Chapter 4

Weathering Processes and Forms

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'See how that huge boulder has been split by the sun and wind.'
Commentary by tour guide, Namib Desert, July 2006.

Introduction

Previous reviews of weathering in deserts (e.g. Cooke et al. 1993, Goudie 1997) have been excellent at identifying the mechanisms considered to operate and the landforms with which they are generally associated. Invariably, however, such reviews – especially if orientated towards students – deal primarily with perceived certainties. In reality, weathering studies continue to be characterized more by uncertainties and gaps in knowledge – especially in deserts. This chapter will therefore attempt to concentrate upon the ongoing development of ideas. The aim is not to be exhaustive or comprehensive, but by focusing on a limited number of underlying themes it hopefully questions some traditionally held views and could stimulate future research.

Background to Weathering Studies in Deserts

The harshness and difficulty of access to many desert environments has meant that, compared to many other climate zones, much research is still at the exploratory stage; the nature of desert environments is incompletely understood and many features have yet to be fully described. Because of a comparative paucity of rigorous fieldwork, many geomorphological studies have also relied heavily upon the uncorroborated reports of early explorers and data collected for other purposes. This is particularly the case for weathering studies where, for example, views on temperature and moisture regimes have been strongly influenced by accounts of 'baking hot days', and the sounds of rocks spontaneously splitting during 'freezing nights'. Whilst such reports were no doubt successful in selling travelogues, they have had the unfortunate side effect of creating a popular view of deserts as places of uniformly extreme diurnal temperature regimes, and the total absence of moisture apart from the very infrequent flash flood. This has in turn influenced perceptions of how processes operate and encouraged a concentration upon the role of extreme events, be they the highest ever rock temperature or the longest period between rainfall. Nowhere is this better illustrated than in the still widespread popular belief in, and the numerous theories that have grown up around the concept of insolation weathering. Thus, as pointed out by Gómez-Heras et al. (2006 p, 237), 'insolation weathering has became 'a dogma that is solidly accepted' (Blackwelder 1933), 'a matter of speculation' (Cooke and Warren 1973), or an 'academic abstraction' (Camuffo 1998). This is despite, as Smith and Warke (1997) identified, a growing acknowledgement amongst researchers that, whilst temperature is a major driver of desert weathering, there are many other environmental agents operating with the means, motive and opportunity to weather rocks in deserts and the high probability that they will 'collude' to bring this about. An area where

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this preoccupation with the effects of environmental extremes has had a particularly significant impact is in the design of weathering simulations, which, in the absence of field studies, have formed a cornerstone of much weathering research.

It must be acknowledged, however, that the last few decades, and even the period since the first edition of this book, have seen a significant increase in the number and objectivity of field studies into weathering in deserts. Coincident with this – at least in the English-speaking world – has been a tendency to concentrate on the investigation of weathering process and their reduction to the basic physical and chemical mechanisms that control their operation. In the specific context of desert weathering studies this can, however, lead to difficulties. First, the majority of *in situ* observations of weathering mechanisms tend to be short-term and of restricted spatial extent. This creates problems as to their long-term significance in regions where few areas have been mapped in detail, and where little is known about medium- and long-term climatic variability. Secondly, process studies focus attention upon features that exhibit measurable change, especially within the one or two years over which the fieldwork for most research projects is carried out. Consequently, features such as cavernous hollows (tafoni) have received much attention, but, until relatively recently (e.g. Migon et al. 2005) few studies have examined the intervening cliff faces or the debris slopes below them. Thirdly, the reductionist tendency has fostered a segregationist approach, wherein specific mechanisms such as insolation weathering, salt crystallization, or freeze–thaw are examined in isolation. This approach has been prevalent in laboratory-based simulation studies, where it is their aim – and their virtue – that they can examine a particular mechanism under controlled conditions. In doing this, however, environmental conditions may be unrealistically simplified and possible synergisms between weathering mechanisms and other geomorphic processes discounted. Finally, in common with other areas of geomorphology, we are only now beginning to question whether our improved understanding of weathering mechanisms can allow us to cross scale barriers and allow the explanation of either the present configuration of most desert landscapes or their long-term evolution. Process studies are only a limited end in themselves. If we are to justify our preoccupation with them we must ultimately breach the spatial and temporal boundaries that separate individual, short-term studies at the microscale from their cumulative impact on the landscape.

Despite the above observations our understanding of weathering processes and environments and the landforms they produce has steadily progressed. Many of the early misconceptions concerning weathering have been dismissed or modified and we are beginning to understand aspects of the fundamental controls upon rock breakdown and some of the key questions that need to be answered. For example, Cooke et al. (1993) listed six observations that they considered should guide our understanding of weathering in deserts.

(a) Weathering processes are likely to be distinctive because of distinctive diurnal and seasonal temperature and relative humidity regimes.
(b) Contrary to popular belief, moisture for weathering is widely available, from rainfall, dew, and fogs.
(c) Relative humidity is often high at night.
(d) Physical processes are probably significantly more important than elsewhere, but the role of chemical processes should not be ignored.
(e) Some of the debris and weathering features seen today may well be inherited from different climatic conditions.
(f) Despite a common emphasis upon weathering landforms, the most important role of weathering is to provide debris for fluvial and aeolian systems.

Perceptive as these observations are, like all generalizations, they are open to question and qualification. It is these qualifications that identify directions for future research.

The Role of Temperature

The Search for Extremes and a New-Found Variability

Discussions concerning the role of temperature in desert weathering have largely concentrated upon the extremes of diurnal variability – normally under summer conditions – and an apparent preoccupation with the search for maximum ground-surface temperatures (Goudie 1989, McGreevy and Smith 1982). These observations have in turn played an essential role in the design of laboratory experiments for the simulation of, for example, salt weathering. The selection of a realistic and appropriate temperature regime for an experimental simulation is, however, vital in ensuring, for example, that salt weathering and not thermoclasty is being investigated (McGreevy and Smith 1982). The use of extreme diurnal temperature regimes therefore raises a number of questions concerning the extent to which they represent weathering environments and, indeed, the realism of using one temperature regime to characterize a complex climatic zone (Jenkins and Smith 1990).

With regard to these points, 'arguably the most important change in our understanding of desert rock temperatures over the last 20 years has been the appreciation of their considerable spatial and temporal variability' (Smith et al. 2005, p.214). This point was examined initially in studies such as that by Smith (1977) in Morocco, which highlighted seasonal variations, and the fact that most deserts have winters. This study also demonstrated the importance of aspect and shading in controlling the timing, duration, and intensity of diurnal heating and cooling patterns. It has, for example, been long known that desert weathering concentrates in so-called 'shadow areas' (e.g. Evans 1970) to produce features such as cavernous hollows (tafoni), honeycombs, and pedestal rocks (Fig. 4.1). Rock surface temperatures and diurnal ranges associated with these features are reduced compared with those on surrounding exposed surfaces (Dragovich 1967, 1981, Rögner 1987, Smith and McAlister 1986, Turkington 1998), and within shaded areas moisture availability appears to be the critical control upon rock breakdown rather than extreme temperatures. In particular, a rapid, early morning temperature should inhibit absorption of any moisture deposited overnight on exposed surfaces and hence restrict the operation of processes such as hydration, solution, and hydrolysis. High rock temperatures may therefore not only be unnecessary but undesirable for many desert weathering processes.

Fig. 4.1 Cavernously weathered boulder, Skeleton Coast, Namibia showing accentuated weathering of the shadow zone, itself created by the cavernous weathering

There is also debate concerning the representativeness of many of the very high rock surface temperatures recorded, even for exposed surfaces. This was reviewed in detail by McGreevy and Smith (1982, 1983), where it was noted that many early measurements, often in excess of 70°C, were made with less sophisticated equipment than now used. Peel (1974), for example, used thermocouples without a 0°C datum, and Williams (1923) used a black bulb thermometer. With the advent of reliable thermistors and, more recently, infrared thermometers, many of the maxima recorded from a wide range of desert conditions are below 60°C (Table 4.1). Where 60°C is exceeded there may also be exceptional circumstance, such as the bitumen-coated road surfaces measured by Potocki (1978). Unfortunately, very high absolute temperatures and temperature ranges have been used in, for example, many salt weathering simulations. Data from similar tests used for assessing building-stone durability have shown that rates of breakdown are closely related to the maximum temperatures to which samples are heated. Minty (1965) showed that when samples of dolerite were cycled in sea water between 105 and 110°C the rate of disaggregation was approximately 40 times greater than when similar samples were cycled between 48.9 and 65.6°C. Other workers have also noted that the type of damage caused during salt crystallization durability tests varies with different heating cycles. Marschner (1978) found that damage was restricted to surface layers in samples heated to 60°C but that deeper cracking occurred in samples heated to 105°C. The critical nature of the 60–70°C range was also demonstrated by Yong and Wang (1980), who found that microcracking could be initiated in granite only after samples were heated above 60–70°C, and by Goudie and Viles (2000) who produced microfracturing in marble by repeatedly heating and cooling it between 25 and 80°C. Thus, as pointed out by McGreevy and Smith (1983), the use

Table 4.1 Selected rock surface temperature (°C) measurements from desert environments

Surface temperature °C	Material	Location (altitude)	Time of year	Source
74.4	Black bulb	Egypt	August	Williams (1923)
62	Limestone	Egypt	August	Sutton (1945)
49	Quartz Monzanite	California	August	Roth (1965)
72.5	Rock	Sudan	September	Cloudsley-Thompson and Chadwick (1969)
78.5	Basalt	Tibesti	August	Peel (1974)
79.3	Dark Sandstone	Tibesti	August	Peel (1974)
78.8	Light Sandstone	Tibesti	August	Peel (1974)
57	Basalt	Tibesti	March	Jäkel and Dronia (1976)
46	Granite	Tibesti	March	Jäkel and Dronia (1976)
48	Sandstone	Tibesti	March	Jäkel and Dronia (1976)
48.1	Limestone	Morocco	August	Smith (1977)
21.1	Limestone	Morocco	January	Smith (1977)
62	Asphalt	Abu Dhabi	Summer	Potocki (1978)
73	Asphalt	Abu Dhabi	Summer	Potocki (1978)
41.0	Basalt	Karakoram Mountains	July/August	Whalley et al. (1984)
33.5	Basalt	Karakoram Mountains	July/August	Whalley et al. (1984)
46.3	Desert Varnish	Karakoram Mountains	July/August	Whalley et al. (1984)
54.0	Sandstone	Karakoram Mountains	July/August	Whalley et al. (1984)
50.8	Limestone	Negev	August	Rögner (1987)
56	Rock	Algeria	June	George (1986)
52.2	Sandstone	Tenerife (2070 m)	June	Jenkins and Smith (1990)
42.5	Sandstone	Tenerife (2070 m)	January	Jenkins and Smith (1990)
35.0	Sandstone	Tenerife (950 m)	June	Jenkins and Smith (1990)
29.0	Sandstone	Tenerife (950 m)	January	Jenkins and Smith (1990)
50.0	Sandstone	Tenerife (50 m)	June	Jenkins and Smith (1990)
41.0	Sandstone	Tenerife (50 m)	January	Jenkins and Smith (1990)

of very high rock temperatures and rapid temperature change in simulation experiments of salt crystallization under 'desert conditions' may enhance breakdown by increasing crystallization pressures, ensuring complete crystallization from solution and perhaps better crystal development, and by instigating effects such as the purely thermal expansion of salts and/or thermal fracturing of the rock itself.

By subjecting rocks to unnaturally high temperatures, especially for salt weathering environments, it may not be possible to limit or identify the mechanisms responsible for any breakdown. There may also 'be the danger of introducing thermal effects which would not be encountered in nature' (McGreevy and Smith 1983, p. 300). Laboratory studies that use extreme temperature cycles to produce insolation weathering (e.g. Rodriguez-Rey and Montoto 1978) must therefore be approached with considerable caution. This precautionary approach has found its way into an increasing number of studies in recent years. and is reflected in a 'less dogmatic view toward the appropriateness of a single temperature/humidity regime and a willingness to tailor regimes to specific environmental conditions' (Smith et al. 2005 p. 215). As they point out, this is demonstrated in the paper by Goudie (1993) in which he used six different diurnal cycles ranging from the extreme 'Wady Digla' regime of Williams (1923) to the 'Negev Cycle'. The latter oscillates between 10 and 40°C and 20 and 100% relative humidity and is considered to more closely represent conditions associated with the regular overnight deposition of dew experienced in this desert (Goudie and Viles 1997).

The use of diurnally based thermal cycles in laboratory weathering simulations does, however, presuppose that this is the primary variable influencing physical breakdown. Yet, as noted by Cooke et al. (1993), there can be significant seasonal variations in temperature. Under winter conditions freezing temperatures may be frequently experienced at rock surfaces and freeze–thaw activity becomes a possibility. Both seasonal and diurnal variability are clearly systematic, but superimposed upon these are additional layers of spatial and temporal variability which, although far less predictable, may be of considerable significance. Reduced air and rock temperature maxima within caverns have already been noted, but to these must be added other aspect-related differences. This applies not only to obvious variations between north- and south-facing slopes, but also between east- and west-facing surfaces. In low latitudes, for example, any overnight moisture should be retained for longer on surfaces facing west. This moisture is then quickly driven off as rocks come out of shadow and are rapidly heated by a sun already high above the horizon (Smith 1977). This observation is corroborated by recent studies in Petra (Paradise 2002) that identified greatest weathering on east- and west-facing walls.

Added to these effects is the amelioration of rock temperature regimes resulting from maritime influences in coastal deserts, and a variety of effects associated with increasing altitude. Unfortunately, there are relatively few studies of rock temperature variations in high altitude desert areas, but those that have been made record surface maxima similar to those observed in low altitude environments (e.g. Whalley et al. 1984). These high surface temperatures frequently occur in conjunction with low air temperatures, and any interruption of incident radiation can result in a rapid drop in surface temperature. This phenomenon was illustrated by Jenkins and Smith (1990) in a study of altitudinal and seasonal variability in daytime temperatures on the island of Tenerife. By continuous measurement of surface temperatures on a standard sandstone block moved between three sites at different altitude, they showed that at an altitude of 2070 m there were numerous short-term fluctuations of 3–15 min duration related to shading by a light cloud cover and wind speed variations. During these fluctuations surface temperatures could drop up to 15°C with temperature gradients sometimes in excess of 2°C per minute. At mid-altitude (950 m) and coastal (50 m) sites, where cloud cover was greater, additional fluctuations of 1–2 h duration were noted which effectively destroyed the daytime element of any generalized diurnal temperature curve. The frequency and intensity of these short-term variations suggests that they could play an important role in disruptive processes such as granular disintegration and the formation of thin surface flakes, not least because recent studies using microthermistors have identified that the most important thermal gradients are produced within the upper few millimetres (Gómez-Heras et al. 2004) rather than the upper few centimetres of exposed stone (Peel 1974).

Previously, short-term fluctuations have either failed to register when desert temperatures were recorded at set intervals (every 15, 30 min, etc.) or have been disregarded in insolation and salt weathering simulations based upon diurnal cycles of heating/cooling and wetting/drying. It is interesting to note, however, that in one of the few sets of experiments to register detectable alteration (using reflectance and microhardness changes) resulting from thermal fatigue, the heating and cooling cycle was of only 15 min duration (Aires-Barros et al. 1975, Aires-Barros 1977, 1978). The efficacy of short-term temperature cycling in producing near-surface salt weathering has also been demonstrated experimentally by Warke and Smith (1994) and by Smith et al. (2005). In the latter experiment blocks of sandstone and limestone, some impregnated with sodium chloride, were subjected to 24,000 heating and cooling cycles of 30 min duration under dry conditions. Only those samples impregnated with salt produced any debris (through surface disaggregation), and whilst the amounts were very small they are possibly more representatative of actual rates of decay under non-accelerated conditions and could indicate the viability of differential thermal expansion as a salt weathering mechanism.

The growth of interest in the localised action of weathering processes has paralled a growing appreciation of the importance of microclimatic controls. Warke (2000), for example, has stressed the importance of the boundary-layer climate and changes at the rock/environment interface in controling weathering, including feedbacks in which intrinsic and extrinsic rock properties influence the conditions experienced. In support of this contention Viles (2005a) has demonstrated, through the use of exposure trials, the importance of microclimate in determing both patterns and rates of weathering in Namibia. This recognition has also stimulated a renewed interest in the role of fatigue failure in generating surface breakdown, in as much as rapid near-surface temperature cycling dramatically multiplies the number of possible ‘fatigue events’ to which exposed rock could be subject. Hall and his co-workers have, for example, recently revisited and championed the possible role of fatigue failure under cold arid conditions in a number of papers. They used high frequency monitoring of temperature to confirm both the existence of repeated, rapid surface and immediate sub-surface temperature change (Hall and Andre 2001), and their importance in generating stress at the microscale that could lead to fatigue failure (Hall 1999). Under extreme conditions of very low air temperatures Hall and Hall (1991) also raised the possibility of any interruption to incoming insolation creating a 'thermal shock' in which the stresses generated occur more rapidly than can be accomodated by the required internal deformation (Yatsu 1988). They similarly drew attention to the possible importance of stresses generated between adjacent crystals with different coefficients of thermal expansion and/or crystallographic allignments (Hall and Andre 2003). This theme was further developed experimentally by Gómez-Heras et al. (2006), who identified a number of papers that have calculated crack resistance energy for grain boundaries as only some 0.4 times the crack resistance energy for bulk elements (Yang et al. 1990, Sridhar et al. 1994; Zimmermann et al. 2001, Weiss et al. 2002). In their experiments they demonstrated measurable temperature differences and rates of temperature change between adjacent crystals/grains for a range of rock types when heated in a controlled ambient environment using infra-red lamps. Albedo was shown to be the main control on individual and overall maximum temperatures, but crystal size was shown to be the major factor determining temperature differences between adjacent crystals/grains. Large differences in crystal/grain size appear to magnify stresses resulting from differential thermal expansion.

Thermal Stress and Rock Fracture

In light of the previous discussion, it is clear that the patterns of thermal stress experienced by desert rock surfaces are exceedingly complex. Recognition of the range of stresses to which rocks are subjected suggests that rarely, if ever, is rock breakdown a function of a single mechanism acting in isolation. Invariably it is a product of two or more mechanisms acting together or in alternation (Jenkins and Smith 1990). The most obvious example of this is insolation weathering itself. It is difficult to envisage any situation in which desert rocks are subjected to temperature fluctuations in isolation. It has already been noted that moisture is invariably available in some form within hot deserts (see also next section). Additionally, all rocks exposed at desert surfaces have a unique stress history which will probably leave them weakened to a greater or lesser extent, and more or less susceptible to either mechanical breakdown or chemical decay.

Sources of pre-stressing in desert rocks are numerous. They include chemical alteration, either under present conditions of limited, but assured moisture availability or inherited from former periods of moister climate; dilatation acting at a range of scales; and previous exposure to processes such a salt and frost weathering. Frost weathering is particularly relevant when one considers that mountains consistently constitute the dominant terrain type in the world's major desert areas. A study by Fookes (1976) identified, for example, 43% of the Sahara desert, 39% of the Libyan Desert, 47% of Arabia, and 38.1% of the deserts in the south-western United States as mountainous. Other studies of desert terrain (e.g. Cooke et al. 1982) also identified a range of mountain–plain models as the most representative of the world's desert landscapes. Within these models most debris mantling alluvial fans and plains is derived initially from adjacent mountain catchments, where moisture may be more readily available and temperatures can frequently fall below 0°C. Debris produced under these conditions is likely to carry with it a memory of inbuilt stresses that can be exploited by other weathering processes more characteristic of hot desert environments *per se* (Warke 2007).

Potentially exploitable weaknesses in the form of microfractures can also be created within the rock mass prior to exposure. These will then be carried over into any debris derived from these rocks. The range of microfractures found within near-surface rocks has been discussed by Nur and Simmons (1970) and Simmons and Richter (1976), and was summarized by Whalley et al. (1982a) as consisting of (a) cracks at grain boundaries produced during ascent to the Earth's surface; (b) stress-induced cracks produced by the principle of non-hydrostatic stress; (c) radial and concentric cracks about grains enclosed by material with different volumetric properties; (d) tube cracks produced by magmatic fluid solution, dislocation, etching, etc.; (e) cracks induced by thermal shocks and gradients; and (f) cleavage cracks. To these crack-opening processes should be added potential weathering lines comprising, for example, certain mineralogical concentrations. The role of crack propogation in determining patterns of weathering has been demonstrated by Kane (1999 – reported in Smith et al. 2005), who also highlighted the accentuation of thermally driven expansion where rocks are bounded or 'buttressed' within a larger rock mass (Merrill 1906). She did this by studying the combined effects of confinement and compression on granite test blocks that were compressed between two metal plates before being subjected to repeated salt weathering cycles in a climatic cabinet. Results showed that the decay of unconfined test blocks progressed steadily through gradual surface disaggregation as salt crystallisation stresses are accommodated by the elastic expansion and contraction of the rock to form a network of near-surface microfractures. In contrast, blocks initially compressed at 900 kNm^2 showed little initial loss. This is because, under compression any microfractures perpendicular to the direction of compression tend to close (Batzle et al. 1980) and inhibit salt penetration. There was some slight, initial loss of material as surface parallell cracks were exploited to produce limited surface flaking, but the rate of loss was no greater than that for unconfined blocks. However, the repeated low-magnitude stressing of the salt weathering cycles eventually propagated a microfracture network into the now inelastic granite through a combination of stress concentration at crack tips (Griffith 1921) and salt crystallisation through ever-widening cracks. The interaction of adjacent fractures can effectively create a localised failure plane (Janach 1977) that is typically at an oblique angle to the direction of compression, and it was the breaking away of angular fragments defined by these planes that eventually triggered the accelerated breakdown of the confined block. To test this interpretation, Kane took one further block and doubled the compressive loading. As predicted this had the effect of delaying even further any noticeable surface loss – through more effectively closing existing fractures into the block – but once it began, breakdown was noticeably more rapid and produced larger fragments than the block under the lower compression.

Understandably, microfractures and lines of potential weathering are normally associated with igneous and metamorphic rocks. Stress relief does, however, occur within sedimentary rocks, where unequal release of confining pressures by erosion of once deeply buried rocks can lead to exfoliation and splitting (e.g. Bradley 1963). Similarly, sedimentary and metamorphic rocks can contain a wide range of different lineations associated with stylolites, cleavage and bedding planes and alignments of grain boundaries and/or voids – even in apparently unbedded sandstones (e.g. Smith and McGreevy 1988).

Weathering susceptibility is also influenced by alteration skins, rock varnishes (see Chapter 6) and case-hardened layers (e.g. Conca and Rossman 1982) which frequently cover desert rock surfaces. Indeed it is not unknown for the presence of a rock varnish, and the consequent reduction in albedo, to be used as a justification for the high rock-surface temperatures recorded in deserts (e.g. Peel 1974). The importance of albedo has been demonstrated by Carter and Viles (2004) who showed, for example, how the growth of black lichen raised both surface and subsurface temperatures. Although, presumably the growth of a light-coloured lichen on a dark substrate would have the effect of increasing albedo (Fig. 4.2, Goudie and Viles 1999). Warke et al. (1996) also showed under experimental conditions how the artificial soiling of stone surfaces markedly increased both maximum surface temperatures and rates of temperature change under simulated insolation – especially in rocks with an initially high albedo. Ironically, however, most weathering simulations that have argued for the importance of albedo, and its role in facilitating extreme temperature variations, have subsequently done little to explore its significance. This is because most have utilised climatic cabinets in which rocks with different thermal characteristics are forced through the same heating and cooling cycle.

Fig. 4.2 Light-coloured lichen covering the upper surfaces of dark-coloured boulders and increasing their albedo, Skeleton Coast, Namibia

In doing this, they negate the very differences in thermal response that are known to occur in nature and undoubtedly contribute to the differential weathering observed in deserts (Warke and Smith 1998). Changes in albedo are also frequently associated with a physical and/or chemical modification of the outer layer of the rock. It would be reasonable therefore to explore the possibility of insolation weathering occurring in the presence of such crusts. Work on other materials such as concrete has shown that durability problems frequently occur because of expansion in materials with different surface layers. This can lead to patterns of rupture redolent of those found on desert rocks (Fig. 4.3).

Albedo is not, however, the only rock property that inflences response to weathering, and other studies have identified the significance of physical and chemical variability as a catalyst to breakdown. Experiments using 'coupled' samples of two rock types exposed to simulated salt weathering conditions have, for example, shown how moisture can be retained for longer at boundaries between different stone types. This can lead ultimately to a concentration of crystallized salt, if available, and enhanced disaggregation (Haneef et al. 1990a, b). Other salt weathering simulations (McGreevy and Smith 1984) have shown how the presence of swelling clays in sandstone samples, coupled with increases in microporosity, can enhance breakdown through flaking. Exploitation of potential cracks comprising aligned grain boundaries and pores in sandstones may be a factor in salt-induced contour scaling of sandstones (Smith and McGreevy 1988). Breakdown following the cyclical wetting and drying of basalts has similarly been attributed to the swelling of hydrothermally derived smectite clays (McGreevy 1982a). Systematic studies of the physical rock properties that influence mechanical breakdown have identified water-absorption capacity, porosity and microporosity, saturation coefficient, and tensile strength as amongst the most important properties (Cooke 1979). Compressive strength, specific gravity, compactness, hardness texture, shape, anisotropy, microfractures, and cleavage can also play a role in determining breakdown (Fookes et al. 1988). Additional factors which are of specific relevance to the role of temperature in weathering are variations in specific heat capacity, and thermal conductivity. These not only influence rock surface temperatures and internal temperature gradients (Kerr et al. 1984, McGreevy 1985), but also the establishment of other internal stresses as rock constituents with varying thermal properties, including additions such as clays and salts, expand and contract differentially.

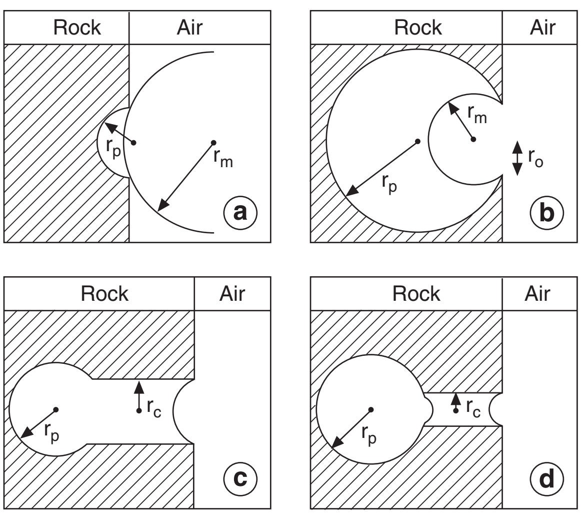

Fig. 4.3 Typical failure patterns in an elastic outer layer on a rigid base due to forces concentrated at the interface. (**a–d**) illustrate force concentrated along the interface, (**e–f**) illustrate force exerted through expansion at a point. (**a**), parallel separation of rigid plate; (**b**), peeling from base; (**c**), bending – plastic failure; (**d**), bending – elastic sudden failure; (**e**), cleavage – sudden failure; (**f–h**), punching or pop-out failure – pattern of failure depends upon brittleness of outer layer. At high brittleness numbers failure occurs suddenly (**f**). At low brittleness material behaves in a ductile fashion (**g**). At intermediate brittleness failure occurs through stable crack growth (**h**) (adapted from Bache 1985)

The presence of physical variability, pre-stressing, and chemical alteration within rocks is one possible explanation why insolation weathering continues to find support from field studies (Ollier 1963, 1984) through observations of split and cracked bouolders (Fig. 4.4). The fact that laboratory simulations using unconfined, freshly cut and polished blocks have historically discounted thermal fatigue may find an explanation in that they simply did not continue for long enough, or that the weathering that did take place was not detected (Goudie 1989). Within surface rocks a wide range of potential weaknesses thus exists that can be exploited during thermally induced surface expansion and contraction. Folk and Begle-Patton (1982) and Rice (1976) have pointed out that the early experiments of Blackwelder (1925, 1933) and Griggs (1936) could have been more successful in producing failure if they had used larger blocks that were in some way confined.

Fig. 4.4 Parallel stone cracking of boulder, Skeleton Coast, Namibia. Such features have commonly been attributed to 'insolation weathering', although such stones are, or have been partially buried and this could contribute to stress concentrations that result in the characteristic fracture pattern

Similarly, such experiments may have produced a more realistic assessment of weathering in deserts if they had used samples characteristic of those exposed by erosion and transport at desert surfaces; rather than freshly quarried blocks selected for their purity, freedom from imperfections and absence of recognizable weathering. Possibly the major difficulty with these experiments is, however, the way in which they attempted to reproduce fatigue failure by subjecting rocks to what were, in effect, repeated thermal shocks.

The Role of Moisture

Variable Availability

The role of moisture in shaping desert environments has for many years appeared anomalous. On the one hand, deserts are perceived as areas of little if any moisture, while on the other we are faced with landscapes in which many elements are of patently fluvial origin. For many years explorers denied these origins by invoking other processes including: 'sculpture due to solar heat shattering the rocks and wind removing the pulverised residue' (Peel 1975, p. 110). But as understanding of fluvial processes progressed through the early part of the last century, alternative explanations became necessary. Many fluvial features were thus interpreted either as legacies from some former climate of more assured rainfall, or as products of infrequent but intense rainfall operating over long timespans. It is only relatively recently that alternative viewpoints have been explored in which moisture is seen as being more readily available in many hot deserts than previously surmised (e.g. Peel 1975, Goudie 1997). In his paper on water action in deserts, for example, Peel referenced Slatyer and Mabbutt (1964) who found that intense, 'catastrophic' rainfall of the kind traditionally associated with deserts is much more common around the margins of deserts. In a survey of various deserts, they found that 50% of the rain that falls usually occurs in gentle showers of moderate intensity. Similarly, Peel drew on observations by Dubief (1965) who showed that mean rainfall intensities over the whole of the Sahara are little higher than those in France. Rainfall of lower intensities is less likely to generate overland flow, more likely to be absorbed by rock and soil, and will most probably play a more important role in promoting a range of weathering processes. In his review of desert weathering, Goudie (1997) tabulated data from a range of world deserts which demonstrate that even in what are perceived to be dry locations there may be an appreciable number of days in which measurable quantities of rain are received. Frequent wetting enhances the efficacy of both mechanical and chemical weathering processes.

Rainfall can also concentrate locally and might lead to enhanced scope for weathering in certain situations. Apart from obvious orographic controls, it seems that there may be a spatial organization to desert storms (e.g. Sharon 1981) related to phenomena such as Rayleigh cells within the atmosphere, although there is no significant long-term tendency for storms to occur at preferred locations. However, local rainfall is not randomly scattered and there is evidence that it can concentrate in valleys which also receive rainfall deflected away from exposed interfluves by higher winds (Sharon 1970). Windward slopes will likewise receive substantially more rainfall than adjacent leeward slopes. Yair et al. (1978) suggested that windward slopes of 20° with incident rainfall of 40° would receive almost twice as much rainfall as opposing valley sides. Similar observations were made in Sinai by Schattner (1961) who noted that granular disintegration of granites was more rapid and penetrated more deeply on rocks that were not exposed to the 'strongest and largest insolation' (p. 254). Instead, disintegration is most intensive on surfaces facing north and west into the dominant rain-bearing winds. South-facing slopes, he found, rarely exhibited intensive disintegration.

The degree to which water is concentrated into certain areas is also dependent upon surface controls, such as the extent of soil and rock cover. Where bare rock dominates, surface runoff is more frequent and extensive, and infiltration concentrates in certain parts of catchments, such as slope foot zones of colluvial accumulation (Yair and Berkowicz 1989). Because of between-catchment differences in hydrological response, Yair and Berkowicz (1989) suggested that climatic parameters such as temperature and rainfall alone provide insufficient indication of ground surface aridity.

The Importance of Direct Precipitation of Moisture

Both Peel (1975) and Goudie (1997) drew attention to the importance of moisture from fogs. Goudie tabulated the relative contributions of rainfall and fog as moisture sources at Gobabeb in Namibia over the period 1963–1984. He found that rainfall averaged 24.5 mm and fog 31.7 mm per annum. More universally appreciated is the contribution that direct precipitation in the form of dew can make to the moisture budget of desert areas (Verheye 1976). Little is known in any detail of dewfall from many of the world's deserts, and most opinions concerning its significance have been based upon pioneering work by Duvdevani (e.g. 1947, 1953) and Evenari et al. (1971) in Israel, especially at the ancient city of Avdat in the Negev Desert. On occasions they reported that dewfall can exceed annual rainfall (28.4 mm as opposed to 25.6 mm in 1962–1963), but over the four-year period 1963–66 mean dewfall at the ground surface was 33 mm from an average of 195 dew nights and, at a height of 100 cm, 28 mm from 172 dew nights. This compares with an average annual rainfall at Avdat of 83 mm for the period 1960–67 (Evenari et al. 1971, p. 30). Clearly these figures suggest that although dewfall is frequent, the amounts per night are small and are a product of low ground surface temperatures and low moisture contents of the contact zone rather than high levels of air moisture (Verheye 1976). A further demonstration of the importance of altitude and aspect in controlling deposition from dew and is to be found in the detailed studies by Kidron, again in the Negev (e.g. Kidron 2000). These show a 2–3 fold increase in average daily dew and fog amounts from sea level to 1000 m above m.s.l., and an increase in the percentage of days with heavy dew and fog (Kidron 1999), as well as demonstrating the importance of surface radiation input and temperature in controlling dew deposition (Kidron et al. 2000a). Moreover, continuous monitoring of dew deposition has shown that in sun-shaded habitats dew condensation can continue even after sunrise and may explain the higher overall dew values and its longer duration in these environments (Kidron 1999).

It is not surprising therefore, that direct precipitation has been invoked as a major moisture source and key component in the accentuated weathering associated with features such as tafoni and other cavernous hollows (e.g. Smith 1978, Rögner 1988a) and in low-lying areas (e.g. Schattner 1961). Recent studies (e.g. Goudie et al. 1997, Viles and Goudie 2007) have also demonstrated the efficacy of direct deposition from fog in promoting rapid rock breakdown in coastal deserts by salt weathering, as well as investigating its impact through linked laboratory simulations (Goudie and Parker 1998, Goudie et al. 2002a). The weathering significance of direct precipitation is further enhanced by its chemical composition. This is largely a function of the aerosols and dissolved salts that it can scavenge from the atmosphere as well as any dry fallout previously deposited on to rock surfaces and mobilized during condensation. The greatest potential for this mobilization is normally associated with polluted urban environments (e.g. Wisniewski 1982, Ashton and Sereda 1982). What observations there are from arid to semi-arid areas have, however, also shown markedly enhanced concentrations of calcium, bicarbonate, and sulphate ions in dewfall compared with rainfall (Yaalon and Ganor 1968). The most likely source for these ions is sea salts, but chemical analysis of airborne dusts blowing from desert areas suggests that they can contain elevated concentrations of numerous minerals. These include a range of clay minerals and soil nutrients (e.g. Wilke et al. 1984) as well as salts (Pye 1987, Table 6.3, p. 130).

Before ascribing too much importance to dewfall, certain other factors must be considered. First, the information we do have is drawn from meteorological observations which are not necessarily relevant to conditions experienced at rock surfaces. Secondly, we have little knowledge of the controls that humidity conditions exert upon weathering (McGreevy and Smith 1982). Is, for example, all moisture that condenses at a rock surface absorbed and, if so, what role does it play in mobilizing salts?

Meteorological observations are normally collected to reflect standard conditions, with figures invariably averaged over a given time period. The figures for dewfall at Avdat (only 80 km from the Mediterranean) are, for example, given in Evenari et al. (1971) as annual averages showing higher deposition at ground level than at 100 cm. Duvdevani (1953) showed, however, that under dry summer conditions dew deposition at various sites in Israel increases from ground level upwards – his so-called 'arid gradient'. It is only during moist winter months, when moisture can be drawn up from below, that dew deposition is greatest nearer the ground.

The mechanism of moisture deposition can itself be complicated. Under clear-sky conditions with rapid radiative heat loss at night, surface temperatures can fall below ambient air temperature. As they do so, condensation will occur at relative humidities progressively below 100%, especially if hygroscopic salts are present (Arnold 1982). Moisture flux from atmosphere to rock surface may, however, take place first in the vapour phase within pores before dew point is reached and before spontaneous condensation of liquid water.

In his review of condensation–evaporation cycles in capillary systems, Camuffo (1984) suggested that condensation begins with an adsorbed film of molecular moisture. This can form at relative humidities below 100% because 'the binding forces between the water molecules and the solid surface are larger than the binding forces between the first adsorbed layer and newly arriving vapour' (p. 152). As relative humidity in the environment increases, further molecular layers can accumulate which may eventually collapse under gravity into larger pores and take on 'bulk-liquid structure'. Such patterns of deposition have been observed within buildings (e.g. Camuffo 1983) under conditions that are otherwise effectively arid. Advantages of this mechanism are that it can promote deep moisture penetration, carrying with it gases scavenged from the atmosphere, and that moisture absorption occurs very frequently (Camuffo 1983). Such relatively small quantities of moisture may also be important in fractured rocks where, if it can form near crack tips, it may generate stresses through the formation of an electrical double layer (Ravina and Zaslavsky 1974).

As rock surface temperatures fall overnight and the relative humidity of air in contact increases, eventually a point is reached where surface condensation occurs. Whether such moisture is absorbed by the rock will depend on pore size and geometry. In the case of a small, hemispheric cavity facing the external environment (Fig. 4.5a) it will remain empty except for the monolayer of moisture until the radius of the meniscus of condensed moisture r_m equals the pore radius r_p, after which the cavity will progressively fill with moisture. This process works in reverse when relative humidity falls and the cavity progressively empties. In the case of large 'spheric pores' which open to the atmosphere via a small opening (Fig. 4.5b), the cavity remains empty until relative humidity reaches a critical level where r_m equals r_p. Thereafter the cavity will completely fill with no increase in relative humidity. If relative humidity subsequently falls, however, the pore will remain filled until a critical relative humidity is reached where r_m equals r_o, the radius of the opening, which then triggers evaporation of moisture. Where pores are connected to the atmosphere via capillaries, the pattern of filling and emptying depends upon the ratio of the pore radius r_p to that of the capillary r_c. Where $r_c > r_p/2$ (Fig. 4.5c) condensation occurs first in the pore and then in the capillary. On the other hand, where $r_c < r_p/2$ (Fig. 4.5d), condensation will occur within the capillary, trapping air within the pore. This trapped air can act as an effective block against further moisture ingress even though relative humidity may rise above 100%. Camuffo (1983) pointed out in fact that in rocks with pores showing this 'ink-bottle effect' the only way to fill inner cavities with moisture would be to submerge them in liquid water, although some moisture may be 'sucked in' if temperature falls and trapped air contracts. Conversely, if temperature were to rise, trapped air could expand and expel water without evaporation.

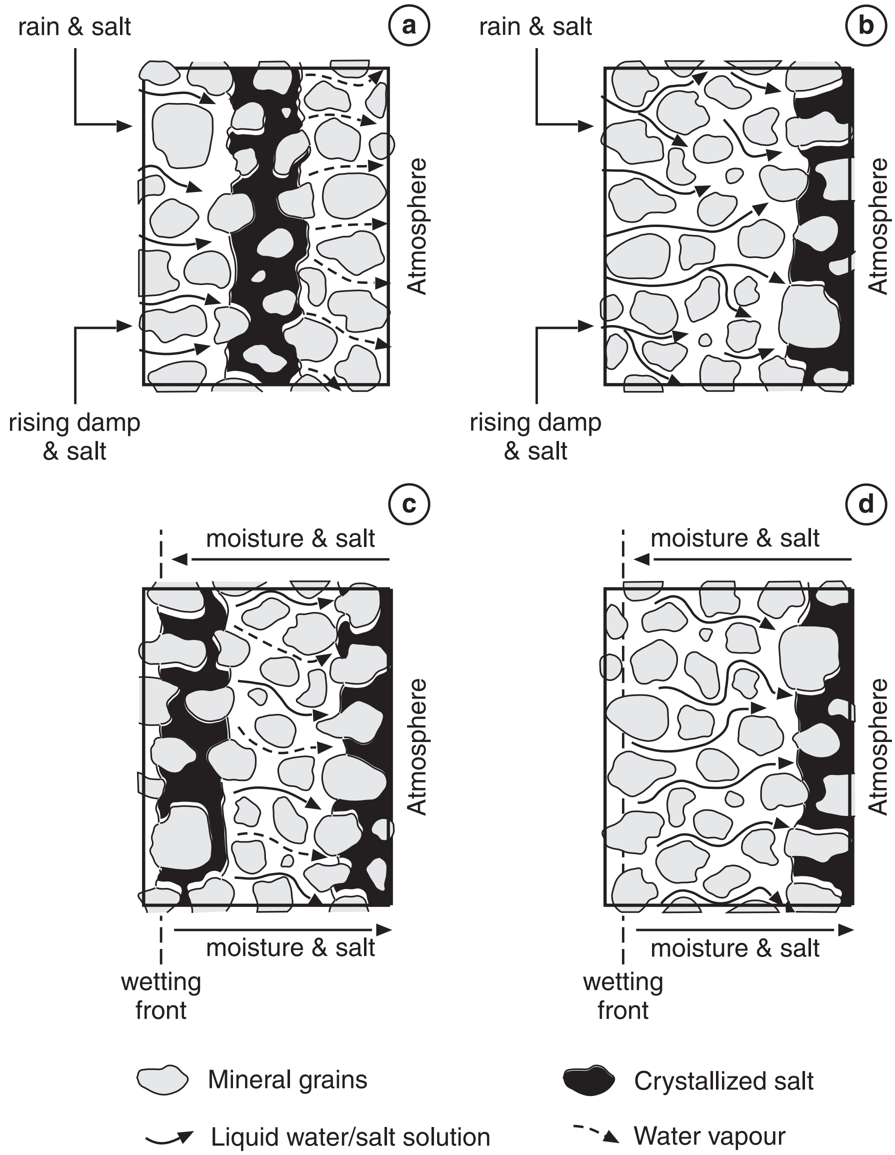

Fig. 4.5 Influences of pore and capillary characteristics upon moisture deposition and condensation at a rock–air interface. r_p = pore radius, r_m = radius of meniscus of liquid water (which increases with relative humidity such that $r_m = \infty$ at RH = 100%), r_c = capillary radius. (**a**) Hemispherical cavity facing large pore or external environment; (**b**) spheric cavity open to the environment through a small hole; (**c**) pore connected to environment by wide capillary ($r_c > r_p/2$); (**d**) pore connected to environment by narrow capillary ($r_c < r_p/2$) (adapted from Camuffo 1984)

Moisture and its Significance for Salt Weathering

Although it is not theoretically required, the norm is for drying to involve evaporation and, as Rodriguez-Navarro and Doehne (1999) have clearly shown, evaporation rate can have a major impact on the crystallization of any salt that is present in solution within the rock. This occurs in two stages (Hall et al. 1984). In the first stage, drying is strongly influenced by both temperature and wind speed. A 10°C rise in temperature will approximately double the drying rate, as will a quadrupling of wind speed. The drying rate in the first stage is also directly proportional to relative humidity. A second phase will eventually be reached when the drying rate declines substantially. During this phase drying is not significantly affected by humidity. Towards the later stages of drying in a porous body 'liquid phase capillary continuity is lost and vapour phase diffusion no doubt becomes the only transport mechanism within the pores' (Hall et al. 1984, p. 18). A similar situation can be reached earlier under conditions of either very rapid surface moisture loss or where the rate of evaporation is greater than the rate of replenishment of capillary moisture from within the rock (Amoroso and Fassina 1983, Lewin 1981). Under these conditions work on building stones has shown that the liquid-to-vapour transition will occur at some depth below the rock surface. This depth can have very significant implications if the moisture contains dissolved salts, which will begin to crystallize out of solution at this depth as a subflorescence (Fig. 4.6a). Growth of crystals at this wet–dry interface has been found experimentally by Lewin and Charola (1979) to cause the outer surface of stonework to separate or crumble (also reported in Amoroso and Fassina 1983, pp. 33–34). A similar suggestion that scaling could be associated with wetting depths was made by Dragovich (1969) working on granites in Australia. If the rate of drying is less than that of moisture replenishment from within the stone (Fig. 4.6b), then evaporation will take place from the surface and an efflorescence of crystallized salt will develop. Under these conditions observations on the behaviour of building materials suggest that disruption will be reduced compared with subsurface crystallization (Amoroso and Fassina 1983, p. 29). There is, however, some evidence from studies of sandstone used in construction that, beneath an outer surface characterised by salt-induced flaking and granular disaggregation, a zone of structural weakening can occur which is rapidly exploited once the outer surface is lost (Warke and Smith 2000). They also observed that, in a mixed salt environment salts within a rock can become segregated on the basis of their solubility, and that it is possible for complex distributions with depth to occur that represent combinations of the scenarios described in Fig. 4.6.

Under desert conditions, salt-rich moisture migrating from within a rock is most likely to derive from rising groundwater. Indeed, groundwater is now recognized by many as a major source of moisture

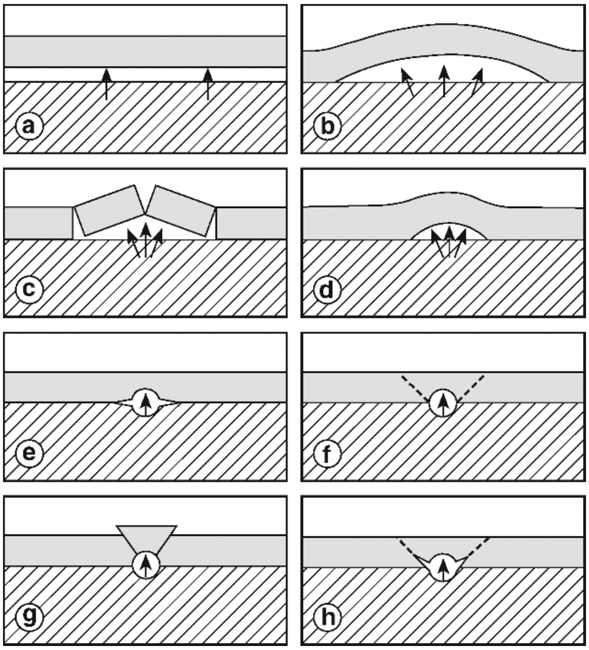

Fig. 4.6 Patterns of salt crystallization within a porous rock: (**a**) subsurface deposition where there is a high rate of surface evaporation and salt solution drawn from within the rock; (**b**) surface deposition where the rate of evaporation is less than the potential rate of outward salt solution migration; (**c**) surface and subsurface salt deposition after evaporation of surface-wetted rock (sodium sulphate, magnesium sulphate); (**d**) surface salt deposition after evaporation of surface-wetted rock (sodium chloride)

and salts in numerous low-lying and/or coastal desert areas. It has, for example, been identified as a major cause of building decay in these areas as well as being instrumental in the formation of many natural salt-weathering phenomena such as tafoni and honeycombs (see Cooke et al. 1982, Chapter 5 for a review). In urban areas from other environments, rising damp is seen to produce characteristic weathering zones above ground level (e.g. Arnold 1982), with maximum deterioration in a periodically dried zone some distance above the ground. Experimental work has also demonstrated the rapid effectiveness of salts derived from rising capillary moisture in causing rock breakdown (e.g. Goudie 1986, Benavente et al. 2001), in tests lasting as short as two weeks and with drying conditions of only 20°C and 60% relative humidity (Lewin 1981), although salt crystallization pressures are likey to be influenced by a range of variables, including pore size and salt-solution interfacial tension (La Iglesia et al. 1997). Under conditions of prolonged saturation it is also possible that salts could migrate through rock via ionic diffusion rather than in solution (Pel et al. 2003). This process could lead to complex, three dimensional distributions of salts throughout a rock mass and also facilitate chemical reactions with rock constituents (Turkington and Smith 2000). Salts deposited from groundwater are often complex mixtures and occur in high concentrations (e.g. Goudie and Cooke 1984). As a consequence, these salts can exploit a wide range of environmental conditions to exert stresses within rocks through repeated expansion and contraction. Complete pore-filling by salts could enhance the efficacy of differential thermal expansion, whereas mixtures of salts acting singly or synergistically hydrate and dehydrate or dissolve and crystallize across a wide range of temperature and humidity conditions.

Most desert environments are not, however, characterized either by an abundance of salts or of moisture. Instead, as already argued, moisture is frequently derived in small amounts from direct precipitation overnight at or near an exposed surface from where it is lost during the following day. Under these conditions large quantities of salt might even prevent ingress of small quantities of moisture by acting in effect as a ‘passive pore-filler’ (Smith and McAlister 1986). Smith and Kennedy (1999) have, for example, shown under experimental conditions how just one cycle of wetting and drying with salt solution can significantly modify subsequent patterns of infiltration and drying. Experimental work has also shown that, where sandstone blocks are sprayed with limited daily applications of weak salt solutions to one face, the subsequent salt distribution depends not just on the drying regime, but is also particularly sensitive to the detailed pore characteristics of the rock (Smith et al. 2005) and the solubility of the salt used (Smith and McGreevy 1988). So that in sandstone blocks treated with 10% solutions of sodium sulphate and magnesium sulphate, salt accumulated at or near the approximate diurnal wetting depth and just below the surface (see Fig. 4.6c) and eventually resulted in surface scaling. In contrast, in a block treated with more soluble sodium chloride, salt only accumulated in a narrow zone at and beneath the stone surface resulting in limited granular disintegration (see Fig. 4.6d). It must be remembered, however, that as salt weathering progresses it can significantly alter the porosity (Benavente et al. 1999, Nicholson 2001) and other physical characteristics of fresh rock. Any correlation between weathering performance and original conditions must therefore be interpreted with care.

Clearly, the effectiveness of any salts precipitated out of solution in causing rock breakdown is dependent on where they crystallize. It must, however, also depend on the manner of their crystallization and the stresses that they can exert from within pores and capillaries on surrounding mineral grains (Scherer 1999), which, in turn can depend on factors such as the solution supersaturation ratio (Rodriguez-Navarro and Doehne 1999, Flatt 2002). The process of crystal growth out of solution itself is considered by many to be the most important mechanism by which stresses are exerted and rock fabric disrupted (Cooke et al. 1993). A situation is frequently envisaged where crystals bridge across a pore and continue to grow against confining pressures provided there is a film of solution at the salt–rock interface (Correns 1949, La Rijniers et al. 2005). Although, as Scherer (1999) has pointed out, the stress generated within a single pore is insufficient to cause failure as it acts on too small a volume. Instead, salts are required to propogate throughout a a zone within the porous medium so that sufficient stress is generated to cause failure by interacting with the larger flaws that ultimately control material strength. Observations of crystallized salts within pores from a range of environments suggest that pore-filling begins with a layer of salts crystallizing around the boundary of the pore. Bernardi et al. (1985), for example, identify authigenic calcite and gypsum growing around ooliths in limestones from the Ducal Palace in Urbino. These salts come into contact first either in capillaries or the narrow necks connecting pores. By concentrating stresses here they could cause disruption without the necessity either for pore-bridging or pore-filling. Indeed, although there are now numerous descriptions of salt efflorescences from many environments and of subfluorescences exposed when blisters are removed, there are few, if any, examples of pore-bridging within sound rock. Long, acicular gypsum crystals have been frequently described growing out of carbonaceous nucleii (e.g. Del Monte and Sabbioni 1984, Del Monte et al. 1984, Del Monte and Vittori 1985). But these are observed growing on the surface of polluted stonework or subsequently under moist laboratory conditions from sources such as fly-ash. More detail on the mechanics of salt weathering can be found in the excellent reviews by Goudie and Viles (1997) and Doehne (2002).

As an alternative to partial pore-filling by clearly defined salt crystals, relatively rapid evaporation under laboratory conditions can fill pores with microcrystalline salts (e.g. Smith and McGreevy 1988). Complete pore-filling of this kind may cause breakdown by crystallization if salts continue to migrate towards the pore, but will also result in stresses due to differential thermal expansion and/or hydration if environmental conditions and salt types permit it (Stambolov 1976, Sperling and Cooke 1980a, b).

The areal distribution of salt accumulation and ultimately salt-induced decay is very dependent upon the wetting and drying characteristics of stone surfaces. Varying pore size and shape characteristics will, as already discussed, lead to differing patterns of moisture condensation, absorption, and loss as well as affecting susceptibility to various salt weathering mechanisms. Variability of this kind has often been used to explain the location of so-called sidewall tafoni (Fig. 4.7) on cliff faces (e.g. Bryan 1925, Mustoe 1983, Turkington 1998). In contrast, basal tafoni are normally ascribed to greater moisture availability in cliff foot zones from either groundwater seepage or enhanced direct precipitation near ground level (Dragovich 1967, Smith 1978, Migon et al. 2005). Once created, any hollows protected from rainfall become sites of preferential salt accumulation (Fig. 4.8) and, it has been proposed, sites of accelerated weathering related to higher average levels of relative humidity and protection from direct insolation (see previous section). This proposition has, however, always required that at some stage moisture is evaporated from within the rock. This may be achieved through insolation reaching the interior of the hollow later in the day or through a general rise in air temperatures and fall in relative humidity. Work on building stone decay has suggested that drying might be accelerated in honeycomb features by turbulent airflow (Torraca 1981, Rodriguez-Navarro et al. 1999), although Huinink et al. (2004) are unconvinced that the same effect is experienced in larger tafoni. Simulations of the environmental conditions experienced in the caverns created as building blocks retreat (Turkington et al. 2002), combined with field and laboratory analyses of the weathered blocks themselves, suggest however, that within caverns environmental cycles tend to be restricted to the outer few millimetres of the stone surface at a scale that is commensurate with the multiple flakes that are characteristic of many cavern interiors (Smith et al. 2002a).

Fig. 4.8 Salt concentration at the back of a tafoni, Twyffelfontein, namibia

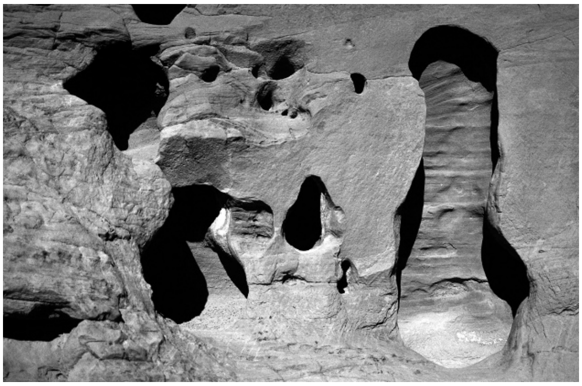

Fig. 4.7 Sidewall tafoni, Valley of Fire, Nevada

It would seem, therefore, that moisture is available for weathering in deserts from a variety of sources. Often this moisture is spatially concentrated, but although it may occur frequently (e.g. as dewfall), it is normally available in limited quantities. The concentration of weathering in shadow zones where moisture is retained, rather than in areas experiencing the highest rock surface temperatures suggests that moisture availability is a critical control on processes such as salt weathering in deserts.

Chemical and Biochemical Weathering

Much debate about chemical weathering in deserts concerns the question of how much can be attributed to present-day climatic conditions and how much is inherited from conditions of greater moisture availability. The best-documented use of the latter argument is the work of climatic geomorphologists such as Büdel, who proposed a climatogenetic interpretation of desert landscapes which combined them with tropical savanna and rainforest areas as a zone of Tropical Planation (e.g. Büdel 1982, pp. 120–85). Of paramount importance in this planation process is prior deep weathering, which he envisaged as having occurred at some stage across the whole zone. According to Büdel, most of the deeply weathered material has been removed from present-day desert areas under a geomorphic regime in which long-term fluvial erosion and aeolian deflation exceeds debris supply through weathering. Because of this, much of the evidence of former deep weathering consists not of weathered profiles but of the gross similarity between, for example, the inselberg and plain landscape of tropical West Africa and that of the Sahara to the north. Morphological evidence of this kind must be approached with caution, especially in light of the great antiquity of the land surfaces in these areas and the likelihood of morphological convergence towards landforms and landscapes conditioned by geological controls. In addition, there are few studies and a relatively poor understanding of what constitutes chemical weathering under hot desert conditions and the extent to which chemical processes are currently active in desert environments.

Contemporary Chemical Weathering

The clearest evidence that chemical weathering occurs in hot deserts is the already stated argument that moisture is readily available. This is supported by the widely held belief that salt weathering constitutes the principal mode of rock breakdown (e.g. Cooke and Smalley 1968, Evans 1970, Goudie 1985), which presupposes a major role for moisture absorption by rocks. Indeed, it is arguable whether certain salt weathering mechanisms are not themselves partly chemical in nature. Thus Winkler (1987a), writing about the distinction between chemical and physical weathering, makes the point that: 'on both natural outcrops and urban buildings the process (weathering), turns out to be more complex, and not readily separable into chemical and physical components. Where can we draw the line between crystallization and hydration pressure, as these are physiochemical processes? There is also evidence of chemical reaction between salts in the masonry and the masonry substance (Arnold 1981) leading to new compounds. Today we still do not understand the physical behaviour and chemical effect of water trapped in capillaries under pressure in most types of stone' (p. 85).

Despite the presumption that desert environments are not conducive to chemical weathering, we can no longer assume that where deeply weathered rock is found it is inherited from moister conditions. In a study of saprolites on igneous rocks in the Negev, Singer (1984) has shown that the climate during their formation has never been moister than semi-arid. Nonetheless, clay contents of 45%, 53%, 11.3%, and 37% were found on andesitic basalt, Na–basanite, microgranite, and microdiorite. Moreover, a wide range of alteration products was identified (Table 4.2), which runs counter to traditional associations of arid conditions with limited alteration of silicate rocks only to smectite clays except under exceptional local circumstances (Barshad 1966, Tardy et al. 1973, Singer 1980). Crystallized iron oxides were also found in the clay mineral fraction of saprolites from basic rocks and argillation seemed to affect mafic minerals preferentially. Significantly, the alteration that Singer observed was not associated with, and thus not dependent upon, significant leaching of the weathered profiles. The presence of supergene mineralisation in the very ancient Atacama Desert has also been interpreted by Clarke (2006) as indicative of long-term weathering under arid conditions, without the possibility of previous periods of more humid climate.

The formation of iron oxides and their movement through rocks by capillary action has also been reported from other deserts. Selby (1977) described red stainings, principally of limonite, but also possibly

Table 4.2 Alteration products of deep weathering in the Negev Desert (Singer 1984)

Parent material	Alteration product
Na–basanite	Dioctahedral smectite
Andesite basalt	Hydroxy: interlayered, high charge smectite
Microgranite	Principally smectite and chlorite with moderate amounts of kaolinite
Microdiorite	Mica: after sericitization of feldspars and osmectite

of goethite, from granites in Namibia associated with scaling features. The staining could sometimes be seen emanating from biotite crystals and a few feldspar crystals showed evidence of moderate kaolinization and sericitization. This, together with some possible chloride formation, was taken to indicate slight chemical weathering. Selby also noted some strong leaching and decomposition of biotites in granites experiencing granular disintegration in the central Sahara. Conca and Rossman (1985) similarly noted haematite coatings on tonalite boulders in California derived from the leaching of biotite, while Osborn and Duford (1981) described streaks of iron oxide running down the sides of sandstone inselbergs in south-western Jordan. They also noted iron-rich weathering rinds up to 1 cm thick on many boulders. The rinds were found to reduce the degree of disintegration when boulders were dropped from cliffs and to offer a degree of protection to the underlying sandstone. A more recent study of these coatings by Goudie et al. (2002b) attributed the case hardening of sections of cliff face to 'iron and other solutions' precipitated by rainwash. Interestingly, they also identified an algal layer just below the case-hardened surfaces. Subsequent detailed analysis of the crusts by Viles and Goudies (2004) found iron, manganese and calcium to be common constituents of the cements within the crusts and and associated them with the presence of cryptoendolithic biofilms containing cyanobacteria and fungi. They are, however, careful to acknowledge that whilst these microorganisms may assist in the formation of the crusts, they are not a necessary component of all such crusts. Similar protection has been assumed by those who see tafoni and other cavernous features as forming through the breaching of weathering rinds. Breaching is followed by excavation of rock behind the rind that was preferentially weakened during rind formation by weathering and outward migration of certain constituents (e.g. Wilhelmy 1964, Winkler 1979). In addition to the surface precipitation of iron compounds, case hardening is also produced by the outward migration in solution and precipitation of calcium carbonate, which may be derived from within the rock or in non-calcareous rocks from lime-rich ground water (e.g. Bryan 1926). Calcite deposition can alternatively weaken rock structure when, for example, it crystallizes within the pores of a quartz sandstone (Laity 1983). The constituents of case hardened coatings can also originate as external additions of wind-blown dusts which are subsequently mobilized by surface moisture. Conca and Rossman (1982) have described thin case-hardened layers on quartz sandstone in Nevada which contain calcite, hydrated calcium barite, and raised levels of kaolinite of possibly aeolian origin.

Behind these crusts the interior of boulders may be weakened by a process termed core softening by Conca and Rossman (1985). They describe how tonalite boulders in Baja California have undergone preferential decay beneath iron-rich surface layers. This decay consists of kaolinization of feldspars (normally along cleavage planes and grain boundaries), a loss of iron from biotite, and the formation of authigenic silica and traces of calcite and gypsum. They proposed that the weathering is accomplished by migrating capillary moisture. This may have an elevated chemical reactivity due to increased dissolved CO_2 and organic acids if it had previously flowed through the upper soil horizons within which the boulders are partly buried. Weakening of rock structure may similarly be accomplished in quartz sandstone by selective silica dissolution under saline conditions, revealed as solutional etching on grain surfaces (e.g. Young 1987). In light of these observations, Young proposed that this dissolution can lead to the development of cavernous hollows. Mustoe (1983) also made the point that such cavities can be produced by chemical weathering in addition to the normally assumed salt weathering. This weathering is possible given that the cycles of hydration and dehydration required for most salt weathering will also encourage a range of chemical reactions leading, for example, to the decay of feldspars. Silica solubility is also enhanced under saline conditions, and Winkler (1987b) described silica coatings on tuffs in Texas, which he attributed to the 'high solubility of silica in a dry and hot west Texas... enhanced by probably alkalic waters' (p. 975).

The increase in solubility of both quartz and amorphous silica with increasing pH is well established (Goudie 1997) and is reflected in the formation of extensive silicate deposits within the evaporite sequences of alkaline lakes (e.g. Eugster 1967, 1969, Eugster and Jones 1968). The corollary of this increase in solubility is that quartz and silicate rocks will be corroded if they come into contact with alkaline lake water. This is illustrated by Butler and Mount (1986) who described corroded cobbles along former lacustrine strandlines in Death Valley. This corrosion manifests itself in the form of pits and honeycombs, and it derives principally from the dissolution of quartz, feldspars, and phyllosilicates in a variety of rock types. Corrosion occurs preferentially along grain boundaries, laminations, cross beds, and foliations and can enlarge vesicles in extrusive igneous rocks. This process can lead to development of a secondary porosity, which increases moisture penetration and encourages further corrosion but can also increase susceptibility to salt weathering if the water is high in total dissolved solids.

Arid Karst and the Role of Biological Activity

In their study of alkaline corrosion, Butler and Mount (1986) observed little corrosion of limestone or dolomite. Again, it would seem that environmental factors, such as high temperatures and paucity of soil and vegetation giving low dissolved CO_2 contents in desert moisture, generally combine to produce little karstic weathering in deserts. Thus the traditional view is that espoused by Jennings (1983) who concluded that carbonate karst decreases with precipitation. This does not mean that karstic phenomena are absent; simply that they are invariably explained away as relics of former climates (Table 4.3). As such they are described as being destroyed by arid processes including aeolian abrasion (Krason 1961), mechanical weathering (Smith 1978), and fluvial incision (Marker 1972). Some features do, however, manage to retain a degree of development under arid conditions. Thus, Hunt et al. (1985) reported active karst on gypsum and anhydrite deposits of the Tripolitanian pre-desert in Libya, and Castellani and Dragoni (1987) described active erosion of solution pipes in south-eastern Morocco. The moisture for the latter erosion condenses as dew and is blown along the limestone surface. The ability of rivulets of coalesced dew to effect solution is further evidenced by studies of building stone, where they can cause serious damage (e.g. Camuffo et al. 1986). It is not likely that dew could initiate features such as extensive solution piping. Instead, the pipes probably formed under conditions of more assured rainfall, but continue to develop slowly at the present time.

Table 4.3 Karstic features described from desert environments

Principal features	Location	Source
Dayas (dolines)	Algeria, Morocco	Conrad (1959, 1969), Clark et al. (1974), Castellani and Dragoni (1977)
Caves	Algeria	Conrad et al. (1967)
Solution hollows	Algeria	Quinif (1983)
Caves/hollows/karren	Morocco	Smith (1987)
Cone karst	Egypt	El Aref et al. (1986, 1987)
Caves	Australia	Grodzicki (1985)
Caves	Morocco	Castellani and Dragoni (1987)

Dewfall does, however, appear to play significant direct and indirect roles in the formation of microkarstic features found in deserts (Smith 1988). Wind-driven dew can form microrills or rillensteine (Laudermilk and Woodford 1932) and plays an important part in the solution facetting of limestone pebbles (Bryan 1929) and micropitting of pebbles buried within the vesicular layer beneath stone pavements (Smith 1988). Other active karst described from arid areas have included rillenkarren (Sweeting and Lancaster 1982) and solution pans (Smith 1987).

In addition to highlighting the role of direct precipitation, karst studies have stressed the predominance of features such as pans and pits (Fig. 4.9) which trap moisture (Smith et al. 2000) and the role of organisms in promoting solution (Smith 1988). The general role of microorganisms in weathering and particularly the erosion of limestone has been reviewed by Viles (1988, 1995) Naylor et al. (2002) and their specific role in desert weathering by Cooke et al. (1993). In deserts emphasis has been placed on endolithic cyanobacteria and blue-green algae in, for example, the formation of weathering pits (Danin et al. 1982, Danin and Garty 1983); on endolithic algae in the 'solubilizing' of minerals, including silicates (Friedman 1971); and on lichen in exfoliation (e.g. Friedman 1982). Similar observations were made by Smith et al. (2000) on limstones in southern Tunisa, who identified the active formation of a microkarst associated with endolithic and epilithic algae responsible for algal boring, plucking and etching (Fig. 4.10), and stressed the importance of microclimatic controls on the weathering environment and the interaction of biological processes with active salt weathering. The effectiveness of this synergistic

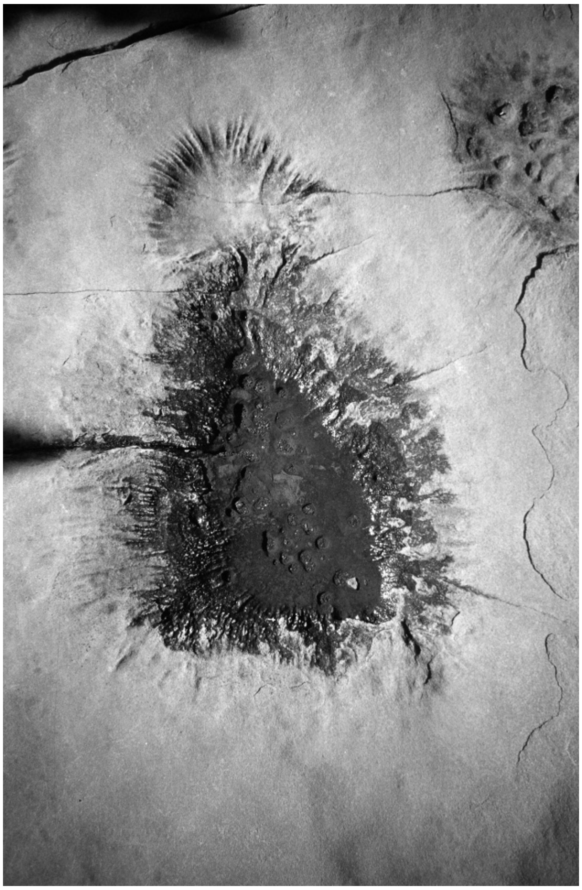

Fig. 4.9 Solution pan, Grand Canyon, characteristic of the active karst and pseudokarst of arid regions that persist because of their ability to trap and concentrate the limited moisture that is available

association has been identified elsewhere where it has contributed to accelerated urban stone decay (Papida et al. 2000). Smith et al. (2000) also drew attention to the occurrence of patches of iron amd manganese

Fig. 4.10 Small scale solution litting of a limestone in southern Tunisia. Each pit supports an epilithic algal community

rock coatings on the limestones. These have been studied in greater detail by Drake et al. (1993), are only found on stable surfaces and are considered to be of possible biogenic origin. Fungi and bacteria occur on both iron and manganese crusts, but rare microcolonial fungi are only found on the latter. On more unstable substrates, microbes were generally found living within, rather than on the surface of the rock. There are also observations which suggest that biological agents can, at least in the short-term, help to stabilize rock surfaces in ways other than facilitating the precipitation of mineralogic crusts (Carter and Viles 2005). For example, Kurtz and Netoff (2001) working in Utah proposed that sandstone surfaces could be temporarily stabilized through the deposition of extracellular polymeric substances and filamentous growths that retarded erosion by wind and water. There is also evidence that, through the retention of moisture, epilithic lichen can reduce the surface thermal stress experienced by rocks such as limestone (Carter and Viles 2003).

The widespread occurrence of microorganisms, especially in deserts such as the Negev, is generally attributed to the high frequency and regularity of dewfalls (Kappen et al. 1979) as a source of essential moisture. Because of the crucial role of dew, especially surface moisture duration (Kidron et al. 2000b), in determining the distribution of microbiotic crusts, the distributions of these organisms show a marked variability with aspect. Lichen, for example, were found by Kappen et al. (1980) to be much more productive on shaded north-facing slopes in the Negev than on drier east-facing slopes exposed to strong insolation from the rising sun. Danin and Garty (1983) similarly found that north-facing slopes were characterized by epilithic lichen, whereas south-facing slopes were inhabited by endolithic cyanobacteria which are more capable of surviving drought conditions (see also Viles 1988). These variations translate through to biological weathering effects, and Stretch and Viles (2002) showed on Lanzarote that lichen grew preferentially on N and NE facing surfaces and that this resulted in a rate of weathering rind formation that was approximately sixteen times that on lichen-free lavas. Such aspect-related variations can also operate at the micro-scale and, for example, Kidron (2002) noted that the the tops of cobbles in the Negev Desert could receive approximately twice as much dew as was precipitated on the sides. Care must, however, be taken in attributing all micro-solutional phenomena to present-day biotic weathering. Danin et al. (1982) found, for example, that although lichen and blue-green algae can produce characteristic jigsaw-like patterns of microgrooves at the boundaries between lichen colonies in Israel, they are only active in Mediterranean areas. Where these grooves occur in arid areas they are interpreted as indicating more humid conditions in the past.

From this very brief overview of chemical weathering, it would seem that there are very few chemical processes that cease completely or are totally excluded under desert conditions. Their rates of action, mode of operation, and frequency of occurrence do change significantly. Limestone solution, for example, does not cease, but no longer produces clearly defined macro- and mesoscale karstic phenomena. Instead it is responsible for a complex variety of microscale features and, it would seem, an undifferentiated slow loss of limestone in solution (Smith 1988).

Weathering of Debris in Deserts

As noted in Cooke et al. (1993), weathering of debris in deserts has tended to be neglected in favour of investigations into rock masses or, at best, boulders of a size that can support features such as cavernous hollows. Yet weathering does not cease with the production of coarse debris, but continues as long as the debris remains at or near the desert surface (Smith 1988). Indeed, weathering of debris exercises an important control upon the overall rate of erosion of many landscapes mantled and protected by stone pavements. Even when such debris is fractured or disaggregated, further comminution may be required before it can be removed by either low energy fluvial erosion or aeolian deflation. A crucial stage in this process is the reduction of sand-sized or coarser debris into silt-sized and finer fragments that are the principal constituents of many wind-borne dusts that blow out of the world's deserts.

Coarse Debris

Clearly, exceptions to the above generalization exist and many experimental studies of salt weathering have used 'debris-sized' blocks of stone (e.g. Goudie et al. 1970, Goudie 1974, Fahey 1986, Cooke 1979). Unfortunately, these studies have rarely been used to explain debris behaviour *per se*. Instead, results have invariably been applied to generalized questions such as relative stone durability, the effectiveness of various salts and salt combinations in causing weight loss, or the nature of weathering mechanisms. Little attention has been paid to patterns of decay or, with some exceptions (e.g. Smith and McGreevy 1983, 1988), to the reproduction of weathering features similar to those observed under field conditions. The effectiveness of salt weathering in causing debris breakdown has been further demonstrated through studies of natural debris (e.g. Goudie and Day 1980) and cut blocks of stone (e.g. Goudie and Cooke 1984, Viles and Goudie 2006, Warke 2007) placed in salt-rich desert environments. Progress has been made despite the problems envisaged earlier in causing mechanical breakdown in small, unconstrained blocks which can themselves expand and contract in response to temperature variations.

Salt weathering is not, however, the only weathering process to operate in deserts, and rock debris should be subject to the full range of weathering processes active in these environments. Such debris could originate in a high mountain environment where it is released from the rock mass by frost-related processes exploiting intrinsic microfractures. It might then experience abrasion and high energy collisions during subsequent fluvial and debris-flow transport and finally be exposed to extreme temperature variations and partial or complete burial in a salt-rich environment such as an alluvial fan. As a result, the debris found mantling many desert surfaces exhibits a wide range of different stress histories, which in turn may explain the variety of weathering patterns observed and the comparative ease with which some debris is seen to breakdown in response, for example, to temperature cycling. The stress history described represents a form of climatic change brought about by spatial relocation. While there has been some work on the combined effects of different weathering regimes on rock breakdown (e.g. frost and salt weathering, Williams and Robinson 1981, McGreevy 1982b, Robinson and Jerwood 1987, Jerwood et al. 1990), only recently has there been any systematic examination of the sequential exposure of rocks to different weathering mechanisms. This is despite the recognition that many weathering mechanisms, such as salt crystallization and wedging by expanding lattice clays, are exploitative in character and only operate in many originally non-porous rocks once fractures or microcracks have been initiated by other mechanisms (e.g. Smith and Magee 1990). Thus, Warke et al. (2006) were able to show how sandstone blocks subjected to alternating combinations of simulated frost and salt weathering underwent a period of internal structural weakening before rapid surface loss was triggerred. The length of this preparatory stage was dependent on the precise combination of frost and salt events and also on the texture/porosity of the sandstone.

A wide range of weathering patterns has been observed on desert debris (see Table 4.4), but it is possible to identify a number of underlying characteristics. First, enhanced moisture and salt availability (via upward migration in low-lying alluvial and lacustrine environments) can produce rapid and thorough disintegration. Secondly, processes such as insolation and salt weathering are enhanced where debris is partially buried within finer material. Partial burial can produce accentuated internal temperature gradients around the exposure and burial boundary, constraint upon the expansion of the buried portion, and possibly enhanced chemical weathering within the subsurface environment. Thirdly, many recent studies have stressed the role of biological agencies in facilitating the solution of limestone and the fixation of iron and manganese in desert varnishes. Finally, contrary to normal expectations, weathering need not be dominated by physical breakdown. On limestone debris, for example, it was noted in the previous section that a range of microsolutional features can be observed which are the most likely representatives of a truly arid karst.

Table 4.4 Weathering features associated with debris on stone pavements

Weathering phenomena	Sources
Parallel stone cracking	Yaalon (1970), Cooke (1970)
Desert varnish	Verheye (1986), Bowman (1982), Dorn (1983)
Solution pitting	Amit and Gerson (1986), Rögner (1988a)
Vesicular weathering	Smith (1988)
Solution facetting	Bryan (1929)
Solution pinnacles	Joly (1962)
Microrills	Laudermilk and Woodford (1932)
Case hardening	Smith (1988)
Shattered rocks	Ollier (1963), Rögner (1988b)
Polygonal cracking	Soleilhavoup (1977), Bertouille et al. (1979)
Dirt cracking	Ollier (1965)
Granular disintegration	Cooke (1970), Goudie and Day (1980)
Honeycombing	Butler and Mount (1986)
Split/cleaved boulders	Whitaker (1974), Ollier (1984)
Weathering rinds	Osborn and Duford (1981)

Fine Material and the Question of Desert Loess

Comminution of boulder and cobble-sized debris on desert pavements produces a variety of material ranging from rock fragments to individual mineral grains and, less frequently, fragmented grains that can progressively form a complete surface cover (Al-Farraj and Harvey 2000). This material is in turn subject to further weathering and/or attrition which can, under suitable conditions, reduce material to fine sand or silt. This is especially the case in low-lying areas such as the Qattara Depression (Aref et al. 2002) and coastal pans (Viles and Goudie 2007) where salts are available in high concentrations and from where large quantities of the ‘dust’ produced can potentially be removed by deflation. Much of this material can be transported great distances with wide-ranging consequences (see, for example, Goudie and Middleton 2001 and Tafuro et al. 2006), but of particular interest within this material are the silt-sized particles that ultimately make up the loess (e.g. Yaalon and Dan 1974, Caudé-Gaussen et al. 1982, Caudé-Gaussen 1984) and loess-like deposits of desert marginal areas (e.g. Smith and Whalley 1981, McTainsh 1984). It is generally considered that the transformation of quartz sand into silt requires either the precise application of disruptive stresses or the general application of very large stresses. Thus glacial grinding has been invoked to explain the loess deposits of high latitudes. Indeed, the apparent pre-eminence of glacial theories for loess-silt formation has meant that, even where silt is blown out of deserts such as the Gobi, it has been proposed that it must have originated initially from surrounding mountains (e.g. Smalley and Krinsley 1978).

There is, however, increasing evidence that a number of weathering and other processes that operate in deserts are capable of producing quartz silt of loess size

Table 4.5 Possible origins of loess-sized quartz silt in desert environments

Process	Sources
Salt weathering of loose granular material	Goudie et al. (1979), Pye and Sperling (1983), Peterson (1980)
Salt weathering of sandstone	Goudie (1986), Smith et al. (1987)
Frost weathering	Moss et al. (1981), Lautridou and Ozouf (1982)
Biotite weathering in granitic profiles	Pye (1985)
Inherited tropical deep weathering	Boulet (1973), Leprun (1979), Nahon and Trompette (1982)
Inherited silica dissolution of dune sands	Little et al. (1978), Pye (1983)
Inherited weathering from temperate soils	Pye (1987)
Silica dissolution in saline environments	Young (1987), Magee et al. (1988)
Inherited glacial grinding	Smalley and Vita-Finzi (1968), Boulton (1978)
Aeolian abrasion	Whalley et al. (1982b, 1988)
Fluviatile abrasion	Moss (1972), Moss et al. (1973)
Fracturing and/or dissolution in duricrusts	Nahon and Trompette (1982)
Microdilation in granitic profiles	Nur and Simmons (1970), Moss and Green (1975), White (1976), Power et al. (1990)

(20–60 μm). These processes are reviewed in Table 4.5 and also by Pye (1987), McTainsh (1987), and Smith et al. (2002b). Some of the proposed processes, including salt weathering and aeolian and fluviatile abrasion, clearly act under hot desert conditions. Others, such as frost shattering, are most likely to be effective in either desert marginal areas or 'climatic islands' within deserts induced by altitude. A further and important group comprises processes which operate under moist tropical or temperate conditions, the effects of which may represent climatic inheritance when encountered in present-day deserts. It is a characteristic of many of the mechanisms listed in Table 4.5 that they operate through exploiting flaws in sand-sized and larger grains. These flaws could be inherited microfractures (Moss and Green 1975), cleavage planes (Wilson 1979), or dislocations formed by tectonic deformation (White 1976). Such flaws are not always necessary, and within sandstones fractures could be initiated by compressive loading of sand grains when salts expand within adjacent pores (Smith et al. 1987). It is difficult to envisage, however, how other processes such as salt weathering of loose grains could be effective in the absence of pre-existing flaws or microfractures.

In keeping with larger debris, therefore, sand-sized and finer material exposed at desert surfaces is invariably the product of a complex stress history. This history may represent sequential exposure to different stress mechanisms through climatic fluctuations, long-distance transport, and/or release from parent rock. Grains may additionally be subject to processes such as frost and salt weathering acting alternatively or in conjunction. The end products of exposure to these stresses are either comparatively flawless core grains of the type found in many dune sands or silt-sized fragments of a calibre prone to deflation and removal from the desert environment.

Conclusions and Future Directions

There are many aspects of desert weathering that require further investigation and refinement. In particular, it is clear that the weathering environments of hot deserts are numerous and varied. We need much more information on the range of environmental conditions experienced before rock responses can be fully understood. Moreover, the distinction must be made between conditions as measured for meteorological purposes and the temperature and moisture regimes experienced at the rock–atmosphere interface. The overall significance of weathering achieved within these environments cannot be appreciated until (a) we improve the precision with which weathering damage is recognized and assessed, and (b) we begin to investigate the overall contribution of weathering not only in the development of individual landforms but also in the long-term evolution of desert landscapes as a whole.

Damage Recognition and Assessment

A criticism levelled at many early simulation studies of rock weathering is that they dismissed weathering mechanisms such as insolation-related fatigue failure, simply because they employed techniques of damage assessment insufficiently sensitive to register changes that might have been expected over the duration of the experiment (see Rice 1976). Similarly, many weathering changes are threshold phenomena, in that specimens may exhibit little surface change while experiencing internal changes that can lead to sudden and catastrophic damage as strength thresholds are breached and, for example, large contour scales break away. As a result, damage assessment by weight loss alone can lead to misleading estimates of relative durability if simulation experiments are of limited duration. Alternatively, there is a temptation to increase the aggressivity of such tests to the point that, while measurable damage is achieved, conditions no longer relate to those encountered in nature. There is a need, therefore, for more precise and consistent methods of assessing surface and internal changes in rock properties consequent upon weathering. Ideally such methods should be applicable under both field and laboratory conditions to permit work to be integrated between the two situations.

Considerable progress has been made in recent years in terms of damage assessment, including the development of a rigorous terminology for defining weathering damage. A major driving force in providing this sytematic information has been 'crossover' research into the decay and conservation of stone monuments within arid enviroments. The leading proponents of this have been Fitzner and co-workers at Aachen who have devised a detailed and extremely comprehensive classification of weathering forms (e.g. Fitzner et al. 1992, 1995). They have in turn applied this to historic monuments in Cairo (Fitzner et al. 2002), in which they tailored an interpretation of weathering form and intensity to the petrological characteristics of the limestones used as well as environmental conditions. Other applications of this approach have included the deterioration of sanstone monuments in Petra (Heinrichs and Fitzner 2000), the weathering of Pharaonic monuments in Luxor (Fitzner et al. 2003a,b) and the use of damage assessment linked to diagnosis to identify a weathering progression at Petra (Heinrichs 2002). Petra has also been used by other researchers as a 'natural environmental laboratory' for weathering studies (e.g. Paradise 2005). The regularity of construction makes such cities, and individual buildings, ideal for the isolation of controls such as aspect on weathering processes (e.g. Paradise 2002). Their known age greatly facilitates studies of weathering rates under different environmental conditions and for different stone types (Paradise 1998) as well as the identification of decay thresholds (Paradise 1995). Elsewhere, weathering rates and surface change over time have been established using a variety of strategies, including the analyses of previous conservation interventions and records (Selwitz 1990, Wüst 2000, Wüst and Mclane 2000), the protrusion of lead lettering (Klein 1984) and a movable frame for plotting surface contours (Sancho et al. 2003). Typically, such studies have also been associated with a forensic analysis of the immediate and underlying causes of decay as the first step towards the selection of appropriate conservation strategies. This includes not just external influences such as the origins and impact of soluble salts (e.g. Wüst and Mclane 2000), but also an emphasis on the importance of geological controls (e.g. Rodriguez-Navarro et al. 1997) that is sometimes absent from or underplayed in geomorphological studies.

The Need for Integration

There has long been a tendency in studies of desert weathering to isolate individual weathering mechanisms and to seek an understanding of weathering phenomena through segregation rather than integration. This tendency has, as indicated in the introduction, reflected a general trend in geomorphological investigations which have become increasingly process orientated and reductionist in character. Like these wider studies, the understanding of desert weathering mechanisms is not an end in itself; ultimately such mechanisms must be set within wide temporal and spatial contexts to understand the landscapes within which weathering occurs (Smith 1987, Turkington and Paradise 2005, Turkington et al. 2005). In doing this, it must be recognized that weathering mechanisms do not operate in isolation from one another and that integration must take place at a variety of levels. There is a need, for example, to integrate across the scale boundaries that separate the detailed understanding of weathering at the nano- and microscales from their role in the evolution of complex landforms and landscapes, especially through its control on debris character and rates of debris supply.

In relation to the up-scaling of weathering process and effects to the landform scale and beyond, Viles (2001) saw the identification of the spatio-temporal scales of weathering phenomena as an essential first step. In pursuit of this she tabulates weathering features in terms of their scale of operation, together with the factors that control weathering and how these change across a range of scales. This work is also an acknowledgement that many of the potential approaches to the crossing of scale boundaries are rooted in the interpretation of geomorphic phenomena as non-linear dynamical systems. This interpretation builds upon the work of Goudie and Viles (1999) in which they demonstrated the applicability of the magnitude/frequency concept to understanding the operation of a wide range of weathering processses. They concluded in turn that the magnitude and frequency of weathering events is largely controlled by climate, but that climate itself comprises different magnitude and frequency distributions including cyclical and non-cyclical events. This theme was later developed (Viles and Goudie 2003) through a specific examination of the links between climatic variability at different scales and geomorphic activity in general.

The non-linearity of the behaviour of weathering systems has been further investigated by Phillips (2005), through an analysis of their stability and instability at four different scales concerned with weathering processes, the allocation of weathering products, the interrelation of weathering and denudation and topographic and isostatic responses to weathering. Following a detailed analysis of these systems, he was led to conclude that instability is prevalent at local spatial scales, except over the longest timescales. At intermedite spatial scales stability is contingent upon the strength of the feedbacks between weathering and erosion – with stability more likely at shorter than longer timescales. Over the largest spatial scales, instability is likely, except possibly over intermediate timescales if weathering and erosion feedbacks are weak. As he points out, the significance of these analyses lies in the tendency for stable systems to experience convergent evolution, whereas instability favours divergent evolution in which original heterogeneities in the landscape tend to be accentuated over time. An understanding of systems behaviour at this level is essential if we are going to progress desert weathering beyond its exploration and descovery phase to the point where the interactions between process, form, material and environment can be modelled and future change predicted.

One area where the challenge of modelling system behaviour has been taken up in recent years is the evolution of cavernous hollows (honecombs and tafoni). As Viles (2005b) has pointed out, there have been a number of recent studies that have sought to model cavern development and raise the question of whether there are common underlying controls that cut across scale, lithological and broad environmental boundaries. For many years it has been proposed that caverns develop to optimise internal environmental conditions that favour the operation of processes such as salt weathering (e.g. Smith 1978), but more recently, Turkington and Phillips (2004) have formalised this concept by proposing that caverns develop as a self-organised response to dynamical instability within the weathering system. Although, as Phillips (1999) notes, self-organisation is a broad umbrella which encompasses concepts that range from internal adjustment towards the maintenance of a steady state; through to the tendency for natural systems to evolve towards a critical threshold. The potential for complexity in the application of this paradigm is demonstrated in their paper by observation of the disparity between caverns, which tend to minimise surface area through development towards a spherical interior, and the tendency for the outcrops on which they occur to progress towards maximum exposed surface as caverns develop. A similar invocation of self-organisation was presented by McBride and Picard (2004), although this time in relation to the initial distribution of salt-rich water within the rock and its potential role in controlling the location of caverns. Lastly, Huinink et al. (2004) use a detailed analysis of salt migration to model the development of a single cavern. As Viles (2005b) observes, the appearance of these three studies in the same journal volume highlights the opportunities and need for integration, not only across scales but also between field and laboratory, between different data sets and, possibly most important, across the mindsets of researchers coming from different backgrounds.

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