute trends are remarkably parallel on near-horizontal surfaces and reflect the flow direction of the highest velocity winds in an area (Fig. 19.4) (Laity 1987). On large curved surfaces facing the wind, flutes or grooves often radiate outward from a central pitted area. Sharp (1949) observed that flutes and grooves are not mutually exclusive, for the surfaces of large grooves are often fluted on a small scale.

Tremblay (1961) characterized three scales of groove development: striae are fine lineations (Fig. 19.10), grooves are of intermediate scale, and channels attain several centimeters in depth (up to 13 cm deep in his study area). None of the lineations exceeds a metre in length, and most are considerably shorter. Whereas fine striae appear to cross a whole outcrop, in detail each comprises a succession of short scoop-like depressions a few centimetres long.

As in the case of flutes, there is little understanding of groove formation. According to Maxson (1940), grooves suggest vortices descending along the rock surface in the wind direction: once formed, the grooves may be modified by saltating grains at air velocities below those critical in the generation of vortices. Schoewe (1932) discovered that sand grains impinging at low angles on hard, smooth surfaces skid instead of rebounding directly into the air and proposed that this action may be related to the development of flutes and grooves.

Etching and fretting

Etching occurs when the composition of a rock mass is not homogeneous, and the wind selectively erodes

Fig. 19.11 Etching and development of incipient lineations and small-scale pits on a heterogeneous layered plaster target placed in the Little Cowhole Mountains for approximately 3 months. The target, placed on the hill crest, is in a bi-directional wind regime. The target is ∼22 cm in length and 10 cm in width

less resistant strata or foliations. In nature, etching is often well developed on layered ignimbrites. Etching also develops in artificial rock targets with layers of different hardnesses (Fig. 19.11).

Fretting develops where there are harder inclusions within the rock material, with projecting points, knobs (Fig. 19.12), and ridges forming a particularly rough surface (Sharp and Malin 1984). The inclusions resist erosion, while the surrounding material is removed by abrasion. This may result in the formation of erosional "fingers" with a visible inclusion at the tip (sometimes referred to as dedos) (Fig. 19.13) or, in some cases, large rounded xenoliths that project out from the overall ventifact surface (Fig. 19.12).

Fig. 19.12 A knobby texture formed on the high-angle windward face of a granite boulder. The large xenolith inclusions are finer grained and more resistant to erosion than the matrix, standing out several centimeters in relief

Helical Forms

Helical forms are uncommon. They begin as shallow grooves, deepen and spiral in a downwind direction, and terminate in a sharp point (Fig. 19.10); range from several millimeters to several centimetres in width and depth; and maintain a consistent form as their scale increases. Helical forms may be found in association with flutes. Observations in both desert and periglacial settings in California suggest they occur where wind velocities are very high, such as within topographic saddles and near hill crests. They develop on the upper face of the ventifact and have been observed in marble, basalt, and granite.

Fig. 19.13 Marble ventifact from the Little Cowhole Mountains, Mojave Desert, California. The finger-like projections (sometimes called dedos) extending from the ventifact result from visible inclusions of resistant material

Abrasion-Hardened Surfaces

Abrasion may create a hardening of the surface that resists subsequent mechanical and chemical weathering. This process may require a silicic form of rock, as it has been observed in volcanic tuffs, granites, and andesites, but is less evident in basalts, and does not appear to be a factor in marbles. Any loss of the protective rind (fracturing or sloughing) allows renewed erosion of the underlying rock, and can result in an exposed abrasion carapace standing free of the main rock (Fig. 19.1).

Processes of Rock Destruction

Three processes have the potential for material erosion in an aeolian regime: abrasion, deflation, and rock wedging. Abrasion results principally from the impact of particles in saltation and is accomplished by sand-sized grains (about 60–2000 μm in diameter) (Greeley et al. 1984). Finer material in suspension is thought by some to erode rock mass at the microscopic level (Whitney and Dietrich 1973). Deflation is the removal of previously weathered material by strong winds and appears to be largely insignificant in ventifact formation. Whitney (1979) proposed erosion by pure air flow alone, but as no ventifacts have yet been recorded in a nonabrasive environment, and as particulate material is generally abundant, the efficacy of erosion by pure air need not be an issue. Rock breakdown by wedging occurs when grains moving at high velocity are packed into cracks, and succeeding particles impact grains that are already wedged. Hall (1989) observed this process in Antarctica, but it has yet to be recorded in warm deserts. Solution processes may play a role in enlarging or initiating pits or in forming rills, but these processes appear subordinate to wind erosion (McCauley et al. 1980).

In some environments the apparent absence of sand has led to a consideration of snow as an abrasive (Teichert 1939, Dietrich 1977b, Schlyter 1989). Although snow at −44°C has a hardness similar to quartz, it has half the specific gravity, so that sand will have two to three times more kinetic energy upon impact (McKenna-Neuman 1993). Furthermore, field experiments do not support abrasion by snow. McKenna-Neuman and Gilbert (1986) showed that poles covered with eight coats of exterior enamel were rapidly stripped and eroded by aeolian sands, but remained perfectly preserved when subject to blowing snow and ice. These observations agree with those of other researchers (Blackwelder 1929, Tremblay 1961, Miotke 1982, Nero 1988), and indicate that sand is the most important abrasonal agent in cold climates.

Nature of Abradant

Dust

Although most researchers attribute ventifact formation to a sandblast action, dust has also been invoked to explain polish and the formation of finely detailed features such as flutes. Higgins (1956) suggested that dust in suspension was the abrasive agent for small multifaceted pebbles embedded in bedrock, but later concluded that the stones were probably not ventifacts (Higgins, personal communication 1988). Sharp (1949) considered that polish on ventifacts may result from material finer than sand. In the Namib Desert, Lancaster (1984) suggested that dust particles are probably more important than sand in creating smooth polished rock surfaces, flutes, and grooves, as the effects of sand laden winds would be akin to the destructive effects of industrial sand blasting processes. Nonetheless, sand was abundant at his field site. Maxson (1940) and Whitney (1978) attributed flutes or lineations to cutting by particles fine enough to follow vortex currents, suggesting that most sand grains travelling by saltation probably pass through small vortices without much change of path.

Wind tunnel experiments by Dietrich (1977a) indicated that wind-blown materials, which are relatively soft or small in particle diameter, can produce minor amounts of erosion. For 455 days, dust was blown continuously against blocks of low hardness (halite and sylvite) and against synthetic periclase. At the end of the experiment, the microtexture was rougher, but the blocks had no measurable weight loss. As most natural materials are harder than those tested, and as dust storms are not common, averaging approximately 22 per year in warm deserts of Africa and the Middle East and less than 5 per year in North America (Middleton 1989), a very long period of time would be necessary to remove the mass essential for ventifact development.

A review by Breed et al. (1989) also posed several arguments for dust abrasion abrasion, asserting that sandblasting can produce facets, but cannot account for delicately-textured erosional markings (pits, flutes, grooves, or helical scores); that ventifacts within the saltation zone lack fine detail such as fluting on their windward faces; and that ventifacts rarely occur in or near sand dune fields, but are common on the stony surfaces of sand-poor regions. Furthermore, ventifacts were said to have more rock mass eroded from their lee areas than from their windward. As a further complication, McCauley et al. (1979) concluded from the negative flow observed to the lee of a ventifact subject to a bubble-generating device (simulating suspended particles), that pits and flutes on all sides of rocks could be explained by wind from a single direction. These points will be discussed further in the following section.

Sand

Despite laboratory evidence that dust may be capable of abrading very soft rock over extremely long time periods, it is likely that sand is the most effective agent of abrasion in ventifact formation. Areas where sand has been identified as an agent of erosion include coastal environments (King 1936, Knight and Burningham 2003), periglacial regions (Powers 1936, Tremblay 1961, Miotke 1982, McKenna-Neuman and Gilbert 1986, Nero 1988, McKenna-Neuman 1990, 1993), and desert regions (Sharp 1964, 1980, Sugden 1964, Selby 1977, Smith 1984, Laity 1992, 1994, 1995). To date, there is no field evidence of ventifacts forming in areas where dust is the sole agent.

The following arguments can be made for sand as the leading agent in aeolian erosion.

(a) Analytical models indicate that sand is a much more effective agent of erosion than dust. The mass of material lost per particle impact is directly proportional to the kinetic energy of impacting grains (Greeley et al. 1984, Anderson 1986). Dust not only has less mass, but is well coupled to the wind, being deflected around rocks and rarely impacting the surface directly. For suspended grains ($D \leq 0.31$ mm), Anderson (1986) showed that particle deflection by the air flow around the ventifact leads to a reduction in the delivery of kinetic energy to the surface. By contrast, saltating sand achieves enough mometum to be decoupled from the boundary layer velocity profile at peak acceleration near the top of its trajectory, with the sand velocity on the order of 50% or less of wind velocity at that height (Bridges et al. 2005). Although the velocity of sand is less than that of dust, the mass, which increases by the cube of particle size, is much greater. Indeed, abrasion experiments using particles ranging in size from 75 to 160 μm show that S_a (the susceptibility to abrasion by rocks) varies with particle diameter D to approximately the third power (Greeley et al. 1982, 1984). Thus, the mass of a 100 μm sand grain has 1,000 times the mass of a 10 μm dust particle. Considering the effects of both velocity and mass, the kinetic energy of sand upon impact is ~50–100 times that of dust. Additionally, large particles are more likely to impact the surface than small particles. Anderson (1986) predicted that the number of impacts from 10 μm particles would only be 10% of that of 100 μm grains. Taking into account both the kinetic energy upon impact and the potential number of impacts, it is likely that the total kinetic energy transferred to rocks by sand would be on the order of 1,000 times greater than that of dust (Laity and Bridges, 2008) . In the field, Sharp (1980) showed that the cutting rate increased coincident with the increased flux of windborne sand.

(b) Experiments also show that the amount of abrasion incurred by a ventifact impacted by particles less than 90 μm in diameter decreases much more rapidly with decreasing particle size and velocity than predicted by kinetic energy considerations or experiments using larger sized particles (Stewart et al. 1981). Two factors account for this. First, where clay particles are present in the entrained material, these particles are transferred to the surface of the target, sheltering it from subsequent abrasion by larger particles (a cushioning effect). Second, there is an apparent kinetic energy threshold for impact fracture, suggesting that the abrasion mechanisms change with kinetic energy.

(c) The present-day absence of sand from a region does not preclude it from consideration as the primary abrasive agent. Mainguet (1972) and Sharp and Malin (1984) emphasized that the movement of sand in aeolian corridors tends to be episodic, so that sites which have been traversed by large quantities of saltating grains now harbour only small accumulations. Also, many ventifacts are relict, and the abrasive sand that carved them has been deposited elsewhere.

(d) In many areas, the distribution of ventifacts is clearly associated with the presence of sand (King 1936, Tremblay 1961, Smith 1984, Laity 1992, 1995 and others). In the eastern Mojave Desert, fossil forms occur in association with deposits of stabilized sand or in aeolian corridors where sand has travelled to nearby sites of deposition. Actively forming ventifacts occur in rock outcrops presently being traversed by fine-grained sands (Figs. 19.3, 19.4, and 19.14). Ventifacts are not found in areas subject to dust influx alone. More than a metre of silt has been deposited in some areas of the Cima volcanic field (Dohrenwend 1987), yet despite long exposure to dust, there is no evidence of abrasion on rock surfaces.

(e) Actively forming ventifacts lying within the saltation zone clearly demonstrate fine details on their surfaces (Figs. 19.3, 19.4, and 19.13). Sharp (1964) described the formation of flutes on bricks in a wind-driven sand environment. In the Little Cowhole Mountains, Mojave Desert, fine sands are seasonally redistributed by the winds, alternately burying and exposing rocks (Laity 1995). All of the marble within the vicinity shows considerable mass removal and planation as well as sharply defined erosional markings (flutes, grooves, or helical forms). Some of the striations are only 1 mm across, separated in parallel arrays by sharp ridges. If sand is not responsible for the creation of these microfeatures, it could be argued that they would be destroyed by the protracted impact of saltating grains.

Fig. 19.14 Abrasion of ventifacts during the passage of a front in the Mojave Desert. Wind speeds reached 23 m s^{-1}, with higher gusts. Abrasion patterns on the hill slopes were complex, as sand was driven from near the base of the slope up towards the crest of the 100-m hill. As the sand moved up and over the hill, ventifacts were alternately exposed by wind scour and buried. Later in the day, the wind reversed and the opposite faces of ventifacts were abraded. Upslope wind acceleration results in maximum erosion near the crest. Note that whereas the windward face is exposed to abrasion, the lee face is commonly protected by a sand tail. The fossil ventifact shown in Figs. 19.8 and 19.9 is located further to the south along the same ridge, at a site presently less subject to sand passage

Saltating sand grains are also carried across the Amboy (Greeley and Iversen 1986) and Pisgah (Laity 1987) lava fields of the Mojave Desert. The flutes and grooves that occur on the subhorizontal flow surfaces are developed by saltating grains on a descending path. Near the surface, where wind velocity approaches zero, dust would be an ineffective agent of abrasion (Anderson 1986).

(f) Polished surfaces and fine features can result from sand abrasion. Actively forming ventifacts of the Pisgah lava flow (basalt) and the Little Cowhole Mountains (marble) demonstrate a high degree of sheen and are macroscopically smooth. Under a microscope, there is abundant sand visible within the cavities and, at higher magnifications, scanning electron micrographs of flute interiors show considerable topographic roughness indicative of repeated chipping. Occasional large impact structures have fresh cleavage facets that show no loss in definition at high magnifications (Laity 1992). Thus, surfaces that appear very smooth to the eye can be very rough at the microscale. This evidence suggests that dust need not be invoked to explain polish or fine features.

(g) Most ventifacts show significant erosion on their windward faces and no erosion on their lee surfaces (Blackwelder 1929, Sharp 1964, McKenna-Neuman 1993). This relationship has been demonstrated unequivocally since 1993 at an instrumented site in the Little Cowhole Mountains, California. An automated weather station records the bidirectional winds, that seasonally move a reversing sand dune across a hillslope covered by marble ventifacts. The abrasion of painted poles, plaster and simulated sandstone targets, foam boards, and balsa wood posts has been related to wind direction throughout this period. Abrasion was recorded only on the windward face: no "lee side" abrasion from negative flow has occurred. Furthermore, in small-scale plaster "rocks," lineations have developed within the saltation zone, paralleling the direction of the highest velocity winds and with the same orientation as adjacent ventifacts.

Large boulders, that are not subject to movement and overturning, also exemplify this relationship. Blackwelder's (1929) study of ∼1 m high boulders showed the windward surfaces to be strongly grooved and drilled, whereas on the leeward face "no such markings were found, but on the contrary the rocks are merely cracked and exfoliated in the ordinary manner" (p. 256). Similarly, McKenna-Neuman and Gilbert (1986) observed well-polished flutes and grooves on the windward face and a lee face typically covered in lichen, with no evidence of abrasion. Regardless of lithology, ventifacts in California show the anticipated relationship between maximum abrasion on the windward face, and a facet that slopes away from the wind. Furthermore, field observations during a strong wind event (up to $23\,\mathrm{ms}^{-1}$) in the Mojave Desert (near Rasor Road) showed no lee side abrasion. Indeed, the deposition of sand on the lee face of some ventifacts acted to protect them from abrasion (Fig. 19.14).

(h) Pits are generally considered to be diagnostic of high-angle impact by sand grains on windward faces (Sharp 1949). Studies of pitted surfaces and wind flow in the Mojave Desert confirm this relationship. Pits are not developed equally on all sides of rocks, but are limited to high-angle windward surfaces (Smith 1984, Laity 1987, 1992).

In summary, laboratory experiments have shown that dust is capable of changing the surface micromorphology of rocks but, in contrast to sand abrasion, mass removal is generally too small to be measured and would be quantitatively insignificant, even over long time periods. To date, no actively forming ventifacts have been identified that develop in an environment dominated by dust alone. Dust has been invoked primarily as an agent in the formation of fossil ventifacts when a sand source is not evident, and to explain the polish and fine flutes and grooves that develop on ventifact surfaces. However, lack of sand in the immediate vicinity of ventifacts is not conclusive, as the movement of sand across landscapes is commonly episodic. Furthermore, ventifacts with polished surfaces and fine features are shown to occur in regions where sand dominates the flux of particles and dust storms are rare. Whereas polish appears to be macroscopically smooth, at the microscale the surface is very rough, with features indicative of repeated chipping and gouging by sand particles.

Formation of Ventifacts

The formation of ventifacts has been assessed analytically, by wind tunnel experiments, and by field examination. Early wind tunnel experiments are discussed in Schoewe (1932). Laboratory experiments have sought to assess ventifact formation by examining different abradants, target materials and slope angles, and wind conditions. Several studies have addressed abrasion rates.

There are several limitations to wind tunnel studies. They are typically conducted in tunnels that are too small to allow natural conditions of grain bounce and do not replicate natural surface conditions. The abradant varies greatly in grain size (Schoewe 1932, Kuenen 1960, Whitney and Dietrich 1973, Dietrich 1977a) and in angularity (Miotke 1982), and seldom represents the size, shape, or sorting characteristics of material abrading ventifacts in the field. Natural rocks are rarely used as the target material as they abrade too slowly. Nonetheless, wind tunnels provide the ability to control parameters that affect aeolian processes, giving insight into fundamental mechanisms. Bridges et al. (2004) conducted wind tunnel experiments using sandstone simulants and foams, with five shapes defined by the angle of the front face relative to the wind (15°, 30°, 45°, 60°, and 90°) and, as a ground truth calibration, placed similar targets at a ventifact site in the Little Cowhole Mountains, Mojave Desert (Fig. 19.11). Intermediate-angled faces exhibited the greatest angle changes, but not the greatest mass losses. Although wind tunnels yield useful information on short term abrasion and feature formation, wind abrasion in the field is more aerodynamically complex and is impossible to fully reproduce in the laboratory.

Process-oriented field studies were pioneered by Sharp (1964, 1980). A long-term field study (1993 to present) in the Little Cowhole Mountains has provided further insights into the formation of abrasional features in a regime of active sand transport (Laity 1995 and unpublished data, Greeley et al. 2002, Bridges et al. 2004). Other field studies have concentrated on aspects of ventifact morphology and on the use of ventifacts in palaeoclimatic interpretations.

Magnitude of Erosion

The magnitude of ventifact erosion is dependent on the susceptibility of the rock to erosion, S_a (as determined by density, hardness, fracture–mechanical properties, and shape of the rock) and on properties of the impacting particle, such as particle diameter D, density ρ_p, speed V, and angle of incidence $\alpha(90^\circ$ = perpendicular impact) (Greeley et al. 1984, Anderson 1986). The mass of material lost per impact, A (Scattergood and Routbort 1983) is

$$A = S_a \rho_p (V \sin\alpha - V_o)^n (D - D_o)^m \qquad (19.1)$$

where V_o and D_o are the threshold particle speed and diameter that will initiate erosion. Abrasion experiments indicate values of $n = 2$ and $m = 3$ (Anderson 1986). The mass of material removed per impact is roughly proportional to the kinetic energy of the impact.

Height of Abrasion and Development of the Erosion Profile

Ventifacts develop within the curtain of saltating sand grains, and their shape and development is dependent on particle fluxes within this zone (Fig. 19.15). Saltating grains follow a path termed the saltation trajectory. Grain velocity increases throughout the path, reaching 50% of maximum velocity near the peak of the trajectory. In general, particles travelling at greater heights have higher velocities, owing to an increase in wind speed above the ground and the longer saltation paths that allow more time for them to be accelerated by the wind (Greeley et al. 1984). Calculated kinetic energy fluxes of saltating particles are greater where liftoff velocities are increased by grain bounce on elastic surfaces rather than mobile beds (Anderson 1986). In deserts, harder surfaces that promote grain bounce include extensive areas of stone pavement, boulder-strewn slopes, surfaces impregnated by late stage calcium carbonate, and exposed bedrock areas, including basalt flows. Such settings are subject to more vigorous erosion (Greeley and Iversen 1985, Laity 1987).

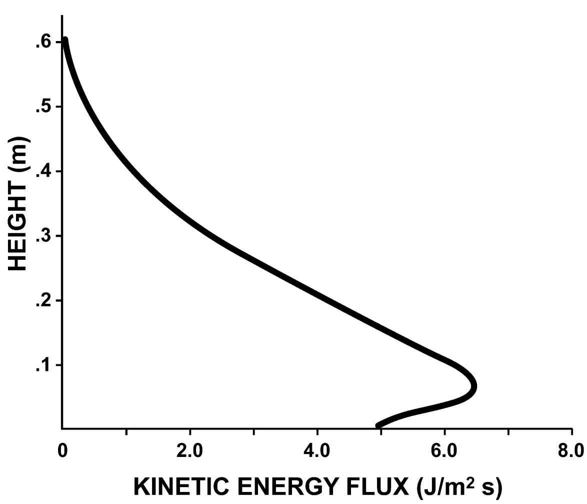

Fig. 19.15 The kinetic energy-flux profile for saltation (from Anderson 1986). The typical erosion profile, developed in fence posts, on rocks, or on yardangs, will be similar to this form, although the height of maximum kinetic energy flux (and hence erosion) above the ground varies according to surface conditions and slope angle. The faces of many ventifacts reflect the lower part of the curve, receding backwards at a nearly constant angle

The range of effective sand abrasion rarely exceeds 1 m above the surface on level ground. Hobbs (1917) observed this limit on a variety of materials in Egypt, including cast-iron telegraph poles, thick adobe walls, and granite knobs, where a lower polished zone intersected the remaining weathered rock along a fairly sharp boundary. “Pedestal rocks” develop as a result of concentrated erosion near the rock base. Ventifacts 1 m in height show sand blast effects to their upper surfaces. On hillslopes, abrasion may occur to a height of several meters (Figs. 19.1 and 19.7).

Within this general zone of abrasion, erosion profiles develop with distinct maxima of mass removal (Sharp 1964, 1980, Wilshire et al. 1981, Anderson 1986). The pattern of erosion is similar for all materials, although the magnitude depends on material properties. The height of maximum erosion is influenced by such parameters as wind speed and the degree of grain bounce. Sharp (1964, 1980) recorded erosion maxima 0.10–0.12 m above a level ground surface in Lucite rods exposed to the wind for 15 years. The height of maximum abrasion shifts upward as wind velocity increases (Liu et al. 2003). Elevated heights also result from greater grain bounce on hard surfaces relative to softer ones, moving upwards from 0.28 m to 0.43 m (Wilshire et al. 1981). On hillslopes, the erosion maxima are also higher, owing to wind acceleration and other effects related to shifting sand base levels: balsa block array maxima on slopes in the Little Cowhole Mountains were at heights of 0.3–0.5 m (T. Boyle and J. Laity, unpublished data).

As a consequence of the increase in abrasion up to the maximum level, many ventifacts ultimately develop semi-planar faces, with the upper part of the abrasion face receding more rapidly than the lower part. Very large ventifacts may also exhibit a surface that slopes away from the prevailing wind, but owing to the greater height and mass of the rock, to a limited time of exposure to abrasion, and to variable rock resistance, such features are usually less well developed than those of smaller ventifacts.

Friction with the ground results in wind speed that is minimal near the surface. A sill of uneroded material at ground level developed on bricks and hydrocal blocks in an experimental plot (Sharp 1964) and on targets in wind tunnel experiments (Bridges et al. 2004). Owing to the complexity of interactions between topography, wind flow, and shifting sands, such a sill is commonly, but not always, observed in nature. On subhorizontal surfaces such as lava flows, saltating grains on a descending path erode flutes and grooves.

Characteristics of the Particle

As discussed earlier, the composition, shape, size and quantity of windblown materials all affect abrasion. Greeley et al. (1984) showed that there is little difference in abrasion by quartz and basalt particles, that ash is less efficient in erosion, and that aggregates may be plastered on to the target, forming a protective coating, and thereby lessening erosion. Laboratory experiments have shown that abrasion increases with particle diameter and with angularity, but field studies have rarely assessed these characteristics. Most abradants are moderately well sorted and in the size range of fine- to medium-aeolian sands (Greeley and Iversen 1986). The sharp-edged particles that develop in polar areas may be particularly abrasive (Miotke 1982).

Sharp (1949, p. 185) characterized the amount of sand necessary for ventifact formation as being “adequate but not too abundant”. He demonstrated a direct correlation between rates of erosion at his experimental plot and the influx of grains (Sharp 1964, 1980). Laboratory experiments show that where blowing sand concentrations are great, rebound effects from the target rock interfere with incoming grains and lessen the abrasion (Suzuki and Takahashi 1981). In the field, too great a supply of sand results in burial of ventifacts (Fig. 19.14).

Susceptibility to Abrasion of the Target Materials

The erodibility of different materials has been assessed experimentally by controlling for impact velocity, impact angle, and impacting particle size and type (Greeley and Iversen 1985). The bond strength of the rock appears more important than its hardness in predicting abrasion (Dietrich 1977a, Suzuki and Takahashi 1981). Experiments by Greeley et al. (1982) indicate that glassy materials such as obsidian will erode quickly for surfaces perpendicular to the wind, whereas crystalline materials such as granite and basalt erode more quickly when surfaces are subparallel to the wind.

The role of the angle of incidence of the incoming grain to the target has been assessed primarily through wind tunnel experiments. For steep angles (∼90°), the abrasion rate is lowered because rebounding grains hit incoming grains, slow them, and result in lower impact energies (Greeley et al. 1982). In the case of shallower angles (∼15°), grains tend to skid along the surface, and hit at the lower velocities characteristic of grains at the end of their trajectories, so that the surface is more likely to be lowered than reduced in angle. Thus, very steep (90°) and very shallow (15°) targets more or less maintained their shape.

By contrast, slopes in the range of 30–60° tend to undergo changes in slope with abrasion. Within this range, the abrasion rate is greatest for higher angled slopes (Bridges et al. 2004). Schoewe (1932) showed that the rate of abrasion on faces sloping at 30° is about one-third as great as faces inclined at 60°. As a result, abrasion rates tend to lessen through time as the facet becomes more inclined. Preliminary field evidence (Bridges et al. 2004) of eroded targets tends to support the wind tunnel results and suggests that, given enough time, targets will evolve to an angle of ∼30°.

The published results on slope angle appear to differ when the ventifacts become very small, although there is little information on this topic. Needham (1937) measured the facet angle for very small ventifacts (up to 1 cm in diameter) and found that most (63%) fell within the range 45–69°, 22% had high angle faces (70–89°), whereas very few (15%) had low angle faces (25–44°). No ventifacts were measured with facet angles less than 25°.

Rate of Abrasion

Natural rates of abrasion are difficult to determine and are probably highly variable through time owing to the many different controlling factors that influence ventifact formation. Wind velocities are not constant, but vary according to season, time of day, and the passage of fronts. Most of the abrasion occurs during periods of high velocity winds, which occur for only a small percentage of the time. Abrasion rates also change because, as discussed above, the rocks gradually wear, lowering the angle of incidence of the impacting grain and the height of the rock. If abrasion is episodic through time, weathering of the rock between erosional episodes may prepare it for abrasion and increase subsequent rates of surface wear. During periods of stability (no abrasion), weathering rinds may develop that cause surface hardening, necessitating higher than normal abrasion to renew erosion of the surface.

Time-dependent particle flux also determines abrasion rates. Even when sand is available, it may move intermittently through an area. A 15-year study of abrasion by Sharp (1980) showed an annual rate of wear 15 times greater during the last 3-year interval than in preceding years owing to an increased flux of windborne material derived from nearby fluvial flooding debris. Megascopically visible effects (polish, pitting, and incipient fluting) developed within 10 months during periods of intense erosion characterized by an abundance of windborne particulate material (Sharp 1980). Over short time periods (seasonally), ventifacts may be buried by sand and be protected. Over longer time periods, the availability of particles may decline through time owing to climatic change.

Rates of abrasion inferred from various ventifact sites range from approximately 0.01–1 mm y^{-1} (Greeley et al. 1984) and the time taken to form ventifacts may range from hours to days to months along storm-exposed coastlines (Kuenen 1960), to dozens or hundreds of years (Sharp 1964, 1980), to thousands of years (Selby 1977). Knight and Burningham (2003) examined coastal ventifacts formed within the past century and estimated abrasion rates of 0.24–1.63 mm y^{-1}. Abrasion is more rapid in Antarctica, primarily owing to higher wind speeds, with Miotke (1982) estimating rates of 5–20 mm y^{-1}. Liu et al. (2003) demonstrated that the abrasion capacity of saltating sand (the ratio of abrasion rate to aeolian sand transport rate) increased logarithmically with wind velocity.

In addition to wind speed, abrasion rates are influenced by particle supply, particle diameter and density, periods of burial, target hardness, and target surface roughness. Long-term observations of modelling foam targets at the Little Cowhole Mountains clearly indicate that softer materials abrade more rapidly than harder ones. Additionally, targets that are initially pitted abrade more rapidly than those that have smooth surfaces (Bridges et al. 2004).

Topographic Influences on the Development and Spatial Distribution of Ventifacts

Local and regional topography affect wind velocity, sand flux, and the direction of wind flow, and thereby influence the location of ventifacts, the orientation of flutes, grooves and facets, and the magnitude of erosion. Two topographic situations that commonly affect ventifact development are (a) wind speed increase through topographic constrictions, and (b) wind acceleration up the windward flanks of hills. At large scales, wind acceleration through constrictions may involve passage through a valley, and at a small scale, through a saddle or dip in an outcrop. A series of eight 100-m transects, laid out at 2-km intervals in a topographic constriction in a structurally controlled valley in the Mojave Desert, California, showed that faceting and grooving affected 70–90% of all exposed cobbles and boulders. In the area downwind of the constriction velocity generally declines as streamlines spread out, and ventifacts are absent (Laity 1987).

Wind also accelerates as it moves up the windward flanks of hills or dunes (Ash and Wasson 1983, Lancaster 1985). The compression of streamlines in the boundary layer causes the wind to accelerate towards the crest of the slope and then to decelerate on the downwind side. This is important geomorphologically because the increase in wind speed over the surface of the hill produces increased sand transport as well as additional surface shear stress (Jackson and Hunt 1975, Lancaster 1985). Mason (1986) found that for a smooth, nearly circular hill rising 70 m above the surrounding terrain, mean velocity 8 m above the surface $u(8)$ was reduced to a minimum of 0.8 $u(8)$ at the base of the upstream face and flow over the summit increased to 2.0 $u(8)$. These observations suggest that flow is reduced on the uphill face, increases on the sides and summit of the hill, and separates on the lee slope. The actual acceleration of wind depends on the height of the hill and the angle of the incident flow. The largest shear stress values, pressure changes, form drags, and wind velocities are recorded when the near-surface wind approaches approximately normal to the topography. The effect of velocity speed-up increases with the height of the topographic obstacle.

The threshold velocity for the movement of sand may be reached as a consequence of this acceleration of wind near the crest. Therefore, an increase in sand transport at higher elevations is anticipated (Fig. 19.14). This effect is particularly marked because the rate of sand transport is proportional to the cube of the excess of wind velocity over the threshold velocity for sand movement (Bagnold 1941). Lancaster (1985) reported that with a 6.5 m s^{-1} wind, sand movement at the crest of a transverse dune is 19.5 times greater than at its base if the dune is 5 m high, and 121 times greater if the dune is 20 m high. Higher wind speeds keep sand active at some hill crests in the Mojave Desert. Fossil ventifacts may be found on the lower hillslopes and active-forming ventifacts near the summits (Figs. 19.2 and 19.3).

Wentworth and Dickey (1935), in their survey of ventifact localities in the United States, noted that many ventifacts occur on the surfaces or margins of mesas where pebbles are exposed to strong, persistent winds from adjacent areas of lower elevation. Field mapping of ventifacts in the Mojave Desert showed that upper hill slopes and crests are favourable sites for ventifact formation and that, in many cases, the intensity of abrasion, as measured by pit diameter and groove width and length, often increases with elevation up the slope (Laity 1987). Local topographically enhanced velocity increase may allow winds of moderate velocity to be effective in abrasion (McKenna-Neuman and Gilbert 1986). Where regional winds are strong, deep grooving and pitting may occur, particularly on large boulders that present high-angle faces to winds (Figs. 19.1, 19.7, and 19.8).

Use of Ventifacts in Determining an Aeolian Regional History and Palaeocirculation

Numerous investigators have shown that wind direction may be determined by reference to the position of the sharpest bounding edge of a facet, by pitting on the face, or by the direction of grooving and fluting (Maxson 1940, Selby 1977, McCauley et al. 1980, Laity 1987, Nero 1988). For faceted ventifacts, the keel is oriented in a large majority of cases at right angles to the wind (Maxson 1940): in the central Namib Desert 93% of ventifacts have facets indicating a dominant wind from the north-east (Selby 1977).

Fossil ventifacts provide an excellent record for palaeocirculation reconstruction. As small ventifacts can change their orientation through time, large stable boulders and outcrops are the best choice for mapping. The relict nature of ventifacts is often indicated by the weathered, dulled and partly exfoliated condition of the rock surfaces (Blackwelder 1929, Smith 1967, 1984) and by the presence of rock varnish (Dorn 1986) or lichens. Grooves and fluting may not cover the entire surface of the boulder, but rather occur in patches where weathering has failed to remove them (Powers 1936). The growth of vegetation surrounding ventifacts, the lack of any apparent wind-blown material, or the stabilization and incipient soil development of aeolian deposits also indicates the fossil nature of ventifacts.

Numerous studies have used ventifacts to infer wind direction and palaeoclimate (Powers 1936, Sharp 1949, Tremblay 1961, Nero 1988, Smith 1984, Laity 1992) In the east-central Mojave Desert, California, three principal flow directions – westerly, northerly, and southerly – were identified from the analysis of mapped grooves (Smith 1984, Laity 1992). Relative ages of ventifacts were assessed by field relationships and an examination of the surface micromorphology of ventifacts. The widespread cessation of abrasion is marked by weathering of the micro-impact structures in grooves and flutes, by the formation of rock varnish, and by the stabilization of sands in the immediate vicinity of ventifacts.

Problems in reconstructing surface palaeowinds by ventifact mapping occur when there has been more than one pulse of erosion and high velocity winds emanated from different directions in succeeding episodes. The most recent winds may erase the imprint of earlier activity, although in rare cases cross-grooves are discernible. Multiple episodes of erosion are best preserved in hilly or mountainous regions, because erosion by earlier winds may be preserved on topographically protected (leeward) slopes.

Yardangs

Yardangs are elongate wind-eroded ridges that develop at a range of scales, from microyardangs (centimetre-scale ridges), to meso-yardangs (metres in height and length), to megayardangs (also called ridge and swale systems) that are tens of metres high and kilometres in length. Although closely related, mega-yardangs and meso-yardangs differ in their aerodynamic form and scale and probably in the relative roles of abrasion and deflation. Mega-yardangs are best developed in more resistant materials (Figs. 19.16 and 19.17), whereas micro- and meso-yardangs commonly form in softer sediments.

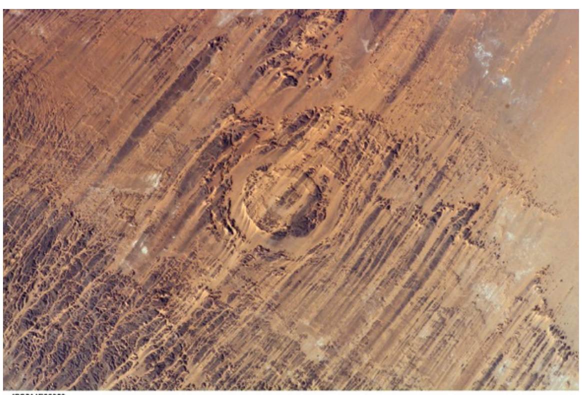

Fig. 19.16 Space Shuttle image of wind-eroded ridges following, in an arc, the deflection of the trade winds around the Tibesti Mountains. The system crosses the Aorounga Crater, Chad, near the centre of the photograph. Source: Image Science and Analysis Laboratory, NASA-Johnson Space Center. 10 Jul. 2006. "Astronaut Photography of Earth – Display Record." http://eol.jsc.nasa.gov/scripts/sseop/photo.pl?mission=ISS014&roll=E&frame=6304

Mega-Yardangs

Mega-yardangs have been recorded in a number of locations on Earth (Goudie 2007) and are of interest owing to their significance in understanding the aeolian history and geomorphic development of the Martian surface. In the Lut basin in southeastern Iran, yardangs up to 80 m high form very elongate ridges with flat to rounded summits, separated by troughs more than 100 m in width (Fig. 19.18) (McCauley et al. 1977). West of the Rio Ica, near Cerros Las Tres Pirámides, Peru are streamlined yardangs up to 1 km in length, developed in clastic Tertiary sediments. Near the town of Mangnai in central China, very large yardangs are eroded into lacustrine deposits at a scale comparable to some aeolian features observed on Mars. Martian yardangs include those of the Medusae Fossae Formation (MFF), spread across the Martian equator in the Amazonis and Elysium Planitiae regions (Bradley et al. 2002). In some areas of the MFF, yardangs reach 150 m in height: elsewhere, yardangs average 10–40 m in height. Jointing may play an important role in establishing yardang orientation in the easily-erodible MFF deposits.

One of the most spectacular mega-yardang systems is that lying on the southeastern flanks of the Tibesti massif of northern Africa (Figs. 19.16 and 19.17) (Grove 1960, Mainguet 1970). Mainguet et al. (1980) consider the Sahara a single vast aeolian unit, with wind being the most active geomorphic agent. The region is divided into sectors where either sand transport or deposition dominates. Sand seas or ergs represent the depositional sector. Zones of sand transport are characterized by landforms of erosion, including

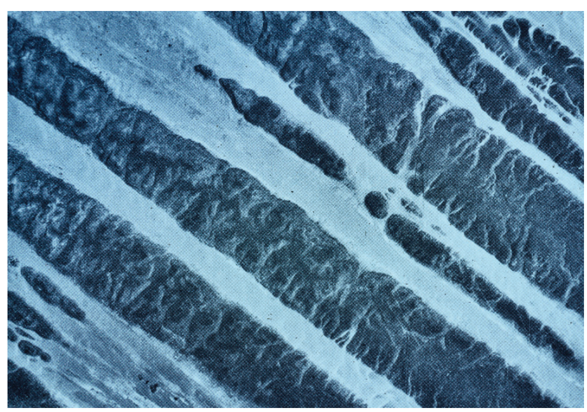

Fig. 19.17 Ridge and swale systems of the Sahara. Vast systems occur in a zone bordering the Tibesti Mountains to the east, south, and west. On aerial images the ridges appear dark owing to rock varnish, and the swales are lighter coloured owing to the presence of sand. In the Bembéché region of Chad, illustrated here, the ridges are very wide (up to 1 km) in proportion to their height (Cliché Institut National Géographique, Paris)

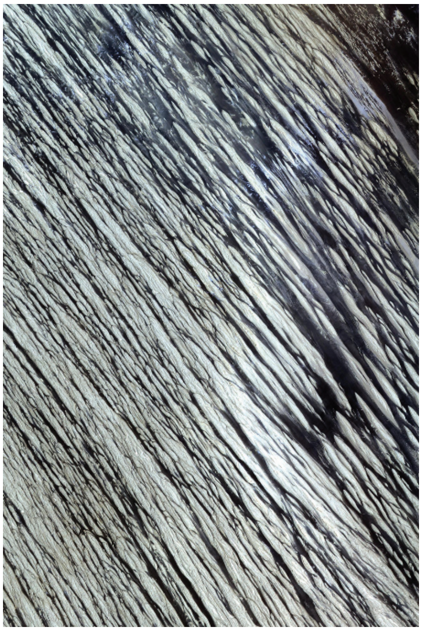

Fig. 19.18 The central portion of the Dasht-e Lut, Iran, is carved into yardangs that exhibit remarkable parallelism and a high length-to-width ratio. NASA Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) image, May 13, 2003

wind-abraded ridge and swale systems, yardangs, and zones of deflation.

Ridges and swales are best demonstrated by vast systems that occur in a zone bordering the Tibesti Mountains to the east, south, and west. On satellite and aerial images the ridges appear dark, owing to the well-developed patina of rock varnish, and the corridors show as lighter-coloured lineations. In orbital views, the systems appear to be continuous for hundreds of kilometres, sweeping in a broad arc around the mountains. On aerial photographs (1:50,000), the discontinuous nature of the systems is evident, with the largest ridges not longer than 4 km (Mainguet 1972). In the Bembéché region, which lies to the north-east of Faya-Largeau, the ridges are very wide (up to 1 km) in proportion to their height.

There are three factors that account for the development of these remarkable features: (a) extensive exposures of sandstone, (b) a dense network of joints that channelizes the wind, and (c) a monodirectional wind, charged with sand. The cover of Palaeozoic and Mesozoic sandstones is largely preserved on the southern flank of the mountains, whereas on the northern flank only isolated patches persist (Hagedorn 1980). The sandstone appears particularly susceptible to erosion, and forms of this type do not appear in basalts, crystalline bedrock, schists or siltstones (Mainguet 1972). In diatomites, the ridges have a scale of development ten times less than that of sandstone.

The circum-Tibesti region is deformed and fractured along two major axes, NE–SW and NW–SE. Wind exploits the fracture system most closely aligned with its own direction, gradually enlarging corridors, the size of which is a function of the deviation between the wind direction and joint trends. Where the coincidence is best, the corridors are, with respect to the ridges, the least enlarged and the most regular in form. Fractures that run counter to the ridge trends are often exploited to form the fronts of ridges or yardings, these appearing to line up in a row.

The general direction of the wind is determined by trade wind circulation, except where it is diverted around major topographic obstacles such as the Tibesti Mountains. Northern Chad, to the east of the Tibesti, is characterized by the constancy of sand-laden winds that blow 8 months out of 12 from the north-east. Maximum velocities are reached during the daytime, when sand is transported at velocities of 6–8 m s^{-1} or greater. The ridges follow, in an arc, the deflection of the wind.

Topographic chanelling between the Tibesti and Ennedi Mountains enhances the Bodélé Low Level Jet (LLJ), which in turn increases the frequency of erosive surface winds (Fig. 19.19). To the lee of the Tibesti, there is a pronounced downslope flow, with a subsidence core at around 1.5 km overlying the Bodélé depression. The large scale topography increases the magnitude of the jet over the region and increases the frequency with which episodes of deflation and abrasion occur. Simulation models suggest that the Bodélé LLJ would have been even stronger during the Last Glacial Maximum (LGM) (Washington et al. 2006).

The amount of sand in the corridors is variable, with some being totally engulfed. Sand abrasion acts preferentially in the corridors and is ineffective on the middle and upper slopes of ridges, which are marked by a deep patina above the basal fringe area of eroded rock. There is no evidence that deflation plays a significant role. Wider swales may be occupied by barchans, whose axes parallel those of the ridge systems and the resultant winds (Mainguet 1972). The barchans are indicative of sand transport through the region and are clearly evident on satellite images.

The channelized topography has a wavelength that, in cross-section, appears to be relatively constant for any given group of ridges and swales (Mainguet 1970). The periodicity is determined to some degree by the

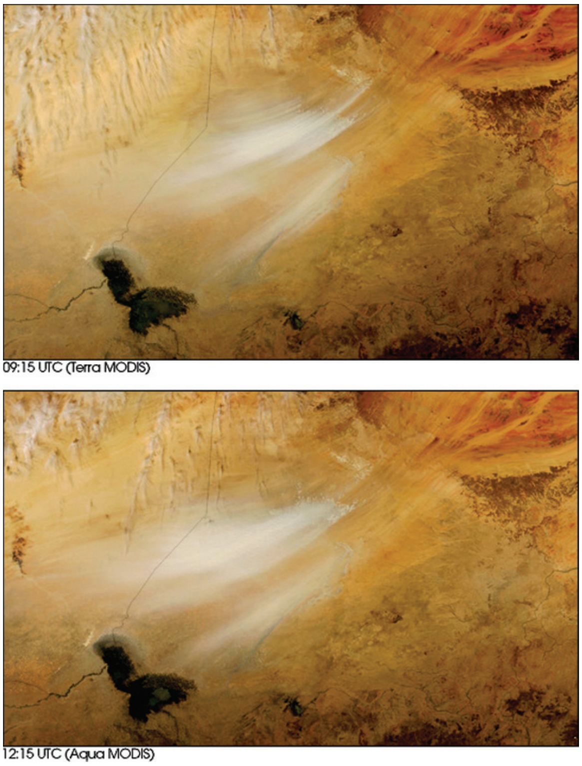

Fig. 19.19 Northeastern Harmattan winds are funnelled and intensified as they pass between the Tibesti and Ennedi Mountains (*upper right corner*) in northern Chad, causing episodes of abrasion and deflation in the Bodélé Depression, one of the most active global dust source regions. The erodible lake sediments are lowering rapidly, with 4-m high yardangs having formed in a period of 1,200–2,400 years. These NASA Moderate Resolution Imaging Spectroradiometer (MODIS) images show a February 11, 2004 dust storm with a 3-h interval between images. In the lower image, the storm is approaching Lake Chad

fracture density, but topography and wind strength also appear to play a role, with the largest ridges found on the more elevated parts of the terrain, and the smaller forms in the lowest. Abrupt changes in scale may occur as an escarpment is surmounted by the wind (Mainguet 1972). In some areas, two families of ridges and corridors may develop, with the larger system being the most elevated and the longest, and the smaller superimposed within the corridors.

An interplay through time between fluvial and aeolian processes is evident. Lacustrine sediments occur in the largest aeolian corridors, apparently emplaced after the initial cutting of the ridges. Deposition was followed by a renewed phase of aeolian excavation. Lacustrine deposits, as well as ravines developed on the slopes of some ridges, suggest alternating wind and water dominance. Fluvial action may also have etched and enlarged many of the joints in the sandstone, thereby allowing wind channelization to occur. At the present time, wind action is dominant, and acts to erase most traces of fluvial activity.

Meso–yardangs

In contrast to ridge systems, yardangs are more streamlined and aerodynamic in form. They are often described as resembling an inverted ship's hull, although in many cases the yardangs are flat-topped. The windward face of the yardang is typically blunt-ended, steep and high (Fig. 19.20), whereas the leeward end declines in elevation and tapers to a point (Whitney 1984) (Fig. 19.21). It should be noted, however, that yardangs take on many different forms (Whitney 1983, Halimov and Fezer 1989).

Measurements of yardang length:width ratios commonly average 3:1 or greater. This elongate form minimizes the drag or resistance to the wind. Yardangs form parallel to one another, typically occurring as extensive fields (Fig. 19.18), and their long axes are oriented parallel to the strongest regional winds. They may occur as tight arrays, separated from one another by either U-shaped or flat-bottomed troughs, or as widely spaced, highly streamlined features on wind-bevelled plains.

History of Yardang Research

Yardangs have received attention as curiosities and geological oddities since the late 1800s. Stapff (1887) gave an account of "aerodynamic landforms" sculpted

Fig. 19.20 Windward face of yardang at Rogers Lake, Mojave Desert, California, showing a blunt end and erosional markings

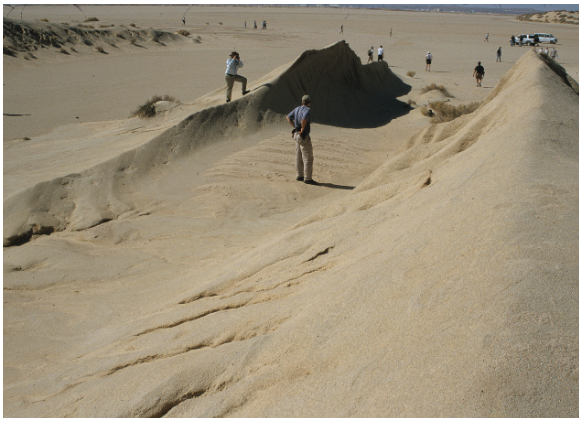

Fig. 19.21 Yardangs at Rogers Lake, illustrating the sub-parallel nature of the long axes and a form that tapers to the lee

out of bedrock in the Kuiseb valley of the Namib Desert of south-west Africa, and yardangs were described by Walther (1891, 1912) in Egypt and by Kozlov (1899) east of Lop Nor. The term yardang was introduced by Hedin (1903, volume 1, p. 350) to describe a labyrinth of clay "terraces" in the Lop Nor (Nur) (Mongolian *nur* means lake) region at the eastern edge of the Taklimakan Desert in China, which the natives of the area called yardang. Hedin (volume 2, p. 139) attributed their formation to initial erosion by running water, and subsequent resculpturing by the wind. More recent interest in yardangs was fostered by the discovery of large yardang fields on Mars (Ward 1979) and by the availability of aerial photography and satellite imagery (Mainguet 1972, Ward 1979, Halimov and Fezer 1989).

Yardangs have not been studied in detail and much remains to be determined about their formation. There is little meteorological information available for most yardang fields. Although it is apparent that most form in association with monodirectional wind regimes, the velocity of winds within the corridors and the frequency of sand-transporting winds are not known. Nor is there any knowledge of the interplay between wind and topography in major yardang and ridge systems. Our understanding of wind flow around individual yardangs is based primarily on laboratory and theoretical determinations (Ward and Greeley 1984, Whitney 1985), rather than instrumentation. The long-term evolutionary development of yardang fields, the role of fluvial erosion and jointing in initial channelization of winds, and the rates of formation are poorly understood. The role of abrasion and deflation in determining the ultimate form of yardangs also remains uncertain, and probably varies according to the rock type.

Factors Affecting the Distribution of Yardangs

Yardangs occur in desert regions on all continents, but they vary greatly in their scale, development, and spatial extent. Overall, they occupy only a very small part of the Earth's surface, as they require conditions of great aridity, nearly unidirectional winds, and, in some cases, a favourable material and some assistance from weathering to form. Nonetheless, aerial photography and satellite imagery show that yardang fields in some regions are of great areal extent. McCauley et al. (1977) and Goudie (2007) provide comprehensive discussions of the Earth's major yardang fields. In Africa, yardangs occur in the Arabian Peninsula, Egypt, Libya, Chad, Niger, the Namib Desert of southern Africa, and possibly along the coast of Mauritania. Asia has several yardang fields, including those of the northwestern Lut Desert, Iran, the Taklimakan Desert in the Tarim Basin (described by Hedin (1903)), and the Qaidam depression of Central Asia. Well-formed yardangs are found along the coastal desert of Peru in South America. They occur as minor groups in North America and Australia. In Europe, relict yardangs are present as a small field in the semiarid Ebro Depression of Spain (Gutiérrez-Elorza 2002). Satellite imagery suggests that yardangs are also widespread on Mars.

Yardangs develop where wind action dominates over fluvial processes, and are thereby limited to extremely arid deserts. In order for the wind to be effective, plant cover and soil development are generally minimal. In the Borkou region of Chad, for example, no vegetation occurs expect where subsurface oasis water is within the reach of plant roots. Many yardang fields develop where strong unidirectional winds occur throughout much of the year (Hobbs 1917, McCauley et al. 1977), but others, such as those of the Lut Desert, develop in regimes of seasonally opposing winds, where one wind is dominant and the other lighter and less frequent.

Yardangs form in a broad range of geologic materials, including sandstones, limestones, claystones, dolomites, granites, gneisses, schists, volcanic ignimbrites and basalts, and lacustrine sediments. The lithologies of many of the major yardang fields are summarized in Goudie (1989).

The role of topography in yardang development has not been fully explored. A number of yardang fields, including those of Lop Nur and the Lut Desert, occur in topographic depressions or are surrounded by mountain ranges that rise to considerable heights above the desert floor. Hagedorn (1980) observed that around the Tibesti Mountains there is an altitudinal zonation of geomorphic processes, with aeolian corrasion dominant at the lowest elevations, decreasing in intensity up to an altitude of 800 m.

Form and Scale Relationships

Wind tunnel experiments and measurement of mature, streamlined yardangs suggest an ideal length-to-width ratio of 4:1 (Ward and Greeley 1984), independent of scale. Yardang form is highly variable, however, owing to variations in lithology, wind strength and direction, and yardang age. The ideal proportions are probably approached only after a long period of erosion. In Peru, well-developed, streamlined yardangs have ratios ranging from 3:1 to 10:1 (McCauley et al. 1977). In the Qaidam Depression of Asia, some yardangs attain a length: width: height ratio of 10:2:1, apparently under conditions of high wind speeds that result from a long fetch (Halimov and Fezer 1989).

Although usually of dimensions measured in a few metres, some yardangs can attain heights of as much as 200 m and be several kilometres long (Mainguet 1968). Ridges in the Qaidam Depression attain 5 km in length (Halimov and Fezer 1989) and megayardangs up to 10 km in length are developed in Holocene basaltic flows in the Payun Matru Volcanic Field in the southern Andes Mountains, Argentina (Inbar and Risso 2001).

The inter-yardang spaces have been variously termed troughs, couloirs, corridors, swales, and boulevards. Where the inter-yarding space is narrow, troughs appear to be U-shaped; as they widen, their bottoms become flattened (Blackwelder 1934). In some cases, flattening may be due to a resistant stratum, such as a hard clay layer. Although attention is often focused on the yardang, most of the geomorphic activity appears to be concentrated in the troughs themselves. The troughs may be totally engulfed in sand, or be only partially sand covered (Mainguet 1972), show low transverse ridges of fine gravel (Blackwelder 1934) or ripple trains that diverge at the head and converge in the downwind direction around the yardang flanks (McCauley et al. 1977). Lag surface of pebbles or even mollusc shells may develop around yardangs and reduce erosion in the corridors (Mainguet 1972, Brookes 2001, Compton 2007). Corridors are commonly occupied by migrating barchan dunes whose major axes parallel those of the ridges and the resultant winds (Gabriel 1938, Hagedorn 1971, Mainguet et al. 1980). The rocks in the corridors often carry numerous marks of aeolian erosion, including longitudinal striations, and shallow erosional basins, metres or tens of metres in length, that occur either as isolated forms or groups (Mainguet 1972).

Processes of Yardang Field Formation

Yardangs are probably produced by a combination of abrasion and deflation and further modified by fluvial erosion, weathering, and mass movement. The significance of each of these processes in determining the ultimate form varies according to climate and yardang lithology and structure.

Abrasion

Abrasion by sand particles is probably the dominant process by which most yardangs form (Fig. 19.22). In many yardang fields, the passages are filled with aeolian sand (Grolier et al. 1980, Halimov and Fezer 1989) or gravel (Blackwelder 1934) that erodes the corridors and lower yardang slopes. Mainguet (1972) and Hagedorn (1971) emphasize the episodic nature of sand transport, so that yardang fields may be temporarily free of drifting sand even within a wind-corrasion landscape. The inter-yardang corridors are zones of transportation and erosion, not deposition.

In addition to influencing the overall form of the yardang, abrasion also affects the micromorphology, fluting and polishing the surface, and affecting its colour. The windward face commonly develops a well-developed re-entrant form. Polish and fluting typically occur to a height of one or two metres (Hobbs 1917, Hagedorn 1971, Grolier et al. 1980).

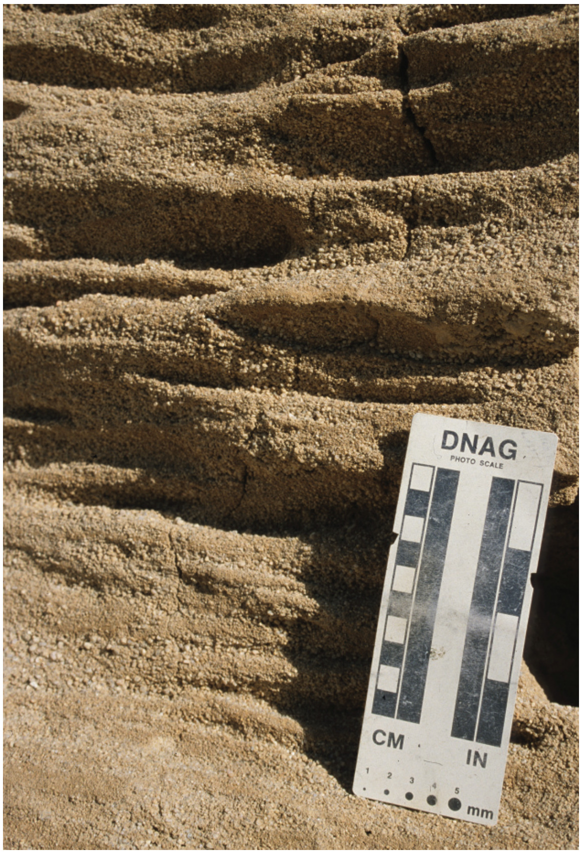

Fig. 19.22 Abrasional detail on the sides of a yardang at Rogers Lake

In sandstone yardangs in Africa, the upper yardang surface is rough, unshaped by the wind, and covered by a dark weathered crust, whereas the lower slopes are smoothed and lighter in colour (Hagedorn 1971, Mainguet 1972). In the southern Namib Desert, crystalline dolomite yardangs are streamlined up to 10–15 m in height (Corbett 1993). Actively forming yardangs are smoothly polished and show fluting, whereas relict forms have rough surfaces colonized by lichen and affected by solutional weathering.

Numerous small-scale features may retard abrasion, including salt crusts in the couloirs, armoring by mollusc shells in pan sediments (Compton 2007), and clay-rich drapes resulting from rainfall (Hörner 1932).

Deflation

Deflation removes relatively loose materials from the yardang surface, including unconsolidated sediments, or grains that are weathered from consolidated or crystalline materials. The dark patina of sandstone yardangs in the Sahara indicates that there is little active removal of material from the ridge summits, and indeed the varnish probably aids in cementing the grains and protecting them from deflation. Similarly, limestone yardangs in Egypt commonly have flat irregular tops that retain weathered surfaces that pre-date erosion (El-Baz et al. 1979).

Deflation may play a more important role in poorly indurated lacustrine material. Formed in fine-bedded white siltstones, yardangs 30–50 m high and up to 1.5 km long on the Pampa de la Averia, Peru, possess smooth, streamlined shapes from base to crest (McCauley et al. 1977). The aerodynamic form is the primary evidence of deflation. As yet, there have been no field studies to confirm that erosion and modification of yardangs take place by this process.

Researchers differ in their opinion as to the relative importance of abrasion and deflation in forming yardangs, even in a small field such as that at Rogers Lake, California (Figs. 19.20, 19.21 and 19.22). These yardangs are carved in moderately consolidated deposits containing beds of fine gravel, sand, silt, and clay (Ward and Greeley 1984). McCauley et al. (1977) felt that they attained their streamlined shape and smooth ridge crests principally by deflation, as evidenced by the lack of small-scale grooving and scouring, and the winnowed appearance of the rocks caused by quartz grains and clay swept away in suspension: abrasion played a minor role, causing some undercutting of the windward end and flanks, and contributing to trough lowering. On the other hand, Ward and Greeley (1984) considered abrasion to be the most important process in the development of the Rogers Lake yardangs, dominating trough formation and initial sculpting of the ridges. Thereafter, deflation increases in importance as it combines with abrasion to maintain the aerodynamic shape. Blackwelder (1934) also considered that the rounded form of low yardangs at Rogers Lake results from abrasion by saltating grains that flow both around and over the forms.

Fluvial Erosion

Running water acts in several ways to aid in yardang field development. At the outset, yardang fields may be initiated along stream courses that are enlarged and modified by the wind. Li Daoyuan (466–527 AD) referred to the formation of the yardangs of Lop Nor, China, as "rills cut by water is [*sic*] blown by wind subsequently" (Xia 1987). As climate changes or streams are diverted, yardang landscapes may be flooded. Hörner (1932) describes a condition at Lop-Nor where "millions of small yardang hillocks and groups of them stick up like islets, while the hollows between them are filled with water (p. 312)." Proximity to piedmont slopes and their periodic streamflow was considered to be critical to the dissection of yardangs in the Lut Desert (Krinsley 1970). Subsequently, the inter-yardang troughs may be fluvially eroded and the yardang slopes gullied. Although fluvial processes often initiate and abet yardang formation, the development of yardangs in hyperarid environments (including Mars) suggest that running water may not be essential to their formation.

The role of running water in yardang modification is most evident in yardangs composed of relatively soft deposits, such as lakebed silts and clays. At Rogers Lake, the concave-upward form of yardang flanks is thought to result from sheet-wash and gullying (Fig. 19.23): in more arid regions, such as coastal Peru, streamlined yardangs have broader crestlines and convex upward flanks. Intense aeolian abrasion destroys evidence of fluvial erosion at the windward ends of Rogers Lake yardangs, with the more stable mid-sections harbouring the largest gullies, and gully development hampered at the leeward by deposition of a porous mantle of aeolian sediment (Ward and Greeley 1984). Small alluvial cones deposited at the gully bases are destroyed by sand blasting (Blackwelder 1934). In the Lut Desert, Iran, winter rains cause intense gullying, earthflows, and solution on the flanks and summits of 60-m-high yardangs formed in playa sediments. The upper slopes are well above the range of effective abrasion and deflation is too slow a process to remove evidence of fluvial dissection (Krinsley 1970).

Fig. 19.23 Fluvial erosion of yardang at Rogers Lake

In areas of the Sahara, the gullies observed on yardang flanks may be relict of earlier, wetter climates, formed when lacustrine sediments were deposited in the inter-yardang corridors. Today, erosion of the deposits and the filling of gullies by sand indicate that wind is the dominant process (Mainguet 1972).

Almost all yardangs are found in very arid conditions, as frequent fluvial erosion destroys the yardang form and vegetation limits wind erosion. An exception appears to be a cluster of approximately 50 relict yardangs in the Ebro Basin, Spain, that still retain their form despite an annual rainfall of ∼400 mm and partially-vegetated flanks (Gutiérrez-Elorza 2002).

Mass Movement

Mass movement has received little attention as a modifying process, although the presence of slump blocks alongside yardangs appears to be quite common, particularly where the nose or base of the yardang has been undermined by abrasion or where the yardang is formed of strongly jointed materials such as sandstone (Hörner 1932).

Weathering

It is likely that weathering plays a role in preparing material for removal of the wind, although such processes are not well documented. As many yardangs are developed in or near to playa sediments, it is likely that salt weathering is particularly significant. Stein (in Hörner 1932) refers to "salt encrusted yardangs" in the Lop-Nor region. Grains weathered free from yardangs by salt weathering are subject to removal by deflation.

Models of Yardang Development

Yardang fields may be initiated in gullies, fractures, and tectonic features that are aligned parallel to the direction of the prevailing wind and become enlarged by this wind (Hedin 1903, Blackwelder 1934, Ward and Greeley 1984). As air enters the passage, the streamlines are compressed and the wind accelerates. Sand that is carried through the passage abrades the bottom and sides of the trough, causing the slopes to become progressively steeper (Blackwelder 1934). During the early stages of formation, wind and occasional fluvial erosion erode the passages more rapidly than the yardangs, causing the passages to deepen and the yardangs to grow (Halimov and Fezer 1989). The effects of abrasion are most evident within a metre of the floor of the trough, decreasing in intensity on the upper slopes of the yardang. The low levels to which abrasion is effective account for the commonly ragged appearance of the crests of yardang ridges.

Blackwelder (1934) suggested that yardang troughs are initially elongate and somewhat sinuous in form, become wider with time, and finally breach the ridges at places of weakness. As abrasion is more intense at the yardang prows, the ridges become shorter and smaller, eventually developing into conical hills, mesas, and pyramids (Halimov and Fezer 1989). Blackwelder (1934) likened the processes that destroy the windward ends of yardangs to those which affect ventifacts or wind-abraded outcrops.

When the yardangs are lowered to approximately 1 m (Rogers Dry Lake) or 2–3 m (Qaidam Basin) in height, sand blasting affects the entire form, and the crests become more rounded and streamlined. In some cases, the hills vanish, leaving wide interspaces. Halimov and Fezer (1989) noted that only the streamlined whalebacks, which appear to be the most aerodynamically adjusted forms, survive for a long period. These progressive changes in yardang form are slowed by two processes: the development of a wind-resistant lag material on aeolian sands (Brookes 2001), and the cementation of yardang walls or crests by crusts variously composed of sand grains, loess, clay drapes, and salt.

Variations in the evolutionary form of yardangs occur because of material differences. Where the wind erodes horizontally layered rocks of varying resistances, tabular forms with protruding shelves or stair-step profiles are formed. On the other hand, inclined or vertically layered sediments will be eroded into ridges characterized by grooves and fins (Blackwelder 1934). Yardangs sometimes develop at angles to the structural grain of the rock. Joints and faults cross wind-shaped forms in the Sahara, and differential erosion in schists may result in a structural grain that runs obliquely to the form of the yardang (Hagedorn 1971).

Rate of formation of yardang fields

The rate of formation and modification of yardang fields is not well constrained, but several studies suggest that development takes less than 10,000 years, with much faster erosion in soft playa sediments than hard rocks. Yardangs at Rogers Lake, Mojave Desert, appear to be eroding by abrasion of the headward end at a rate of 2 cm y^{-1}, and by lateral erosion caused by both abrasion and deflation at 0.5 cm y^{-1} (Ward and Greeley 1984). The attachment of some yardangs to the original shoreline deposits suggests that this small yardang field is still evolving. Halimov and Fezer (1989) calculated that small yardangs in the Qaidam Basin formed in less than 1,500–2,000 years, based on dated pottery in the silty sediment. Yardangs in the Lop-Nor region, China, are also relatively young, and still retain vestiges of dead vegetation (copses) that are remanent of earlier times when the Tarim River flowed near the abandoned town of Lou-Lan. Hörner (1932) estimates that they formed within the past 1,500 years by the action of drifting sand. In the eastern Sahara, Haynes (1980) estimated that the development of mature yardangs, several meters in height, took place over several thousand years in soft playa muds, facilitated by a change to hyperarid conditions and a drop in the water table. In the Bodélé Depression, Chad, 4 m high yardangs in diatomites probably formed within 1,200 to 2,400 years (Washington et al. 2006).

There has been very little research on the rates of landform erosion in harder materials. Basaltic yardangs, 2–3 m in height, formed in less than 10,000 years in the Payun Matru Volcanic Field, Argentina (Inbar and Risso 2001). Rates of wind erosion in bedrock megayardangs, such as along the coast of Namibia or in the Sahara, are poorly constrained, although the "erasure" of large parts of the landscapes suggests that very long time periods are required. Millions of years may be involved, particularly in deserts that had their origin before the Pleistocene (Goudie 2007).

Wind Tunnel Simulations of Yardang Development

Ward and Greeley (1984) conducted wind tunnel experiments to simulate yardang development. Natural rock samples require an impractical duration of exposure to abrasion before significant erosion occurs, and therefore synthetic sediments were used, and soap bubbles substituted for sand. Several forms (such as mounds and cubes) were moulded and subjected to uniform winds. The samples evolved in a common sequential order, which resulted in final length-to-width ratios of about 4:1. Erosion proceeded from the windward corners, to the front slope and crestline, to the leeward corners, and finally to the leeward slope. Abrasion dominated at the windward end of the yardang, whereas deflation and reverse flow were more important near the middle and at the leeward end. Rates of erosion were greatest at the beginning of the experiment and diminished as the form became more streamlined.

Streamlined shapes are developed by modification of the original form. Where the original form is broad, erosion decreases its width, and elongation may occur by deposition of a tail. If the original form is more elongate, the dominant change will be a decrease in length.

Small-Scale Aeolian Grooving

Small-scale channelling is a feature less often mentioned in the literature. Worrall (1974) described channelling on nearly horizontal surfaces in the Faya district of Chad. The grooves are remarkably parallel and regularly spaced, but shallow (a few centimetres in depth and width) and discontinuous. They occur in homogeneous diatomite and on saline crusts and are formed by saltating sand grains. Grooving has also been observed by this author on clay surfaces of the Mojave River bed, traversed by blowing sand; on dry playa surfaces where saltating grains have created dust events; and on frozen sand dune surfaces at the Coral Pink Sand Dunes, Utah.

Desert Depressions

The role of wind erosion in the excavation of both small- and large-scale desert depressions is difficult to determine, as closed depressions may form by a number of different mechanisms, often acting in combination. These include (a) block faulting, (b) broad shallow warping, (c) crater lake development in volcanic areas, (d) chemical weathering and solution, (e) zoogenic processes, and (f) wind erosion.

The origin of very large enclosed basins, such as the noteworthy depressions of the Western Desert of Egypt (Siwa, Qattara, Baharia, Farafra, Dakhla, and Kargha), has been subject to several interpretations. Several of these basins have floors that lie below sea level (e.g., Kharga at 18 m BSL and Qattara at 143 m BSL). Depressions occur along the boundaries of northward-dipping strata and are bounded to the north by escarpments and to the south by gently rising valley floors. The depressions are often cited as textbook examples of deflation, with the depth of erosion limited by the groundwater table which forms a base level (Ball 1927). Knetsch and Yallouze (1955) invoked tectonic action in their models of depression formation. Said (1960) excluded the possibility of a tectonic origin, noting the absence of faults in the regions, which indicates that uplift was not accompanied by any significant tensional stresses.

The role of wind action is suggested by the conformity of depression locations with areas of thinner and more easily breached limestone capping. However, the sheer volume of the depressions, most notably the Qattara depression (20,000 km^2), suggests that the wind alone could not have excavated the material. Said (1981, 1983) showed that the southern part of the Western Desert experienced a wet climate through most of the Tertiary and Quaternary, punctuated by brief episodes of aridity. Albritton et al. (1990) proposed that the Qattara Depression was originally excavated as a stream valley, subsequently modified by karstic activity, and further deepened and extended by mass wasting, deflation, and fluviatile processes. Salt also played a role, by weathering and preparing materials for deflation. In a self-enhancing process, the creation of low areas alllows water and solutes to accumulate, and the crystallization of salts weakens sedimentary cements, allowing further deflation and salt accumulation in the lowered basins (Haynes 1982). Thus, the research to date suggests that the depressions probably had a polygenetic origin, with wind erosion playing a major role only during arid phases of the Quaternary.

Remote sensing has revealed that smaller depressions, termed pans, are widespread. They are especially well developed in southern Africa, on the High Plains of the U.S.A, and in western and southern Australia (Goudie and Wells 1995). Pans may develop in interdunal basins, as palaeodrainage depressions aligned along former river courses, and by excavation on the floors of former pluvial lakes (Goudie 1999).

Research into the origin of pans indicates that a single mode of genesis is unlikely, with several predisposing factors for their development. A dry climate is important, as is a vegetation-free surface that enhances wind flow and permits deflation. During periods of enhanced aridity, the water table lowers, allowing deepening of the basin (Haynes 1982, Holliday 1997, Langford 2003). Pan growth requires materials susceptible to deflation, and depressions form preferentially in poorly consolidated sediments, shales, and fine-grained sandstones. Space Shuttle photography indicates that pans are important source areas for dust storms (Middleton et al. 1986).

Several feedback mechanisms enhance deflation. Water that accumulates in the depressions evaporates to leave salts, which further retard vegetation growth and make sediments more susceptible to deflation by comminution of debris through salt weathering. (Haynes 1982, Goudie 1989). As the depressions enlarge, they become more attractive to grazing animals who, attracted by the lack of cover and availability of water and salt licks, further disrupt the surface and render it more prone to erosion (Goudie 1989).

Inverted Topography

Inverted relief develops when previously low areas of the landscape, such as river channels, are left high standing by erosion in a later phase of topographic development. This process is not commonly described in the aeolian erosion literature. Hörner (1932) remarks that in the yardang landscape of the Lop-Nor basin, China, former rivercourses have become inverted and are now marked by ridges and remnant hillocks. This inversion is attributed to the comparative resistance of the silty riverbed with respect to the other soft, young sediments in the region.

Integrated Landscapes of Aeolian Erosion and Landform Hierarchies

Although desert depressions, yardangs, and ventifacts may develop in isolation from one another, they are formed by similar processes and, as such, are often found in close proximity, often as part of an integrated aeolian system of erosion, transportation, and deposition. An example of such a landscape is found in the southern Namib Desert, where sand derived from the coast moves through an elongate deflation basin (20 km wide and 125 km long, including subbasins lowered to below sea level by salt weathering and deflation) to merge to the north with the Namib Sand Sea. (Corbett 1993). The system conveys sand in barchan trains from the coast to the sand sea, forming ventifacts, depressions, and yardangs en route. Changes in sea level dictate sediment availability which, in turn, affects the development of erosional features (Corbett 1993, Compton 2007).

Landscapes of aeolian erosion are also strongly associated with dust production. The role of saltating sand grains in producing dust is well established. Sand grains break up soil aggregates by ballistic impact, overcoming the otherwise strong cohesive forces associated with small particles (Gomes et al. 1990, Shao et al. 1993). This is nowhere more apparent than in the Bodélé depression, presently considered to be the world's greatest dust source (Fig. 19.19). An analysis of this landscape suggests that erosion is part of a system of processes that must be viewed over a palaeotimescale. Enhanced deflation by the Bodélé Low Level Jet (LLJ) during the Last Glacial Maximum (LGM) lowered the surface, creating a depression that was populated during wetter Holocene phases by diatoms associated with Mega-Lake Chad. This highly erodible material is conservatively lowering at a rate of 0.16–0.31 cm yr^{-1}, allowing 4-m high yardangs to form within 1,200–2,400 years (Washington et al. 2006).

Aeolian bedforms arise from a two-way interaction between the surface and the airflow, involving a transfer of material and modification of form until a dynamic equilibrium is reached. Within the erosional landscape, hierarchies of landforms develop, from small scale aeolian grooving, to ventifacts of varying scale, through micro-yardangs, to normal yardangs, to megayardangs. As in other landform systems, in an aeolian erosional landscape there are large numbers of small features and very few large ones. Evans (2003) notes that lineation and streamlining are often associated with scale-specificity. On a local scale, yardangs within a field are all similar in size, but between different areas, their scale varies greatly, owing to differences in materials, time span, and process activity and intensity. Crest spacing is broadly similar within a field, such as the 20–40 m spacing in Holocene diatomites in Chad and the 1.6 km spacing of Palaeozoic sandstone megayardangs (Mainguet 1970, Evans 2003). For ventifacts, the scale hierarchy also exists: numerous ventifacts with small-scale features are found on the plains and lower hillslopes, but feature scale increases with altitude and sandblast activity and intensity, with the largest forms occurring on selected hilltop locations.

Yardangs have a microrelief that is quite different from the general form: in other words, small yardangs are generally not found to piggyback on larger structures. Erosional features such as flutes on yardangs are similar to those found on nearby ventifacts, emphasizing the similarity of process at the smaller scale.

Conclusions

Landforms of aeolian erosion vary in scale from vast systems of ridges several kilometers in length and up to one kilometre in width to small grooves only millimetres in amplitude. None the less, they exhibit many forms in common. Small faceted ventifacts exhibit shallow grooves; as ventifacts increase in size the grooves and flutes become more fully developed and dominate the form; outcrops in the immediate vicinity may be similarly abraded, as are the lower slopes of some yardangs; near-horizontal surfaces may be fluted and grooved; and ultimately small-scale aeolian grooving (Worrall 1974) may be transitional to yardangs and ridge and swale systems.

The two primary mechanisms of aeolian erosion are deflation and abrasion. Deflation contributes to the formation of large-scale features such as depressions and streamlined yardangs composed of easily erodible materials. Abrasion by the impact of sand grains is the primary process by which ventifacts form and is a major contributor to the development of yardangs and ridge systems in indurated material. Erosion by suspended grains has not been documented in the field, with particles commonly swept around an obstacle rather than striking it directly.

Landforms of wind erosion have received much less attention than those of deposition and are still quite poorly understood. Fundamental questions as to the relative role of abrasion and deflation in the formation of yardangs, the mechanism of microfeature formation on yardangs and ventifacts, the interaction of yardang and ridge systems and wind flow, and the age, evolutionary history and rates of formation of erosional landforms remain await further process-oriented research. Such studies will provide an improved basis for understanding the climatic and geomorphic history of arid regions.

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